# EPSC 355: Sedimentary Geology

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# Contents

# Chapter 1

# Introduction to Sedimentary Geology

## 1.1 Background

These course notes are intended to serve the dual purposes of 1) providing me with a set of teaching notes that I can update and work from every time I teach the course, and 2) offering students a detailed set of pre-typed notes so that they can spend less time dictating what I say in class and more time concentrating on what I say. I began these course notes in 2007 and I update them every time I teach the course. In doing so, I introduce new material and detail, but also new errors. I envision these course notes eventually approaching the form of a rudimentary textbook in sedimentary geology designed around the way I teach the course. However, at present, they are still somewhat rough and somewhat dependent on a few primary sources, namely textbooks by Boggs (2011), Tucker (2001), and Bridge and Demicco (2007), and a set of extremely useful MIT Open Courseware notes developed by John Southard (*http://ocw.mit.edu/courses/earth-atmospheric-and-planetary-sciences/12-110-sedimentary-geology-spring-2007/lecture-notes/*). I endeavor to give credit where credit is due, and where feasible, design my own tables and figures, but invariably have omitted some references, for which I apologize. Because many of the figures are copyrighted, these notes are intended only for your personal educational use. As the reader of these course notes, you are welcome and encouraged to offer corrections and suggestions to improve them for future years.

## 1.2 What is Sedimentary Geology?

Sedimentary Geology encompasses any field of science dealing with sediments or sedimentary rocks — that is rocks formed by Earth surface processes that begin with weathering on land, followed by erosion, and finally the physical settling of grains or chemical precipitation of minerals from air, water, or ice. Sediments (unlithified) and sedimentary rocks (lithified) are archives of information about environmental, tectonic, and biological conditions that prevailed at the time they were laid down; even their very existence 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. Hence, they represent one arc in the continuous and long cycle of the shaping and reshaping of Earth's landscape, as envisioned by James Hutton.

## 1.3 Why Sedimentary Geology?

I am a sedimentary geologist, so obviously I have a penchant for sedimentary rocks. I hope at least a few of you will as well. As you will see, they are both valuable for reading the history of the Earth and for storing water and other resources. The are also widespread, so to the extant that humans want to control their environment, they most know something about sedimentary processes and sedimentary rocks.

### 1.3.1 Distribution of Sedimentary Rocks on Earth

Sedimentary rocks only make up about 5–10% of the volume of the continental crust, but they cover 75% of the surface of the Earth. Much of this cover is in the oceans, but of course, they are widespread on the continents as well, covering, e.g., huge swaths of southern Quebec.

### 1.3.2 Sedimentary rocks have tremendous economic and social importance

- Petroleum (oil + gas)
- Coal
- Iron, potash, uranium, phosphate, copper, zinc...
- Heavy minerals (for example zircons)
- Water (i.e. aquifers)

### 1.3.3 Earth History

Sediments and sedimentary rocks archive the history of the Earth. Much of what we know about terrestrial environments and regional climate change is derived from sediments deposited in lakes and rivers. On the other hand, sediments deposited in the oceans may reflect both regional and global conditions. Deep sea sediments, which are systematically cored through the Integrated Ocean Drilling Program (IODP), have provided extensive records of global environmental change over the past 200 millions. It is through the study of sedimentary rocks that geologists track the tempo of oxygenation of the atmosphere and oceans and the long-term decline in $CO_2$ concentrations. Much of what geologists have deduced about Earth's environmental history has been accomplished through geochemical studies of rocks, in particular rocks that are *precipitated* from seawater, such as carbonates, and organic-rich mudstones, which tend to scavenge metals from seawater.

Most information about ancient life on Earth derives from sedimentary rocks, which alone are able to preserve fossils. Hence, any palaeontologist that works with fossils has also to work with sedimentary rocks. Not only are the fossils preserved in sedimentary rocks, but also the rocks bear witness to the environments in which organisms lived and died. The fossil record was the original evidence of evolution. At the same time, evolution has fundamentally changed the sedimentary record. Not only to organisms make sediments, but they also irrigate sediments in search of food and protection. This *bioturbation* wipes out original sedimentary textures and structures and also influences how chemicals cycle.

Indeed, it has recently been proposed the the origin of bioturbation animals imposed an important negative feedback on oxygenation of the environment through decreasing the potential for burial of organic material and recycling of phosphorous (Boyle et al., 2014).

If the incontrovertible evidence of ancient mountain belts lies in their metamorphic or igneous roots, the more detailed history of mountain building and destruction is best deduced from the sedimentary record. Furthermore, sedimentary basins are a direct manifestation of tectonics. Hence, sedimentary rocks are also an invaluable source of data for reconstructing tectonic evolution.

### 1.3.4 History of Mars

The Mars Science Observatory (Curiosity Rover) was launched in November 2011 and landed in Gale Crater in August, 2012. The principal mission of the rover is to assess whether Mars currently or has previously harboured environmental conditions that could have sustained microbial life—that is, is it or was it habitable. Given that simple life may well have inhabited environments that are also prone to sediment accumulation, such as pond and lakes, Curiosity is effectively a Martian sedimentary geologist. In fact, the project scientist on the mission, John Grotzinger, is a well known sedimentologist who has specialized in Precambrian strata and Earth history. One of the field projects for the Curiousity is to log a section of what appears to be a thick stack of strata within the central uplift of Gale Crater.

- Evidence for water
- Evidence of life
- A preserved analogue for early Earth?

## 1.4 Stratigraphy and Mapping

### 1.4.1 Subdisciplines and relatives of sedimentary geology

- Process sedimentology
- Petroleum geology
- Marine geology
- Stratigraphy
- Paleontology
- Paleoclimatology
- Paleoceanography
- Earth history
- Tectonics
- Geomorphology

Figure 1.1: Nicolas Steno (1638–1686), the first stratigrapher.

## 1.5 Concepts in Stratigraphy

Geology has only been a scientific discipline since the early 19th century. However, curious intellectuals contemplated the deeper meaning of sedimentary rocks long before that. Aristotle, Leonardo Da Vinci, and Nicolas Steno all famously endeavoured to understand sedimentary rocks, with varying degrees of success. Steno (1638–1886), a Danish anatomist, was certainly the most successful geologist of these early dilettantes; for his efforts, he is often regarded as the proto-father of geology.

### 1.5.1 Steno's laws and sedimentary contacts

**Steno's Laws**

- Law of superposition
- Law of original horizontality and lateral continuity
- Law of cross-cutting relationships
- Law of inclusion

**Sedimentary contacts**

- conformable
- unconformity
- angular unconformity
- disconformity
- nonconformity

You may come across the term *diastem*, which is the geologically fancy way of implying an interval of non-deposition or local erosion, without any major change in depositional

Figure 1.2: Siccar Point, Scotland: The angular unconformity between folded Silurian greywackes and the Devonian Old Red Sandstone, first recognized by James Hutton and John Playfair in 1788 and cited as proof of the geological evolution of Earth.

environment. In this sense, it is something like a minor disconformity.

James Hutton, typically considered the father of modern Geology, developed the concept of *Uniformitarianism*: the geological record of past processes can be understood in terms of processes that are actively occurring on Earth today. This philosophy opposed the notion of *Catastrophism* that the Earth was shaped by a series of catastrophic events, such as the "Great Flood." Many geologists today would accept that both actualistic and catastrophic processes were responsible for features of the geological record, but unsurprisingly, you will find geologists at both ends of the spectrum.

### 1.5.2 Lithostratigraphic subdivisions

*Stratigraphy* 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). We will discuss stratigraphy in much more detail later in the course, but because stratigraphy underlies much of the field of sedimentary geology (and will be alluded to regularly in class), it is useful to introduce for now lithostratigraphic units. 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

### 1.5.3 The age of sedimentary rocks

Knowing the age of sedimentary rocks is important, not just for understanding geological history, but also for many practical reasons, such as exploring for minerals. There are various different ways of establishing ages of sedimentary rocks.

- Relative dating
- Correlation
- Absolute (radiometric) dating
  - (U-Pb, $^{14}C$, Rb-Sr, Re-Os, Lu-Hf, K-Ar)

### 1.5.4 Correlation

Most stratigraphers are deeply preoccupied by *correlation*, which is the science (or art) of connecting geographically separated rock units either within a single sedimentary basin or between different sedimentary basins based on their similar age or lithological character. It is typically more useful to correlate samples by age than by lithology alone, and many tools can be employed in *chronostratigraphy.* Indeed, many of these tools are the bases for distinct subdisciplines whose principal intent is to establish time lines that can be used for correlation and calibration o the sedimentary record. We will discuss correlation more in the section on stratigraphy later in the course.

### 1.5.5 Other types of stratigraphy

- Biostratigraphy
- Chronostratigraphy
- Magnetostratigraphy
- Chemostratigraphy
- Seismic stratigraphy
- Sequence stratigraphy

## 1.6 General classification of sedimentary rocks

In the most basic way, the sediments in sedimentary rocks can be distinguished as having one of four origins. We will further subdivide each of these rock types during the course.

- Siliciclastic (many detrital silicate minerals and rock fragments or clasts)
- Biogenic/Organic
- Chemical
- Volcanigenic

# Chapter 2

# Weathering: the Source of All Sediments

## 2.1 Introduction

*Weathering* is the decomposition of rocks on Earth's surface into particles (sediments) and dissolved constituents. It occurs primarily at the interface of the atmospheric and lithospheric systems, but the cyrosphere, hydrosphere, and biosphere are also implicated in weathering. Weathering is logically viewed as the first step in the production of sedimentary rocks, because it is the ultimate source of all sediments—both physics and chemical. The physical product of weathering is sediment grains, commonly referred to as clasts or detritus (Bridge and Demicco, 2007). Sediments grains derived from physical break-down of the land surface are referred to *terrigenous.*

Weathering involves physical, chemical, and biological processes, usually in concert with one another, which result not only in freeing up the constituent parts of rocks for transport (erosion), but also in physical and chemical sorting through the production of secondary minerals, release of labile elements, and retention in soils of immobile elements (e.g. Fe and Al). To the first order, the products of weathering are controlled by mineralogy, which determines the initial size, shape and solubility of the constituents of rocks.

Weathering is also intricately interconnected with Earth's long-term climate, mainly through the *silicate weathering feedback*, which regulates atmospheric $CO_2$ concentrations over timescales of >100's of thousands of years. Here we will only discuss subaerial weathering, but submarine weathering is also an important process in modulating the chemistry of the oceans.

Soils can be thought of as the reaction chamber where a significant component of subaereal weathering takes places. Quite simply, soils are the physically, chemically, and biologically altered mantles of material that exist between bedrock and the atmosphere. Detailed discussion of soil development and soil types is beyond the scope of this course. However, in sedimentary geology, *paleosols* are of particular interest insofar as they provide windows into past earth surface environments.

## 2.2 Physical weathering

### 2.2.1 The freeze-thaw cycle

The most efficient physical mechanism for breaking down rocks in the freeze-thaw cycle, through which freezing water in cracks exerts pressure and the bounding rocks. This mechanisms is of course only active where temperatures drop below freezing—that is in temperate climates. Large aprons of *skree* at the base of high mountain slopes are testament to the effectiveness of freeze-thaw cycles in splintering fragments of bedrock. Once these have fallen from the mountain, they are more prone to chemical and biological weathering due to their increased surface area and exposure to rainfall and flowing water.

Freezing water in pore spaces can also lead to disintegration of rocks. The famous *nids de poule* in Montréal are an excellent example. Asphalt rests on a bed of porous gravel. When those pores are saturated with water, which then freezes, it causes the overlying pavement to dome upwards and crack. Once this process has started, it does not take long for a pot hole to form as cars drive over it and additional freeze-thaw cycles continue to break it up.

### 2.2.2 Pressure release and jointing

Another important and more widespread mechanism for fracturing bedrock is the result of decreased overburden pressure as material above is removed by erosion. The decreased pressure results in the generation of joints that are roughly parallel to the land surface. Once these joints open, water, roots, and other agents of weathering can make their way in and contribute to the physical breakdown of the rocks. *Exfoliation*, is initiated by pressure release jointing and leads to layers of isotopic rocks like granite peeling off like onion skins.

Joints also result from cooling, as in the case of lavas, and tectonics. These too contribute the eventual physical breakdown of bedrock. Isotropic rocks that have three sets of roughly perpendicular joints (common in sills and dikes, for example) may weather out as spheroids. *Spheroidal weathering* is the result of the preferential chemical and physical abrasion of the corners of cubic rock fragments formed by such jointing.

### 2.2.3 Other physical weathering processes

Other physical processes, such as insolation, wetting and drying, and the crystallization of salts, such as halite and calcium sulfates, all contribute to weathering, although to a lesser known extent. In theory, insolation can lead to cracking of a rock due to repeated heating and cooling, resulting in expansion and contraction of a rock. However, the overall change in volume is typically small. The exception is during large fires, where rocks can crack and explode due to heating. Wetting certain minerals, such as anhydrite and clays, can lead to significant expansion, which like the in situ precipitation of salts, can also crack rocks.

## 2.3 Chemical weathering

Chemical weathering involves alteration to the chemistry and mineralogy of rocks and results in the loss of mass through dissolution. It is largely driven by acid-base and oxidation-reduction (redox) reactions which take place in wetted soils. As should be expected, chemical weathering is heavily dependent on the availability of water, temperature, and concentrations of dissolved gasses. It is the ultimate source of bioavailable nutrients and plays a central role in biogeochemical cycles (Putnis and Ruíz-Agudo, 2013).

### 2.3.1 Congruent dissolution

*Congruent dissolution* occurs where a particular constituent dissolves completely. Many common minerals dissolve in water due to the high dielectric constant of water. This is known as *hydrolysis*. The dissolved minerals will exist in solution in water as their constituent ions. You are familiar with the equilibrium constant, $K$, which for a simple scenario of a solution A and mineral B, which dissolves into ions C and D is defined as

$$K = \frac{[C]^c[D]^d}{[A]^a[B]^b} \tag{2.1}$$

Where we are considering the solubility of a mineral in water, we can simplify this to the solubility product, $K_{sp}$:

$$K_{sp} = [A]^a[B]^b \tag{2.2}$$

which is typically defined at standard temperature (25°C) and pressure (1 atm). Highly soluble minerals such as halite have relatively high $K_{sp}$ and insoluble minerals, such as quartz, have vanishingly low $K_{sp}$.

In practice, the theoretic concentration of the ionized constituents of a dissolved mineral differ from what is actually chemically available in solution. Hence, geochemists usually consider the *activities* (effective concentrations) rather than actual concentrations. The activity of an ionic substance is defined as

$$a = m\gamma \tag{2.3}$$

where $m$ is the molality (number of moles of the ionic substance in a kilogram of water) and $\gamma$ is the *activity coefficient*. The activity coefficient is usually $<1$ but approaches 1 in dilute solutions. As a consequence, dilute solutions (such as fresh rain water) are much more effective at dissolving minerals than concentrated solutions. This an extremely important consideration in quantifying chemical weathering rates.

### 2.3.2 The role of carbonic acid

Some minerals, such as halite, dissolve readily in pH neutral waters. However, the dissolution rates for most minerals are much higher at lower pH. For example, carbonates will dissolve gradually in pure, neutral waters, but their dissolution is significantly accelerated in the presence of dissolved $CO_2$:

$$CaCO_3 + H_2O + CO_2 \Leftrightarrow Ca^{2+}(aq) + 2HCO_3^-(aq) \tag{2.4}$$

When this dissolved calcium carbonate reprecipitates, it will release back its $CO_2$ molecule to the atmosphere. Hence, dissolution of carbonates is not a mechanism for reducing atmospheric $pCO_2$. On the other hand, the weathering of silicate minerals does remove $CO_2$ permanently. *Silicate weathering* is hence defined as the hydrolysis of silicate minerals through reaction with carbonic acid. For example:

$$Mg_2SiO_4 + 4CO_2 + 4H_2O \rightarrow 2Mg^{2+} + 4HCO_3^- + H_4SiO_4 \quad (2.5)$$

similarly, for Ca silicates (but written in a different way)

$$Ca_2SiO_4 + 4CO_2 + 4H_2O \rightarrow 2Ca^{2+} + 4HCO_3^- + H_4SiO_4 \quad (2.6)$$

You can see from these simple equations, that these reactions produce Mg and Ca cations and the bicarbonate anion. These ultimately go on to form carbonate, while the dissolved silica can be incorporated into opal or chert. These equations are the basis for the *silicate weathering feedback*, whereby higher $CO_2$ concentrations in the atmosphere result in a higher temperatures and increased rainfall, which combined increase rates of hydrolysis of silicate minerals, which tend to reduce the amount of $CO_2$ in the atmosphere, hence cooling climate and slowing down the rate of these reactions. The silicate weathering feedback is responsible for regulating Earth's climate over geological time scales, but at shorter (say human) time scales, it is too slow to play a large role in modulating atmospheric $CO_2$ concentrations.

### 2.3.3 Incongruent dissolution

Not all silicate minerals weather the same. Specifically, aluminosilicate minerals are more complicated and tend to weather *incongruently*, except where intense chemical weathering occurs. That is, their chemical weathering produces both a solute and a new less soluble secondary mineral. For example, the incongruent weathering of a potassium feldspar:

$$2KAlSi_3O_8 + 2H_2CO_3 + 9H_2O \rightarrow Al_2Si_2O_5(OH)_4 + 4H_4SiO_4 + 2K^+ + 2HCO_3^- \quad (2.7)$$

This produces both a clay (kaolinite), which is less soluble than the original feldspar, and dissolved silica and carbonate. Indeed, it the dissolution of feldspars and not quartz that yields most silica into solution in groundwater. In addition to consuming $CO_2$ and producing solutes, incongruent hydrolysis produce clays, which are a major constituent of sediments and themselves play an important biogeochemical role.

### 2.3.4 Oxidation-reduction

Oxidation and reduction (redox) reactions are less important as a means of breaking down the continents, but are nevertheless important in biogeochemical cycles, regulating atmospheric oxygen concentrations, and increasingly in bioremediation. Oxidation reactions are most important for reduced iron and sulfur, which commonly coexist in sulfide minerals. Significant ferrous iron ($Fe^{2+}$ is also released from the dissolution of iron silicate minerals. In both cases, rather than form clays, as may happen to weathered Mg silicates during incongruent dissolution, the iron is typically rapidly oxidized and incorporated into iron oxyhydroxide minerals:

$$Fe3+ + 3H_2O \Leftrightarrow Fe(OH)_3 + 3H^+ \quad (2.8)$$

The resulting amorphous iron oxyhydroxide is highly insoluble, but with time, it tends to convert into even more insoluble goethite, then hematite.

Table 2.1: The main weathering processes and their resulting weathering and sedimentary products (adapted from Boggs, 2011).

| **Weathering process** | **Weathering products** | **Example** | Depositional product |
|---|---|---|---|
| physical weathering | Particles | silicate minerals and rock fragments | sandstone, conglomerate, mudstones |
| chemical weathering | | | |
| *congruent solution* | solutes | $HCO_3^-$, $SO_4^{2-}$ | limestone, chert, evaporites |
| *hydrolysis* | solutes | silicic acid, cations, $HCO_3^-$ | limestone, chert, evaporites |
| | secondary minerals | clay minerals | mudstones |
| oxidation | secondary oxides | Fe oxyhydroxides, Mn oxides | accessory minerals |
| | solutes | silicic acid, $SO_4^{2-}$ | Chert, evaporites |

The oxidative weathering of pyrite, in addition to producing iron oxides, also produces sulfate:

$$FeS_2 + \frac{15}{4}O_2 + \frac{7}{2}H_2O \Leftrightarrow Fe(OH)_3 + 2SO_4^{2-} + 4H^+ \tag{2.9}$$

This is the ultimate source of sulfate in seawater and in in calcium sulphate evaporites. Hence, it is an extremely important reaction to sedimentary geologists, geochemists, and home builders!

Reduction reactions also take place, but these require a reducing environment, which is mostly commonly generated in presence of organic matter. This mostly takes place in sediments, and hence these reduction reactions will because further in the context of early diagenesis.

## 2.4 Clay minerals

As mentioned, clay minerals are a common product of chemical weathering and a major constituent of sedimentary rocks. They are important for a host of other reasons; they are repositories of trace elements, including nutrients, and for they readily adsorb organic matter and ions. Clay minerals can be separated into four classes: kandites, smectites, illites, and mixed-layer clays. These are largely made up of two basic building blocks: sheets of silica tetrahedra and sheets of alumina octahedra (an aluminum atom surrounded by 6 OH groups). The mineral consisting just of a sheet of aluminum and hydroxide groups is known as *gibbsite*. *Brucite* is the equivalent mineral with a $Mg^{2+}$ cation in place of an $Al^{3+}$ cation in the centre of the octahedron.

### 2.4.1 Kandites

*Kandites* are two-layer clays (alternating octahedral and tetrahedral layers held together by *van der Waals* bonds), of which kaolinite is a common example. Kaolinite is a common alteration product in soils and forms most readily under warm, humid conditions due to alteration of feldspars and other aluminosilicate minerals. Serpentine is a two-layer clay in which the octahedral layers are made up of brucite.

### 2.4.2 Smectites

*Smectites* are three-layer clays (such as montmorillonite), comprising two tetrahedral layers separated by an octahedral layer. $Mg^{2+}$ and $Fe^{2+}$ substitutions into the $Al^{3+}$ sites in the octahedra and $Al^{3+}$ substitutions for $Si^{4+}$ in the tetrahedral layers result in negative charge imbalances in the individual smectite minerals, which attracts water and exchangeable cations, such as $Ca^{2+}$ and $Na^{+}$. Consequently, smectites have significant ion exchange capacity and the ability to expand or contract, depending on availability of water. Smectites tend to form during the alteration of poorly drained mafic rocks, such as basalt flows.

### 2.4.3 Illites

*Illites* are also three-layer clays in which the interlayer charge imbalance is compensated by $K^{+}$. Vermiculite is like illite, except $Mg^{2+}$ occupies the interlayer site. In chlorite, the octahedral layer brucite is sandwiched between the three layer structures instead of cations. Illite and chlorite have poor cation-exchange capacity and do not tend to shrink and swell.

### 2.4.4 Mixed-layer clays

This diverse group of clay minerals comprises alternating layers of smectite, illite, and chlorite structural layers.

# Chapter 3

# Physical Properties of Sedimentary Rocks

*Supplemental reading: Boggs, Chapters 1, 3, 5; Tucker, Chapters 2 (11–20), 3 (92–102); Bridge and Demicco, Chapter 3*

## 3.1 Introduction

The first thing a geologist does when trying to understand or document sedimentary rocks in the field is to describe the physical characteristics of those rocks. Colour is an obvious physical property and should, of course always be described. When describing colour, one should describe the colour of both weathered and fresh surfaces of the rock, as well of both the grains and matrix or cement (where visible) that make up the rock. Although colour can be misleading to the extent that is prone to change during *diagenesis* and weathering, it nevertheless may be an important clue to the mineralogical or elemental constituents of the rock. And often colour is diagnostic in identifying certain sedimentary units in the field.

Typically more reliable than colour in describing a sedimentary rock are its textures (such as grain size and sorting) and its physical composition. These should be described systematically whenever describing a rock and should be consistent, unlike colour, which can be whimsical.

## 3.2 Textures of Siliciclastic Sedimentary Rocks

The texture of a sedimentary rock refers the geometry of the grains comprising that rock. The texture (and composition, which will be discussed in the next section) of a sedimentary rock reflects 1) the mineralogy and grain size of source rocks from which the sediments were derived, 2) the weathering and erosion processes that produced and transported the sediments, 3) the processes by which the sediments were deposited, and 4) alteration to these sediments after burial (*diagenesis*, which will be discussed in a subsequent chapter). These basic features are important for reconstructing depositional environment and provenance, hence geological history. These are also the principal features that are used to describe and classify sedimentary rocks.

- Grain size

- Grain morphology
- Grain Fabric
- Textural Maturity
- Compositional Maturity

### 3.2.1 Grain size and grain size parameters

Grain size is the most fundament physical property of sedimentary rock, and yet it is not necessarily straightforward. How do we define the size of a grain? Right, this is not so easy because grains are generally irregularly shaped. Theoretically, we might use the *nominal diameter* of a particle, which is defined as the diameter with the same volume as that of a grain in question. However, this is not a particularly practical approach, and we can hardly measure every grain of a sediment to determine patterns of grain size.

In practice, one useful way to define grain size is operationally, through sieving, which selects grain size based on the largest dimension in the least cross-sectional area of the grain (Southard, 2007). This method works reasonably well for silt and coarser sediments, but not for very fine-grained sediments. But again, it is not always practical, for you cannot readily sieve a lithified sandstone. So in fact, we usually just make a judgement call, perhaps with the assistance of a grain-size chart.

- Wentworth Table
- Size class and mm scale most practical for descriptive purposes
- phi ($\phi$) scale useful in mathematical treatments: $\phi = -\log_2(D/mm)$, where D is the diameter of the particle.

**Characterization of grain size:**

- mean grain size
- mode
- median grain size
- sorting
- skewness
  - most sediments plot out as gaussian

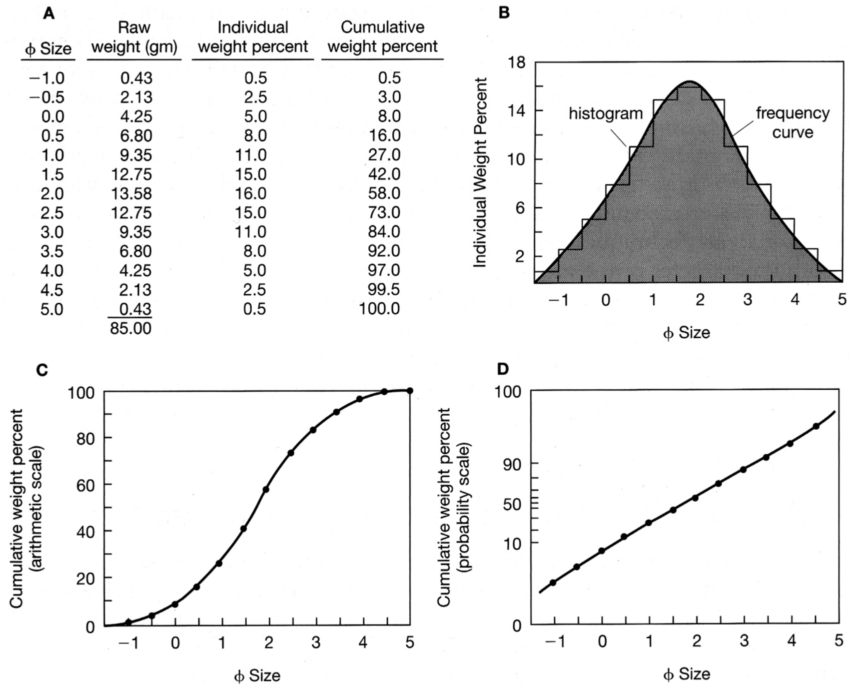
The most meaningful way to calculate the mean grain size is to measure the *logarithmic mean*,

$$M = \frac{\Sigma(fx')}{N} \tag{3.1}$$

where $f$ is the frequency (usually measured as weight percent) of a given sieve range, of which x' is the midpoint, and N is the total number of size intervals. However, there are

| | Millimeters | µm | Phi (φ) | Wentworth size class | |
|---|---|---|---|---|---|
| | | | -20 | | |
| | 4096 | | -12 | Boulder (-8 to -12φ) | |
| | 1024 | | -10 | | |
| | 256 | | -8 | | |
| | | | | Pebble (-6 to -8φ) | |
| | 64 | | -6 | | |
| | 16 | | -4 | Pebble (-2 to -6φ) | |
| | 4 | | -2 | | Gravel |
| | 3.36 | | -1.75 | | |
| | 2.83 | | -1.50 | Gravel | |
| | 2.38 | | -1.25 | | |
| | 2.00 | | -1.00 | | |
| | 1.68 | | -0.75 | | |
| | 1.41 | | -0.50 | Very coarse sand | |
| | 1.19 | | -0.25 | | |
| | 1.00 | | -0.00 | | |
| | 0.84 | | 0.25 | | |
| | 0.71 | | 0.50 | Coarse sand | |
| | 0.59 | | 0.75 | | |
| 1/2 | 0.50 | 500 | 1.00 | | |
| | 0.42 | 420 | 1.25 | | |
| | 0.35 | 350 | 1.50 | Medium sand | Sand |
| | 0.30 | 300 | 1.75 | | |
| 1/4 | 0.25 | 250 | 2.00 | | |
| | 0.210 | 210 | 2.25 | | |
| | 0.177 | 177 | 2.50 | Fine sand | |
| | 0.149 | 149 | 2.75 | | |
| 1/8 | 0.125 | 125 | 3.00 | | |
| | 0.105 | 105 | 3.25 | | |
| | 0.088 | 88 | 3.50 | Very fine sand | |
| | 0.074 | 74 | 3.75 | | |
| 1/16 | 0.0625 | 63 | 4.00 | | |
| | 0.0530 | 53 | 4.25 | | |
| | 0.0440 | 44 | 4.50 | Coarse silt | |
| | 0.0370 | 37 | 4.75 | | |
| 1/32 | 0.0310 | 31 | 5 | | |
| 1/64 | 0.0156 | 15.6 | 6 | Medium silt | |
| 1/128 | 0.0078 | 7.8 | 7 | Fine silt | |
| 1/256 | 0.0039 | 3.9 | 8 | Very fine silt | Mud |
| | 0.0020 | 2.0 | 9 | | |
| | 0.00098 | 0.98 | 10 | | |
| | 0.00049 | 0.49 | 11 | Clay | |
| | 0.00024 | 0.24 | 12 | | |
| | 0.00012 | 0.12 | 13 | | |
| | 0.00006 | 0.06 | 14 | | |

Figure 3.1: The Wentworth grain size scale, after Blair and McPherson (1999).



A

| ϕ Size | Raw weight (gm) | Individual weight percent | Cumulative weight percent |
|---|---|---|---|
| −1.0 | 0.43 | 0.5 | 0.5 |
| −0.5 | 2.13 | 2.5 | 3.0 |
| 0.0 | 4.25 | 5.0 | 8.0 |
| 0.5 | 6.80 | 8.0 | 16.0 |
| 1.0 | 9.35 | 11.0 | 27.0 |
| 1.5 | 12.75 | 15.0 | 42.0 |
| 2.0 | 13.58 | 16.0 | 58.0 |
| 2.5 | 12.75 | 15.0 | 73.0 |
| 3.0 | 9.35 | 11.0 | 84.0 |
| 3.5 | 6.80 | 8.0 | 92.0 |
| 4.0 | 4.25 | 5.0 | 97.0 |
| 4.5 | 2.13 | 2.5 | 99.5 |
| 5.0 | 0.43 | 0.5 | 100.0 |
| | 85.00 | | |



Figure 3.2: Methods to display grain-size data (from Boggs, 2011).

also useful shortcut methods for calculating the mean, as shown below, which also have the benefit of not counting the tails in the distribution, which are often hard to measure accurately and can distort your calculations.
Graphic Median and Mean Median

$$M = \phi_{50} \tag{3.2}$$

Graphic Mean

$$M_z = \frac{\phi_{16} + \phi_{50} + \phi_{84}}{3} \tag{3.3}$$

Inclusive graphic standard deviation

$$\sigma_i = \frac{\phi_{84} - \phi_{16}}{4} + \frac{\phi_{95} - \phi_5}{6.6} \tag{3.4}$$

Inclusive graphic skewness

$$SK_i = \frac{\phi_{84} + \phi_{16} - 2\phi_{50}}{2(\phi_{84} - \phi_{16})} + \frac{\phi_{95} + \phi_5 - 2\phi_{50}}{2(\phi_{95} - \phi 5)} \tag{3.5}$$

**Sedimentological meaning of grain size**

The size of grains in a sediment is controlled mainly by

- The nature of the weathered material the produced the sediment
- The history of the sediment during transportation and deposition

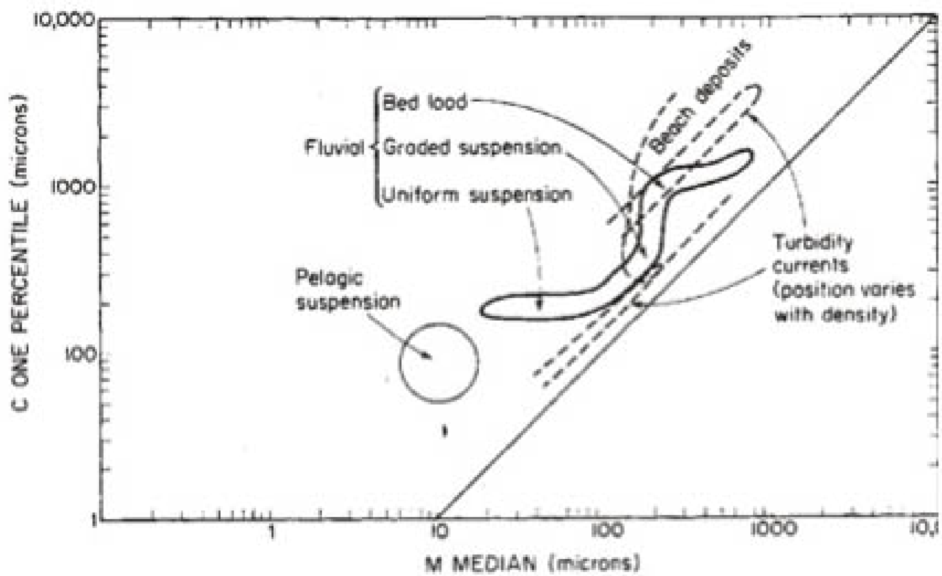



Figure 3.3: A C-M diagram (C = the 1 percentile on a cumulative weight percent plot; M= median grain size), displaying how grainsize data can be used to discriminate between depositional processes and environments. From Passega (1964).

Table 3.1: Measure of sorting by the standard deviation of grain size distribution

| **standard deviation** | **degree of sorting** |
|---|---|
| $<0.35\phi$ | very well sorted |
| $0.35–0.50\phi$ | well sorted |
| $0.50–0.71\phi$ | moderately well sorted |
| $0.71–1.00\phi$ | moderately sorted |
| $1.00–2.00\phi$ | poorly sorted |
| $2.00– 4.00\phi$ | very poorly sorted |
| $>4.00\phi$ | extremely poorly sorted |

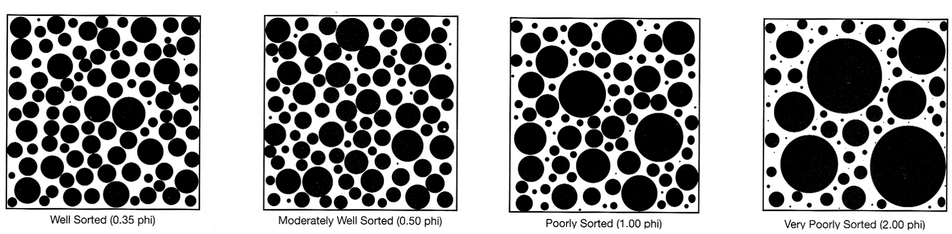



Figure 3.4: A visual tool for estimating grain size sorting (from Boggs, 2011)

For example, a source area that was dominated by shales would yield only fine-grained sediment far enough downstream that any fragments had abraded. This example leads us to a common observation in river systems, which is that sediment size decreases downstream.

Table 3.2: Ranking of grain size skewedness.

| | **skewness** |
|---|---|
| >+0.30 | strongly fine skewed |
| +0.3 to +0.10 | fine skewed |
| +0.10 to -0.10 | near symmetrical |
| -0.10 to - 0.30 | coarse skewed |
| <-0.30 | strongly coarse skewed |

This feature, known as *downstream fining* is a result of many factors (Southard, 2007), which include

- storage of larger grains upstream
- progressive abrasion with transport downstream
- fracturing of some of the coarser particles (due, e.g., to repeated collisions)
- weathering of sediments, both within the river and in temporary storage places (bars, floodplains, etc.)
- dilution by finer-grained sediment populations derived from tributaries

### 3.2.2 Grain morphology

#### Shape or sphericity

One way to describe e the shape of a grain is to compare it to a sphere. The *sphericity* of a grain is defined as the ratio of the diameter of the largest inscribed circle to the nominal diameter. Hence, Sphericity varies from 0 to 1, with 1 corresponding to a perfect sphere. The drawback of this technique is it does not tell us much beyond whether a grain resembles a sphere or not.

Another approach is to compare the three axes of a grain: L, I, and S, where L, the long axis is the longest line that can be drawn through a grain and S is the shortest that can be drawn perpendicular to that axis. I is then the remaining, intermediate axis. Qualitatively, we can described grain shape based on the relative sizes of these axes:

- equant or sphedoidal (a basketball or a cube): L $\approx$ I $\approx$ S
- oblate or discoidal (a frisbee or hamburger): L $\approx$ I > S
- prolate or rod (a football or sausage): L > I $\approx$ S
- blades (triaxial): L > I > S

#### Roundness (or angularity) of grain corners

Roundness is defined as the *ratio of average radius of curvature of edges and corners to the radius of the largest inscribed sphere*. Given this definition, grain roundness can be measured quantitatively, but in practice, this is rarely done because it is tedious and of limited value. This measure can be simplified to the radius of curvature of the single sharpest corner of a grain. Nevertheless, it is more common in most circumstances to describe roundness descriptively, commonly using some sort of visual chart for reference.

- very angular
- angular
- sub-angular
- sub-rounded
- rounded
- well-rounded

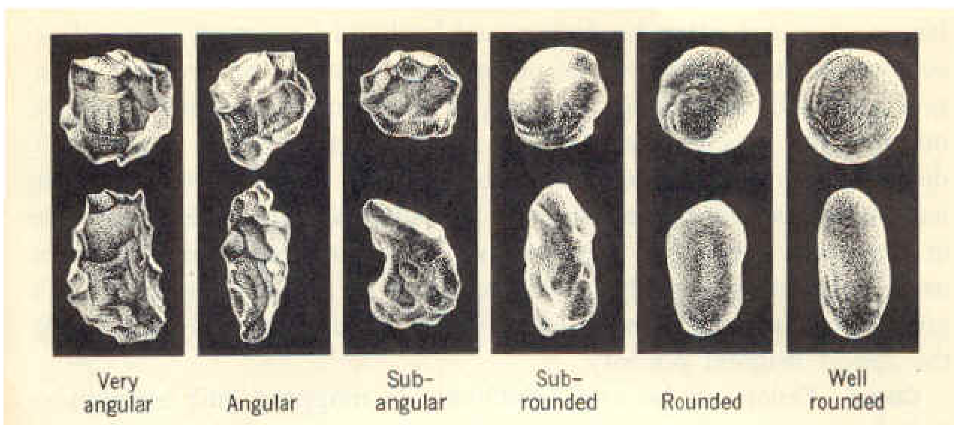



Figure 3.5: Roundness and sphericity of grains (Pettijohn et al., 1987).

In general, degree of roundness and sphericity are a function of duration of transport and reworking. However, there are various other factors that influence grain shape and form:

- mineralogy
- degree of weathering
- post-depositional modification of grains

### 3.2.3 Grain surface features

The surface texture of a grain typically reflects processes acting on that grain prior to or during erosion.

- frosting
- pitting
- striations
- fractures

### 3.2.4 Grain fabric

The *grain fabric* of sedimentary rocks is essentially how the grains are arranged with respect to one another. Grain fabrics are typically easy to pick out in a coarse-grained sediment, but in sandstones, fabric usually is determined under the microscope. A grain's fabric says something about the original process by which it was deposited and how and when it was influenced by diagenesis.

- Orientation
- Packing
  - Cubic vs. Rhombohedral
  - effect of sorting on porosity
- Grain versus matrix supported
- Type of grain contact
  - point
  - linear
  - concavo-convex
  - sutured

Grain fabric is also important in determining the *porosity* (volume of void spaces) and *permeability* (ease with which a fluid can flow through those void spaces) of a sediment. Although these are important physical characteristics, they will be discussed in more detail in the section on diagenesis.

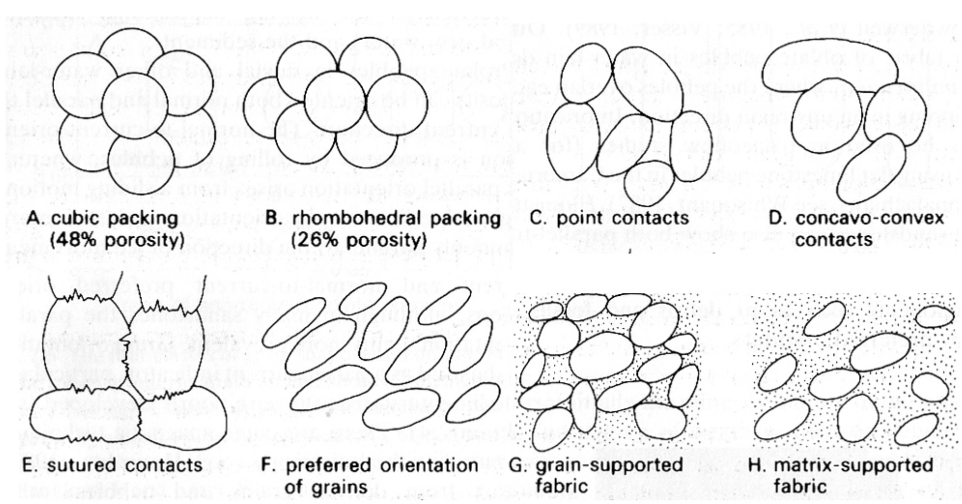



Figure 3.6: Examples of grain fabrics, from Tucker (2001).

## 3.3 Composition of Sedimentary Rocks

The composition of a sedimentary rocks is the mineralogical (or geochemical) makeup of the rock. Siliciclastic sedimentary rocks are made up of constituents that include minerals and rock fragments. The *framework grains* of a siliciclastic rock are the dominant, larger original grains that are in contact with one another. The *framework mineralogy* of a sandstone is the ensemble of mineral types (or rock fragments) that make up the framework grains. In contrast, the *matrix* is the fine-grained material (commonly defined as less than ∼40 microns) that occur along with the coarser framework grains. The common framework minerals include quartz, feldspar, and muscovite. Minerals that make up less than 5% of the rock are known as *accessory minerals*, the most important of which are the so-called *heavy minerals* (or *heavies*, for short).

Matrix minerals are those that are found in the space between the larger framework minerals. These may have been co-deposited with framework grains, filtered into pore spaces following deposition, or may have formed diagenetically (e.g., through alteration of feldspars). Distinguishing between framework and matrix grains is somewhat arbitrary and can be difficult in poorly sorted sediments. Soft larger grains may even become squished and deformed during compaction to the point that they become indistinguishable from matrix.

### 3.3.1 Quartz

Quartz is arguably the most important mineral in sedimentary rocks and is certainly the most abundant (∼25% of all sediments; Southard, 2007), although it decreases in abundance with decreasing grain size. Quartz is so abundant for two main reasons:

- It is an important constituent (10-20%) of average igneous and metamorphic rocks
- It is highly resistant to both physical and chemical weathering

Quartz is robust because it very pure, has no cleavage planes, and is stable under a high range of temperature and pressure conditions (Southard, 2007). The silica that is released by chemical weathering is derived from other silicate minerals and minimally from quartz, which dissolves only under specific conditions (high temperatures and/or high pH). Whereas quartz breaks down during weathering and erosion, this is largely along crystal boundaries and fractures; the extent of physical abrasion of quartz during erosion, although evident in the rounding of quartz grains, is small. Therefore, quartz obeys mass balance during weathering and erosion: the mass of quartz leaving the source area equals the mass of quartz deposited in the sedimentary basin (Southard, 2007).

Certain other minerals, namely zircon, are also rugged and survive multiple sedimentary cycles, but are insufficiently abundant in the crust to be a significant component in sedimentary rocks (even though they may be extremely useful for certain purposes, such as radiometric dating and provenance studies).

- Most quartz gravel and coarser-sized quartz is derived from the weathering of vein quartz
- Most sand-size quartz is monocrystalline, although polycrystalline quartz also occurs and is a useful indicator of source rocks

- There is little clay-size ($< 2\ \mu$m) quartz

### 3.3.2 Feldspar

Feldspar is the second most abundant mineral in siliciclastic rocks, comprising ~10–12% (Southard, 2007). Although reasonably common, its abundance in sandstones is highly variable, and feldspar is commonly insignificant. The abundance and type of feldspar in sedimentary rocks is also distinct from crystalline source rocks, with sedimentary rocks containing only about 20% as much feldspar as crystalline crustal rocks. Calcic plagioclase is relatively rare, sodic plagioclase slightly more common, and orthoclase, the most abundant feldspar, followed closely by microcline. It has been shown that feldspar does not abrade much faster than quartz, and hence that chemical weathering is the main factor controlling the breakdown of feldspar (Nesbitt and Young, 1996). These patterns in feldspar abundance are the result of a combination of factors:

- Feldspar is fundamentally less chemically stable
- Feldspar has cleavage surfaces (unlike quartz)
- Like other silicate minerals, Ca-rich feldspars (Ca-plag) weather more rapidly than Na-rich feldspars, which in turn weather more rapidly than K-feldspar.

The mineralogy and abundance of feldspar in a siliciclastic rock is therefore an important indicator of the sedimentary environment in which it formed. The classic interpretation, driven by the basic rules of silicate weathering, is that abundant feldspar in a sandstone reflects weathering and erosion in a cool, dry environment. However, while a reasonable interpretation, it is incomplete and hence largely incorrect. In addition to temperature and water, the main ingredient in the chemical breakdown of silicate minerals is time. If a feldspar grain is delivered from crystalline source to sedimentary sink in a hurry, it has little time to breakdown. Therefore, topographic relief (i.e. tectonics) probably plays the single most important role in the preservation of feldspar-rich sedimentary rocks.

### 3.3.3 Mica

Micas occur in sandstones both as large detrital grains and as matrix, in which case the mica may be either detrital or post-depositional in origin. Of the micas, muscovite is the most abundant, being more resistant to chemical and physical degradation than biotite and chlorite. Muscovite is common and moderately abundant (a few percent) in sandstones (thus normally an accessory mineral). Its flaky morphology results in it commonly settling with the finer-grained sediment fraction. The result is that micas are typically concentrated on bedding plans, increasing the tendency of mica-bearing rocks to break along bedding planes and making these bedding plans shiny.

### 3.3.4 Heavy minerals

Heavy minerals make up a small fraction (maybe 1% on average) of sedimentary rocks, but are relatively important for economic reasons and for the purpose of tracing source regions of sedimentary rocks. Most common sedimentary minerals have a density of about 2.65 $\frac{g}{cm^3}$. Heavy minerals are loosely defined as the fraction of minerals with a density $> 2.9$ $\frac{g}{cm^3}$. As such, they are a diverse and not necessarily related group of minerals.

- Opaque minerals include magnetite, hematite, and sulfides
- Non-opaque minerals include zircon, apatite, tourmaline, garnet, and rutile

Relatively high density mafic silicate minerals such as olivine, pyroxene, and amphibole are rare in sedimentary rocks due to their chemical instability during weathering.

### 3.3.5 Rock fragments

Rock fragments (or *lithics*) are detrital components in sedimentary rocks that are large enough to be distinguishable from mineral grains. Obviously, rock fragments cannot be smaller than their constituent grains; hence, coarse-grained crystalline rocks (e.g. granites and gneisses) do not contribute much in the way of lithic fragments to sand-sized and finer sediments. Rock fragments make up 10–15% of the average sandstone and are useful in determining source area of sediments (see subsequent section on Q-F-L classification), although of course insufficient in determining all source lithologies since many rocks will decomposes into their mineral constituents and leave no direct trace of their original lithology. Common rock fragments include

- Volcanic rocks
  - Mafic volcanics not common because they are chemically unstable
  - Felsic volcanics relatively common, but may be difficult to differentiate from impure chert
- Metamorphic rocks—mainly foliated rocks (slate-phyllite-schist)
- Sedimentary
  - Shale clasts are not uncommon, but are mechanically unstable. Hence, where they are found, they indicate a nearby shale source.
  - Chert is the most common rock fragment due to its ubiquity and resistance to weathering. It can be hard to distinguish from proper quartz grains.

### 3.3.6 Matrix

The fine-grained (mud) fraction that occurs in interstitial spaces is referred to as the *matrix*. The matrix (roughly $< 0.04$ mm) may include a combination of the same minerals forming the framework (quartz and feldspar) and other fine minerals, most notably clay minerals. The distinction between framework and matrix minerals may be hard to establish in matrix-rich rocks. It is also difficult to distinguish detrital matrix minerals from authigenic minerals. Clays are the most common matrix minerals minerals, and of these, illite, smectite, and kaolinite are the most abundant in sedimentary rocks.

# Chapter 4

# Classification and Provenance of Siliciclastic Rocks

*Supplemental reading: Boggs, Chapters 1, 3, 5; Tucker, Chapters 2 (11-20), 3 (92-102); Bridge and Demicco, Chapter 3*

## 4.1 Coarse-Grained Sediments

The large grains (>2 mm in diameter) in a coarse grained sedimentary rocks are commonly referred to as *clasts*. Conglomerates, like sandstones, often contain a matrix, only this matrix can be much coarser, since the pore spaces between the clasts can be much coarser. It is important to note the nature of this matrix and particularly important to note whether the clasts are rounded or not and whether the rock is grain-supported (and if so, by the clasts or by sand, e.g.). If you can make these observations about a coarse-grained rock, you can often make some useful, first-order interpretations about the depositional process. Hence, the first order of business in describing a coarse-grained rock is to define it based on the rounding versus angularity of clasts, degree to which it is clast-supported, and the nature of the matrix. Here are the main relevant terms.

- *Conglomerate*: generally, rounded clasts in grain-grain contact
- *Sandy conglomerate*: has a sandy matrix
- *Paraconglomerate*: grain-supported, but a muddy matrix
- *Breccia*: angular grains in grain-grain contact (the clasts have clearly not travelled far, if at all, in flowing water)
- *Diamictite*: Coarse grains (rounded or angular) 'floating' in a fine-grained (usually muddy) matrix

Next you should describe the conglomerate based on the nature of the clasts. Arey the quartzose, for example (e.g., a *quartz pebble conglomerate*)? Are the clasts intraformational, flat (tabular), oriented (i.e. imbricated)? Are there multiple clast types, in which case it can be classified as *polymictic*. Obviously, you can make out the lithology of clasts in a conglomerate, which is automatically an excellent indicator of where the clasts came from.

### 4.1.1 Genetic classification

- Epiclastic
- Volcanic
- Cataclastic
- Solution
- Impact Breccia

## 4.2 Sandstone and Wacke

A sandstone or *arenite* is a rock formed primarily of sand-sized grains in grain-to-grain contact. A wacke also contains sand-sized grains, but has a muddy matrix. By some definitions (as in Fig. 4.2.1), a wacke must contain at least 15% mud. The term *greywacke* is often used to refer to a texturally and compositionally immature wacke (that is, contains abundant, poorly sorted and rounded lithic and/or feldspar grains), usually inferred to have been deposited as a turbidite. A wacke is like a finer-grained equivalent of a diamictite, though the latter, by definition, is matrix-supported.

### 4.2.1 Quartz-Feldspar-Lithics (QFL) and matrix classification

The QFL classification (Fig. 4.2.1) is a practical scheme for classifying sands (arenites) and wackes consistently, and in a way that other geologists should understand. I recommend using this scheme for all field, hand and thin-section descriptions of sandstones and wackes. Note that there are many different versions of this classification scheme; the one here seems to strike a nice balance between detail and generality.

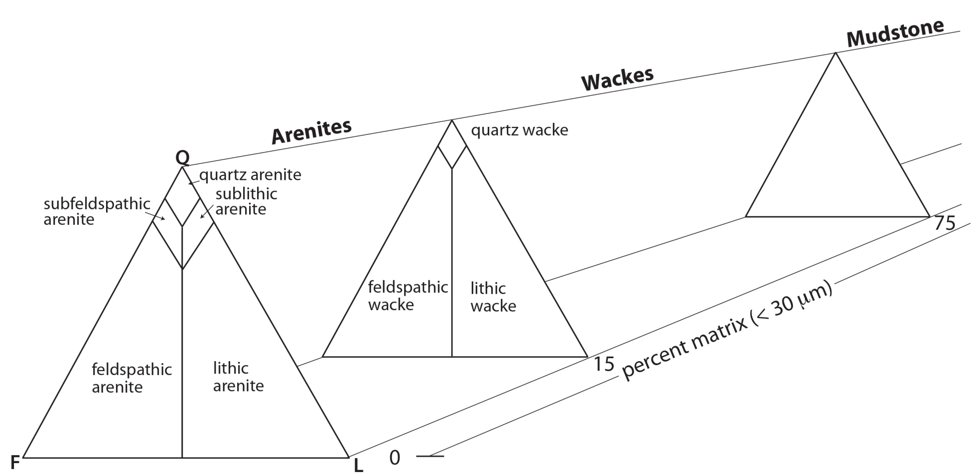



Figure 4.1: Sandstone/wacke/mudstone classification using the Quartz-Feldspar-Lithic Fragments (QFL) ternary diagram with an additional mud dimension. Modified from (Southard, 2007).

### 4.2.2 Sandstone maturity

Maturity is a rough measure of how much time sediments in a sandstone have spent being eroded and *reworked.* For example, fluvial sandstones are often somewhat immature, whereas beach sandstones tend to be very mature. For extra confusion, there are two separate ways of describing maturity, but generally, a *mature* sandstone is regarded to be a well-rounded, well-sorted quartz arenite.

- Compositional
- Textural

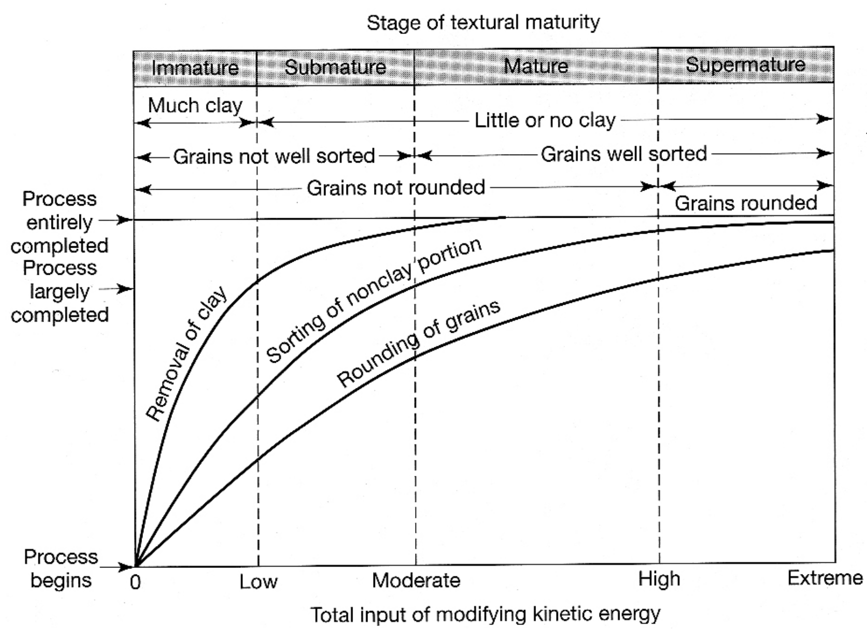



Figure 4.2: Graphic depicting the evolution of the textural maturity of a package of sediments. From Folk (1951).

## 4.3 Fine-Grained Sediments

Siltstone, shales, and mudstones, in combination, comprise up to 70% of all sedimentary rocks by some estimates. They all consist of some combination of silt and clay, although formally, none of them need to contain any actual clay minerals - they may simply be clay-sized quartz or feldspars. But normally, shales do contain clay minerals since these are common low-temperature alteration products of micas and feldspars. You will encounter significant disagreement on what to call rocks that contain clay and silt; for example, in some lexicons, siltstone does not really exist and what I would call a siltstone is referred to as a mudstone or shale. Below and in Table 4.3 is a reasonable way to classify them.

- siltstone: $> 2/3$ silt
- claystone: $> 2/3$ clay

- mudstone: intermediate between siltstone and claystone

Each of these rock types may be further classified based on whether it is laminated or bedded.

Table 4.1: Classification of siltstones, shales, and mudstones (>50% grains <0.062 mm). Modified from Boggs (2011).

| | % clay-size constituents | 0-32 | 33-65 | 66-100 |
|---|---|---|---|---|
| Non-indurated | Beds $>$ 10 mm | Bedded silt | Bedded mud | Bedded claymud |
| | Laminae $<$ 10 mm | Laminated silt | Laminated mud | Laminated claymud |
| Indurated | Beds $>$ 10 mm | Bedded siltstone | Mudstone | Claystone |
| | Laminae $<$ 10 mm | Laminated siltstone | Mudshale | Clayshale |
| metamorphosed | Degree (low) | Quartz argillite | Argillite | |
| | of | Quartz slate | Slate | |
| | metamorphism (high) | Phyllite and/or mica schist | | |

**Shale composition**

- Fine-grained quartz and feldspars
- Clays (e.g. kaolinite, smectite, illite, chlorite)
- authigenic minerals
  - sulfides
  - carbonates
- organic carbon

### 4.3.1 Shale chemistry

The chemistry of shales closely approximates the average chemistry of the upper crust. For this reason, you will often find the trace element concentrations (particularly the REE's) of sedimentary rocks normalized against average shale compositions, such as PAAS (post-Archean Australian shale) and NASC (North American shale composite).

## 4.4 Provenance

The source of sediments in siliciclastic sedimentary rocks is typically referred to as *provenance*. The provenance can be an important indication of tectonic setting and climate, and

Table 4.2: Concentration of the major elements in shale compared to the average upper crust. Adapted from Boggs (2011).

| element | Shale | Upper crust |
|---|---|---|
| Si (%) | 27.5 | 30 |
| Al (%) | 8.8 | 7.83 |
| Fe (%) | 4.72 | 4.17 |
| Mg (%) | 1.5 | 1.64 |
| Ca (%) | 1.6 | 3.15 |
| Na (%) | 0.59 | 2.54 |
| K (%) | 2.66 | 2.56 |
| C (%) | 1.2 | 0.023 |
| Ti (%) | 0.46 | 0.33 |
| P (%) | 0.07 | 0.086 |
| Mn (%) | 0.085 | 0.077 |

so is often an integral part of studying sequences of siliciclastic sedimentary rocks. With coarse-grained rocks, such as diamictites and conglomerates, it is relatively easy to determine the provenance, because you can see the actual rock types that were being eroded. However, in this case, they usually are not far-traveled. Finer-grained sediments are derived from farther afield, and muds, in particular, integrate a large source area. Whereas petrological methods are the traditional way to study provenance, a host of geochemical techniques are now widely used.

### 4.4.1 Framework mineralogy

The mineralogy of sedimentary rocks (mainly sandstones and wackestones) reveals a significant amount about the source area (*provenance*) of its sediments. For example, a sandstone rich in zircons would likely comprise sand grains that have been recycled many times, while a sand with olivine and pyrexene grains would represent a first generation sand from a reasonably nearby mafic source.

Often, sedimentologists are primarily concerned with determining the tectonic setting of the sedimentary basin they are studying. Dickinson et al. (1983) developed a widely used and still applicable model for discriminating tectonic environment based on QFL diagrams alone (Fig. 4.4.2). As one might suspect, a sandstone comprising mainly quartz likely represents an old, recycled source, whereas a sandstone with abundant lithic fragments might represent a nearby volcanic arc.

While still a common tool, this method of sedimentary petrology has been supplemented, and commonly replaced by geochemical techniques that include basic elemental geochemistry of sedimentary rocks (which can be quite effective in discriminating between mafic and felsic source terranes) and increasingly sophisticated isotopic techniques.

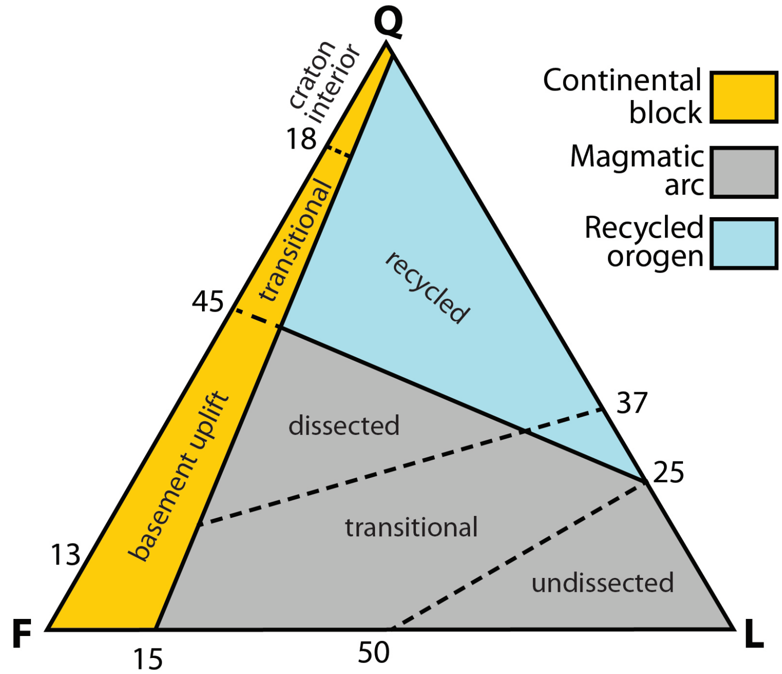



Figure 4.3: The classic Quartz-Feldspar-Lithic (QFL) ternary diagram to determine tectonic provenance, from Dickinson et al. (1983).

### 4.4.2 Geochemical fingerprinting

#### Elemental ratios

A practical and reasonably useful means of tracking source terranes in sedimentary rocks is through elemental ratios. Certain elements, namely Al, La, Sc, Zr, and Ba are resistant to alteration during erosion and after deposition. For this reason, they all correlate strongly with one another and can be used to track detrital input in rocks. Because La and Sc behave differently during magmatic differentiation (La is an incompatible element, and Sc is compatible), they are fractionated between mafic and felsic igneous rocks. Consequently, La/Sc ratios can be used to discriminate between mafic versus felsic source regions, with higher ratios indicating more felsic derivation, and lower ratios a more mafic source (McLennan, 2003). Zr/Sc and Th/Sc can be applied much the same way. Zirconium, in addition to being incompatible, is also highly conserved during sedimentary recycling because it is mainly hosted in the highly resistant mineral zircon (see below). Consequently, significant Zr enrichment in sandstones reflects sediment that has passed through many erosional and depositional cycles and the concentration of Zr.

#### Chemical index of alteration

Chemical weathering of the continental crust preferentially removes labile elements such as Ca, Na, and K from soils and concentrates other less mobile elements, such as Al and Ti. Thus, measuring the relative concentrations of these elements in sediments is a means of evaluating the degree of chemical weathering of source terrains, and by inference, paleoclimate. The chemical index of alteration (Nesbitt and Young, 1982) is defined as

$$CIA = 100 \times \frac{Al_2O_3}{Al_2O_3 + CaO + Na_2O + K_2O} \tag{4.1}$$

where CaO excludes the carbonate fraction of the rock. The CIA of a specific lithology (ideally mudstone) is expected to be highest in warm, humid climates, when labile cations are lost to solution, and lowest during cool, dry climates when they are retained in soil profiles until removed mechanically (Nesbitt and Young, 1982, 1996).

**Isotopic methods**

- Zircon geochronology (U-Pb)
- Pb isotopes
- Hf and O isotopes of zircons
- Nd isotopes

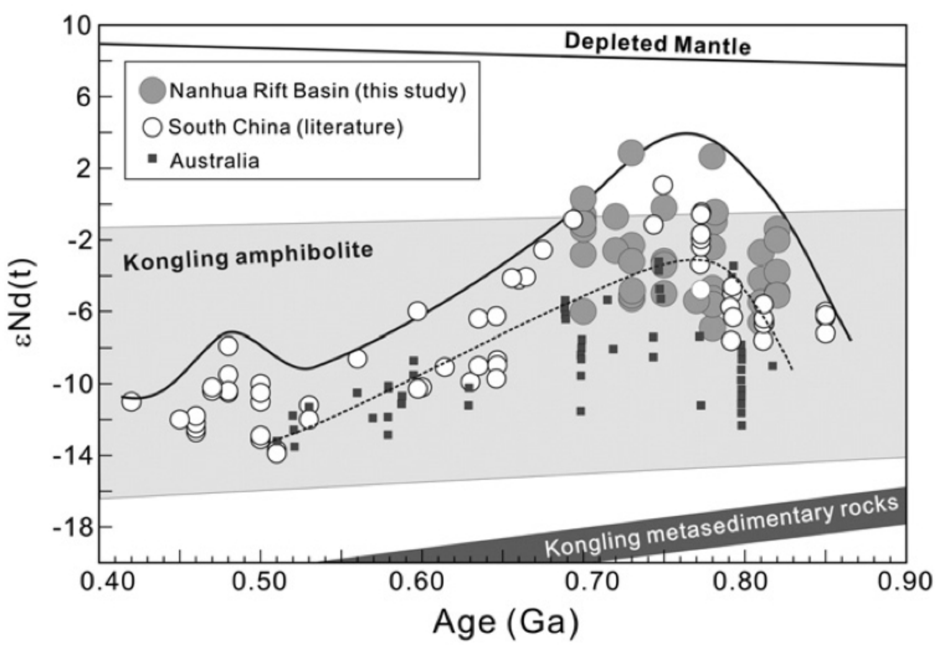



Figure 4.4: Plot of epsilon Nd values versus stratigraphic age from Neoproterozoic sediments in South China (Wang et al., 2011). The positive hump reflects a dominantly mafic provenance around 750 million years ago.

# Chapter 5

# Carbonates

Additional Reading: Tucker, Chapter 4 (110–150); Boggs, Chapter 10 (159–195); Tucker, M.E. and Wright, V.P., 1990. Carbonate Sedimentology, Blackwell (Oxford), 482p.

## 5.1 Introduction

Carbonate rocks include *limestones* and *dolostones*, which together make up about a quarter of all sedimentary rocks. They are also important petroleum reservoirs, accounting for up to 40% of all reservoirs. Carbonates account for 90% of the reservoirs in the Gulf area, where most of these carbonates were deposited in the tropical Paleotethys an Tethys seaways.

Carbonates are considered to be *chemically precipitated*, insofar as the are generated from ions in solution (mainly in seawater). However, since the early Phanerozoic, most carbonate sediments have also been biochemically precipitated by marine invertebrates. For the most part, our discussion of carbonates is of limestones, since most dolostones, although abundant in the sedimentary record, formed through diagenetic alteration of limestones. As you will discover, with the exception of reefal carbonates and other *bound* carbonates, such as stromatolites, carbonates are deposited much like siliciclastic sedimentary rocks.

## 5.2 The Mineralogy and Chemistry of Carbonates

Calcium carbonate ($CaCO_3$) is the dominant carbonate mineral, but there are many other sedimentologically important carbonate minerals. All major carbonate minerals, with the exception of aragonite, are rhombohedral (Table 5.1). The three principal cations in carbonates are $Ca^{2+}$ (0.99 $\AA$), $Mg^{2+}$ (0.66 A$\AA$), $Fe^{2+}$ (0.74$\AA$). Although aragonite is abundant in modern sediments (see below), it is not in the ancient sedimentary record. The reverse is true of dolomite. Because $Fe^{2+}$ easily replaces $Mg^{2+}$ in the dolomite crystal lattice, dolomite commonly contains some iron. This explains why dolostones commonly weather tan or brown, whereas limestones usually weather grey.

There are also a number of major and trace elements that commonly occur in carbonates In the modern ocean, primary carbonates include aragonites and low- and high-magnesian calcites. *High-magnesian calcites* have >4% $Mg^{2+}$, where the $Mg^{2+}$ substitutes for $Ca^{2+}$.

Table 5.1: The principal sedimentary carbonate minerals

| **Mineral** | **Formula** | **Crystal system** |
|---|---|---|
| Calcite | $CaCO_3$ | Rhombohedral |
| Aragonite | $CaCO_3$ | Orthorhombic |
| Dolomite | $CaMg(CaCO_3)_2$ | Rhombohedral |
| Siderite | $FeCO_3$ | Rhombohedral |
| Ankerite | $CaFe(CaCO_3)_2$ | Rhombohedral |
| Magnesiste | $MgCaCO_3$ | Rhombohedral |
| Rhodocrosite | $MnCO_3$ | Rhombohedral |

Table 5.2: Elements and compounds that substitute into carbonates

| **cations** | **cations** | **anions** |
|---|---|---|
| Sr | Mg | $SO_4^{-2}$ |
| Mn | Fe | $B(OH)_4^-$ |
| Co | Zn | |
| Cu | Na | |
| Li | K | |
| Rb | Ba | |

However, this is distinct from dolomite, which is a different mineral altogether.

Aragonite is a metastable polymorph of $CaCO_3$, with a distinct crystal structure (Table 5.1). It is precipitated by many organisms and is the preferred primary seawater $CaCO_3$ precipitate in the modern ocean. But this was not always the case through Earth history. Rather, the preferred primary carbonate mineral has varied throughout the Phanerozoic, most likely as a function of variable Mg/Ca ratios in seawater (Stanley and Hardie, 1998). It has been argued that the variation in Mg/Ca in seawater is a result of fluctuations in the average rate of mid-ocean spreading, but this is likely not the only mechanism.

High Mg concentrations tend to impede nucleation of low-Mg calcite, and therefore encourages precipitation of high-Mg calcite and aragonite, or even *vaterite* in some case. These are *metastable polymorphs* of carbonate, and these follow the *Ostwald step rule*, which states that where a solution is supersaturated with respect to metastable minerals, these will precipitate first then subsequently morph into more stable phases. In the case of carbonates, aragonite and high-Mg calcite are, they tend to convert to low-Mg calcites during burial and diagenesis and/or into dolomite.

If one assumes that carbonates are precipitated in equilibrium with seawater, then the chemistry of carbonates may be used as a proxy for seawater chemistry. However, as just stated, the original minerals tend to change polymorphs after precipitation, and each polymorph has different affinities for different trace elements and may experience different isotopic fractionations (for example in O and C). Nevertheless, carbonate minerals are one of the main proxies we have for ancient seawater composition, and so are a popular target for study.

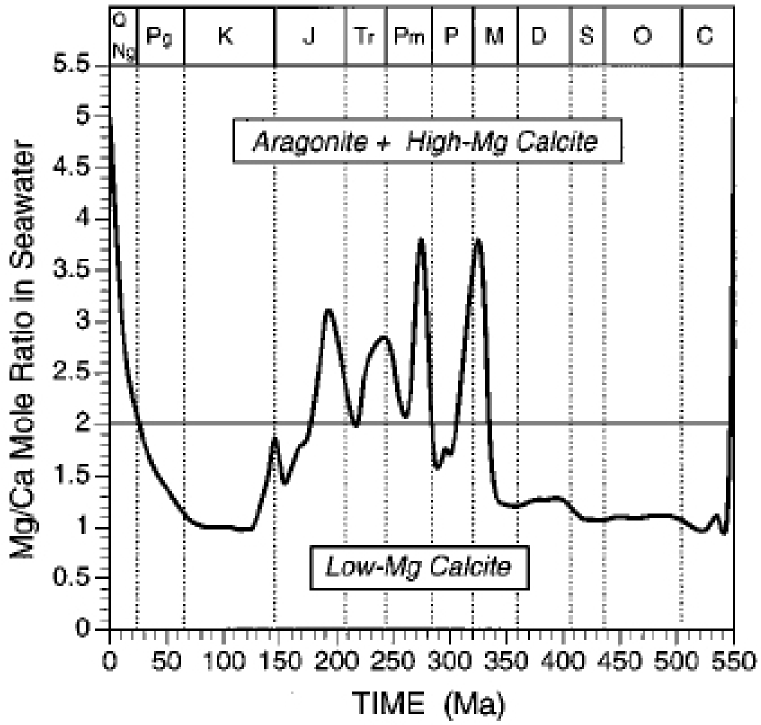



Figure 5.1: Aragonite seas of the Phanerozoic. From Stanley and Hardie (1998).

## 5.3 Controls on Carbonate Precipitation

The dissolution and precipitation of $CaCO_3$ is controlled chiefly by pH, which in the oceans and most lakes, is controlled mainly by $CO_2$ concentrations.
The easiest way to envision the chemistry of carbonate precipitation is that for every mole of $CaCO_3$ precipitated, a mole of $CO_2$ is released

- $Ca^{2+} + 2HCO_3^- \leftrightarrow CaCO_3 + CO_2 + H_2O$
- This explains why $CaCO_3$ solubility *decreases* with increasing temperature
- Also explains why precipitating limestones is not a solution to global warming

Two principal inorganic mechanisms of removing $CO_2$ from the system drive carbonate precipitation:

- increasing temperature (decreases solubility)
- decreasing pressure
  - wave agitation of surface waters
  - upwelling

Saturation state, $\Omega$ of calcium carbonate is defined as

Table 5.3: Some key geochemical proxies for seawater in carbonates

| **system/ratio** | **significance** |
|---|---|
| $^{87}Sr/^{86}Sr$ | continental weathering vs. hydrothermal inputs |
| $\delta^{13}C$ | carbon cycling/perturbations |
| $\delta^{18}O$ | temperature and ice volume |
| $\delta^{34}S$ | sulfur cycling/perturbations |
| $\delta^{11}B$ | pH |
| $\delta^{44}Ca$ | source of weathering input |
| Sr/Ca | temperature |
| Mg/Ca | temperature |
| Cd/Ca | biological productivity |

$$\Omega = \frac{[Ca^{2+}] \times [CO_3^{2-}]}{K^*_{SP}} \tag{5.1}$$

where $K^*_{SP}$ is the stoichiometric solubility product of calcium carbonate (carbonate or aragonite). In the modern surface ocean, the saturation state of aragonite $\Omega_a$ is about 2, and calcite ($\Omega_c$) about 3.

In the surface ocean, calcite is the thermodynamically stable calcium carbonate mineral. However, with few exceptions, calcium carbonate precipitates as aragonite. This due to a variety of factors, which include *inhibitors* that impede calcite nucleation and growth and kinetics. Magnesium and sulfate ions both inhibit calcite precipitation more than they inhibit aragonite. Dolomite, which simply does not precipitate widely in normal seawater, is highly supersaturated (by a factor of 10 more than calcite and aragonite). One likely important reason is the kinetics of dolomite crystal growth. Dolomite is a highly ordered mineral, and hence requires time to crystallize properly. Aragonite and high magnesium calcite are less picky.

Organisms have less trouble than inorganic precipitates when it comes to precipitating shells and skeletons. For this reason, calcium carbonate-secreting organisms can live in cold waters at high latitudes, where the saturation state is much lower. Despite the competitive advantage organisms have, it is clear that inorganic precipitation in the oceans does occur.

- *Whitings* are clouds of water turned milky because of suspended carbonate mud. Whereas many marine geologists have argued that these muds originiate from the breakdown of older carbonate particles, such as the disintegration of calcifying green algae, geochemical evidence does point to a primary precipiate origin for at least a component of the carbonate mud in whitings (Shinn et al., 1989).

- *ooids* form in highly agitated, supersaturated seawater, either through direct precipitation of aragonite needles radially from the outer surface of carbonate grains or the accumulation of already precipitated aragonite needles tangential to the outer surface of the grain (Fig. 5.6.1).

- Direct precipitation of aragonite cements on the seafloor (common during certain intervals of Earth history, reflects extreme supersaturation with respect to $CaCO_3$).

Organisms play a key role in mediating $CaCO_3$ precipitation

- Direct precipitation to form skeletons, shells, or other hard parts
  - Aragonite: many molluscs, some corals, calcareous green algae
  - High-Mg calcite: crinoids, benthic foraminifera, red algae
  - Low-Mg calcite: planktonic foraminifera, coccoliths, brachiopods
- Photosynthetic uptake of $CO_2$
  - mainly by cyanobacteria and small phytoplankton
- Bacterial mediation
  - export of $Ca^{2+}$ from cell, coupled to uptake of $CO_2$ or $HCO_3^-$
  - results in calcitic encrustation outside cell walls
- Bacterial decay of organic matter

## 5.4 Carbonate Dissolution in the Deep Ocean

Because the surface ocean is supersaturated with respect to carbonate, once carbonate precipitation occurs, the carbonate in the surface ocean will remain in tact. However, much of the carbonate that forms in the surface ocean is in the form of fine *tests* which form in the open ocean. Carbonate saturation in the deep ocean decreases as the result of

- increased pressure
- decreased temperature
- remineralization of organic matter (release of $CO_2$)

At a depth known as the *lysocline* (approximately where saturation state reaches 1), the rate of dissolution of carbonate increases sharply. This depth is usually several thousands of meters in the tropical oceans. Below the *carbonate compensation depth* (or *CCD*), as it is commonly abbreviated), carbonate completely dissolves. The depth of the CCD is around 4-5km and varies somewhat between the ocean basins. In the Pacific basin, it is around 4.5 km, while in the Atlantic, it is closer to 5 km. This difference between the ocean basins owes largely to the age of the deep water masses: the older the water mass, the more dissolved $CO_2$ it has accumulated from decaying organic matter. The CCD also varies with latitude: it is shallower at the high latitudes than in the low latitudes. It has varied in depth considerably over Earth history. For example, Zachos et al. (2005) demonstrated a significant shoaling of the CCD during the *Paleocene-Eocene Thermal Maximum*, presumably due to acidification of the oceans owing to large addition of $CO_2$ to the ocean-atmosphere system.

## 5.5 Carbonate Accumulations

We will cover carbonate depositional environments in a subsequent chapter. For now, suffice it to say that there are three principle types of carbonate sediment accumulation:

- carbonate oozes
- organic carbonate build-ups (reefs)
- platformal carbonate sands and muds

Carbonate oozes occur exclusively in the deep ocean, but above the CCD. *Chalk* is the lithified equivalent of a carbonate ooze. Although Cretaceous chalks are relatively widespread in northwestern Europe, carbonate oozes are generally not well represented in the stratigraphic record because they typically rest on ocean crust, which tends to be subduct. Indeed, for this reason the deep sea is only a temporary storage area for carbonate minerals. The principal denizens of carbonate oozes are the tiny shells (tests) of *foraminifera*, *coccoliths*, and tiny gastropods (*terapods* and *heteropods*).

Carbonate reefs, sands, and muds accumulate on *carbonate platforms*, which are tectonically stable, shallow water environments where carbonate sedimentation prevails. The three fundamental requirements for carbonate platforms to develop are

- Shallow waters
- Warm waters
- Lack of siliciclastic input

## 5.6 The Constituents of Carbonates

Calcite in ancient carbonates comprise four main components

- Skeletons and shells (biogical)
- Carbonate grains (silt-size or larger)
- Microcrystalline calcite (*micrite*) - carbonate mud
- *Sparry calcite* - coarse-grained calcite (recrystallized)
- *Microbial laminites* (e.g. stromatolites)

### 5.6.1 Carbonate grains

To a large extent, carbonates consist of detrital grains, just like siliciclastic sediments. *Allochem* are carbonate grains that have been transported, for example by *traction currents* or gravity flow deposits. *Autochems*, such as corals and stromatolites, are precipitated and deposited in place.

- Carbonate clasts— fragments of previously deposited carbonates

  - *extraclasts*—outside the depositional basin
  - *intraclasts*—nearby, semi-consolidated sediments

- Skeletal particles—whole fossils or fragments of fossils
- Ooids—coated, concentrically laminated carbonate grains with a distinct nucleus
  - pisoids (>2mm in diameter)
- Other coated grains
- Peloids—grains composed of micrite and containing no internal structure
  - fecal pellets - small, oval, rounded
  - rounded mud clasts
- Aggregate grains are individual grains cemented together to form a larger, composite grain
  - grapestones
  - lumpstones
- Oncoids are grains with irregular, overlapping microbial coatings

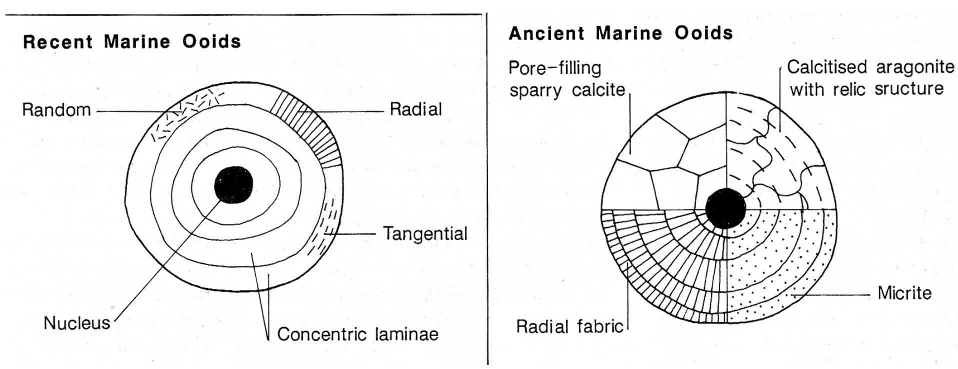



Figure 5.2: Ooid fabrics. From Boggs (2011).

### 5.6.2 Microcrystalline carbonate

Very fine-grained carbonate mud, or *micrite* is a common component of ancient carbonates. However, there is significant confusion and ambiguity regarding this term. Micrite is generally accepted to constitute the fine-grained matrix carbonate rocks as well as the fine-grained component of individual carbonate grains. However, micrite is also commonly distinguished from other fine-grained carbonate components:

- *calcisiltite*: silt-sized carbonate grains
- *microspar*: $> 4\mu$m and $< 30\ \mu$m carbonate component
- *micrite*: $< 4\mu$m

The term *calcilutite* refers to a mixture of the finer and coarser grained components (the equivalent of 'mud' and, in fact, one way in which people define micrite!).

In modern environments, primary micrite mud is found in the form of aragonite needles, 0.001-0.005 mm long that precipitate directly in lagoons. However, this rather restricted source of mud cannot account for all of the micrite in the sedimentary record. It is commonly thought that micrite mud is largely derived from the breakdown of disintegration of calcareous algae such as *Halimeda* and corraline algae (Tucker and Wright, 1990) and *epibionts*, such as sea grass (Flügel, 2004).

It is likely that much ancient carbonate mud was also derived from the breakdown of larger carbonate components and grains, such as shells, stromatolites, and microbialaminite. However, much remains unanswered about the source of carbonate mud in the geological record (Knoll and Swett, 1990).

### 5.6.3 Sparry calcites

Sparry calcite or *sparite* consists of visible crystals of calcite. As such, it is less opaque than micrite. It commonly forms as a cement or by replacement of micrite or carbonate grains.

## 5.7 Carbonate Classification

The textural classification of carbonates is based principally on the relative abundance of the different constituents of carbonates. The inimitable Bob Folk introduced the first modern classification scheme for carbonates (Fig: 5.3) which recognized three fundamentally different components in carbonates: micrite, spar, and*allochems*, which includes the carbonate grains discussed in a previous section (literally, chemical precipitates that come from elsewhere).

Many other classification schemes are used. In fact, it seems as if every carbonate sedimentologist has developed her own scheme, base on the rocks or sediments she works on. However, one of the most widely applied approaches is that of the Dunham scheme (Fig. 5.4), which emphasizes the importance of grains versus carbonate mud. My own preference is to combine the two, such that I use the Dunham scheme, but identify the types of grains where possible. For example, an *ooid packstone* or *intraclast grainstone.*

- other carbonate lithologies
  - *coquina*—mechanically sorted and abraded, grain-supported, poorly-cemented carbonate consisting of fossil debris
  - *chalk*—soft, fine, limestone composed of calcite shells of microorganisms
  - *marl*—mixture of clay and calcium carbonate

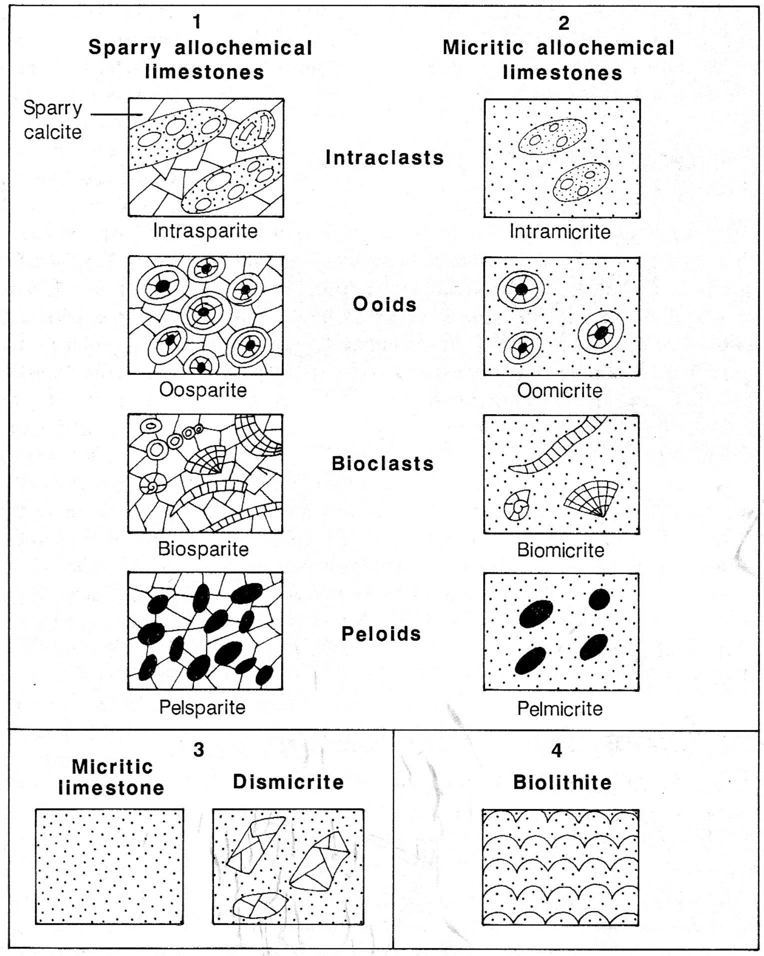



Figure 5.3: The Folk (1959) carbonate classification scheme.

### 5.7.1 Stromatolites

Grotzinger and Knoll (1999) defined *stromatolites* as "attached, lithifed sedimentary growth structures, accretionary away from a point or limited surface of initiation." This definition is rather generic, and is specifically written to include both inorganic and biogenic structures with a definite, non-planar laminated morphology with an element of topographic relief. Most definitions of stromatolite include a microbial component, and might begin something like "organically grown, laminated structures..." This latter style of definition for stromatolites is based on the fact that most stromatolites in the geological record are organic, or at least, organically mediated. That is, they grow as a result of the interaction between *microbial mats* (microbial veneers composed mainly of cyanobacteria) on the seabed and ambient depositional processes.

This definition of stromatolites also allows it to encompass planar-laminated, microbial textures, often referred to as *cryptalgal* laminations, implying that microbial mats were involved in the lamination, but did not manage to get enough of a foothold to develop any positive relief. These nearly flat-laminated microbial laminations are now commonly

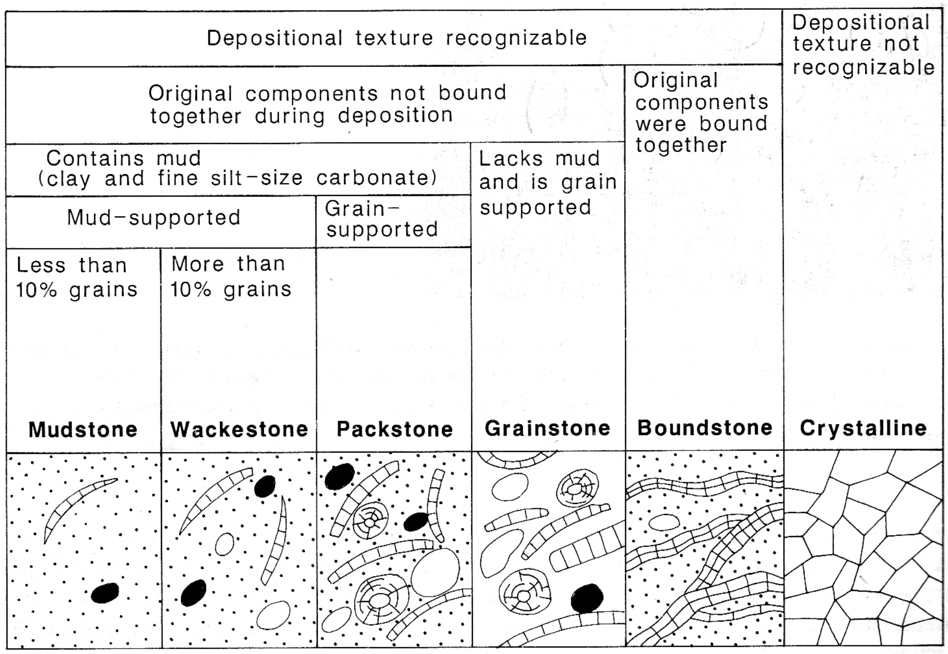



Figure 5.4: The Dunham (1962). classification scheme

Table 5.4: The principal carbonate lithologies

| name | description |
|---|---|
| mudstone | >90% micrite |
| wackestone | mud-supported, >10% grains |
| packstone | grain-supported, micrite matrix |
| grainstone | grain-supported, sparite cement |
| rudstone | coarse, grain-supported |
| boundstone | original components bound together |
| crystalline | recrystallized, original texture obliterated |

called *microbialaminites*. They can be hard to differentiate from laminated micrites, but two features can usually help. First, all microbial laminations tend to be irregular and vary in thickness over short length scales. Second, microbial laminations often contain structures called *fenestrae*, which are bedding-parallel cavities that form as the result of gas generation by microbial mats or the subsequent decay of those mats. The preservation of fenestrae in ancient carbonates attests to the early lithification of the microbialaminites (Flügel, 2004).

Organic stromatolites form by one of two main processes, or a combination of these two processes: trapping and binding of sediments (onto the sticky, mucilaginous surface of the microbial mat), and direct carbonate precipitation within the microbial mat. Stromatolites formed principally by the former processes tend to occur in high energy environments, where abundant sediments are swashed around and can stick onto the microbial-coated

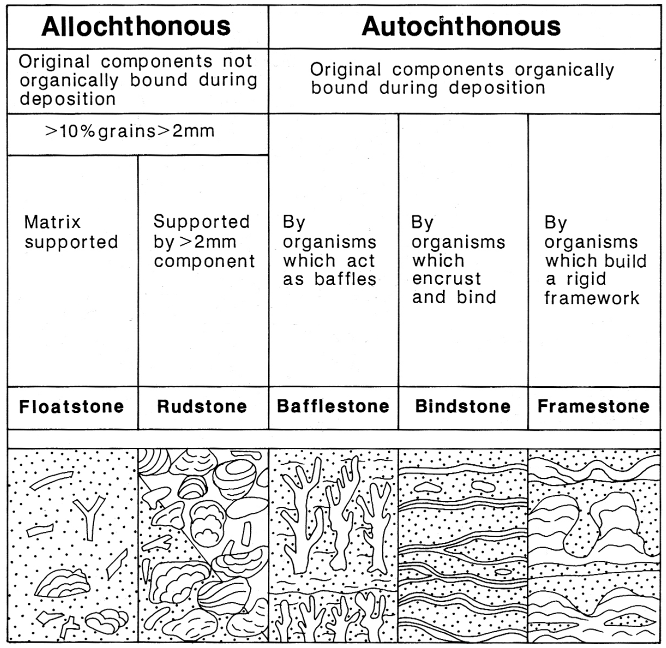



Figure 5.5: Classification scheme for reefal limestones. From Tucker and Wright (1990).

stromatolite structure. In the latter case, this carbonate may precipitate during active growth of the algae or cyanobacteria comprising the mat, through modification of ambient seawater chemistry, or during microbial decay of the mat after burial. Stromatolites often form spectacular shapes and structures, particularly in Proterozoic carbonates. Most stromatolites are recognized as forming columnar or bulbous protrusions from the seabed. They range in size from millimeters to meters in width or height. Some stromatolites have significant *synopiic relief*—that is, the total relief on a single lamination—implying that they stuck out well above the sea floor, while others, including in cases very large stromatolites, may have very low synoptic relief. A laterally continuous bed of stromatolites is referred to as a *biostrome*, whereas an isolated stromatolite mound bound in other sediments is termed a *bioherm*. Bioherms can occur within siliciclastic sediments, attesting to the fact that they do grow by direct precipitation of carbonate, at least in some instances.

Stromatolites take on a wide range of morphologies (Fig. 5.7.1). It is rather common practice to name a stromatolite morphotype or *morphospecies* in the same way fossils are named (with a genus and species name), implying that stromatolites are fossils. Indeed, stromatolites have been used as biostratigraphic tools, particularly in Russia and Australia. However, while organically-formed stromatolites are indeed a relict of bygone life, the use of stromatolites for biostratigraphy is highly controversial, as implied by the title of the Grotzinger and Knoll (1999) article: *Stromatolites in Precambrian Carbonates: Evolutionary Mileposts or Environmental Dipsticks?* The thesis of this paper, and the opinion held by many Precambrian sedimentologists, is that stromatolite morphology is in fact controlled more by their environment of deposition than by evolutionary biology. In any case, it is apparent that stromatolites form in a range of environments, and that certain features characterize stromatolites in each of those environments. However, at the same time, because environments have evolved through Earth history, it is also true that certain types of stromatolites are especially common during certain intervals in Earth history.

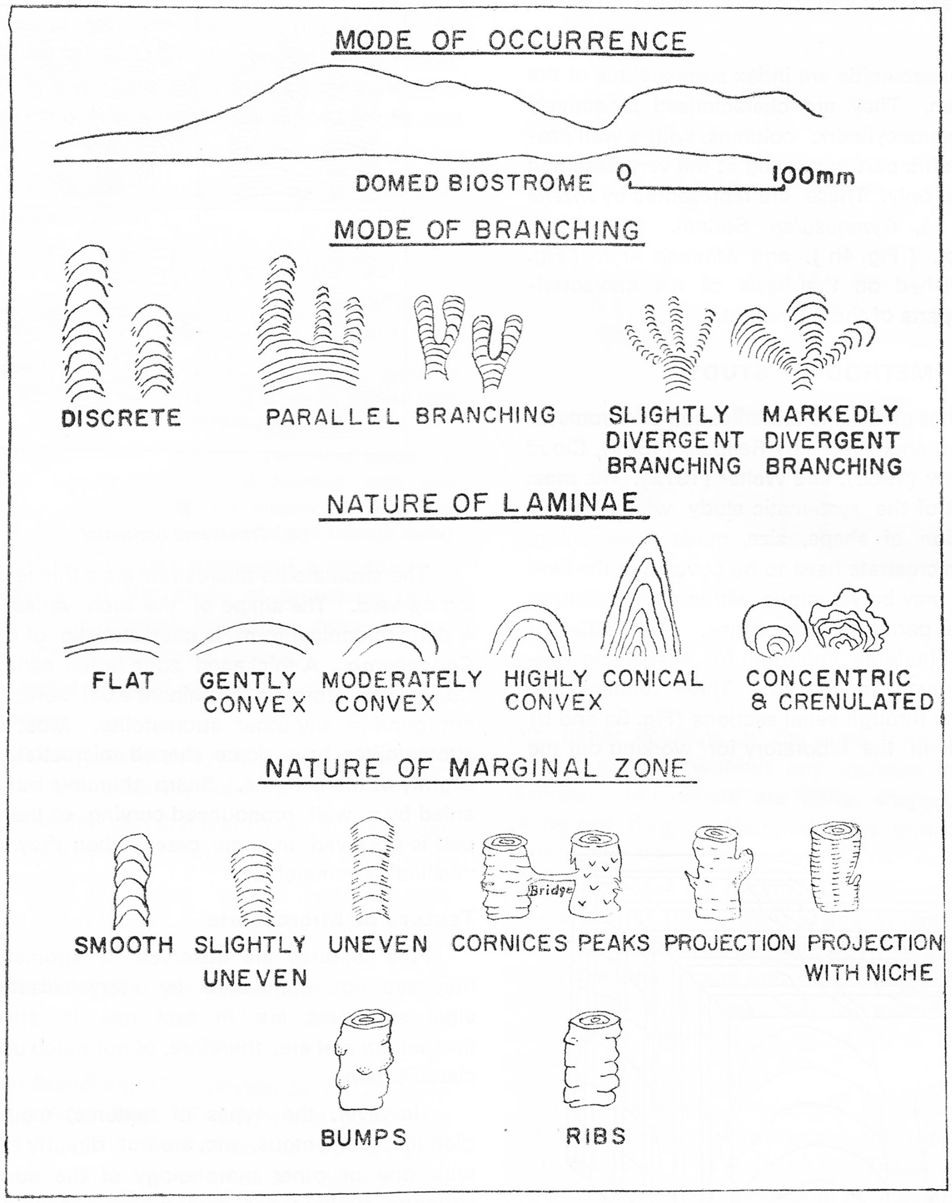



Figure 5.6: Descriptions of various morphologies common in stromatolites. From Walter (1976).

### 5.7.2 Paleokarst

*Karst* develops in terranes underlain by carbonate where groundwaters above the water table dissolve out cavities. Caves and sinkholes are common features of karstification. *Paleokarst* is ancient karst terrane, which is essentially just a subaerial unconformity developed on carbonates. They are recognized by the development of solution features, many of which follow joints (*grikes*) and other structural or sedimentary weaknesses, as well as *solution breccias*. Paleokarst surfaces are often highly irregular with cavities filled by discoloured breccias or other sediments.

### 5.7.3 Hardgrounds

*Hardgrounds* are surfaces or horizons that form as the consequence of synsedimentary cementation at or just below the sediment-water interface. Hardgrounds form where sedimentation rates are extremely low or nil, and commonly form at times of *maximum flooding*. Hardgrounds are typically encrusted by *sessile* benthic organisms that take advantage of the hard substrate.

# Chapter 6

# Other Precipitated Sedimentary Rocks

Additional Reading: Tucker, Chapters 5–8; Boggs, Chapter 7

## 6.1 Evaporites

*Evaporites* are precipitated (sedimentary) saltrocks that initially formed by the evaporative concentration of saline solutions. The formation of evaporites requires both accommodation space for them to accumulation and suitable hydrological conditions for brines to form and remain close to the earth surface. To be preserved in the sedimentary record, they must experience burial in an environment that is not exposed to sufficient undersaturated fluids to dissolve the salts (Warren, 2006).

Of the many evaporite minerals (Table 6.1), only halite, gypsum, anhydrite, and the common carbonate minerals are significant in evaporites that form from modified seawater. Typical evaporite minerals in continental environments are quite different, as described below. Generally, the occurrence of marine evaporites in a sedimentary sequence is a useful paleogeographic indicator, and a compilation of known, large evaporite occurrence through Earth history has shown that they cluster right where they should, in the arid sub-tropics (Evans, 2006).

Table 6.1: The principal sedimentary carbonate minerals

| **Mineral class** | **Mineral name** | **Chemical formula** | **Rock name** |
|---|---|---|---|
| Chlorides | Halite | $NaCl$ | Halite (rock salt) |
| | Sylvite | $KCl$ | |
| | Carnalite | $KMgCl_3 \bullet 6H_2O$ | |
| Sulfates | Anhydrite | $CaSO_4$ | Anhydrite |
| | Gypsum | $CaSO_4 \bullet 2H_2O$ | Gypsum |
| | Kieserite | $MgSO_4 \bullet 2H_2O$ | |
| Carbonates | Calcite | $CaCO_3$ | Limestone |
| | Magnesite | $MgCO_3$ | |
| | Dolomite | $CaMg(CO_3)_2$ | Dolomite |

Evaporating 1000 m of normal seawater yields approximately (and theoretically) 15m of evaporite minerals

- at ± 50%, Calcium carbonate (Mg/Ca increases)
- at ± 20%, Gypsum (Mg/Ca increases)
- ± Dolomite (due to high Mg/Ca ratios)
- ± 10%, Halite
- Mg, K salts

Of these, the chlorides should by far dominant the evaporite budget. However, in practice, this is not what we see in the sedimentary record, and this is because evaporites do not usually form by the simple evaporation of a fixed volume of water. Rather, there is usually some continuous inflow of seawater to the basin, supplying fresh ions for precipitation. At the same time, evaporation forms brines, which because they are more dense, may flow back out of the basin, either below inflowing seawater or as a diffuse flow through permeable sediments or bedrock below.

Non-marine evaporites are characterized by a different suite of evaporative minerals

- In general, more bicarbonate and Mg
- Less sulfate and chloride
- Many non-marine evaporites still contain gypsum, anhydrite, and chlorides, but also
  - hydrated Na carbonates
    - e.g. trona—$Na_3H(CO_3)_2{\bullet}2H_2O$
  - Na sulfates
    - e.g. glauberite—$Na_2CaSO_4$
  - hydrated Mg sulfates
    - e.g. epsomite—$MgSO_4{\bullet}7H_2O$

### 6.1.1 Primary and early syn-sedimentary evaporites

Carbonates are typically the first minerals to form during evaporative concentration of salts in solution. This concentration typically occurs within the margin part of a basin, which is where slopes are lowest and where environments and hence facies can change abruptly with minor changes in water depth (Warren, 2006). As a consequence, primary evaporite minerals are commonly interbedded with other facies, sometimes cyclically. They are also prone to exposure and desiccation. Tepee structures in microbialaminites are an excellent example.

### 6.1.2 Common evaporitic lithologies and textures

Gypsum and anhydrite are the most volumetrically important evaporite minerals

- Gypsum is the usual original evaporite mineral in marine environments. It may form in a variety of environments, from within sediments, creating small lathes, to directly on the sea bed, where it forms swallow-tail twinned *selenite* crystals. Gypsum may also precipitate out of the water column and settle to the seafloor, where it may form laminated gypsum deposits (in a quiet environment, such as a deep silled basin or a lagoon). Gypsum may also be reworked by bottom currents are gravity flow process.
- Mostly anhydrite in the geological record
- At high surface temperatures (>45°C) after burial
  - gypsum → anhydrite
  - 38% loss of volume
- Commonly associated with distorted textures due to mineral growth within the sediments (*enterolithic*)
- Principal anhydrite textures:
  - *nodular anhydrite*—irregular lumps in salt, carbonate, matrix
  - chickenwire anhydrite—elongated, coalescing polygons
  - laminated anhydrites—interbedded with dark, organic-rich laminae (often dolomitic)
  - gypsum casts

Halite forms analogous deposits to gypsum. Along the strand line and in sabkha settings, it may form *hoppers* of salt crystals that displace mud or sand. Where it nucleates directly on the seafloor, it generates a distinctive chevron structure. And like gypsum, it may also precipitate directly from seawater and settle to the basin floor forming laminated halite deposits.

Gypsum, halite, and other evaporite minerals are easily dissolved, particularly during meteoric diagenesis. One result is that evaporite-rich sediments often contain *collapse breccias*, which form *in situ*. Because dissolution of gypsum and anhydrite releases $Ca^{2+}$ to the pore waters, it often involves *calcitization* of dolomites (which, recall, are often associated with evaporites).

Gypsum is also commonly replaced by chert (a variety referred to as *length-slow chalcedony*) during early diagenesis. Where that chert replaces individual gypsum nodules, it is referred to as *cauliflower chert*.

### 6.1.3 Evaporite deposits

Evaporite deposits are widespread in the geological record and geographically, although most formed during the Phanerozoic in sub-tropical latitudes. Most of these are marine evaporites, meaning the source of the ions was seawater, even if evaporite mineral precipitation occurred from highly evaporation-modified waters. Many major evaporites were

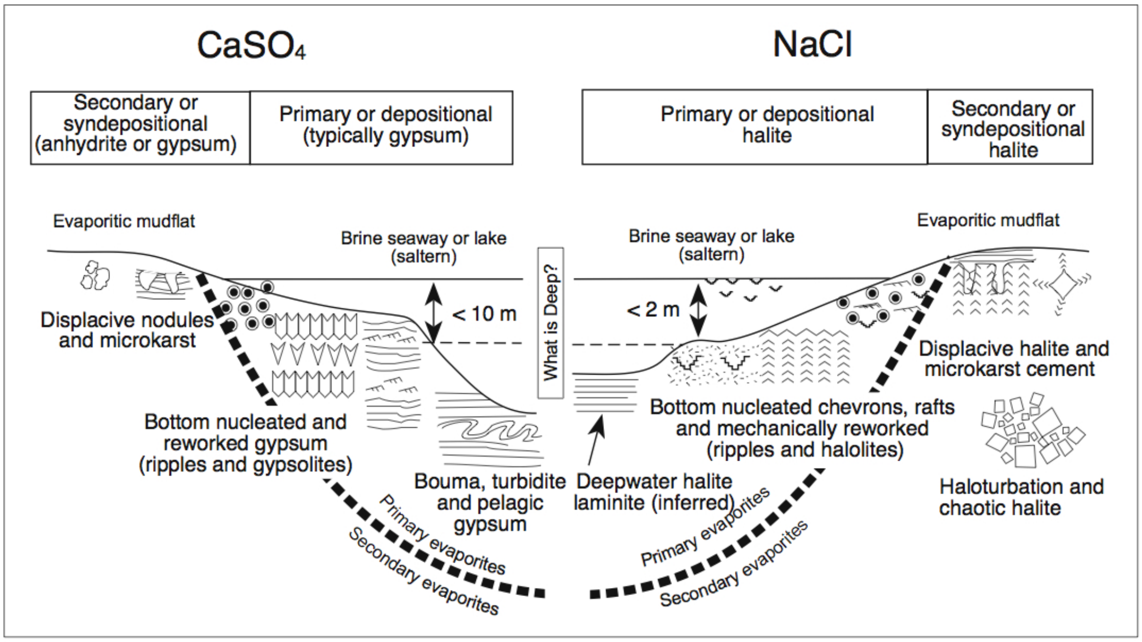



Figure 6.1: Primary and syndepositional evaporite textures and settings. From Warren (2006).

deposited during the Permian, including in the famous Zechstein basin in Germany and Deleware Basin in West Texas. Probably, this peak in evaporite deposition was the result of Pangea's low-latitude paleography, which provided many net evaporitic basins suitable for deposition of vast evaporites.

Some evaporites are very thick, meaning that they cannot form from the simple evaporation of a body of seawater. They also do not usually contain the whole suite of evaporite minerals one would expect from such a scenario. Hence, it is clear that most thick deposits occur within partially restricted basins that experience net evaporation. But this alone is insufficient to explain why they often contain no halite and rarely contain non-halite chloride salts. The explanation is presumably that dense brines flow out of the basin, either through direct outflow from the basin, or as diffuse flow through a permeable barrier (say, a barrier island). The greater the brine reflux (meaning, the more chloride exits the basin), the greater the tendency towards gypsum precipitation. Less reflux would favor halite precipitation.

Commonly, evaporite minerals and are associated with shallow water deposits, either siliciclastic or carbonates (almost always dolostone). These likely formed in a *sabkha*, which is low-lying supratidal area along a hot and dry coast. In this sort of setting, storms provide episodic seawater, which seeps into the sediments, then rises again during evaporation, during which time evaporite minerals are deposited. Hence evaporite minerals are commonly associated with desiccation cracks, microbialaminites, and tepee structures.

## 6.2 Evaporite Depositional Environments

### 6.2.1 Depositional environments of evaporites

Evaporites form in both subaerial and shallow subaqueous environments

- Subaerial
  - sabkhas
  - salt flats
- Subaqueous
  - *salinas*—saline, coastal lakes
  - lagoons
  - silled basins

### 6.2.2 Evolution of evaporites

Evaporites are by their very nature highly reactive, and as such most evaporite textures are secondary, related either to modification upon burial or uplift (Fig. 6.2.2). For example, calcium sulphates almost always begin as gypsum, but during diagenesis and burial, may either be replaced calcite or chert or converted to anhydrite, which entails dehydration and loss of volume. However, ghosts of original evaporites may be preserved within an overall sequence of altered or overprinted evaporite minerals. Like carbonate minerals, evaporites tend to experience the most destructive and pervasive alteration to original textures and chemistry either during early diagenesis or late during uplift (Warren, 2006).

A distinct feature of evaporites is that upon burial, they can flow in a ductile fashion. This flow can occur at any depth during burial, and widespread flow (halokinesis) drives salt tectonics.

## 6.3 Siliceous Sedimentary Rocks

Siliceous sedimentary rocks are fine-grained, dense, hard rocks composed of microcrystalline $SiO_2$ (usually 5-20 $\mu$m, even when recrystallized). They are a relatively common, but volumetrically minor component of the stratigraphic record. Due to long-term evolution of the marine Si cycle, controlled in large part during the Phanerozoic by silica-secreting organisms, the style and locus of chert deposition has shifted. Whereas in the Precambrian, sedimentary chert occurred mainly in banded iron formation (see below) and by early diagenetic replacement of shallow-water carbonates, chert now forms mainly in deep water environments through the accumulation of fine-grained silica tests.

- *Chert* refers to the various types of precipitated $SiO_2$ rocks
  - quartz
  - chalcedony is finely crystalline silica consisting of radiating fibers.

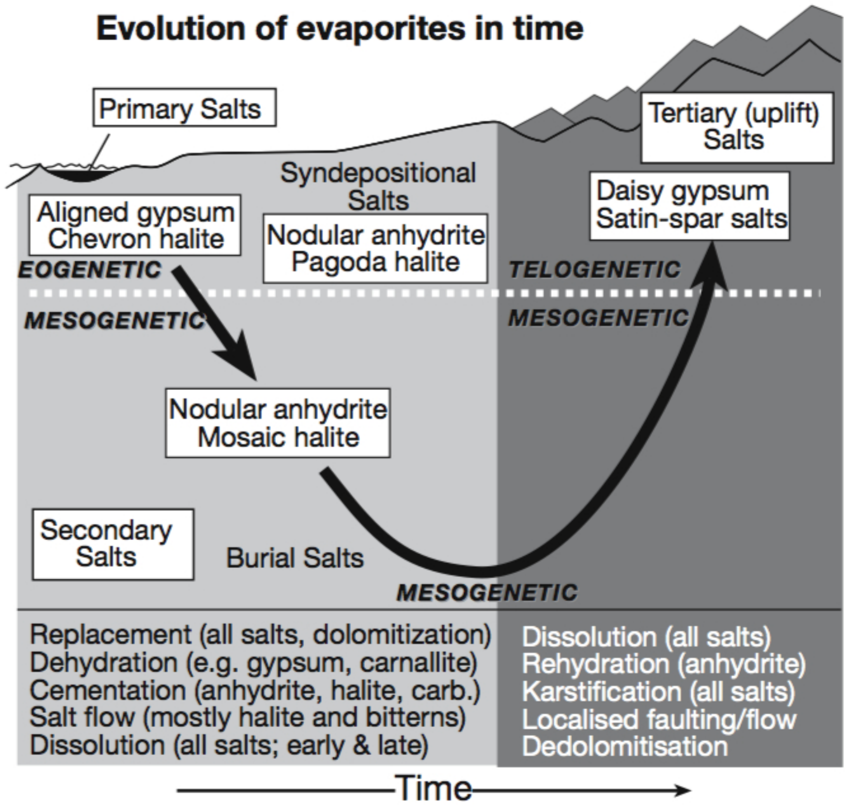



Figure 6.2: Evolution of evaporites from burial and early diagenesis through deep burial and uplift. From Warren (2006).

  - opal is an amorphous silica with water. It forms through the alteration of volcanic ash and, more importantly, is precipitated by siliceous organisms

- Composed mainly of microcrystalline quartz
- Tests of siliceous organisms made up of amorphous, Opal-A
- Several varieties of chert
  - jasper is chert that contains a few percent iron oxide, giving it a characteristically red color
  - porcelanite is dull, finely porous chert
  - sinter is porous silica that forms at hydrothermal vents and in hot springs

Flint is just the non-scientific term for chert, but is usually applied to dark grey chert that occurs as nodules or in beds.
Morphologically, chert can broadly be grouped into two categories

- Bedded cherts commonly form in deepwater environments where they avoid being otherwise diluted by detrital sediments or carbonate. The main source of the siliceous sediment are diatoms (which form diatomite), radiolaria (ribbon chert), and sponge spicules (spicularite).

- Nodular cherts are spheroidal or irregular lenses or layers of chert that form during diagenesis. They are common in shelf carbonates, and particularly important in the Proterozoic.

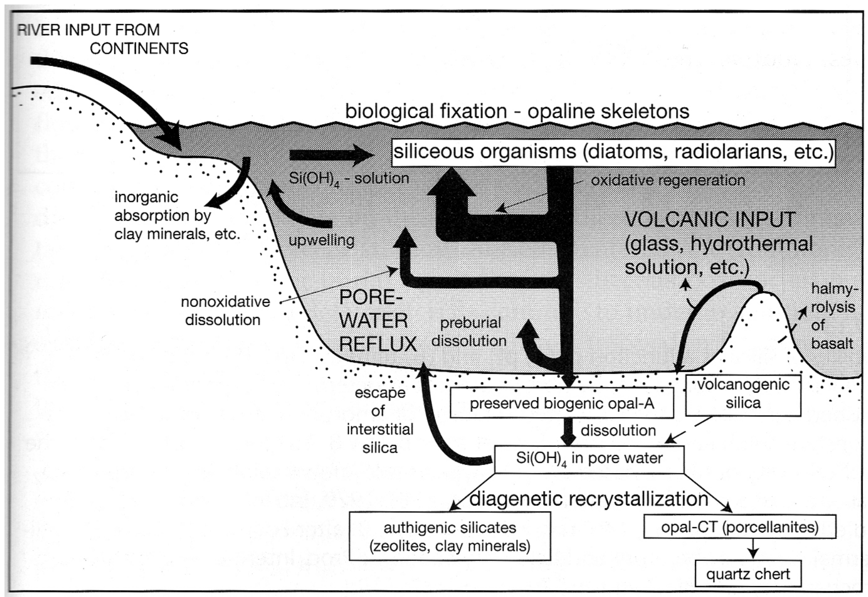



Figure 6.3: The marine silica cycle Boggs (2011).

- Silica sources
  - riverine
  - dissolution of siliceous tests
  - pore water flux
  - volcanic input
- Silica solubility and precipitation
  - solubility increases at high pH, high temperature

## 6.4 Iron-bearing rocks (iron formation)

Iron-rich rocks are those that have at least 15% iron by mass

- Ironstone
- Iron formation
  - banded-iron formation (BIF)
  - granular iron formation
- pyritic black-shales

Table 6.2: The iron-bearing minerals in iron-rich rocks.

| Mineral class | Mineral name | Chemical formula |
|---|---|---|
| Oxides | Goethite | $FeOOH$ |
| | Hematite | $Fe_2O_3$ |
| | Magnetite | $Fe_3O_4$ |
| Silicates | Chamosite | $3(Fe, Mg)O \bullet (Al, Fe)_2O_3 \bullet 2SiO_2 \bullet nH_2$ |
| | Greenalite | $FeSiO_3 \bullet nH_2O$ |
| | Glauconite | $3(Fe, Mg)O \bullet (Al, Fe)_2O_3 \bullet 5SiO_2 \bullet 3H_2$ |
| | Minnesotaite (iron talc) | $(OH)_2(Fe,Mg)_3Si_4O_{10}$ |
| Sulfides | Pyrite | $FeS_2$ |
| | Marcasite | $FeS_2$ |
| Carbonates | Siderite | $FeCO_3$ |
| | Ankerite | $Ca(Fe,Mg)(CO_3)_2$ |

- manganese crusts-nodules

Banded-Iron Formation (BIF) is the most important of the iron-rich sedimentary minerals.

Proterozoic BIFs subdivided into two main types:

- Algoma type
  - thin lenses
  - common in Archaean greenstone belts
  - hydrothermal
- Superior type
  - thick, extensive
  - Archaean - Palaeoproterozoic, again in Neoproterozoic

## 6.5 Phosphorites

Phosphorites are economically important accumulations of calcium phosphates. They are also important and intriguing rocks in the study of Earth history, for although P-rich rocks of many ages occur, they Sedimentary phosphorites contain at least 15% $P_2O_5$

- principal mineral is fluorapatite: $Ca_5(PO_4)_3F$
- contain a variety of other phosphate minerals
- a significant amount of carbonate substitutes for phosphate (francolite)
- the general term for all of these minerals is *collophane*
- Occurs in various forms
  - bedded

  - nodular
  - pebble beds
  - guano

Textures of phosphorites commonly resemble those of carbonates

- grains can occur as peloids, intraclasts, skeletal fragments, etc.
- bioclastic phosphorites composed of skeletal fragments and other biogenic phosphorite components (fish bones, sharks teeth, coprolites)
- nodular phosphorites comprise irregular nodules in organic-rich sediments
  - occur most commonly on upwelling margins
  - commonly diagenetic in origin (P derived from the decay of organic matter)
  - these nodules are sometimes reworked into pebble-bed phosphorites

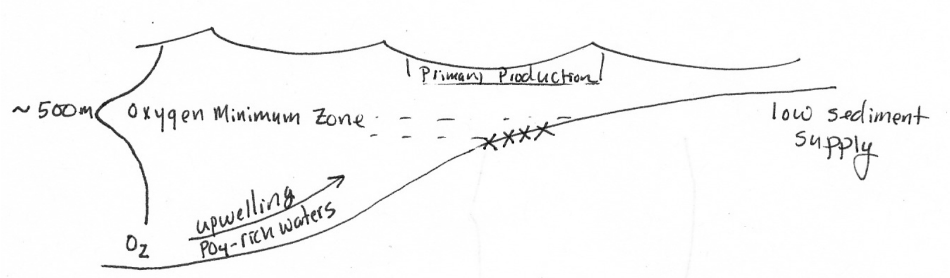



Figure 6.4: Schematic diagram showing the formation of phosphorites on upwelling margins where the oxygen minimum zone intersection the upper continental slope.

# Chapter 7

# Porosity and Diagenesis

Additional Reading: Tucker, Chapter 2 (50-63), Chapter 3 (97-109); Boggs, Chapter 5; Bridge and Demicco, Chapter 19)

## 7.1 Introduction

*Diagenesis* is effectively the process by which sediments are turned into rocks. It begins as soon as sediments are deposited and continues through deep burial. Diagenesis encompasses a broad range of physical, chemical, and biological processes, many of which are interrelated. Diagenesis effectively includes any alteration to sediments occurring after deposition and prior to metamorphism (although note that the 'boundary' between the two is fuzzy). The two important processes are compaction and lithification, which involve changes to texture, structure and mineralogy of sediments, and an increase in bulk density. However, both biological and biological breakdown and alteration of sedimentary organic matter are also important processes, particularly in organic-rich sediments. Many other post-depositional process also conspire to effect the final lithology, porosity, chemistry, and mineralogy of sedimentary rocks.

Unsurprisingly, diagenesis is heavily influenced by the local *geothermal gradient*, which controls the increase in temperature with burial depth. Diagenesis is particularly important in petroleum geology, because it controls the porosity and permeability of rocks, as well as alteration of organic matter and generation of hydrocarbons (which will be discussed in more detail in a later section).

Diagenesis involves many processes, many of which may act in concert at any given time

- Mechanical compaction
- Cementation
- Dissolution
  - pressure solution
- Isochemical alteration (that is, no change in chemistry). Recall *Ostwald's step rule*, which states that unstable polymorphs will gradually transform into stable polymorphs. Similarly, *Ostwald ripening* is the gradual increase in grain size following initial crystallization. This occurs almost universally during diagenesis.

  - *Recrystallization* is simply the dissolution and recrystallization of a mineral phase. The new crystals are typically larger.
  - *Polymorphism* is the conversion of one mineral polymorph into another, such as aragonite into low-Mg calcite.
  - *Neomorphism* is stepwise (gradual) recrystalization, which again usual entails coarsening of crystal size. However, *grain diminution* may also occur, particularly where grains such as peloids or ooids are *micritized*.

- Non-isochemical alteration
  - clay-mineral reactions
  - magnesian calcite replacement
  - dolomitization
  - chertification (or silicification)
  - redox reactions

- Biological reactions are driven by microbial metabolism.
- Thermal alteration of organic matter
- Hydrocarbon generation and expulsion

Because diagenesis encompasses such a broad range of processes under highly variable conditions, it is frequently separated into three (or more) processes.

- *Early diagenesis* occurs in the upper sediment column and largely biologically mediated.
- *Burial diagenesis* spans a broad range of P-T conditions and includes many different processes, including cementation, dissolution, clay transformation, and petroleum generation and expulsion.
- *Exhumation diagenesis* occurs during uplift of the sediment column and often involves meteoric waters.

## 7.2 Porosity and Permeability

*Porosity* is defined as the total void space in a sediment, whereas *permeability* is the ability of fluids to flow through a rock, which means it depends on the porosity and the interconnectedness of that porosity. Most sediments have an initial porosity of between 30 and 70%, with sands and gravels at the lower end of the range and mudrocks at the higher end. The porosity in carbonates is highly variable; whereas carbonate arenites may have initial porosities similar to a quartz arenite, reef carbonates tend to have very high porosities. And porosity is more prone to changes resulting from cementation and dissolution in carbonates.

The porosity in sediments or sedimentary rocks can have several origins:

- Primary porosity is the original porosity imparted on a sediment during deposition, influenced by

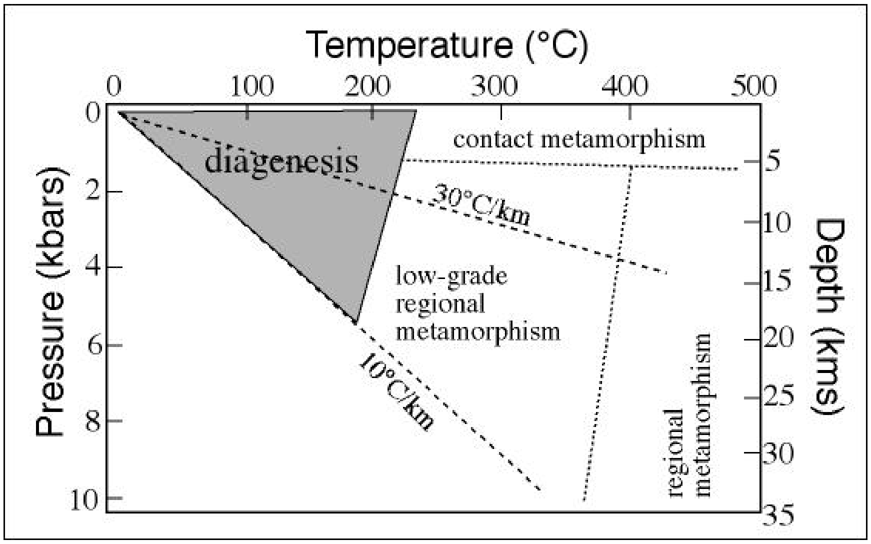



Figure 7.1: Pressure-Temperature diagram depicting the continuum from burial, through diagenesis, to low-grade metamorphism. Note dashed lines represent the typical range in geothermal gradients. After Boggs (2011).

  - sorting of grains
  - composition of grains, e.g. lithic vs. feldspar vs. quartz grains
  - degree of compaction

- Secondary porosity develops after deposition
  - Dissolution of grains and cements
  - Fracturing

The development of secondary dissolution porosity can significantly alter the chemistry and mineralogy of a rock (e.g. when feldspar or kaolinite are dissolved out of a sandstone at depth). In the case of carbonates, it is also particularly important in hydrocarbon reservoirs.

## 7.3 Compaction

Compaction is driven by hydrostatic and lithostatic pressure and results in dewatering, reorientation (increased packing) of grains, and a decrease in porosity. In a sedimentary basin

- *Hydrostatic pressure* is the pressure generated by the overlying water column (including in the pore spaces)
- *Lithostatic pressure* is the result of the weight of the sediment column
- Fluid pressures in pore spaces may exceed the hydrostatic pressure where pore spaces are not connected to the overlying water column; indeed high fluid pressures are characteristic of sedimentary basins

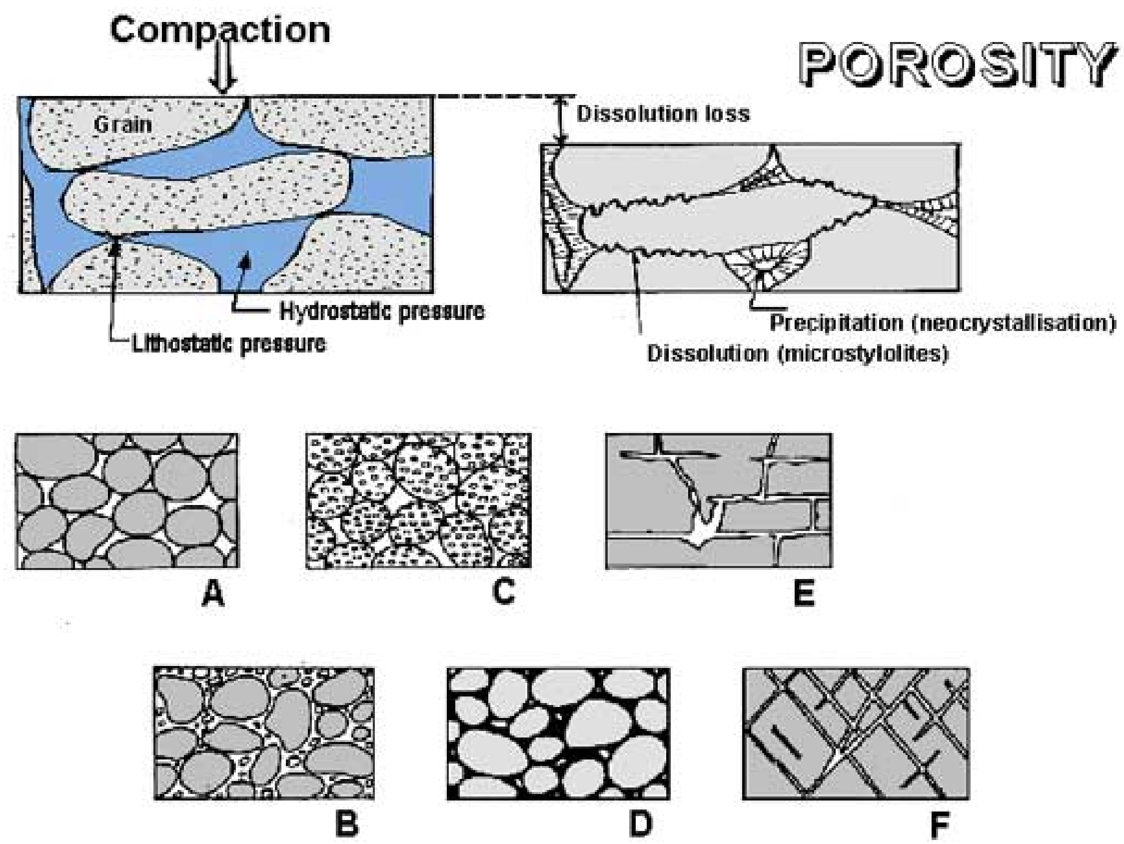



Figure 7.2: Primary and secondary controls on porosity. The top shows the process of pressure solution and subsequent cementation of primary porosity. A. Primary porosity in a well-sorted sediment. B. Primary porosity in a poorly-sorted sediment. C. Primary and secondary (micro-) porosity in a well-sorted sediment. D. Filling of primary porosity by secondary cement. E. Secondary porosity developed as a resulted of fracturing and dissolution.

Increases of pressure may have a variety of effects on sediments:

- May lead to deformation due to high overburden pressures (e.g., salt domes and mud diapirs)
- Fracturing occurs in week grains
- Indentation of grains may occur between grains of different strength
  - Concave-convex grain boundaries
- Pressure solution (chemical compaction) occurs between grains of similar strength
  - Sutured grain boundaries
  - Styolites

As a general rule, increased burial and compaction will lead to a progression from grain-grain, to linear, and then sutured grain contacts. The degree of physical compaction is

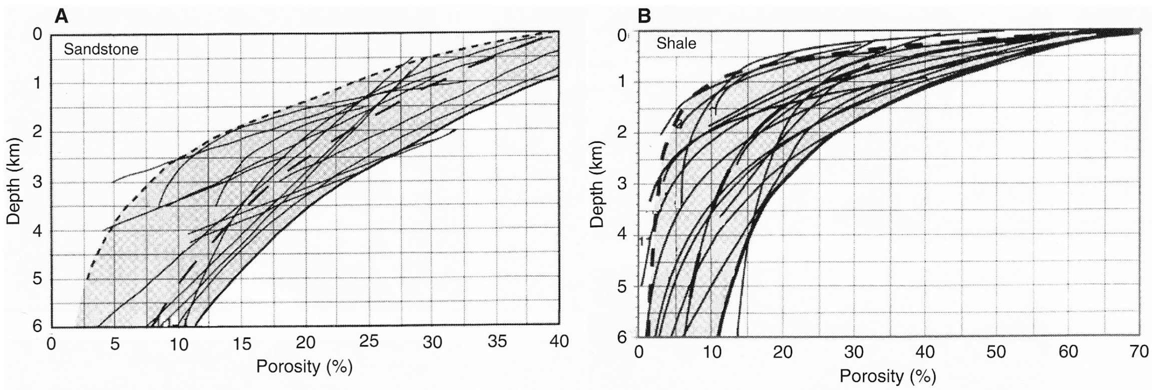



Figure 7.3: Evolution of porosity in sandstones and shales with progressive burial (Giles, 1997). Physical compaction is the principal driver of decreased porosity in the upper 500 m .

closely related to cementation because this fills the pore spaces and hence decreases the extent to which physical compaction accounts for porosity loss.

## 7.4 Pore Fluids

Most pore spaces in sediments are filled with fluids. Pore fluids have four principal origins

- *Meteoric* — derived from rain and snow
- *Connate* — derived from the original depositional water
- *Thermobaric* — derived from breakdown of hydrated minerals
- *Juvenile* — magmatic in origin

Meteoric waters tend to be acidic and oxidizing because they are derived from the atmosphere where they dissolve nitrous oxide, carbon dioxide, and volcanigenic sulfur. Organic acids in soils may contribute additional acicity. Due to the low pH of meteoric waters, they tend to be highly reactive.

Connate waters begin with the composition of the waters in which the sediments were deposited but evolve as the result of diagenesis. In general, connate waters are not as reactive as meteoric waters because they spend more time in contact with sediments, and in the case of precipitated sediments, are similar in composition to those from which the minerals initially crystallized. *Compactional waters* are connate fluids that migrate through the sediment as the result of compaction.

*Thermobaric waters* are high temperature, highly pressurized waters derived mainly from the dehydration of clay minerals in the deep basin. Because connate waters tend to have high solute concentrations, they are typically denser than meteoric waters.

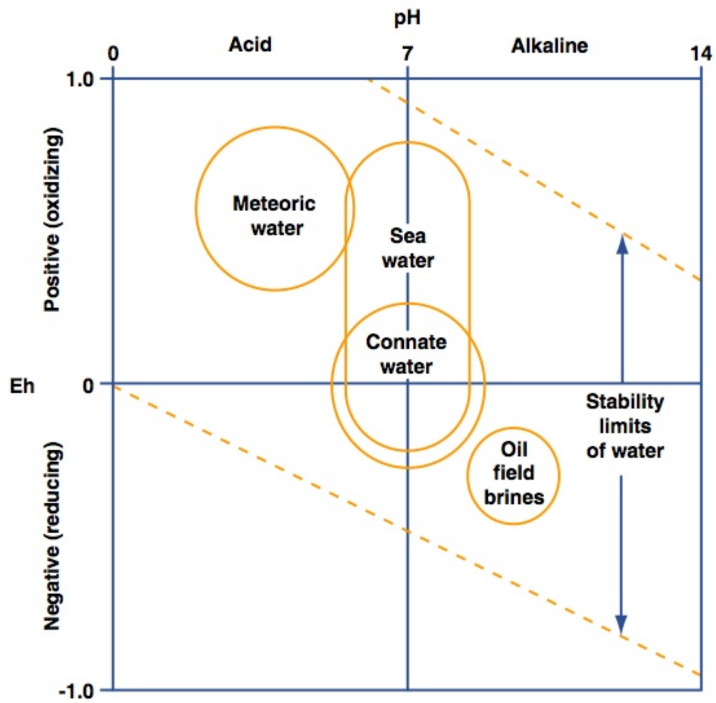



Figure 7.4: Types of water filling pore spaces (from Southard, 2007)

Pore fluids migrate as the result of pressure gradients. Meteoric flow is simply groundwater flow, driven by the head of the groundwater. It is important at shallow depths, in basins that have been uplifted or otherwise exposed by changing sea level, or adjacent to large mountain ranges, where the topographic head can cause them to displace the connate fluids. Compactional flows tend to be most important during early burial and are quite localized. However, at greater depths, heating of the fluids (due to the geothermal gradient) can drive convective flow. As you might imagine, warmer fluids are more reactive and tend to leach minerals, and as they migrate upwards to cooler parts of the basin, may redeposit these minerals. Convective flow is hence very important in diagenesis.

## 7.5 Sandstone Diagenesis

Sandstones tend not to have a large organic carbon fraction and are dominated by quartz. Consequently, the main processes to affect sands during diagenesis are compaction, which has already been discussed, and cementation.

### 7.5.1 Cementation

Cementation is the chemical or biogeochemical precipitation of minerals into the pore space of sediments. Cementation may take place at any stage of diagenesis and is controlled by the mineralogy and chemical compositions of the sediments, the chemistry of the pore waters (including pH and Eh conditions), and the P-T conditions of the sediments. Partial or relatively early cementation that is concentrated in regions produces *nodules* or *concretions*.

Quartz and calcite are the two most important cement minerals in sandstones. The solubilities of both minerals are unaffected by Eh but strongly affected by pH in an inverse way. Consequently quartz dissolution and calcite precipitation, and vice versa, are often intimately linked. The change in solubility with increasing temperature is also different for quartz and calcite. Therefore, with increasing burial and temperature, quartz will often begin to dissolve, while a patchy calcite cement will begin to precipitate (assuming there is sufficient calcium carbonate in solution). Irrespective of which mineral cements pore spaces, *it requires extremely large volumes of fluid given the low concentration of solutes in pore waters.*

**Silica cementation**

- Pressure solution (and reprecipitation)
- Quartz overgrowths on quartz grains (syntaxial overgrowth)
  - Commonly, the original grain can be discerned by an iron-oxide or clay coatings
  - However, where there is no coating, the quartz grains can be interlocking, giving the appearance of a metamorphic grain
  - Source of silica often pressure dissolution
  - Also derived from dissolution of more soluble or finer-grained silicate minerals
- Chalcedony is microcrystalline quartz and is a component of silica cement
- Commonly derived from opal (a hydrated silica), which is unstable at burial P and T

**Other types of cementation**

- Carbonate
  - Calcite
    - *Poikilotopic* cements entirely encase other grains.
    - Drusy calcite mosaics. *Drusy cements* are blocky crystalline cements that grow away from a boundary and protrude into an empty space.
  - Dolomite (Ca,Mg carbonate)
  - Siderite (Fe carbonate)
  - Commonly occur as concretions
  - Commonly displace grains
  - Often triggered by organic matter degradation
- Feldspar
  - As overgrowths of feldspar crystals
  - May explain why some feldspar crystals appear to be very angular
- Clays
  - kaolinite

  - illite

- Zeolites—hydrous aluminosilicates
  - particularly important in sediments with a large volcanic component
- Iron oxides
  - Commonly coat mineral grains and fractures
- Other less common cements include evaporite minerals, barite, and pyrite

### 7.5.2 Recrystalization and mineral replacement

Recrystalization is the reorganization of an existing mineral, which typically involves a change in crystal size and orientation. Recrystalization may preserve primary sedimentary structures or destroy them. This process is most common in carbonates and other precipitated rocks. Mineral replacement involves a change of chemistry and usually is accomplished by the gradual precipitation of a new mineral as a preexisting mineral dissolves. The process my retain the morphology of the original crystal.

### 7.5.3 Glauconite

*Glauconite* is a greenish, Fe- and K-rich mica that forms on a substrate of local organic matter (producing a reducing microenvironment) during early diagenesis, typically in outer shelf to upper slope settings that are *sediment-starved* (Meunier and El Albani, 2007). The glauconite may be reworked, resulting in concentration of the grains.

## 7.6 Mud Diagenesis

Very early diagenesis of muds is heavily influenced by biology due to the bacterial breakdown of organic mater in the upper sediment column. Up to about 500 or 1000 m, physical compaction is important. At about 1000–2000 m, hydrocarbon generation (in organic-rich muds) begins. Clay diagenesis is important later in burial.

### 7.6.1 Bacterial degradation of organic matter

Early diagenesis is particularly important in muds because contain relatively elevated concentrations ($\sim$1% on average) of organic matter (mainly the remains of algae and cyanobacteria). The bacterial degradation of this organic matter, which occurs during early diagenesis. Because the breakdown of organic matter involves oxidation of organic matter (the electron donor) at the expense of some oxidant (the electron acceptor), bacterial diagenesis of muds entails significant redox changes within the sediments and pore waters. Microbial consortia within the sediments tend to utilize the most energetically favorable electron acceptor (oxidant) before switching to a less favorable one, meaning $O_2$ is first consumed in pore waters, followed by $NO_3^-$. However, kinetics and availability of the electron donors also plays a role, particularly in the case of Mn and Fe (Nealson, 1997), which though energetically favourable, tend not to be used up as quickly as sulfate.

**The Principal oxidants**

- $O_2$
- $NO_3^-$
- $MnO_2$
- $Fe_2O_3$
- $SO_4^{2-}$
- Fermentation

The early diagenesis of muds is also heavily influenced by bioturbation, which in addition to destroying original sedimentary structures and layers, irrigates the sediments, allowing oxidants to penetrate deeper into the sediment column.

### 7.6.2 Porosity and compaction

Muds commonly contain a significant proportion of clay minerals, which are gregarious minerals that commonly gather in *floccules* or pellets. *Flocculation* is the formation of clay aggregates, which occurs because clays minerals have negative face charges and positive edge charges. The types of cations strongly influences whether clays will flocculate; $Ca^{2+}$ promotes flocculation, hence floccules are important where clay minerals are delivered to marine environments (e.g. estuaries). Floccules have a very open (*'house-of-cards'*) structure which result in high porosity and a typically a large amount of porewater in clayey sediments. Because of this high initial porosity, muds undergo significant compaction during the first 500 to 1000m of burial. This physical compaction reduces the porosity by at least half and can result in the development of *fissility* due to the flattening of clay sheets (the house of cards collapses).

### 7.6.3 Clay alteration

Early diagenesis is dominated by the alteration to organic matter and mechanical compaction and dewatering, although alteration of clay minerals (clay diagenesis) can also be important. At greater depths and temperatures, clay diagenesis is extremely important. Recall that clays can be subdivided into four main classes: two-layer kandites, three-layer smectites, three-layer illites and chlorites, and mixed-layer clays. Important for this discussion, smectites have a charge imbalance, which is compensated for by Ca and Na cations in the interlayers. Illite is also a three-layer clay, but the charge imbalance is compensated by K cations.

At about 3 km depth, smectite begins to alter to intermediate illite-smectite, then to illite (Curtis, 1985). This conversion is coupled to the alteration of K-feldspar, that releases the $K^+$ ions required to form the illite. Silica is also released by this dissolution, and the exchange of $K^+$ into the cation exchange sites between the clay layers expels water and other cations, mainly $Na^+$ and $Ca^{2+}$. The sodium may drive the albitization of any available plagioclase, whereas the released water contributes to overpressure in the pore fluids. This process takes place within the heart of the oil window and is closely coupled to hydrocarbon generation. The production of $CO_2$, organic acids, and other gases changes porewater pH and this influences clay formation and alteration (Curtis, 1985). In turn, increases in pore fluid pressures as the result of the release of water from interlayer sites contributes to the expulsion of hydrocarbons.

Illite subsequently alters to chlorite, passing first through smectite-chlorite or vermiculate-cholorite

## 7.7 Carbonate Diagenesis and Porosity

Carbonates are uniquely susceptible to early diagenesis for a combination of reasons. First, the primary marine carbonate minerals (aragonite and high-Mg calcite), are inherently unstable and prone to dissolution and recrystallization. Second, most carbonates are deposited in shallow marine environments, where meteoric fluid flow can be significant, particularly at times of low sea level. Meteoric waters, or mixtures of meteoric and connate waters in the subsurface will tend to dissolve carbonate sediments, in particular the unstable aragonite and high-Mg phases.

Primary porosity in carbonates tends to be destroyed quickly, due to the tendency of aragonite and high-Mg calcite to dissolve during early diagenesis, and low-Mg calcite, which is soluble, to precipitate. This process often results in subsequent cementation in pore spaces (see below). Carbonates are often quite porous, but a large part of this porosity is secondary porosity which develops during subsequent dissolution during burial or uplift diagenesis. The principal reservoir rocks in the Persian Gulf are carbonates with significant secondary porosity.

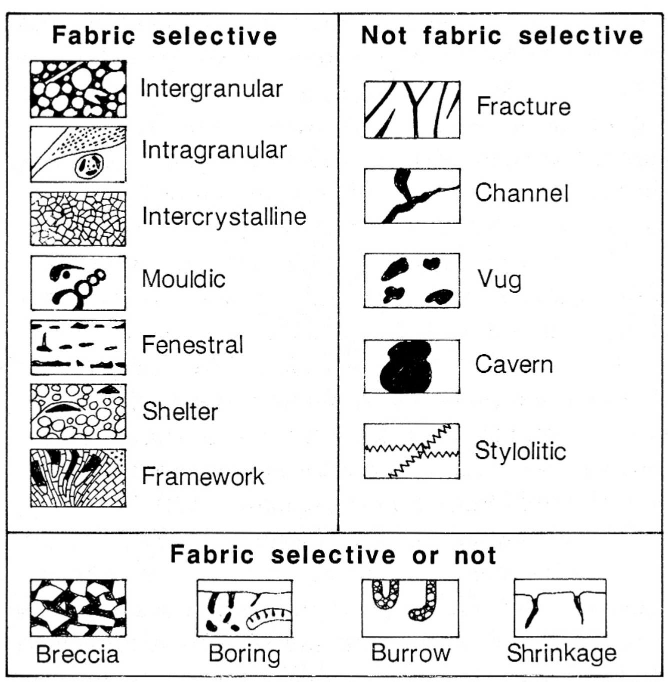



Figure 7.5: Types of porosity in carbonates. From Tucker and Wright (1990).

### 7.7.1 Cementation

#### Non-marine cements

The early dissolution of unstable carbonate phases and subsequent reprecipitation of low-Mg calcite results in the development of *rim cements*, which encrust exposed grain surfaces. These rim cements typically form in the phreatic zone and help account for early lithification of carbonates. This style of cement tends to be blocky. Cements may also form in the *vadose* zone, where rainfall or other episodic wetting wets the pore spaces. In these cases, the cements are referred to as *pendant* cements, and may have a *microstalactitic fabric*. Other vadose cements include *meniscus* cements, which form near grain contacts and are curved due to the surface tension acting on the lining of pore fluid. Beach dunes made of sand-sized shell fragments are often stabilized quickly by this type of cementation.

#### Early marine cementation

Early marine cementation is also common, and occurs shortly after deposition of many carbonate sediments. Marine cements commonly occur as *drusy* (that is, with crystals that project into the open pore space) magnesian calcite or aragonite. Aragonite cements tend to form fine, fibrous fringes, whereas magnesian calcite is blockier. Both cement types tend to be *isopachous*, that is, they have the same width all around the pore cavity. Isopachous cement point unambiguously to precipitation in a water-filled void. *Botryoidal* cements consist of coalescing spheres of radiating cement and are common in cavities in reefs (e.g., the Permian Reef of West Texas).

#### Deep marine cementation

Carbonate cementation may proceed in the deep burial realm. One driver of deep cementation is the generation of warm, saline fluids in shallow, restricted environments, which may migrate downwards, displacing cooler, fresher pore waters occupying sediments deposited in a more open marine realm. Such fluids may pervasively cement carbonate sediments (or siliciclastic sediments) at depth.

Pore fluids may also convect in the marine realm, particularly along slopes or banks, in a process known as *Kohout convection*. Here, fresh, cool seawater may penetrate the sediments and then convect as it warms geothermally. This process may drive deep dissolution, followed by reprecipitation of carbonate at shallower depths. Even without dissolution, it is an effectively way of pumping large amounts of carbonate-saturated waters through pore spaces.

Another source of carbonate cement at depth is pressure solution, just as in quartz cementation. *Styolites* are a common feature in carbonate sediments and in places are pervasive, reflecting a large degree of compaction and pressure solution, which would have generate cementing fluids.

### 7.7.2 Replacement

Components of carbonate sediments, such as ooids, shell fragments, or early cements commonly display a different grain size, and sometimes different mineralogy, from that which they had when they were formed. This can be recognized where *ghosts* of the primary mineral remain. Although in some cases this is the result of complete dissolution of a mineral or grain and reprecipitation of a new mineral, very commonly it is clear that this transformation must have taken place in small steps. That is, dissolution and precipitation of the new mineral took place effectively simultaneously in a process known generally as *neomorphism.* Typically, the new mineral is coarser grained than the primary mineral. Neomorphism can be distinguished from cementation insofar as the new minerals reflects the original grain rather than

## 7.8 Dolomite and Dolostones

A *dolostone* is a carbonate rock with at least 50% of the mineral dolomite. Whereas dolostones are widespread in the geological record, particularly in the Precambrian, where they account for the majority of carbonate rocks, their abundance decreases in the Paleozoic, they are rare in the Cenozoic, and almost non-existent (with a few notable exceptions), in the modern world. In large part, this is because dolomite does not precipitate in seawater, despite the fact that it is highly supersaturated in seawater. Most Mesozoic and older dolomite is well ordered, ideal dolomite, which to date has not been synthesized at temperatures below 100°C. Nor do any organisms make their shells or skeletons out of dolomite. This is the so-called *dolomite problem* (Hardie, 1987), which has vexed sedimentary geologists for half a century. Hence, it is effectively irrefutable that dolomite is predominantly a diagenetic replacement mineral rather than a primary carbonate mineral, thus the term *dolomitization.*

There are two general groups of dolostone. *Fine-grained* dolostones and *coarse-grained* dolostones. In both cases, the dolomite replaces the original aragonite or calcite minerals, but in the case of fine-grained dolostones, this replacement probably occurs early (i.e., early diagenesis), resulting often in good preservation of original limestone depositional fabrics. One way to detect very early dolomitization is in the composition of intraclasts (formed frequently in shallow water deposits) compared to their matrix.

Quite a lot of dolostone forms during burial diagenesis. It is coarser grained than the early diagenesis, owing to formation at higher temperatures and over longer periods of time. This sort of dolomitization tends to destroy original fabrics (as well as fossils).

Later dolomitization may occur as a result of focused or diffuse fluid flow through a carbonate. For example, one often sees dolomitization associated with faults and fractures. These very late dolomitization fronts are often highly destructive and often quite coarse grained.

### 7.8.1 Dolomitization models

Because dolomite is difficult to make experimentally, geologists have struggled to generate robust models for how dolostones form naturally. Nevertheless, a variety of models have been proposed for relative earlier (as opposed to deep burial or exhumation-related dolomitization), and a few have caught on and probably account for much of the dolostone we see in the geological record. In general, these models require a mechanism of concentrating Mg cations in solution and pumping large volumes of that solution through carbonate sediments.

#### Seepage reflux

In restricted settings, $Mg^{2+}$ concentrations are heightened, while $Ca^{2+}$ concentrations are lowered due to precipitation of calcite and possibly evaporites. Mg-rich brines may then flow gravitationally through calcium carbonate sediments below or laterally, altering the $CaCO_3$ to dolomite in the process. An excellent example of where this might occur is in reef-lagoon systems.

#### Brackish-water dolomitization

Because seawater is much more saturated with respect to dolomite than calcite, mixing seawater with fresh water can lower the saturation state of calcite below unity, while leaving dolostone supersaturated. Hence, the mixture of meteoric waters with saline connate waters in the subsurface may generate a zone where dolomitization preferentially occurs. This mechanism would be particularly robust during a drop in sea level, which provide groundwaters more head to penetrate carbonate sediments.

#### Microbial influence

It has been widely observed that many shallow water dolostones are intimately associated with microbial mats. It is also known that sulfate ions and hydrated Mg ions tend to inhibit dolomite crystallization. A variety of experiments have shown that microbial influences can trigger dolomite precipitation, and similar processes may account for certain modern dolomite occurrences. The precise mechanisms remain poorly known, but probably have something to do with removal of sulfate and/or other kinetic inhibitors (Vasconcelos and McKenzie, 1997). Another type of experiment has shown that methanogens can trigger dolomite precipitation by locally highly concentrating dolomite saturation states (Kenward et al., 2009). Although it seems unlikely that microbially influenced dolomitization alone can account for the abundance of dolomite in the geological record, Burns et al. (2000) have argued that secular evolution in seawater chemistry, most notably oxygen and sulfate levels, may have played an important role in governing the extent of microbial dolomitization.

#### Deep burial dolomitization

Dolomitization may also occur during deep burial, where it commonly produces *saddle* or *baroque* dolomite, which is distinguished by it coarse, milky crystals and undulose extinction (Flügel, 2004). One possible source of $Mg^{2+}$ is in the compaction of mud rocks and the conversion of smectite to illite. Deep burial dolomitization is also associated with hydrocarbon migration and the generation of epigenetic sulphides (e.g., Mississippi Valley type mineralization) (Flügel, 2004). Deep burial diagenesis is highly destructive of initial fabrics.

# Chapter 8

# Fluid Flow

Additional reading: Boggs, Chapter 2; Bridge and Demicco, Chapter 5 (121-157)

## 8.1 Fluid flow and the shear stress of fluids

Most sediments transport occurs within fluids, and the interaction between a fluid and the sediment it carries influences the style of sedimentation and the types of structures and bedforms that occur. Unidrectional, turbulent water flow is the most important agent of erosion, sediment transport, and sediment deposition on the land surface and is also important in the ocean and in lakes (Bridge and Demicco, 2007). Air and ice flow and gravity flows are also important agents in erosion and sediment transport and play major roles in shaping the surface of the earth.

Fluid flow is a complicated subject and will only be addressed in this class in a cursory manner. The physics of fluids does, however, form the foundation of the entire field of process sedimentology, which underpins any study of the sedimentary record. If at first the equations give you nightmares, do not worry. You will not be expected to derive these equations or even memorize the different components. Rather, you will only be expected to understand how the equations relate to natural sedimentary processes.

### 8.1.1 Viscosity

The two most important physical properties of fluids are their *viscosity* and *density*.

- *Viscosity* is the resistance of a fluid to deformation
- *Density* is the mass of fluid per unit volume of fluid. This property influences the strength of forces involved in fluid flow.

The relationship between shear stress, strain rate, and viscosity is given by the following useful equation:

$$\tau = \mu \frac{du}{dy} \tag{8.1}$$

- $\tau$ = The *shear stress* (acting on the fluid)
- $\mu$ = The *dynamic* (or *molecular*) *viscosity*

- $\frac{du}{dy}$ = The strain rate

Here, shear stress results from a force that is applied parallel to a fluid surface.

Different types of fluids are defined by a combination of

- The relationship between shear stress and strain rat in the fluid
- The yield strength of the fluid

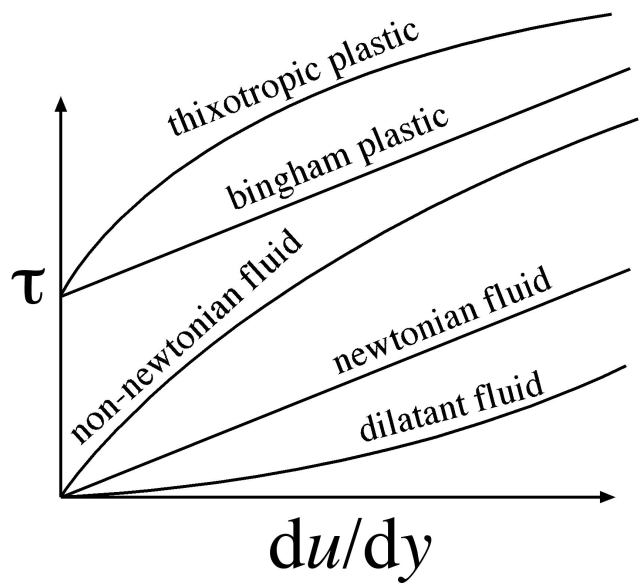



Figure 8.1: Types of fluids can be defined in terms of their viscosity and how viscosity changes with varying applied shear stress. A fluid that does not begin to deform immediately with the application of shear is said to have a *yield strength*.

### 8.1.2 Laminar versus turbulent flow

Fluids obey Newton's first law, meaning that because they are always acted upon by friction, they will tend to come to rest unless some force is available to overcome the force of friction. Two main forces that drive fluid motion are gravity and pressure gradients.

*Laminar* flows are flows in which water particles follow straight paths and where the velocity at a given point in the water remains constant (*steady*). *Turbulent* flows occur when the particle paths are curved and cross each other (typically in swirling motions). Velocities are highly variable in turbulent flows when considered at short temporal and spatial scales (and thus inherently *unsteady*), but over longer time and spatial scales, turbulent flows may appear steady. Turbulent flows contain *eddies*, which are rotating swirls in a flow (i.e. non-linear flow paths). Eddies come in all length scales below the height (or width) of a flow and are constantly changing in shape, grading into one another, and appearing and disappearing.

Flows that maintain a constant velocity and cross-sectional area along their flow path are regarded as *uniform*, whereas those that change velocity and cross-sectional area are *non-uniform*. Anybody who has looked into a stream probably recognizes that most fluid flows that erode or transport sediments are turbulent. Laminar flows do also play an important role in geology, and this include most groundwater flows and lava flows.

**Channel flows**

Most of the fluid flow in which we are interested is considered channel flow, meaning that the base and sides of the flow are in contact with an immobile boundary. The force that makes the fluid flow is gravity (a body force) and it in balanced by the frictional force at the edge of the channels that resist that flow. This frictional force is equal to and opposite the boundary sheer stress. This *boundary (or bed) shear stress* is the force a fluid exerts on the boundary (or bed) of a flow, and it can be defined as

$$\tau_0 = \rho g h S \tag{8.2}$$

where,

- $\rho$ = the density of the fluid
- $g$ = gravitational acceleration
- $h$ = the flow depth
- $S$ = the slope

The boundary shear stress is important in sedimentology because it determines whether erosion or deposition takes place. The *boundary layer* is the zone near the base of a flow where the flow is slowed down by the frictional resistance of the bed.

- The *no-slip condition*: fluid in contact with a solid boundary has exactly the same velocity as that solid boundary

Laminar flows and turbulent flows have distinct velocity profiles. Whereas laminar flow velocities vary gradually (think a half parabolical) with depth, turbulent flows have a more uniform (time averaged) velocity through most of the profile and a sharp velocity gradient just above the bed. This profile is due to the effect of eddies, which largely average flow out over most of the profile, but are prevented from moving vertically near the base of the flow. As a consequence of this restriction, laminar (viscous) sheering becomes increasingly important closer to the bed.

- The *viscous sublayer* is the thin zone (say a few millimeters to a few centimeters thick) where the viscous sheer is more important to the flow than turbulence.

A *dynamically smooth* flow is a flow in which the roughness of the bed is constrained to the viscous sublayer. A *dynamically rough* flow is one in which the roughness protrudes from what would nominally be the viscous sublayer if it were a dynamically smooth flow. The surface roughness of the bed is large enough that there is no continuous zone of viscous flow, even though viscous flow may dominate on the scale of flow around the larger roughness elements on the bed.

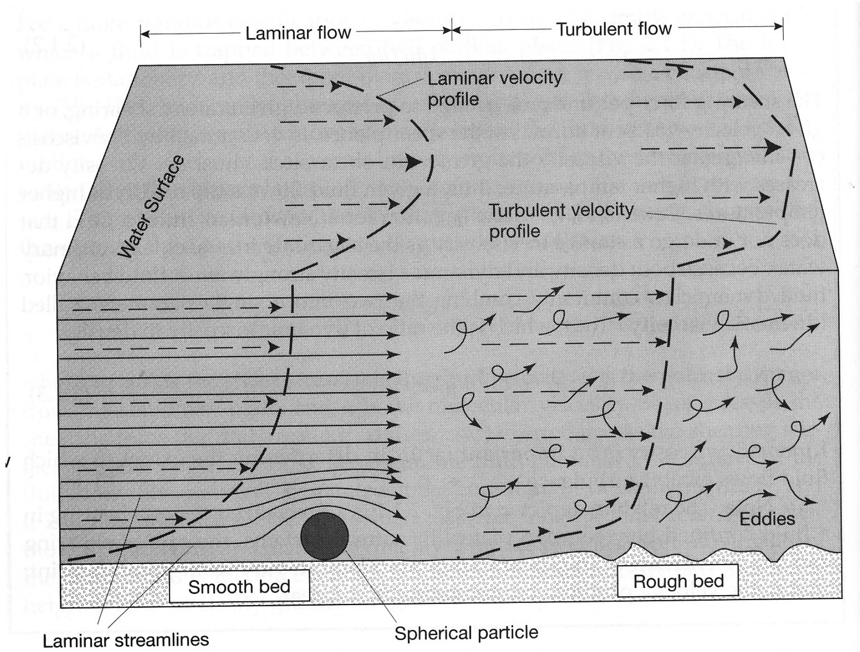



Figure 8.2: Laminar versus turbulent flow, from Boggs (2011).

### The Reynolds Number

The *Reynolds Number* is a dimensionless number that reflects the relative importance of inertial versus viscous forces acting on a fluid, hence whether it will tend to be laminar or turbulent.

$$R_e = \rho \frac{UL}{\mu} \tag{8.3}$$

- $\rho$ = the density of the fluid
- U = is the average velocity of fluid flow
- L = is the characteristic length scale (e.g. water depth)
- $\mu$ = the viscosity of the fluid

*Laminar flow* occurs where the Reynolds number is low; the transition to turbulent flow occurs at $500 < R_e < 2000$. The threshold for turbulent flow is controlled mainly by the surface roughness of the bed over which the fluid is flowing.

## 8.1.3 The Froude number

The Froude Number ($F_r$) defines whether a flow is *subcritical* or *supercritical.* This parameter is important in determining the types of bedforms that occur, as will be discussed in the next chapter. A subcritical, or tranquil flow is one in which wave velocity exceeds flow velocity, with the result that waves can travel upstream. Conversely,in supercritical, or rapid flow, flow velocity exceeds wave velocity.

$$F_r = \frac{U}{\sqrt{gL}} \tag{8.4}$$

- $U$ = mean velocity of flow
- $L$ = water depth
- $g$ = gravitational acceleration

## 8.2 Oscillatory flow

*Oscillatory* flow is a current that reverses velocity periodically with time. Reversing tidal currents are one example of oscillatory flow, but here we are concerned with the much higher frequency oscillatory flows caused by *surface gravity waves* on the surface of lakes and the ocean. These wind-generated waves are formed by the shear stress of the wind acting on the water surface. You will probably recall from previous science courses that the three factors that control the size of waves are the velocity of the wind, the distance over which the wind has blown on the water surface (*fetch*), and the amount of time it has blown.

In a perfect train of waves, the *crest* (highest point) and *trough* (lowest point) would be the same height, the waves would be symmetric, and the period between waves would be fixed. The *wavelength* is the distance between adjacent crests (or troughs), and the height of the waves is measured as the distance from the trough to the crest. As a wave passes, an article on the surface of the water, such as a bottle or a boat, traverses a circle, moving in the direction of wave propagation as the crest passes, and the opposite direction as the trough passes. Water below the wave surface moves in similar orbits, but the magnitude of these orbits decreases with depth. In deep-water waves, the magnitude of these orbits decreases to zero-point, below which there is no motion. In shallow-water waves, these orbits also flatten, and where orbit intersects the bottom of the water body, it is polarized into an oscillatory flow whose velocity varies sinusoidally with the passing of the waves. The maximum velocity, $U_m$, is a function of the wave period, $T$, and orbital diameter, $d_0$:

$$U_m = \frac{\pi d_0}{T} \tag{8.5}$$

If you have ever been on or near the ocean or a large lake during a storm, you will have noticed that the geometry of the waves is far from simple and symmetric. The highly irregular waves that form under these circumstances are the product of shifting wind directions or the intersection of different wave trains propagating in different directions. As you would expect, they give rise to complex flows on the bottom of the water body. Because storm waves are larger, the resulting flows can penetrate much deep water depths. In the ocean, storms waves can generate oscillatory flows to depths of 80 to 100 m.

# Chapter 9

# Sediment Transport

Additional reading: Boggs, Chapter 2; Bridge and Demicco, Chapter 5 (121–157)

## 9.1 Sediment transport

Sediment transport has to do with the forces that act on particles, the thresholds for particle movement, and the types of movement of particles. In this section, we will consider how particles are entrained in a flow, how they move through the flow, how they move along a bed, and how they settle from a fluid.

### 9.1.1 Forces acting on particles

Two types of forces are exerted on a sediment grain resting on a bed below a flowing fluid (Fig. 9.1.1).

- *Viscous forces* act tangential to the solid surface of the grain and result from shear of the fluid as it flows over the grain
- *Pressure forces* at normal to the solid surface of the grain and result from variations in pressure within the flow

These forces vary from point to point along the surface of the sediment grain and the nature of the forces depends critically on the diameter of the grain relative to the thickness of the viscous sublayer. For example, flow around a grain embedded within the viscous sublayer is smooth, whereas flow around a large particle is rough and the result is *flow separation* and turbulence in the wake of the flow around the grain.

Now, picture a sediment grain sitting at rest in on a bed in flowing water. This grain is subject to a variety of competing forces, the nature of which depends strongly on whether or not the grain is entirely encapsulated within the viscous sub-layer or not:

- $F_D$ is the drag force. It results from the boundary shear stress and acts parallel to the bed, downstream
- $F_L$ is the hydraulic lift, which is directly vertically and is the result of the *Bernoulli effect*

- $F_G$ is the gravitational force
- $F_C$ is the cohesion of the particle to the bed, resulting from friction, electrostatic clay-mineral interactions, vegetations, etc.

The drag force and the lift force together comprise a *resultant fluid force* which is directed up and downstream. Because of the effects of turbulent flow, the strength of these forces can vary by factors of four or more over less than a second (Southard, 2007).

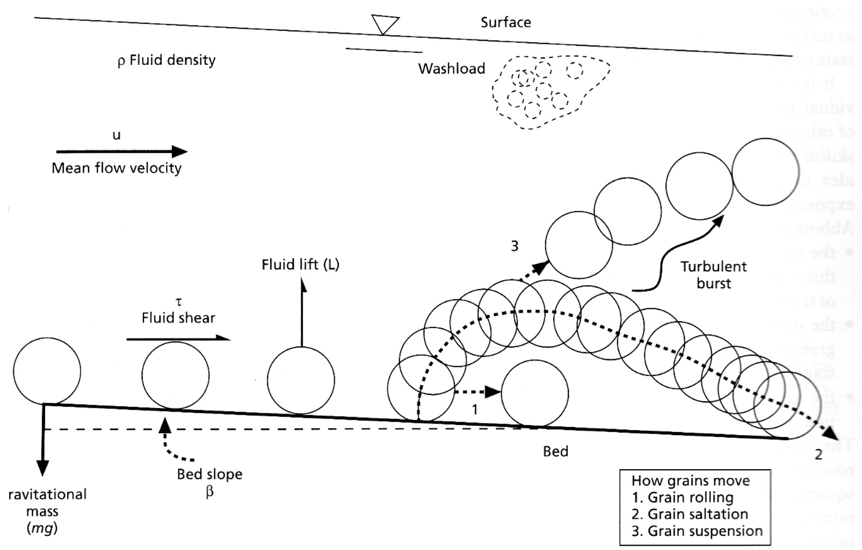



Figure 9.1: Sketch of the various types of particle motion in flowing water, from Leeder (1999).

### 9.1.2 Entrainment of particles

Stationary particles on the bed below a flowing fluid will begin to move downstream when velocity, hence boundary shear stress, surpasses some critical *threshold of movement*. This threshold is in fact a function of the boundary shear stress, fluid viscosity, particle size, particle shape, density difference between particle and fluid, and the flow velocity. Typically, smaller grains begin to move first; however, very fine-grained sediments don't obey this simple rule due to the effect of cohesiveness between clay-sized particles (Fig. 9.1.2). Effectively, grain movement should begin where

$$F_D + F_L > F_G + F_C \tag{9.1}$$

What we would like is a simple graph that shows this critical threshold in terms of flow velocity as a function of grain size. However, and not surprisingly, the critical threshold for grain movement is hard to define and in detail quite complicated. In sedimentology textbooks, this threshold is now commonly expressed in the *Shields diagram* in terms of dimensionless shear stress (instead of velocity) versus grain diameter normalized by thickness of the viscous sublayer (so again, dimensionless)(Fig. 9.1.2). This somewhat complicated version is designed to span all reasonable combinations of grain size and fluids. But because sediment motion often occurs on bedforms or other irregularities on bedding surfaces and typically begins with erratic pulses, it is useful to consider incipient grain movement as stochastic (i.e. at a given flow velocity, there is a certain percent change of grain movement beginning). That is, generally speaking, as flow velocity increases towards some critical threshold, the probability of a grain moving increases, but this motion often occurs in pulses until a high velocity is reached.

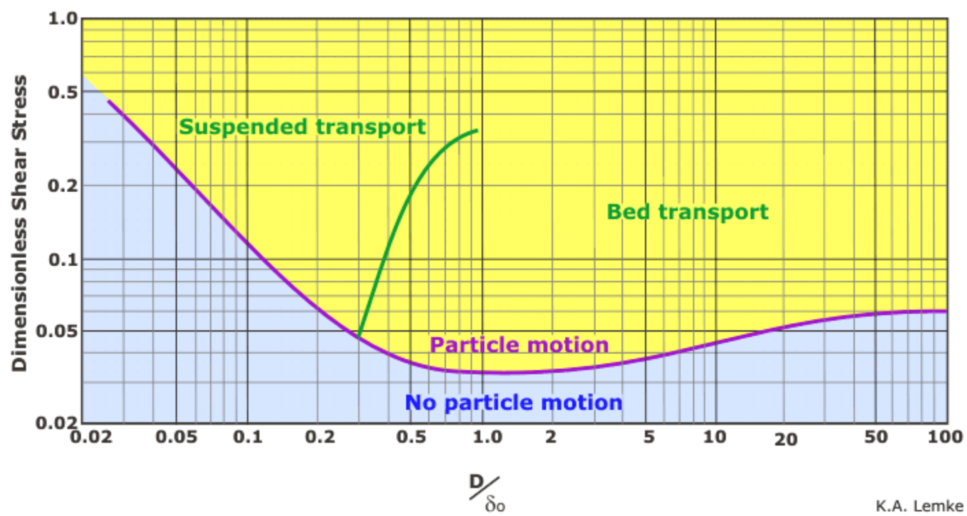



Figure 9.2: A version of the Shields diagram from K.A. Lemke. D is grain diameter and $\delta_0$ is thickness of the viscous sublayer. Note that the dimensionless shear stress increases for very fine grained sediments ($\delta_0$ otherwise being held constant).

Once a particle is entrained in a flowing fluid, the forces required to keep it in motion are decreased (because $F_C$ is no longer a factor). Additionally, turbulence flow and its vertical component of shear stress may help keep a grain entrained in a fluid (but may also help 'push' a grain back down). Of course, there are many other factors to consider in terms of grain entrainment, such as particle size, shape, and density, as well as bed surface roughness.

### 9.1.3 Particle movement in flowing water

In a general way, sediment (i.e. in a river or on the sea floor) can be transported either in a *suspended load* or the *bed load*. The *suspended load* in rivers and other flows includes the *wash load*, which is permanently in suspension and consists of the finest grain-fraction. The bed load is the fraction that travels along or very close to the bed and is therefore less affected by turbulence. Whereas the amount of sediment transported in suspension generally far exceeds that of the bed load in most sediment-transporting flows, it is the bed load that leaves the more legible imprint in the sedimentary record in the form of bedforms and sedimentary structures (Southard, 2007).

**Bed load**

Particles transported as part of the bed load move by *traction*

- rolling
- sliding
- hopping

If you were able to observe the bod load of a stream in slow motion, you would probably see that particles tend to roll and hop downstream for a while, come to rest, and then restart their downstream trajectory (Southard, 2007). This irregular motion is due to that fact that once a grain has started to move (i.e. overcome the critical threshold), the force required to keep it moving is diminished and it will tend to continue moving until it encounters strong resistance or a sheltered spot along the bed. An additional and probably important factor is the dispersive effect of collisions with other grains. The still somewhat theoretical downstream movement by dispersive forces is known as a *traction carpet.*

**Suspended load**

There is no clearcut distinction between suspended and bed load. Conceptually, the suspended load can be regarded as that part of the sediment load that spends little or no time in contact with the bed. Of course, a given grain may move back and forth between the suspended and bed load, thus blurring this definition.

Particles are more susceptible to transport in suspension when the vertical component of turbulent flow exceeds the settling velocities of the particles. And recall that turbulent flow and average velocities increase upwards in the flow, meaning that a particle that makes the jump from bed to suspended load is in a good position to remain in suspension for some time. However, turbulent flow is whimsical, and a strong downward flow may easily return a lofty suspended particle to earth.

**Sediment transport rate**

The sediment transport rate is notoriously difficult to quantify, but you can think of it, to the first order, as a very steeply increasing function of flow strength. The implication is that during floods, rivers may discharge much higher volumes of sediments and that as flows wane, sediment will be deposited extremely rapidly. Similarly, in deserts where where rainfall tends to be rare but intense when it occurs, huge amounts of sediment can be transported despite prevailing dry conditions.

### 9.1.4 Particle movement in air

Sediment transport by wind is broadly similar to that by water, but because of the much greater ratio of density between sediment and air compared to sediment and water, sediment grains in air have much greater relative inertia. Consequently, even though greater fluid velocities are required to initiate motion, once a particle in air is in motion, it is much less susceptible to the effects of local fluid turbulence than in water. The suspended load of sediments in an air mass is known as the *dust load.*

**Saltation and creep**

*Saltation* is a type of sediment movement in which particles launch steeply from the bed surface and then gradually return to the bed surface. Saltation is the main mode of sediment transport when strong winds blow over a sandy substrate. Probably you have witnessed saltation at a beach or on sand dunes before. If you knelt on the sand and stared across the surface, you would see a hazy layer of sediments near the bed, tapering off upward, which is formed by saltating grains, and a second type of transport known as *creep*, where sediments are moving along the bed surface.

Both saltation and creep are heavily dependent on the effect of grain collisions on the bed surface. The ability of these grain-grain collisions to cause a grain to saltate is dependent on the angle of collisions, hence the transfer of linear momentum. It stands to reason that drag and lift forces are also important, despite the much lower viscosity of air.

### 9.1.5 Settling particles

Three forces act on a particle settling through a fluid: drag force ($F_D$), buoyant force ($F_B$), and gravitational force ($F_G$.) The *terminal fall velocity*, that is the velocity at which the particle stops accelerating downward, is reached when

$$F_D = F_G - F_B \tag{9.2}$$

This equation leads to **Stoke's Law** for the maximum velocity ($V_g$) of a settling spherical grain:

$$V_g^2 = \frac{4gd}{3C_D}\frac{(\rho_s - \rho_f)}{\rho_f} \tag{9.3}$$

where $\rho_s$ and $\rho_g$ are the density of the grain and fluid, respective, $g$ is gravitational acceleration, $d$ is the diameter of the grain, and $C_D$ is a drag coefficient. $C_D$ is takes into account the type of flow around the grain; where the flow is turbulent and separates, it decreases the pressure behind the grain, hence decreasing the flow velocity (decreases $C_D$). The drag coefficient is related to the *grain Reynolds number*:

$$Re_g = V_G d\frac{\rho_f}{\mu} \tag{9.4}$$

For small spheres ($Re_g < 1$), the turbulent effect is negligible and Stoke's law simplifies to

$$V_g = \frac{1}{18}\frac{(\rho_s - \rho_f)gd^2}{\mu} \tag{9.5}$$

Stokes law can be used, for example, to understand how normally graded bedding forms: the larger the grain diameter, or $Re_g$, the faster the $V_g$. Hence, coarser grained (or denser) particles settle faster.

## 9.2 Concentrated-Sediment Flows

Our discussion of sediment transport thus far has been focused on *dispersed-sediment flows*–flows in which sediment concentrations within the fluid are low enough that the fluid can still shear readily. These include most river, tidal, storm, and wind currents and typically give rise to well stratified sediments. *Concentrated-sediment flows*, on the other hand, occur where sediment concentrations are very high, commonly altering the fluid properties of the flow. These commonly but not always yield poorly stratified sediments. The most common type of concentrated-sediment flow is a *gravity flow*. This is a flow in which sediments effectively move under their own weight. *Sedimentary-gravity flows* occur when grains become separated and dispersed, reducing internal friction below some critical point, which allows them to flow.

### 9.2.1 Hyperconcentrated flow

Hyperconcentrated flows are moderately concentrated mixtures of sediment and water, such as is found in the Yellow River. Whereas these are undoubtedly important geologically, they are as yet poorly understood.

### 9.2.2 Debris flows

*Debris flows* and *mud flows* are slurries, or mixture of water and mud, that are triggered on relatively steep slopes ($> 10°$)

- Often triggered by melting rain, earthquakes, and melting ice
- Behave as a Bingham plastic
- Subaerial debris flows particularly common in volcanic regions
- Subaqueous debris and mud flows typically associated with glacial outwash and turbidity currents
- Debris flows result in a poorly sorted deposit with a mud matrix.
- Debris flows can occur without a mud matrix (alluvial fans in arid climates)

### 9.2.3 Turbidity currents

*Turbidity currents* are turbulent water masses that initiate on the bottoms of a sloping sea or lake bed and incorporate sediments into suspension.

- Flow downhill due to density contrast with ambient water
- Initiated by turbulent flow, which scours the sea floor (or lake bed), entraining sediment into the flow
- Newtonian fluid
- Often initiated by sediment failure, which can be triggered by
  - earthquakes
  - storms
  - heavy bed load flux

- Commonly occur in deep water, e.g. on continental margins near the heads of submarine canyons or near the toes of large deltas, where they occur as surges (i.e., events)
- May continue for 100s of kilometres
- Result in laterally persistent, tabular, normally graded beds, often with a distinct sequence of sedimentary structures and textures (Bouma Sequence)
- May also occur as steady flows, typically in an area where sediment laden rivers discharge into a sloping lake or sea bed. Steady flows tend to lack the highly erosive head in the turbidity current.

Turbidity currents are commonly separated into low density versus high density flows, where low density turbidity currents comprise 20-30% sediment, and high density flows>30%. Low density turbidity current give rise to more classical normally graded beds, whereas high-density flows produce coarser-grained and perhaps less well sorted beds. Turbidity currents produce *turbidites*. The quintessential turbidite forms a single, fining-upward sequence of five (a–e) sediment types, known as a *Bouma sequence*: (a) massive sands; (b) laminated sands; (c) rippled sands and silts; (d) laminated silts; (e) laminated mud.

### 9.2.4 Liquified flows

*Liquified flows* occur when pore fluids injected from underlying sediments disperse grains.

- Loss of grain-to-grain contact, thus cohesion—usually in wet sands
- Triggered by earthquakes or rapid sedimentation on slopes of $\geq 3°$
- Non-newtonian, high viscosity fluid
- Stop very abruptly
- Form massive, poorly laminated sediments, sometimes with fluid escape structures

### 9.2.5 Grain flows

*Grains flows* occur in cohesionless sediments in which flow is maintained by dispersive pressure created by grain-to-grain collisions. You have likely seen grain flows in action if you have walked along the crest of a sand dune or run down the lee side of that dune.

- Subaerial or subaqeous; interstitial fluids unimportant
- Non-newtonian flows with no turbulence
- Occur where sediments are lying near the *angle of repose*
- Commonly seize up quickly
- Form thin (< several centimeters thick), crudely reverse-graded beds
- Common on the lee side of sand dunes and on the edge of carbonate banks

### 9.2.6 Slumps and slides

*Slumps* and *slides* involved the movement of large, coherent masses of sediment downslope. These are typically triggered by some event that decreases cohesion of sediments, such as heavy rains in a subaerial setting and earthquakes or fluid flow events in a subaqueous setting. These are non-turbulent and do not constitute sediment transport, as is it usually regarded, although it does move large masses of sediment.

# Chapter 10

# Stratification, Flow Regimes, and Bedforms

Additional reading: Tucker, Chapter 2 (25-32); Boggs, Chapter 4 or Tucker, Chapter 2, Bridge and Demicco, Chapter 5 (157-185)

## 10.1 Introduction

This lecture is intended to take you from the more theoretical approach to fluid flow and sediment transport covered in the previous lecture to the actual observable physical features of sediments deposits from flowing water and air. It will cover the basics of stratification and bedforms and serve as an introduction to sedimentary structures.

## 10.2 Stratification

*Stratification*, which is just the layering that results from sediment deposition, occurs at many scales and is by far the the most important sedimentary structure (Southard, 2007). It is brought about by changes in physical, chemical, or biological conditions that control sediment deposition. Stratification in sedimentary rocks is usually obvious and one of the first things one describes about rocks. Other times it is subtle. But always it is useful to identify, because both the processes that brought about the stratification and the physical features the define it tell us something about the environment in which the sediments were deposited.

- *Beds* are tabular or lenticular layers of rock with some sort of compositional, textural, or lithological unity. They are highly variable in thickness, but by definition, >1 cm thick.

- *laminations* are distinct and continuous layers less than 1 cm thick

- *Bedding planes* are the bounding surfaces of beds or laminae.

Stratification may be caused by any of a number of differences that you might readily notice on the outcrop:

- erosion/truncation
- grain size
- grain composition (or abundance of organic matter)
- color or shade difference
- differential weathering
- concentrations of particles, such as shells or pebbles
- sedimentary structures (including preferred orientation in grains or clasts)

They may be erosive, manifested by a change in lithology (e.g., individual carbonate or sand beds are commonly separated by a small amount of mud, known as *mud partings*), or *amalgamated* where two beds of the same type are bound (perhaps cryptically) by a surface that represent a hiatus in deposition.

Table 10.1: Descriptive terms for bed-lamina thickness

| **bed/lamina thickness** | **descriptive term** |
|---|---|
| 0-3 mm | thin lamina |
| 3-10 mm | thick lamina |
| 1-3 cm | very thin bed |
| 3-10 cm | thin bed |
| 10-30 cm | medium bed |
| 30-100 cm | thick bed |
| $>$ 100 cm | very thick bed |

Beds and laminations typically represent some sort of depositional event or deposition by a specific physical process. Bedding planes represent the time between between events or different processes. For example a turbidite is a graded bed that results from a single turbidity current. The same turbidite might grade distally into a graded lamina. Or a cross-bed may represent the migration of a train of subaqueous dunes across a stream channel. However, it is also more complicated, because stratification may be hierarchical, in that beds commonly show their own *internal lamination*, reflecting a higher order fluctuation in depositional processes.

### 10.2.1 Stratification/bedding features:

#### Parting

*Parting* is the tendency for rocks to split evenly along bedding planes. *Mud partings* are concentrations of mud or shale between beds. Beds may part into *flags* (<1 cm), *slabs* (1-10 cm), or thicker blocks. The extent of parting is partly a function of non-sedimentary controls on a rock type, such as the degree of metamorphism and weathering.

**Other features**

- Bands
- Lenses
- Parallel bedding
- Other types of bedding
  - Wavy
  - Discontinuous
  - Curved
  - Curved and non-parallel
- Graded bedding
  - *Normal grading* (fining up within a bed) is very common in sedimentary rocks and can be generated by a variety of processes. However, it usually represents decreasing strength of flow and is predicted by Stoke's law for settling particles.
  - *Inverse grading* may reflect the progradation of certain bedforms or certain types of gravity flows, such as grain flows on the lee side of sand dunes.
- *Massive bedding* in sands or coarser grained sediments is bedding that shows no distinct grading, laminations, or structures.

## 10.3 Bedforms and Flow Regimes

*Bedforms* are geometric elements that form by the interaction of flowing water or air over a deformable, cohesionless sediment bed. The type of bedforms that occur depends on the rate of flow, the depth of flow, and sediment size. Recall that the rate and depth of flow are variables in the dimensionless Froude number ($F_r$), which effectively governs whether or not the surface gravity waves above the bedforms are in phase with the bedforms or not. Flows where $F_r < \sim 1$ are referred to as the *lower flow regime*, and where $F_r > \sim 1$, *upper flow regime*.

The principal bedforms of micro- to mesoscale that form in unidirectional flowing water are ripples, dunes (large-scale ripples), and antidunes (larger bedforms will be discussed later in the class). The spacing and three dimensional geometry of these structures reveals a lot about their depositional environment. Plane bedding occurs both at modest flow velocities in coarser sediments and at high velocities in finer sediments.

### 10.3.1 Lower flow regime

**Ripples**

*Ripples* form when the bed shear stress slightly exceeds the critical threshold to initiate bedload motion, so long as the flow is smooth and subcritical ($F_r < 1$). Ripples typically

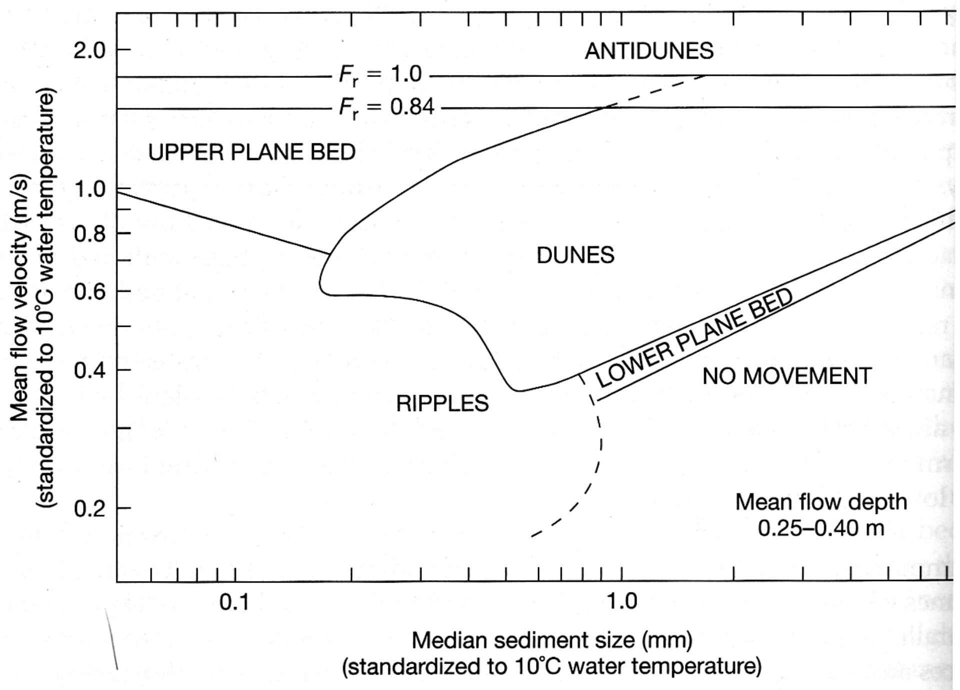



Figure 10.1: Bedforms in unidirectional flow conditions form in either the *lower flow regime* or *upper flow regime* (from Boggs, 2011). The flow regime is controlled by the sediment size, flow velocity, and water depth. Not shown in this figure is water depth, which also plays a role in what bedforms occur under different conditions.

form in a coarse silt- to medium sand-size sediment, but recent experiments have demonstrated that ripples may also form in muds where the mud flocculates to form pellets under similar hydrological conditions to silt-sand ripples (Schieber et al., 2007). The initiation of ripples on a planar bed is the result of variations in turbulent flow velocity, hence sediment transport rate, at the threshold of particle motion. The result is jerky movement of grains, which generates a series of parallel ridges and depressions parallel to the flow (*parting lineations*). Subsequent turbulent downflows (*sweeps*) redistribute the grains on the ridges into a new mound perpendicular to the flow, thus initiating a bed defect that will quickly develop into a ripple.

Ripples are typically 10 to 15 cm long, spaced at 5 to 20 cm, and range in height from about 1 to 3 cm. They only form where the grain Reynolds number ($Re_g$) $< 10$; hence, they do not form in coarser sediments ($> 0.7$ mm). Ripple length, L, is proportional to the diameter of the mean grain size ($L \approx 1000d$) and independent of flow depth.

**Lower flow regime plane beds**

Plane beds occur in the lower flow regime where $Re_g > 10$ (i.e., coarser-grained sediments). Similar to the case for ripples, bed defects a few grain diameters high form when flow exceeds the critical threshold velocity. But instead of transitioning into ripples, these defects simply propagate downstream in low relief bed waves. The length of these bed

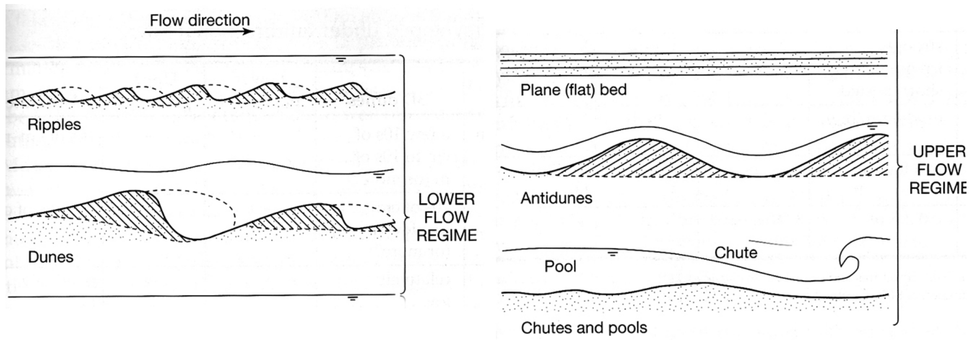



Figure 10.2: There are four principal types of *bedforms*: ripples, dunes, plane beds, and antidunes. From Blatt et al. (1980).

waves is proportional to flow depth, which suggests that they are, in fact, precursors to dunes that form at greater flow velocities.

**Dunes**

*Dunes* are larger bedforms with the similar morphology to ripples. They form at higher flow velocities in fine- to very coarse sand and have wavelengths ranging from under a meter to 10s or even 100s of meters. Broadly, dunes have heights > 4 cm and lengths > 0.6 m (Allen, 1982), but in fact, this definition excludes the bedforms that originate in the transition from the lower plane bed (Bridge and Demicco, 2007). Both ripples and dunes are rarely preserved whole in the sedimentary record, but the *cross-stratification* that forms as they are deposited often is preserved and can be an excellent indicator of paleoflow and hydrodynamic conditions. Dunes are sometimes referred to as *megaripples* or *sand waves* in the literature, although proper sand waves are distinct structures that will be discussed later in the course.

One difference between dunes and ripples is that the lee side of dunes is not necessarily at the angle of repose due to scouring by separated flow. Dunes may also be the substrate to other bedforms, including smaller dunes, plane beds, and ripples. In the troughs between dunes, ripples may have many different orientations, reflecting the complexity and 3-dimensionality of flow on the lee side of the dunes.

### 10.3.2 Upper flow regime

**Upper flow regime plane beds**

With sufficiently high current velocities, the high bed shear stress and sediment transport rate effectively shear out the bedforms. Like lower flow regime plane beds, the upper flow regime plane beds are not truly planar, but rather have a low-relief, asymmetric profile. Upper flow regime plane beds give rise to planar lamination that are defined by difference in grain size, and probably result from the migration of the low-relief bedforms.

**Antidunes**

Under supercritical flow ($F_r$ >1), bedforms include antidunes, chutes-and-pools, and rhomboid ripples. *Antidunes* are symmetric and low-amplitude in morphology and are in phase with surface waves. Antidunes commonly migrate upstream (hence their name) through erosion of their troughs and deposition on their crests. Whereas antidunes are reasonably common in rivers, they are not so common in the sedimentary record, probably because they are obliterated by decreasing flow velocities.

## 10.3.3 Ripple morphology

**Current ripples**

Ripples form during sediment transport in both water and wind. *Assymetric* ripples, commonly referred to as *current ripples*, occur during unidirectional transport, whereas *symmetric* ripples form under oscillatory flow. Ripples are not static - they migrate downstream. The principle mechanisms in the transport of sediments through ripples are *saltation* and *creep*. In short, grains migrate up the *stoss* side of a ripple and avalanche down the *lee* side, resulting in foresets lying at the *angle of repose*.

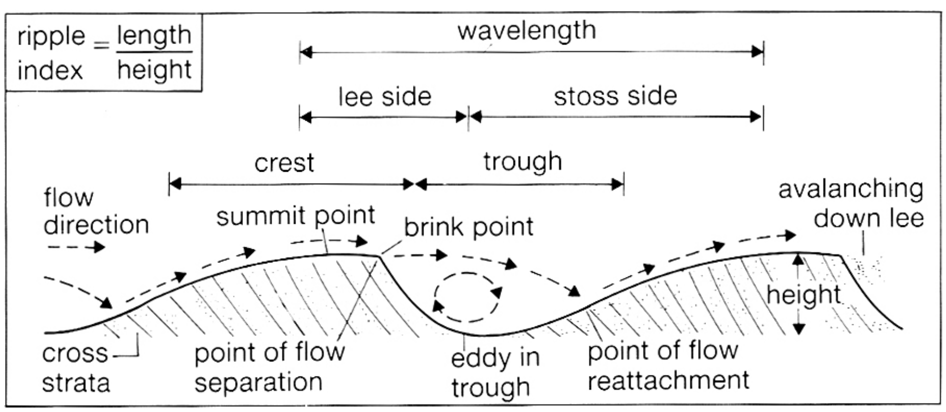



Figure 10.3: Asymmetric ripple morphology from (Tucker, 2001).

The *crests* of ripples in plan view may have any of a number of shapes. Those with straight crests are said to be two dimensional and form in currents with little lateral motion. A common environment of such straight-crested ripples is in the surf zone on beaches. The remaining types of ripple crests reflect three dimensional ripples and form in flow with a degree of lateral motion, in addition to the dominant downstream component. Ripples still migrate in the dominant, downflow direction but are geometrically more complex.

- straight
- sinuous
- catenary
- lingoid

- lunate

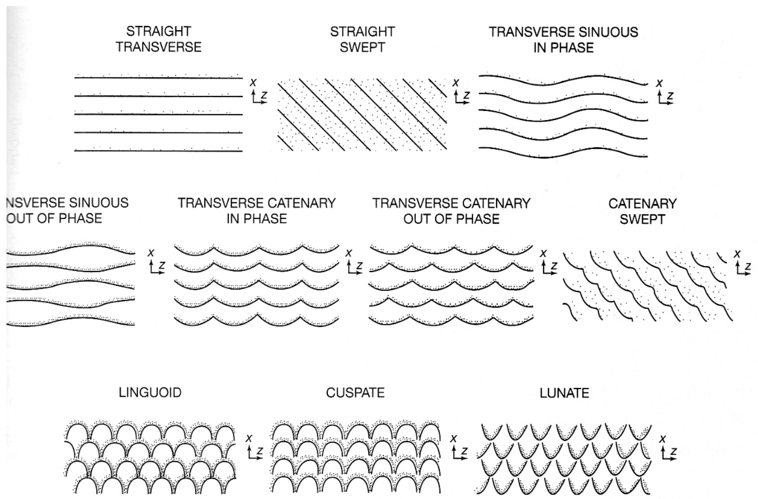



Figure 10.4: Types of ripple crests (Boggs, 2011).

*Climbing ripples* are ripples that form when deposition rates are high, resulting in excellent preservation of ripples that actually 'climb' in the downflow direction.

- subcritical
- critical
- supercritical

The angle of climb of climbing ripples depends on the ratio of bed aggradation and the speed of ripple migration.

*Starved ripples* form where there is a minimal amount of available sediment for deposition and are hence 'encased' in finer grain sediments (that settled out following the event that led to the formation and migration of the ripple).

Interbedded muds and ripple cross-laminated sands or silts form distinctive bedding types typically environments with fluctuation flow conditions (mostly tidal flats and channels), distinguished the relative abundance of sand/silt to mud:

- Flaser bedding
- Lenticular bedding
- Wavy bedding

**Oscillation flow and symmetric ripples**

Ripples may form under oscillatory flow as well as unidirectional flow. Such ripples are known as *oscillation* or wave ripples and form as a result of flow from alternating directions where the orbital flow paths below a passing wave intersect the bedding surface (*shallow water waves*). Where sediment deposition is occurring and there lateral migration of the ripple, it forms *chevron* interlaminae, that are sort of woven together from two opposite directions, the result being a nearly symmetric, *trachoidal* ripple, with a steep crest that is perpendicular to the direction of wave propagation. Where active deposition is not occurring, the oscillatory motion may cause scouring of the underlying bed, also generating symmetric, wave-ripple forms, even if the lamination within these structures is not chevron. Symmetric ripples typically have lower ripple indexes than current generated ripples, and water depth, wave period, and fetch can be estimated based on the ripple index, ripple steepness, grain size, and grain density (Allen, 1981). As a general rule, *wave base*, the depth to which surface (oscillatory) waves can entrain sediment is about half the wavelength of the waves (Reading and Collinson, 1996).

### 10.3.4 Cross-stratification

*Cross-stratification* is a bedding structure in which layers locally intersect bedding planes at an angle. It is an easy-to-recognize structure, is generally aesthetically pleasing, and lends itself to straightforward interpretations of the general physical processes responsible for deposition. Hence, it is of singular importance and interest in sedimentology and no surprise that geologists are attracted to cross-stratification like flies to fly paper.

Cross-stratification occurs at many scales. Typically, *cross-lamination* refers to thin cross-stratified beds or laminae (think ripples) (or *cross-bedding* refers to beds several cms to decimeters thick that contain cross-strata. *Large-scale cross stratification* roughly refers to beds that are several decimeters to meters thick.

Cross-stratified deposits are typically arranged in *sets* of conformable laminae separated from adjacent sets by *truncation surfaces* or other non-depositional surfaces. A *set* is a succession of at least two laminae separated from other sets or beds by a truncation surface or other bounding surface. The laminae are most commonly concave up or planar.

Cross-lamination or cross-bedding normally forms during the migration of ripples or dunes. The sloping beds in cross-bedded sediments are referred to as the *foresets*. The bed or set that the foresets intersect below is known as the *bottomset*. Importantly, the geometry of the preserved cross-stratification boundary is largely determined by the geometry of the troughs of the ripple or dunes, rather than the crests, since this crests are almost always eliminated by the migrated troughs. As a consequence, the height of the cross-strata is always lower than the height of the bedforms that created it.

The actual pattern of cross-bedding and the bounding surface between cross sets can be very complex in 3D flows (those in which the fluid path is sinuous), giving rise to curvilinear bottomsets. See the website **http://walrus.wr.usgs.gov/seds/bedforms/index.html** for excellent graphical representation of these variations.

- Planar cross bedding (dip $\geq 30°$)

  - tabular sets

- Trough cross bedding
- Herringbone cross bedding
- Reactivation surface
- Lateral accretion surfaces

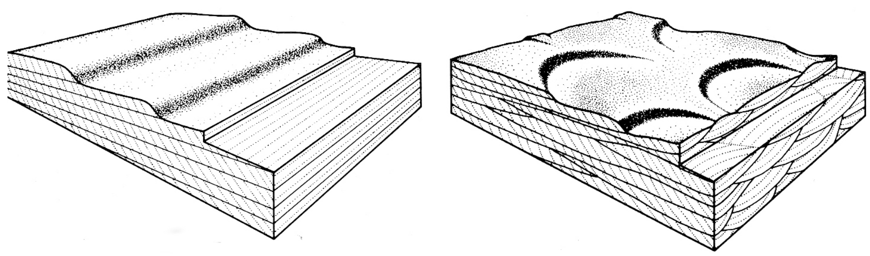


Figure 10.5: The difference between planar and trough cross bedding (Tucker, 2001).

Sometimes, you get lucky and see a planar view of cross-lamination, parallel to the overall flow direction, in which case you may see *rib and furrow* structure, in which the furrows are concave structures that record the migration of a 3-D trough and the ribs separate the migrating trough sets. This structure is excellent for determining paleoflow direction, which is in fact often difficult with 3-D bedforms because we are only allowed two-dimension views at the whim of the local joint orientation or roadcut, which frequently is insufficient for determining paleocurrent direction.

### 10.3.5 Combined flow

*Combined flow* occurs when oscillatory and unidirectional flows interact. This happens, most commonly, under stormy conditions, where coastal set-up drives a unidirectional flow that interacts with the oscillatory flow generated by the storm waves.

#### Hummocky cross stratification

The most common sedimentary structure formed under combined flow conditions is *hummocky cross-stratification* (HCS), which is somewhat difficult to define precisely, but is a unique type of cross-stratification that is commonly isotropic, meaning the orientation of the cross strata is effectively randomly distributed. Other distinctive features of HCS include

- curved, fanning laminations
- low angle lamina intersections
- sharp, often erosional, lower bounding surfaces
- concave up cross-sets that grade up into convex up cross-sets, and vice versa

- long wavelengths (∼0.5 - 2.0 m)

For a long time, the actual conditions under which HCS formed where roughly known from intuition, but they had not been observed to form naturally or in experiments. This has finally changed, and it is has now been determined experimentally that cross-beds form under long wave period, moderate velocity oscillator flow combined with a weak unidirectional flow in rapidly aggrading sediments of silt to fine sand grain size (Dumas and Arnott, 2006). These conditions are met most typically in the mid-continental shelf, just above what is mostly commonly referred to as *storm wave base*, which the wave base attained during storms. The distinction between storm wave base and *fair-weather wave base* is perhaps not so clear cut as we'd like to think, since the frequency with which passing waves on a continental shelf will effect a particular depth is a governed by a power law (where the probability of very deep wave base is infrequent and usually occurs during storms) (Peters and Loss, 2012). Nevertheless, storm weather wave base, as the cut off between where HCS forms and does not form (Sageman, 1996), us a useful reference point and is typically about 15-50 m for a strong storm. However, the storm weather wave base is highly variable, depending on the geometry and restriction of a basin (Peters and Loss, 2012).

Storms generate larger, long wavelength waves and an underflow that flows from the shore towards the shelf due to set up of the storm surge. A slight increase in the unidirectional current velocity results in anisotropic HCS, and consequently, one might expect to encounter isotropic HCS nearer storm wave base, and anisotropic HCS somewhat shoreward of wave base.

### Swaley cross stratification

*Swaley cross-stratification (SCS)* is similar to HCS but forms in fine to medium sand, and is distinguished by its "series of superimposed concave upward shallow scours about 0.5-2.0 m wide and a few tens of centimeters deep" (Leckie and Walker, 1982, p. 143). The angle of cross-laminations is typically lower and basal surfaces are erosive (thus it is defined more by swales than hummocks). The experiments of (Dumas and Arnott, 2006) have demonstrated that it occurs under conditions found at shallower depths than HCS, closer to *fair weather wave base*, where erosion and sediment transport is more important. This finding confirms evidence from the sedimentary record that SCS occurs between HCS and typical shore face facies in *shallowing-upward* sequences (Arnott, 1992).

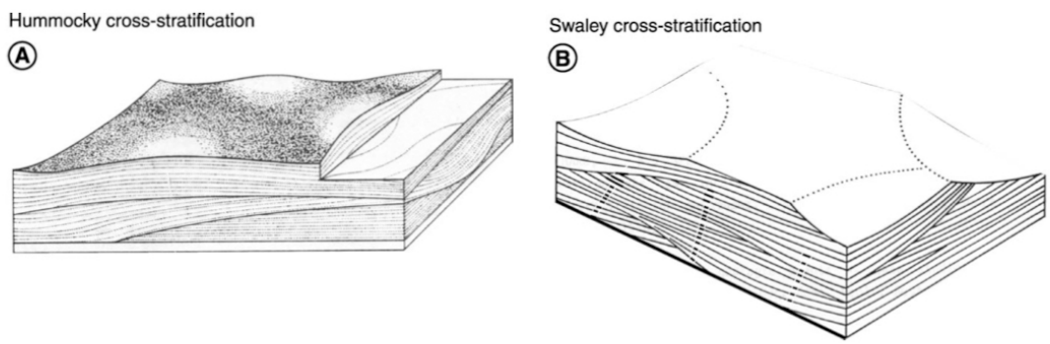



Figure 10.6: Three-dimensional depiction of laminations and bounding surfaces in hummock cross-stratification (HCS) and swaley cross-stratification (SCS) (Dumas and Arnott, 2006),

# Chapter 11

# Other Sedimentary Structures

Additional reading: Tucker, Chapter 1 (1-9), Chapter 2 (21-26); Boggs, Chapter 4

## 11.1 Introduction

Sedimentary structures are macroscopic features of sedimentary rocks. They are formed by a wide variety of processes, both primary (during or soon after deposition) and secondary (during diagenesis or subaerial exposure). Being able to recognize sedimentary structures in the field is satisfying and extremely useful because they are key to interpreting the environment of deposition. Cross-lamination, parting lineation, imbrication, and other structures may be used to infer palaeo-flow direction. Some structures may indicate "way up," which can be particularly useful in structurally complex terranes. This lecture is a natural continuation of the discussion of stratification and bedforms in the previous section, and, in combination with practical exercises and field experience, should render you capable of identifying prominent sedimentary structures in the field and interpreting their setting of deposition. We will not cover all sedimentary structures here and others will arise subsequently in the course, particularly when we cover carbonates. Like much of geology, learning to identify and utilize sedimentary structures confidently requires spending time in the field.

## 11.2 Bedding Plane Structures

### 11.2.1 Large erosional structures

Erosional structures tend to involve shifting or unusually strong current action (e.g. during a storm event) and result in erosion of the sediment bed.

- Channels (common in fluvial and tidal environments)
- Scour and fill structures (common in sandstones), very rapid events

### 11.2.2 Sole marks

*Sole marks* are geometric sedimentary structures that occur as the result of erosion on the base of beds as the result of

- strong currents (*flute marks*)
- mechanical disruption (*tool marks*)

### 11.2.3 Other bedding-plane structures

- Groove casts
- Current crescents
- Parting lineation
- Pebble imbrication

## 11.3 Soft Sediment Deformation

Soft sediment deformation occurs penecontemporaneously with deposition and involves folding and contortion of bedding of unconsolidated or semi-consolidated bedding due to gravity.

### 11.3.1 Structures that result from gravitational or seismic instability

- Convoluted bedding
- Load casts
- Ball and pillow structures
- Flame structures
- Synsedimentary folds and faults
- Dish and pillar structure

### 11.3.2 Other penecontemporaneous structures

- Rain imprints
- Halite casts
- Mud cracks
- Mud-clast breccia
- Syneresis cracks

## 11.4 Secondary Structures

### 11.4.1 Structures that form during early burial and diagenesis

- Sandstone dikes and sills are tabular sand bodies that result from injection of liquified sands into surrounding sediments.
- Concretions are an early diagenetic structure resulting from partial cementation of a sediment, commonly in the form of a spherical to oblate structure that stands out (or preferentially erodes out) when the rock is later exposed to weathering.
- Sand crystals
- Styolites

## 11.5 Sedimentary Structures in Carbonates

Carbonates contain many of the same sedimentary structures found in siliciclastic sedimentary rocks. For example, ripples and trough cross-bedding are relatively common in *grainstones*, which like sandstones, are typically deposited in relatively high energy environments. Carbonates also contain certain unique sedimentary structures.

- tepee structure
- molar tooth structure
- *Styolites* from by pressure solution, where insoluble residue (mainly organic carbon and fine detrital material, such as clays) is concentrated in irregular seems that are typically parallel to the bedding surface (where burial causes the pressure necessary to dissolve carbonate grains). Styolites are not unique to carbonates, but are more conspicuous and spectacular in carbonates than siliciclastic sediments.

## 11.6 Biogenic Structures

Organisms can create a variety of structures on and in sediments, such as trails, bores, and burrows through the burrowing, feeding, locomotion, and other activities (Boggs, 2011). These activities may largely obliterate older sedimentary structures and stratification and replace them with their own biogenic structures, or *trace fossils*. The study of trace fossils (*ichnology*) is important both in sedimentology, where they can be used to infer sedimentary environments, and in paleontology, where they can often be linked to specific groups of organisms and used to reconstruct habitat and life habits. As you might expect, the phenomena of *bioturbation*, which a type of trace fossils than involves disruption to existing sediments, was of minimal importance through most of Earth history, and rose to its prominent place in sedimentary geology in the latest Proterozoic, with the evolution of macroscopic animals.

Trace fossils, which are indicate the activity of organisms without being a true body fossil, can be subdivided into three classes of structures:

- bioturbation structures (e.g. burrows, tracks, and root penetration structures)

- bioerosion structures (e.g. borings, bitings)
- excrement (coprolites)

Trace fossils are commonly subdivided into *ichnogenera*, based on inferred characteristics of the organisms that generated them. For our purposes in sedimentology, the more useful approach to trace fossils is to lump them into *ichnofacies*. Ichnofacies are assemblages of trace fossils often found together that are typical of specific sedimentary environments. In the marine realm, there are six reasonably well established ichnofacies (Fig. 11.6:

- *Trypanites*: borings on rocky coasts and hardgrounds
- *Glossifungites*: burrows and dense burrows in semi-consolidated sediments
- *Skolithos*: Vertical burrows, U-shaped or cylindrical, in soft beach sands
- *Cruziana*: Mixed horizontal, vertical, and inclined burrows and highly diverse tracks in soft sands and muds in the *sublittoral* zone (that is, close to shore, but just below the low tide)
- *Zoophycos*: Simple to somewhat complex, horizontal to slightly inclined feeding and grazing structures in mud in the offshore to slope environment
- *Nereites*: Complex, horizontal crawling and grazing traces in slope-abyssal muds

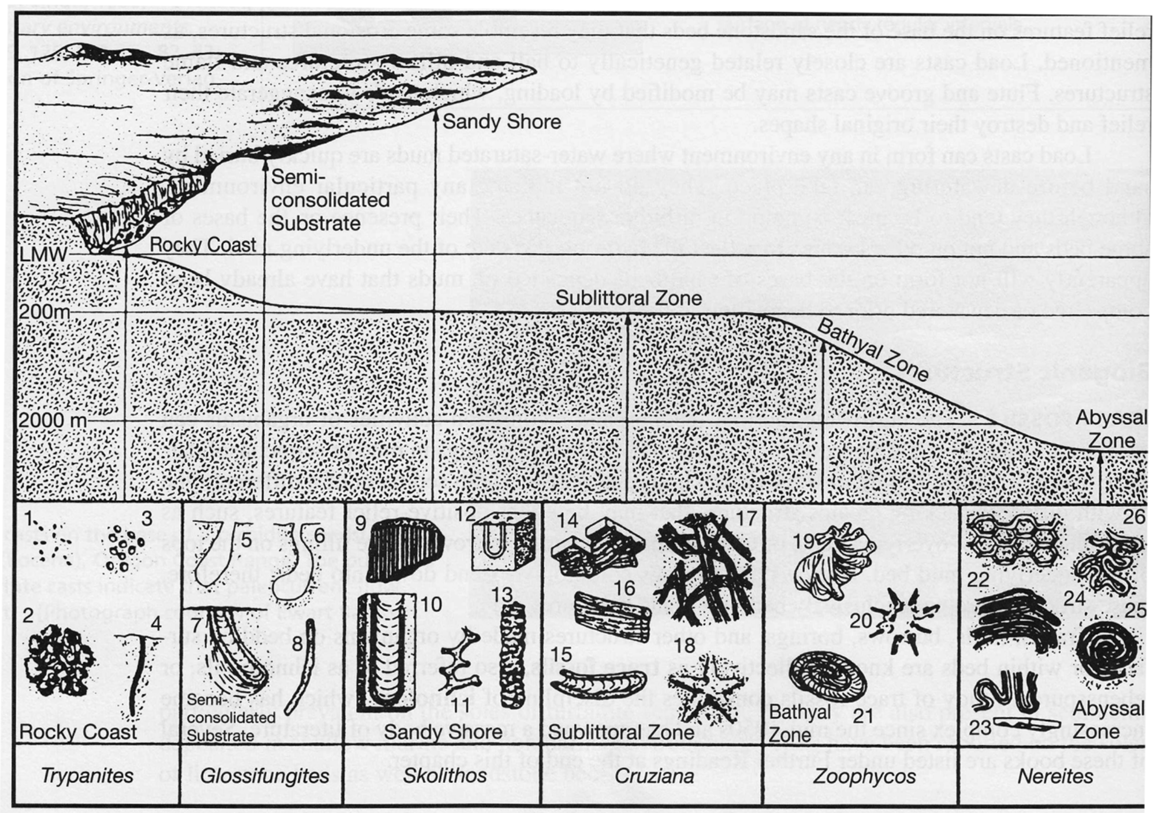



Figure 11.1: The six ichnofacies that commonly occur in the marginal marine to deep marine realm (from Boggs, 2011).

# Chapter 12

# Stratigraphy and Introduction to Depositional Environments

## 12.1 Introduction

Through the first half of this course, we have focused on sedimentary processes and sedimentary rocks as you might seen them in hand specimen or at the outcrop scale. However, most geologists that work on sedimentary rocks are interested in how these rocks occur over larger scales. Hence, most sedimentologists are also stratigraphers and have plenty of work given that the majority of the Earth's surface is covered in sediments and sedimentary rocks. *Stratigraphy* is the description, correlation, and interpretation of stratified rocks. Given this definition, stratigraphy includes the study of volcanic strata, but in this case, would tend to be more focused on the correlation and geometry of volcanics rocks and less on their petrogenetic evolution.

Stratification implies detectable difference in layers of sediments. These are inevitable in the deposition of sedimentary rocks given that conditions are never constant on the Earth's surface over long time scales. Many different processes give rise to stratification, but fundamentally, it is produced by

- variations in continuous depositional processes through time, due, for example, to changing environmental conditions or sediment source areas.
- brief and often catastrophic depositional events that interrupt slow, continuous deposition. *Event* deposition is most commonly initiated by storms, floods, earthquakes, and tsunamis.

*Turbidites* are the quintessential *event beds* and effectively represent a blink of a geological eye. That is, for anybody trying to use the stratigraphic record as a temporal yardstick, it is fundamentally important to take the time scale of event bed deposition into account. To understand this, imagine a deep, quiet water environment in which *pelagic* muds are accumulated impossibly slowly (say, a few mm's per thousand years). If you had a few ages on this interval of muds, it would be reasonably straightforward to produce a time-depth model for the section and convert stratigraphic thickness to time. However, if this deposition were punctuated by a turbidity current, which deposited a turbidite that was 10 cm thick, then you would have to take this into account in your chronology. Otherwise, given the slow background sedimentation rates, you would attribute several tens of thousands of years to a bed that took only a few minutes to deposit.

You can probably imagine a few other process that would result in an event bed, but in case none come to mind immediately, here are a few ideas

- A basalt flow that spills out over the land surface
- A storm the deposits sands on the middle to outer shelf, where sediment is otherwise muddy
- A flood that results in a layer of gravelly sediment
- A large volcanic eruption that deposits a thick layer of ash
- A debris flow diamictite deposited in response to a surging glacier

Despite the fact that most sedimentologists recognize that a large part of the sedimentary record comprises event beds, we are still guided by one of the underlying principals in geology: the present is the key to past. That is, we understand how the ancient Earth operated by witnesses processes active today. This is a reasonable enough maxim, but it is not exactly right. Bridging the ancient sedimentological record with active processes suffers from the geologic version of the Heisenberg Principle. Event deposition is extremely difficult to observe, and even where the process can be observed and quantified, the product is difficult to described. Ironically, the products of such processes are much easier to see and describe in the ancient record, but the process cannot be quantified accurately.

## 12.2 Sedimentary Sections and Facies

You already have some experience with sedimentary sections. A *local section* is a single sequence of sedimentary strata at the outcrop scale. It can be logged and described, just as you did in the field, and the stratigraphic log provides a one dimensional (spatially speaking) view of the sedimentary strata in that region. A single line of section will usually provide a decent picture of the local overall sedimentary succession, because lateral variations in *facies*, that is the suite of lithological and structural features of sedimentary rocks (usually tied to an interpretation of the environment in which those sediments were deposited), are typically not as great as vertical (temporal variations).

In spite of what I just said in the previous sentence and what you have learned from Steno's laws, sedimentary rocks do vary not only in time, but also spatially. The rate of change (as a function of stratigraphic height or lateral distance) is of course highly variable, depending on depositional environment. It might be very low in deep sea sediments, but will typically be relatively high in continental deposits, in particular in basins adjacent to areas of of high topographic relief. The consequence is that rarely do you encounter classic layer-cake stratigraphy, where single, lithological units can be reliably *correlated* between outcrops and across the basin. Rather, specific lithological units tend to be *diachronous*, because given sedimentary facies appear are spatially restricted. Imagine a beach facies. This sedimentary environment may only be several 10s of meters in width. Hence, if you see beach sands in one local section and can trace out a time boundary within this unit laterally (for example, and ash bed), you would find that it would transition laterally into other facies. So for example, as you moved distally in the basin, you would end up in the sublittoral facies, then eventual in the offshore facies. However, you could probably follow the beach sand laterally, it would just either drop or rise in the sedimentary section relative to the fixed time boundary. That is because sedimentary environments tend to migrate spatially in time, because sedimentary environment is a function of multiple factors, such as sediment input rate, changes in relative sea level, and the rate at which accommodation space is produced.

### 12.2.1 Transgression

*Retrogradation* is the shoreward migration of sedimentary environments. This occurs as a consequence of *transgression*, which fundamentally requires a rise in local sea level (otherwise known as an increase in *base level*. That is, there is more space for sediments to be deposited .

### 12.2.2 Regression

*Progradation* occurs when sedimentary environments step out towards the sea. This reflects a *regression*. Regressions may happen for the opposite reason that transgressions, happens, that is, a loss of accommodation space due either to falling sea level or regional uplift. When regression occurs due to loss of accommodation space, it is called a *forced regression*. However, progradation can also happen simply as the consequence of build-up of sediments as occurs at the mouths of active deltas, even if global sea level is rising slowly. *Normal regression* thus occurs where sedimentation rates exceed the rates of production of new accommodation space.

### 12.2.3 Walther's Law

Retrograding and prograding shorelines give rise to stacked sedimentary facies. When we see a succession of facies in sedimentary section (uninterrupted by unconformities), *Walther's Law* tells us that this reflects sedimentary environments that occurred adjacent to one another. Hence, we can use Walther's Law to reconstruct depositional environments and whether rocks were deposited during a transgression or regression. And even though it should now be evident that the inherit instability of many depositional environments in time makes correlating sedimentary strata based on lithology alone is prone to error (assuming the goal is to correlate time-equivalent strata), these same fluctuations yield patterns that are often correctable (for example, sharp transgressive or regressive surfaces).

### 12.2.4 Sedimentary contacts

We have already learned about the major sedimentary contacts with respect to unconformities. But often contacts between two formations (or members, or beds) are not depositional hiatuses or erosional surfaces, but rather record a shift from one sedimentary regime to another, either transitionally or abruptly. The main type of contacts between rock units are

- gradational by continuous change
- gradational and interbedded
- abrupt change in style or type of deposition
- abrupt by unconformity

## 12.3 Cyclicity and Sea Level Change

One common feature in sedimentary sequences is that that the strata appear to be arranged in repeating, or at least similar packages. For example, you may see a whole series of meter-scale fining-upward packages. The repetitions may occur at a variety of scales, from less than a meter, to hundreds of meters. Where these repetitions show similar thicknesses or systematic variations in thickness, they may be cyclical, implying that they are controlled by some time-dependent parameter. However, it is important not to confuse rhythmic-appearing patterns with true cyclicity. For example, a succession of turbidites in a deep-sea environment may look rhythmic (indeed, fine turbidites are commonly referred to as *rhythmites*), but in fact these represent stochastic depositional processes. If you tried to do a time series analyses on turbidite beds, you would find that they would not behave very well.

Cyclicity is most commonly the product of change in sea level. As a reminder, three principal factors give rise to sea-level (or better yet, base level) change:

- Vertical tectonic or geodynamic movements of the lithosphere: subsidence or uplift
- Compaction of the sedimentary column
- Global changes in sea level–*eustasy*

Water depth is different from sea level, and is strictly the distance between the water surface (the geoid) and the sea bed. Water depth can decrease, even as global sea level rises, if sedimentation rates more than compensate for that rise in sea level. Nevertheless, sudden rises in sea level tend to lead to transgressions and sudden drops tend to drive regressions.

Intervals of high sea level (following the peak transgression), are referred to as *highstands*. Intervals of low sea level, following sea level lows, are referred to as *lowstands*.

## 12.4 Sedimentary Depositional Environments

A *depositional environment* is a unique geographic setting characterized by a distinctive set of physical, chemical, and biological conditions. The term can be applied rather broadly, such as comparing the marine environment to the terrestrial environment. But more often, sedimentologists distinguish environments at a smaller scale, so a deltaic environment versus a beach environment.

Most environments that we can observe today are not depositional environments. Rather, they are erosional environments. Hence, depositional environments require that sediments accumulate over the long term (long enough to be preservable in the sedimentary record). Depositional environments, as we tend to define them, have lots of overlap, since individual environments do not typically exist in isolation. The way we distinguish them is not always based on any obvious physiologically defined limits or boundaries. Some depositional environments are common; others are rare. Probably, we do not know all of the possible depositional environments that had existed through Earth history.

We commonly lump depositional environments within broader categories:

- Continental
  - alluvial
  - fluvial
  - aeolian
  - lacustrine
- Marginal marine
  - beach
  - delta
  - lagoon
  - estuary
- Open marine
  - neritic (sublittoral continental shelf)
  - continental slop
  - abyssal (deep marine)
- Carbonate platform
  - reef (or back-reef, reef, and fore-reef)
  - bank
  - ramp

You already know about sedimentary facies. Individual sedimentary facies are typically allied with a depositional environment, but in fact, a given depositional environment will commonly have several separate facies (depending, of course, on how they are defined, which is highly arbitrary). Also, some facies might be be quite similar, such as cross-bedded medium sands in a tidal channel versus on a point bar, and if seen in isolation, may easily be misinterpreted.

In this case, it might be useful to define a *facies association*, which is a group of sedimentary facies, that together point more obviously to a specific depositional environment. For example when I map out 1) supritidal microbialaminite facies, 2) interclast wackestone, and 3) laminated black limestone facies in close physical association, I might reasonably interpret these to have been deposited adjacent to each other in a back reef environment. However, any individual facies could conceivably have been deposited in a different carbonate platform environment.

# Chapter 13

# Continental Deposits

Additional Reading: Boggs, Chapter 8 (241-275); Tucker, Chapter 2 (65-71); Bridge and Demicco, Chapters 13, 16

## 13.1 Introduction

Continental (or terrestrial) depositional environments can be broken down into fluvial, aeolian, lacustrine, and glacial systems (although glacial deposits also occur in the marine environment and all glacial deposits will be discussed in a separate chapter).

Continental sedimentary deposits comprise only a small percentage of all sedimentary rocks, but are important in parts of the geologic record that record important paleoenvironmental and paleobiological information. Some specific terrestrial sedimentary systems of significant geological interest include

- Tertiary fluvial systems of the Rocky Mountains - Great Plains
- Tertiary lacustrine deposits in Wyoming and Colorado (Green River Formation)
- Jurassic aeolian sandstones of the Colorado Plateau and southwestern Africa
- Permo-Carboniferous glacial deposits of Gondwanaland

## 13.2 Fluvial Environments

Rivers are the conduits by which most sediments are transported from the continents to the oceans. For this reason alone they are of central importance to sedimentary geology. Fluvial deposits contain important resources, including petroleum, coal, placer minerals, and of course, water. Rivers are also heavily exploited by humans, for irrigation, transportation, and drinking water. Consequently fluvial systems have been intensely studied by engineers and geologists alike.

Rivers form anywhere continental crust is above sea level, and hence are ubiquitous. The typical river flattens out along its course and increases in volume, and hence, sediment load. Deposition in the fluvial environment (note that the term *alluvial* is often used interchangeably with *fluvial*, but more specifically, refers to a river that is cutting through its own sediments) occurs where sedimentation outpaces erosion. Significant thickness of fluvial sediments may accumulate adjacent to regions of active uplift, such as in the basins that separate the mountain ranges int he *Basin and Range* province in the western United States.

The fluvial environment comprises many subenvironments related to river and stream processes, but can broadly be broken down into two subenvironments:

- Alluvial Fans
- Rivers

### 13.2.1 Alluvial Fans

*Alluvial Fans* are fan- or cone-shaped (in plan view), convex-up (in longitudinal cross-section), amalgamations of course sediment deposited where steep-gradient streams spill out into a broad, shallowly-dipping, environment.

- Common in areas of high relief (i.e. mountain ranges)
- Often form downward of major fault scarps
- Spectacular examples occur in extended terranes in arid environments
- But also occur in both arid and humid environments of high relief that experience periods of intense rainfall
- *Bajadas* are coalesced alluvial fans
- *Fan deltas* occur where an alluvial fan builds into a standing body of water

Modern alluvial fans are easily identified by their morphology

- arcuate in longitudinal section
- lense-shape, convex-up in transverse cross-section
- dominated by gravelly to bouldery deposits
- lobes and distributary channels common and fan-out downslope
- down-fan decrease in grain-size and bed-thickness, increase in sorting (grade into lake or river deposits

In the sedimentary record, they are most easily identified by the coarse-grain sizes, abrupt facies changes, and common close association with faults.

The fist-order process responsible for alluvial fan deposition is the sudden spread of stream-flow upon exiting a high-relief stream channel. This commonly occurs during storms or seasons of high rainfall or snowmelt.

- stream-flow on alluvial fans

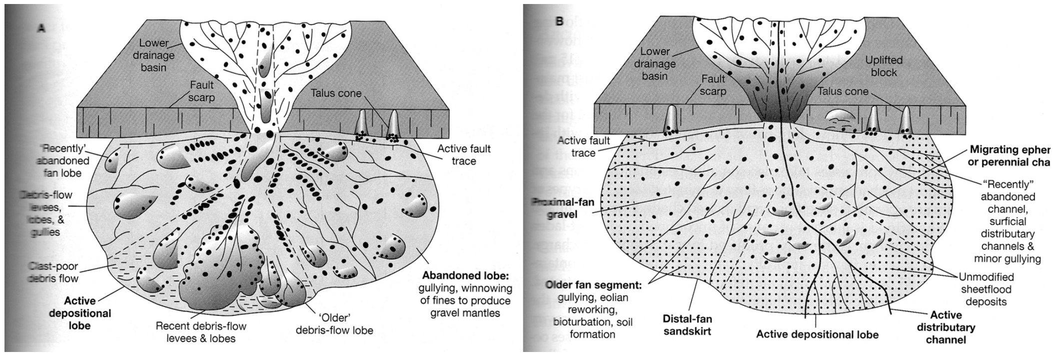



Figure 13.1: Alluvial fans can be subdivided into debris-flow and stream-flow dominated end-members. From (Boggs, 2011).

  - *sheetflood* flow is rapid, broad, unconfined, shallow, sediment-laden flow
  - *incised-channel flow* occurs in channels incised into the upper fan.

- debris- and mud-flow on alluvial fans
  - more common on steeper slopes of fans
  - form very poorly sorted deposits that freeze up very quickly

- alluvial fans further modified by vegetation, winnowing, pedogenesis (often, the formation of calcretes), landslides

In my experience I have found that alluvial fans are not uncommon in the sedimentary record, but usually only a fragment of the fan is preserved. Tectonic activity often generates accommodation space adjacent to uplift, which is both the perfect environment for creating an alluvial fan and for burying it and preserving it in the sedimentary record. Conglomerates that are interpreted to have been deposited in an alluvial environment are often referred to as *fanglomerates*. However, they can be difficult to distinguish from coarse braided river deposits. Distinct features to look for include

- abundance of coarse-grained, discontinuous units
- including debris flow deposits (e.g., diamictites, commonly with winnowed tops)
- lateral facies changes in direction of paleoslope, from coarse to fine
- distinct wedge-shaped geometry (from mapping)

### 13.2.2 River Systems

In general, rivers are more important as conduits for sediment transport than as sites of permanent deposition. However, significant river systems are preserved in the sedimentary record. Morphologically, rivers can be described according to their

- *sinuousity* - degree of deviation from a straight line
- number of channels (single or multiple)

- degree of braiding (that is, channel subdivision by bars and islands)

In practice, rivers are often classified as either *braided* or *meandering*. In general, there is a continuum from alluvial fans, through braided rivers, to meandering rivers. However, the underlying controls on why a river is braided versus meandering is only partly known. The major contributing factors are believed to be

- Magnitude and variability of stream discharge
- Channel slope
- Channel roughness
- Type of sediment load (bed load versus suspended load)

But these factors are really all inter-related:

- grain size $\propto$ channel slope
- channel roughness $\propto$ grain size
- degree of braiding $\propto$ water discharge/given slope
- sinuosity increases with width/depth ratio in single channel rivers
- sinuosity decreases with width/depth ratio in multiple channel rivers

A reasonable simplification is that

- braided rivers occur in higher relief regions with an abundant supply of coarse sediment
- meandering rivers occur in lower slope, downstream parts of the river system dominated by medium- to fine-grained sediments

### Sedimentary processes in river systems

River systems consist of three main elements: *channels, bars, and floodplains*. Sediment transport and erosion occurs mainly in channels, coarse-grained deposition occurs in bars, and fine-grain deposition occurs in floodplains. The route of channels is in large part set during period of high discharge. Downstream flow around channel bends leads to *helical flow*:

- erosion around outer bend
- sediment transport along stream channel to inner bend, where they accumulate laterally
- *point bars* are attached to the river bank and show lateral accretion surfaces, cross-bedding, and an overall fining-upward grading

*Braid bars* occur mid-channel, mainly in braided rivers:

- lateral accretion

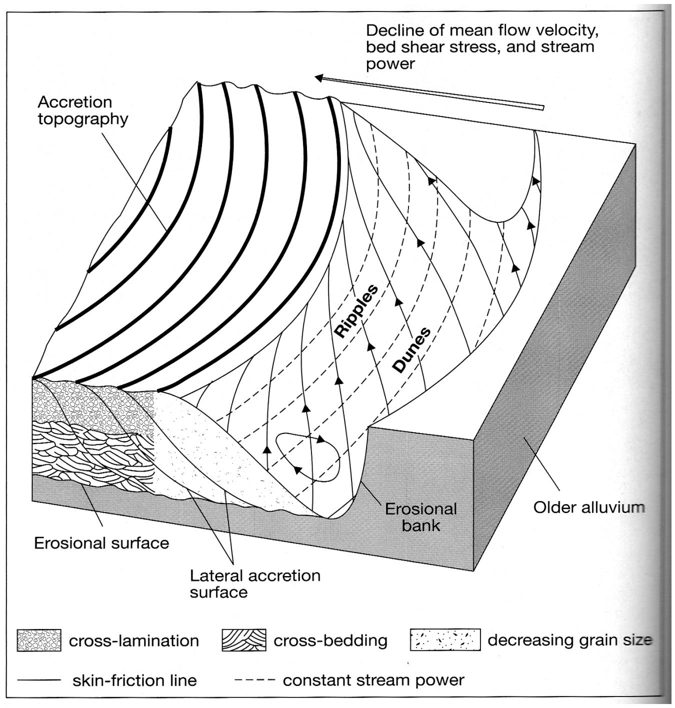



Figure 13.2: Schematic diagram of helical flow around a bend. From Leeder (1999).

- downstream migration

*Floodplains* are areas of lowland around a river that are inundated during floods. Deposition also occurs in the floodplains of rivers. Indeed, in an aggrading river system, deposition in the flood plain and in the channel must be in equilibrium. Deposition during flood events includes

- *natural levee* deposits on eroding banks of loops: stratified sands overlain by mud
- flood basins fill with fine grained muds that settle out of suspension: commonly bioturbated
- *crevasse-splay* deposits occur where flood waters breach natural levees; they produce distinctly fining-upward sequences not dissimilar to a bouma sequence
- *avulsion* is the abandonment of channels; this may lead to creation of *oxbow lakes*, with then fill with fine-grained, laminated muds

General characteristics of river deposits

- mainly sands and gravels (bars) and muds (floodplains)
- moderate to poor sorting
- compositionally immature sediments
- fining-upward beds common (point bars, crevasse splays)

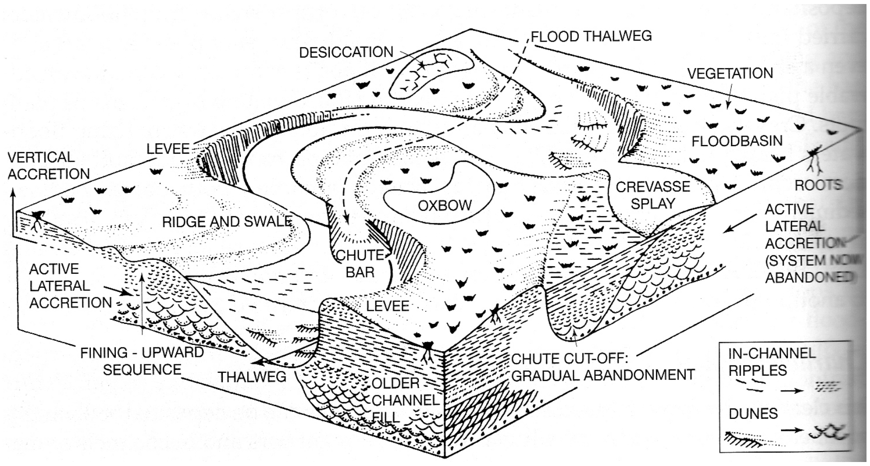



Figure 13.3: Schematic diagram of the elements of a meandering-river system. From Walker and Cant (1984).

- channel lag deposits
- abundant traction bedforms and structures (trough cross-beds, ripple marks)
- unimodal paleocurrent indicators (often with high dispersion)
- stacked coarse channel/bar deposits, separated by muds
- sharp facies changes

With sufficient data, braided streams can be distinguished from meandering streams. Braided streams are dominated by deposition in the channel and bars, hence braided river depositions consist mainly coarse grain sediments deposited by traction currents (Figs. 13.2.2, 13.2.2). Individual channels *elements* are stacked and flood plain muds form lenses between channel elements. In meandering river systems, floodplain deposition of fine-grained sediments is important and channel elements tend to be separated by floodplain muds.

## 13.3 Arid Environments

When we think of deserts, we tend to envision dust storms and sand dunes. Indeed, aeolian (wind) processes are the most important in many desert environments (leastwise, in producing sedimentary deposits that are preserved in the stratigraphic record). However, there are many subenvironments in deserts:

- alluvial fans
- ephemeral streams
- ephemeral saline (*playa*) lakes
- *deflation surfaces* (sediments blown away)

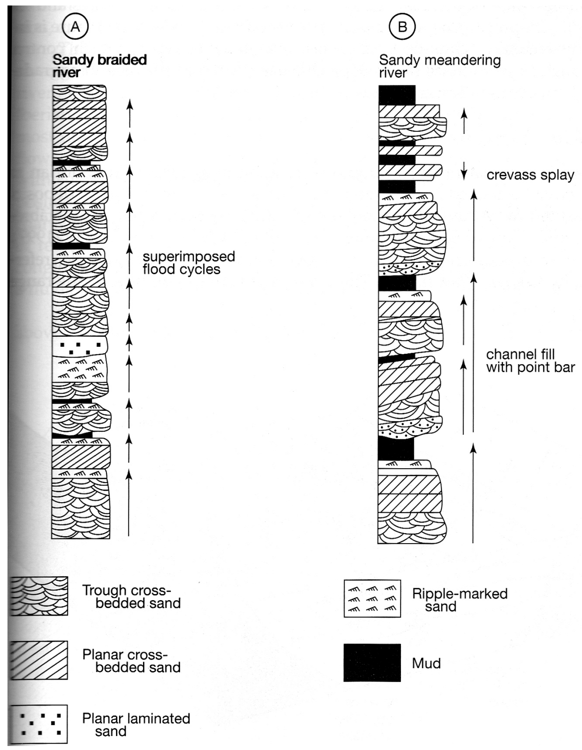



Figure 13.4: Idealised stratigraphic profiles through braided and meandering river sediments (Boggs, 2011).

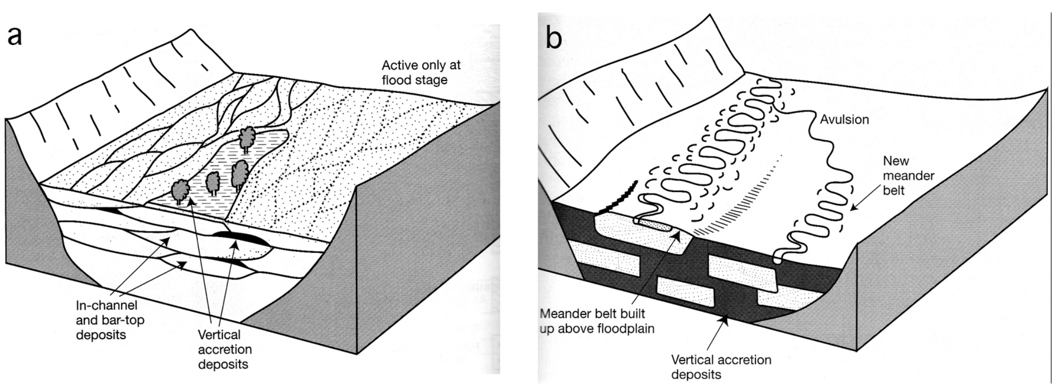



Figure 13.5: Architecture of braided-river (a) and meandering-river (b) systems (from Boggs, 2011).

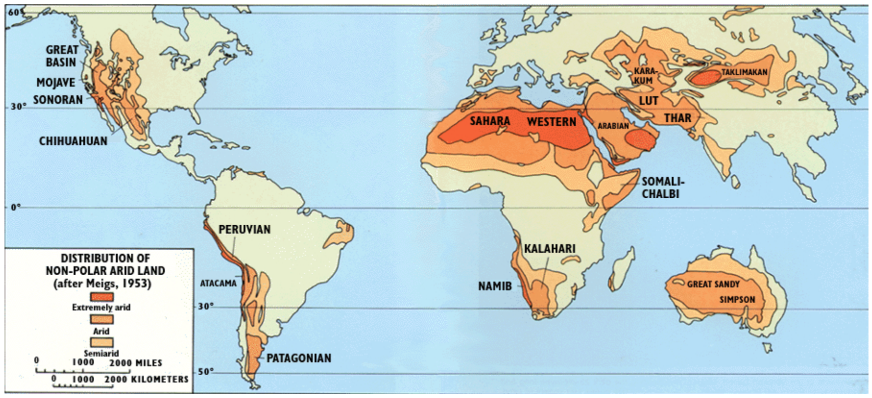



Figure 13.6: Global distribution of deserts.

- sand dune field
- sand seas (*ergs*)

Deserts cover about 30% of Earth's land surface (Fig. 13.3) and the entire surface of many other planets. by a definition of desert based on net precipitation, include much of the Arctic and Antarctic. However, eolian features only develop in significant ways in regions where a protective cover of vegetation (or ice) is sparse and winds can pick up and transport dust and sand (Brookfield and Silvestro, 2010). Hence, for the purpose of this section we are concerned here with the hot and dry variety of desert, which is concentrated in the mid-tropics (10-30° latitude).

### 13.3.1 Non-aeolian depositional influences in deserts:

- light vegetation, sporadic, but often intense rainfall → high erosion rates
- flash floods (including sheet floods) common
- evaporite deposition in playas

### 13.3.2 Aeolian depositional processes in deserts:

Wind can effectively transport loose sand and fine-grained sediments through traction, saltation, and suspension. Separates fines (>0.05mm) from courser sediments. Wind is an important agent of sediment transport in

- deserts
- periglacial environments
- coastal environments
- open ocean (sometimes main source of pelagic sediment)

Wind generates three types of continental deposits:

- *Loess* accumulations of dust (silt)

- deflation pavement
  - *ventifacts* are stones that have been faceted by impact wind-blown sand grains
- moderately- to well-sorted sand deposits
  - ripples
  - dunes

Wind ripples are somewhat analogous to those formed in water. Grain motion (up to granular size) is largely the result of a cascade of grain-grain interactions, which disperse kinetic energy

- Wavelengths of 0.02–2.0 m
- Heights of 0.005–0.1 m $\rightarrow$ rippled indices of 8–50
- Lack clear internal cross-lamination, due to migration by saltating grains rather than traction. That is, they do not build and migrate via grain flows in the same way that subaqueous ripples do
- Often display bimodal grain-size due to concentration of coarser grains on ripple crest and finer grains in troughs

Aeolian dunes are much more diverse morphologically than water-lain dunes

- accrete in down-wind direction by a combination of
  - accretion (creep, saltation) on the stoss side of dune
  - grainflows off the lee side of dune (generate inversely-sorted laminae)
  - backflow on the lee side (often migrating ripples up the lee-side)
- In the simplest scenario, form stacks of large-scale cross-beds from the lowest part of the original bedform, separated by bounding surfaces
  - Reactivation surfaces
  - Interdune surfaces
- More complex dunes yield an interwoven structure of large-scale cross-beds

### Pin-stripe lamination

A distinctive feature of aeolian deposits is *pin-stripe lamination*, which is defined by alternating layers of fine-grained (silt) and coarser-grained (fine–coarse sand) (Fryberger and Schenk, 1988). Pin-stripe laminations occur in several environments. The most distinctive and common pin-stripe laminations are the product of translatent deposition during lateral migration of wind ripples. Here, finer-grained sediments are concentrated in the troughs and coarser-grained sediments in the ripple crests. The ripples themselves are rarely preserved, but the as the ripple migrates over the trough, it caps the finer-grained sediments with the coarser ripple sands, setting up inversely graded coupled of silt and sand (Fig. 13.3.2).

Figure 13.7: Pin-stripe lamination in translatent strata in the Cretaceous Etjo Sandstone, Nambia.

Pin-stripe lamination also occurs on the lee side of dunes, by a completely different process. Here, fine-grained sediments settle in between grain fall avalanche events, often capping curved erosional surfaces, and are capped by the coarser-grained sediments involved in the grain falls. Where grain falls consisting of both a a finer and coarser component occurs, inverse-grading, which gives rise to pin-strip lamination, may cal so occur due to the process of *kinetic sieving*, which causes coarser grains to percolate upwards within the flow.

## 13.4 Lacustrine Environments

Lakes cover 1-2% of the Earth's surface and form in all environments - even underneath the Antarctic ice sheet

- The largest modern lakes occur in
  - Rift basins
  - Intracratonic basins
  - Glacial settings
- Lakes may be either *open* or *closed* hydrologically
- May be fresh to hypersaline

### 13.4.1 The main physical processes that affect lakes:

- Wind
  - deposition of dust
  - set-up of waves

  - currents

- River inflow
  - plumes
  - density underflows
  - lake margin currents

- Atmospheric heating
  - Density differences → stratification
  - Lake overturn
  - Freezing

*Varves* are rhythmic, alternating light (coarser) and dark (finer, more organic-rich) lake sediments characteristic of cold climates

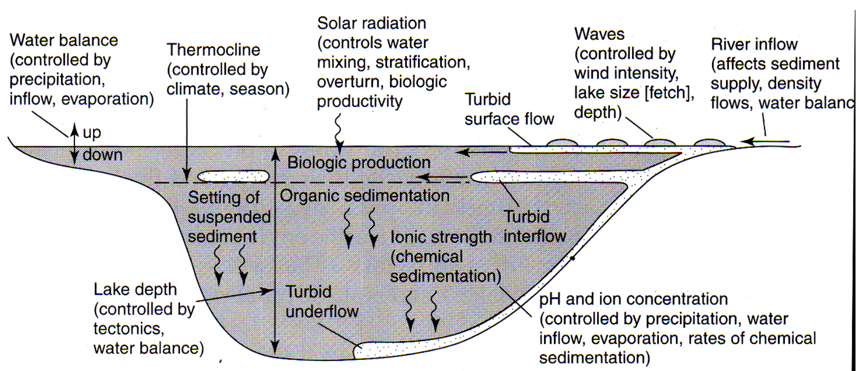



Figure 13.8: Schematic diagram of the sedimentary processes active in lakes. From Boggs (2006).

### 13.4.2 Chemical and biological processes

- A wide variety of chemically precipitated sediments may form in lakes:
  - carbonates
  - evaporates (including unusual minerals not typical of marine evaporites)
  - phosphates
  - cherts

- Chemistry is heavily influenced by evaporation versus precipitation and may be highly variable

- Often anoxic bottom waters

Biological processes are important in lakes, but similar to processes in the oceans

- May generate organic-rich sediments (good petroleum source rocks)

### 13.4.3 How to distinguish lake sediments?

- No evidence of tidal currents
- Sequence of evaporite minerals (or lack thereof)
- Association with other continental environments

# Chapter 14

# Coastlines: Open Shallow Marine Processes and Deposits

Additional Reading: Tucker, Chapter 2 (71-81); Boggs, Chapter 9; Bridge and Demicco, Chapter 15

The *marginal marine* environment lies at the boundary between the continental and open marine depositional realms. It is a relatively narrow zone, but disproportionately important in terms of both modern processes (particularly as the affect humans) and in terms of its representation in the geological record. Furthermore, the movement of this zone shoreward (retrogradation) and oceanward (progradation) is at the heart of sequence stratigraphic analysis.

The marginal marine environment is dominated by tidal, river, and wave processes, which deliver and redistribute sediments along the margin and out to sea. The relative strength of each of these processes determines which one controls depositional patterns and morphologies of of the coastal environment (Davis and Hayes, 1984), and coast-lines can be broadly subdivided by the most important of these three processes.

## 14.1 Storm-Dominated Coastlines

Storm-dominated coasts occur where hurricanes and other large storms pass intermittently. Even where these events are rare, the power of such storms to redistribute sediment on the continental margin is such that those deposits dominate the character of the depositional environment.

Between storms, mud deposition prevails on the continental shelf. Storms act to redistribute coarser sediments from the coast onto the shelf through the strong undercurrents that arise from the storm surge, which is the coastal flooding that occurs as a result of strong winds and low pressure. Consequently, mid-shelf muds tend to be interbeddd with silts and fine sands, which commonly show hummocky cross stratification (HCS) (Fig. **??**). Shoreward, HCS gives way to swaley cross stratification (SCS), an ultimately to shoreface sands.

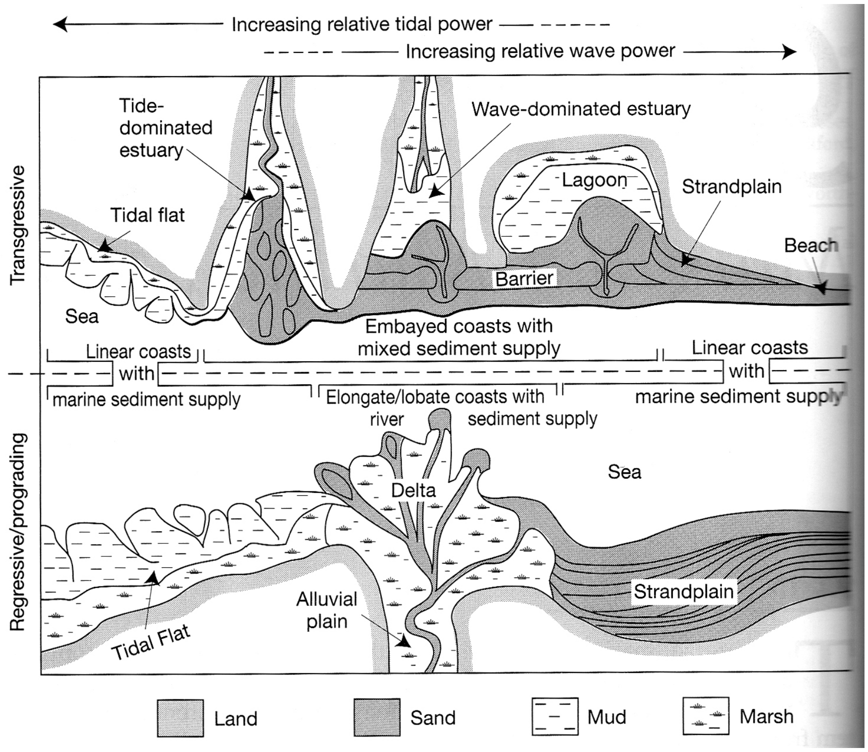



Figure 14.1: Sketch showing the principle marginal-marine depositional environments (from Boggs, 2011).

Storms tend to whip up significant sediment into suspension, which generates turbid waters. A *nepheloid layer* is a layer of water (up to 100's of meters thick) with suspended sediment, making more dense than surrounding water sick that it flows slowly offshore.

## 14.2 Tide-Dominated Coastlines

### 14.2.1 Tidal Flats

Tidal flats occur mainly on coastlines with large tidal ranges in areas of low-relief, and on the fringes of estuaries, interdistributary bays, and lagoons, as well as behind barrier islands. The do not form on coastlines subject to large storms.

- Marshy and muddy to sandy
- plains dissected by a network of channels
- inundated during high tide
- subdivided into
  - subtidal zone (sands)
  - intertidal zone (mixed sands and muds); flasher, lenticular, and wavy bedding common

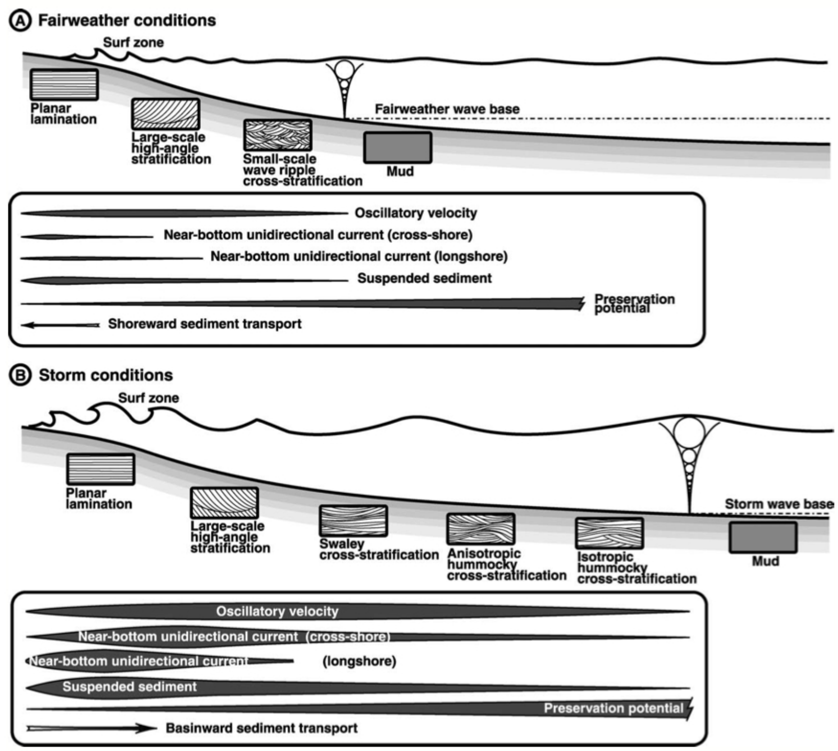



Figure 14.2: Sketch of expected sedimentary facies on a coastline during fair weather and storm weather, (from Dumas and Arnott, 2006).

  - supratidal zone (mostly muds, marshes, intermittently flooded)
    * In arid environments, *sabkhas* occur and are sites of evaporite deposition

- may be zones of either siliciclastic or carbonate deposition

Sedimentary processes and characteristic features of tidal flats

- tidal-channel flow (bi-directional currents, reactivation surfaces, tidal rhythmites)
- abundant erosional contacts
- tidal rhythmites or *tidalites*
- mixed sands and muds
  - lenticular bedding
  - wavy bedding
  - flaser bedding
- cracked muds
- evaporite minerals
- tepee structures (in carbonates)

Most tidal flat deposition occurs as the result of lateral accretion of tidal flat and point bars associated with meandering tidal channels. Where tidal flats prograde, they give rise to a *fining-upward* sequences, where sub tidal sands pass upward into mixed sands and muds, then muds, which may be bioturbated, cracked, or disrupted due to migrating channels or the occasional storm.

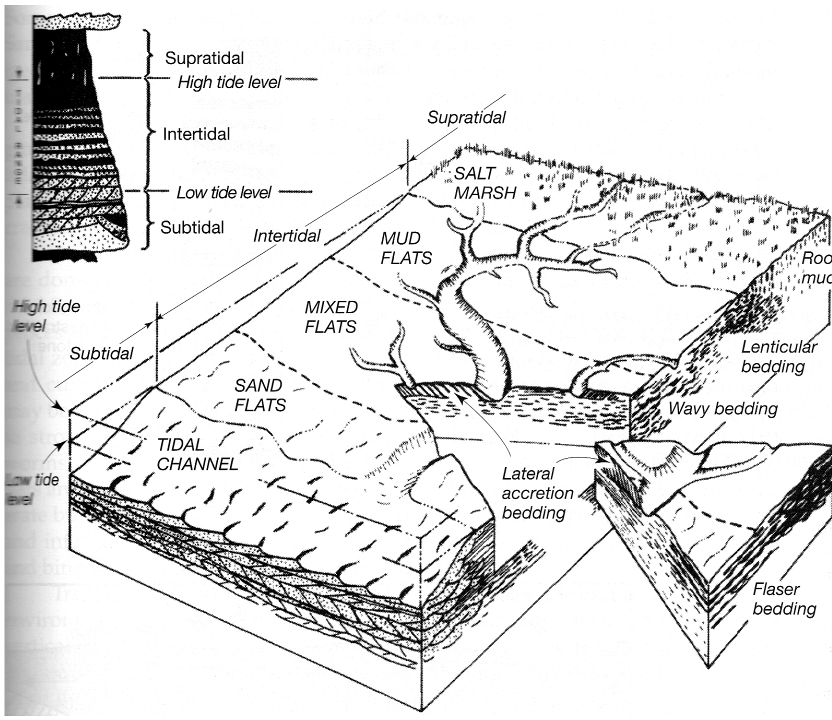



Figure 14.3: Schematic diagram of a tidal flat system, from Dalrymple (1992). Note the change in facies from the tidal channel to the supratidal zone.

## 14.3 River-Dominated Coastlines

### 14.3.1 Deltaic Systems

Deltas are distinguished morphologically as proturbances of the shoreline where rivers enter the oceans (or large lakes).

- Form where high sediment-load rivers enter the ocean
- Particularly common on passive margins (low tectonic activity)
- Main source of sediment to the marginal-marine environment
- Host significant petroleum reservoirs (form natural petroleum systems)
- grain-size from mud to gravel
- may be fluvial-, tide-, or wave-dominated

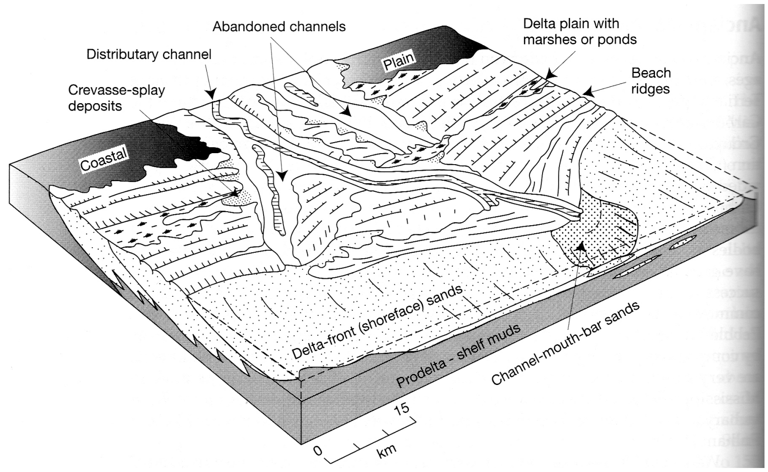



Figure 14.4: Sketch showing the important elements of a deltaic system (Boggs, 2011).

### 14.3.2 Fluvial-dominated deltas

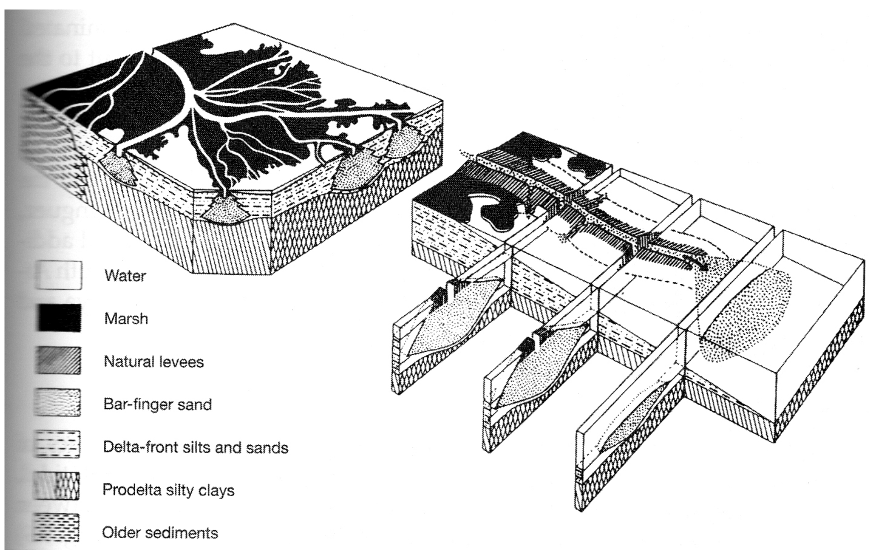



Figure 14.5: Architecture of the Mississippi River Delta, a classic fluvial-dominated delta system (bird-foot delta). From Boggs (2011).

Important components of a fluvial-dominated delta:

- distributary channels
- interdistributary bays
- natural levees
- crevasse splays
- bars

  - bar back
  - bar crest
  - bar front
  - distal bar
  - prodelta

Fluvial-dominated deltas are those in when the sediment discharge from the river largely controls the locus of sediment deposition and the delta morphology.

- *jets* are discharges of river water and sediment into the ocean
- **homopycnal flow** is discharge with an equal density to that of seawater
  - common in coarse-grained systems
  - rapid mixing and abrupt deposition of sediments
  - rapid-seaward spreading due to friction with the seabed
  - deposit distributory mouth bars that generate y-shaped bifurcations in the channels
  - responsible for forming classic *Gilbert-type* deltas, with bottomset, steep foreset (10-20°), and topset *sigmoidal* geometry
  - gravity flow redeposition important
- **hyperpycnal flow** is discharge with a greater density than seawater
  - more common in lakes (which have lower density)
  - jet flow moves along bottom of seabed, forming density currents
  - turbidites common
- **hypopycnal flow** is discharge with a lower density than seawater
  - flows into seawater as a buoyant plume
  - rapid deposition of coarser sediment at channel mouth, form seaward-fining, lobate-shaped bars
  - carries fine-grained sediment in suspension, some of which *flocculates* to form sediment aggregates which then drop from suspension
  - generate large, shallowly-dipping delta-front regions
  - most important river outflow in marine basins

### Tide-dominated deltas

Tide-dominated deltas occur where tidal currents are stronger than fluvial currents

- Large tidal range
- Form funnel-shaped distributary channels
- Channel sands reshaped into a series of linear ridges, perpendicular to the shoreline
- Vast tidal-flats

**Wave-dominated deltas**

Wave-dominated deltas occur where wave action decelerates river outflow and redistributes sediments parallel to the shoreline.

- Commonly constricted or deflect mouth of the river
- Distributary-mouth deposits reworked into beaches, barrier bars, and spits
- In a mature system, the delta front consists of a series of coalesced ridges parallel to the shore (*strandlines*)

**Sedimentology of Deltas**

All deltas can be subdivided broadly into various physiographic zones

- Subaerial - generally a larger area
  - upper delta plain (above high tide): river channels, swamps, floodplains, freshwater lakes
  - lower delta plain (intertidal): distributary channel system, inter-distributary bays, tidal channels
- Subaqueous
  - delta front (0-10m depth): sand and silt
  - prodelta slope: silt and mud

Deltas in the stratigraphic record:

- Active delta-building leads to a distinct vertical succession of facies
  - Coarsening up: shelf muds to mouth bar and channel sands, beach and dune sands
  - Fining up into interdistributary bay muds
- Delta lobes migrate, which can interrupt this sequence, lead to erosion of the abandoned delta lobe

### 14.3.3 Estuaries

Estuaries are drowned valley systems receiving sediment and water from both fluvial and marine sources. There are several types of estuaries:
Estuaries may be either wave-dominated or tide-dominated

- wave-dominated estuaries
  - separated from open ocean by barrier islands
  - sand restricted to the outer estuary
  - low energy middle zone
  - abundant mud deposition - flocculation of clays
  - often stratified

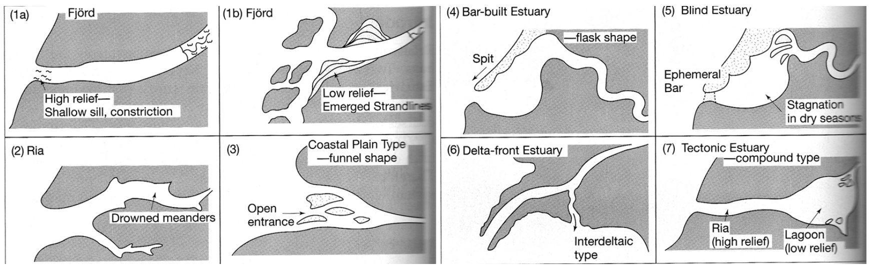



Figure 14.6: The principal estuary types (Boggs, 2011).

- tide-dominated estuaries
  - open to ocean
  - elongate sand bards and tidal channels
  - tidal flats and salt marshes
  - well-mixed

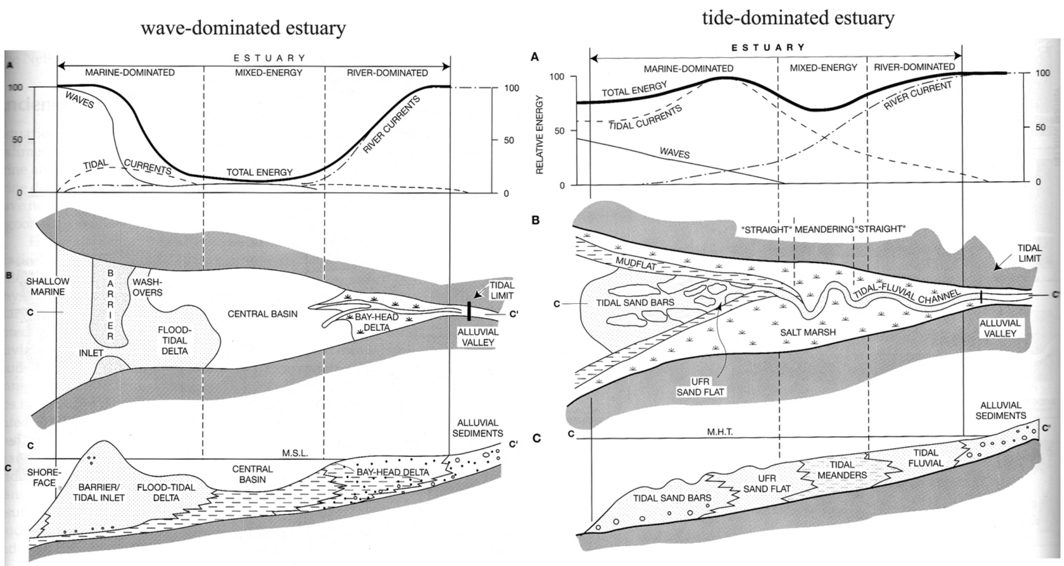



Figure 14.7: Wave- versus tide-dominated estuaries. From Boggs (2011).

## 14.4 Wave-Dominated Coastlines

Wave-dominated coastlines
Beaches tends to be narrow and discontinuous, interrupted by headlands, cliffs, and other marginal-marine systems *Barrier Islands* are separated from the mainland by lagoons, estuaries, or marshes

- Enormously sensitive to environmental change

  - Subtle changes in sea level → significant translation of beach
  - Also vary seasonally in many climates

- Economic importance:
  - Modern beaches: placer gold, platinum, diamonds
  - Ancient beaches: petroleum reservoirs, uranium

- Can comprise any of a number minerals/components

### 14.4.1 Beach morphology

Beach morphology:

- Backshore: landward from high-tide
  - back beach dunes

- Foreshore: intertidal zone

- Shoreface: low-tide level to *fairweather wave base (10-15m)*

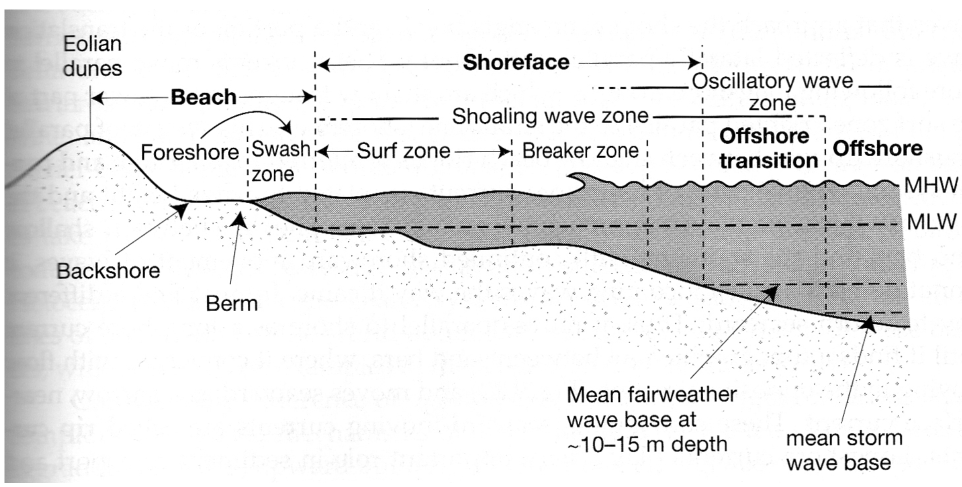



Figure 14.8: Cross-sectional profile of a typical beach system (Boggs, 2011).

### 14.4.2 Sedimentary processes on beaches

Wave processes:

- Dominated by wave action
- shoaling zone
- breaker zone
- surf zone

  - translational waves
- swash zone
- backwash
  - heavily mineral deposition
- shore parallel transport
- low-moderate energy waves - net deposition on waves
- high energy waves - net erosion

Wave-induced currents

- Longshore currents
  - longshore troughs and bars (ridge and runnel system)
  - transport considerable sediment
- Rip currents - near surface

### 14.4.3 Barrier Islands

Barrier island systems consist of three separate environments:

- The sandy barrier islands themselves, which is similar to the beach environment
- The back-barrier lagoon or estuary
- Tidal channels cut through the barrier islands and connect the back-barrier environment to the open ocean

Barrier systems form by a variety of different processes:

- Building and emergence of a long-shore bar
- Later progradation of beaches around a coastal promontory
- Segmentation of a spit parallel to the coast
- Take-over of pre-Holocene beach systems (i.e. palaeo-shoreline)

Beach systems consist mainly of sand deposited in long, narrow bands, but may include pebbles, gravel, and cobbles:

- The backshore is dominated by aeolian and storm wave processes
  - dunes
  - washover deposits
- The foreshore is dominated by swash and backswash
  - gently dipping, parallel laminae (can be both seaward and landward dipping
  - heavy mineral laminae

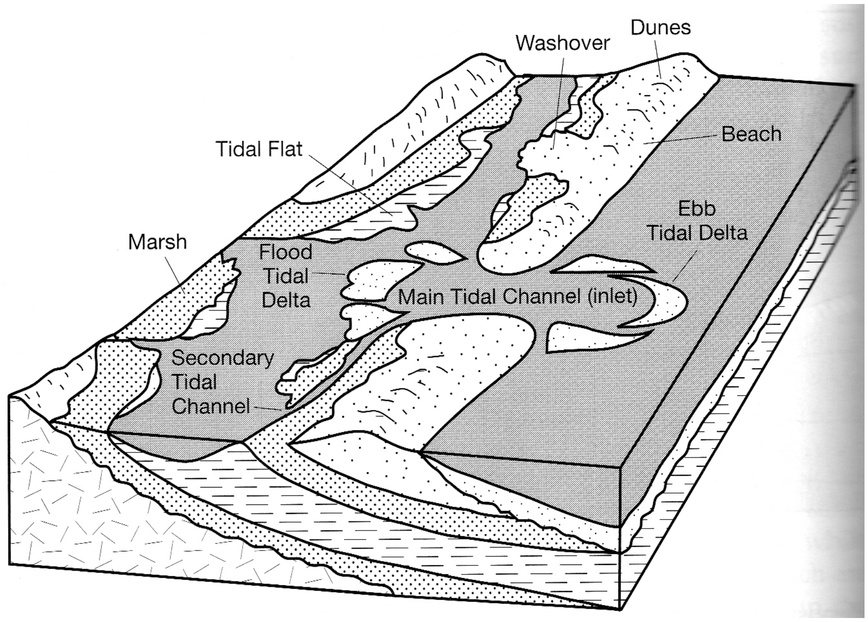



Figure 14.9: The barrier island system comprises an array of subenvironments (Boggs, 2011).

  - can include ripple and dunes migrating parallel to the shoreline

- The shoreface comprises a broader environment than the backshore or foreshore
  - Multi-directional trough cross-bedding
  - Trace fossils
  - Planar lamination
  - Oscillation ripples

### 14.4.4 Lagoons

Lagoons are semi-restricted bodies of water separated from the open ocean typically by a spit or barrier island system, or reef. Lagoons are typically low-energy, and often restricted environments

- Hydrology and sediment flow is heavily influenced by tidal processes
- Wind and high evaporation rates can also be important
- tidal strength regulates degree of restriction of a lagoon
- may be anoxic or dysoxic
- sediments derived mainly from tidal channels and spill-over from the open ocean, plus wind

Characterised by fine-grained sediments

- finely laminated muds and silts, often organic-rich
- carbonates common in tropical environments
- evaporites and dolomite occur in restricted lagoons

The Coorong is a textbook example of a choked lagoon.

Lagoons are relatively straightforward to identify in the stratigraphic record

- Fine, planar-laminated, possibly evaporitic sediments
- Black shales and organic-rich limestone common
- Bound by coarser-grained beach/barrier/tidal sands

# Chapter 15

# Deep Water Processes and Deposits

Additional Reading: Tucker, Chapter 2 (81-88); Boggs, Chapter 10; Bridge and Demicco, Chapter 18

## 15.1 The Continental Shelf

### 15.1.1 Physiography

The open marine environment lies oceanward of the zone dominated by shoreline processes. This environment can be subdivided into four main environments:

- shelf (usually the continental margin)
- shelf margin (or shelf break)
- slope
- deep basin

Other physiographic features of the open marine environment include

- continental rise
- abyssal plain
- deep-sea trenches

The *neritic zone* is defined as extending from the shoreline to the continental margin

- Encompasses the shallow marine environment
- Includes shallow epeiric seas
- average depth of 130m, depth range of 5-1000m
- shelf width of 2 to 1500km
- major zone of deposition and sediment *bypass* (i.e. a thoroughfare for students)

- the modern continental shelf is a relict of the Late Pleistocene, when sea level was lower by ∼130m
- multiple physical processes responsible for sediment transport and deposition

### 15.1.2 Physical Depositional Processes

Various different physical processes are responsible for sediment transport and deposition on continental shelves

- tidal currents
- wave action/storms
- wind
- oceanic currents
- density currents

Wind-generated currents

- Swells and storms
  - increase depth which is affected by waves (up to 200 m)
  - coastal set up and beach erosion
  - seaward transport of sediments
- Nepheloid flow
- Geostrophic currents

Tidal processes are best known from the marginal marine environment, but tidal currents also affect the continental shelf, typically generating structures oriented perpendicular to the shoreline.

Oceanic currents include the thermohaline circulation and the largest of the geostrophic currents (boundary currents)

- The Gulf Stream
- Kuroshio
- The Argulhas current
- The thermohaline circulation

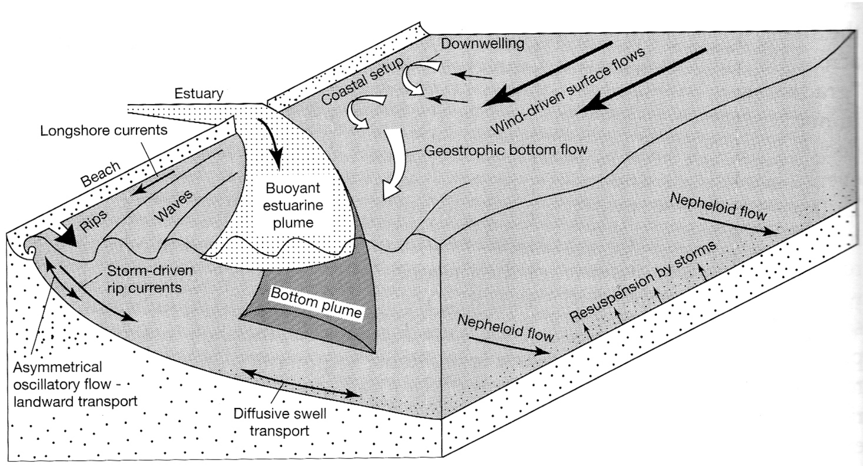



Figure 15.1: Schematic representation of processes responsible for moving sediment on continental shelves. From Boggs (2011).

## 15.2 The Slope

The shelf margin (or shelf break) is the transition from the continental shelf to the slope

- occurs at an average of 130m depth
- may represent any of a number of structures
  - A draped basement high
  - A major boundary fault relict from rifting
  - Edge of a major delta
  - A carbonate bank or reef

The slope is the region between the shelf and the deep ocean

- Average slope of 4°
- Can vary from 2° to 45°
- Typically several 10s of kilometers in width
- Incised by submarine canyons

The principal sedimentary processes on continental slopes gravity-driven mass flow processes

- Slumps
  - breccias
  - folds, detachment surfaces
- Debris flows
  - diamictites

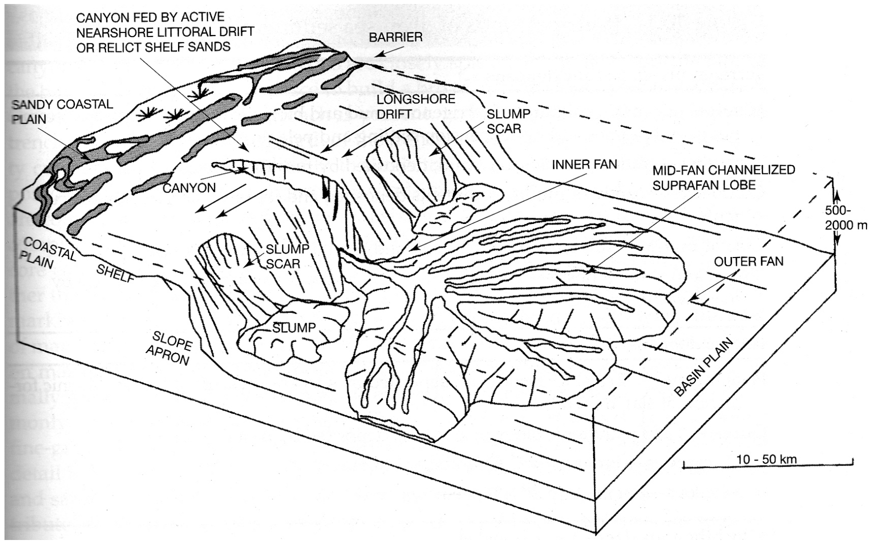



Figure 15.2: Major depositional features of the slope and deep basin environment. From Boggs (2011).

- Turbidity currents
  - Turbidites
  - Subaqueous fan complexes
  - Bouma sequence
  - Rhythmites
- Grain flows
- Contour currents
  - Contourites

## 15.3 The Deep Ocean

The deep ocean environment:

- The continental rise (gently slopes towards the abyssal plain)
- The abyssal plain
- Mid ocean ridges
- Deep-sea trench

Sediments in the deep-ocean realm are dominantly fine-grained

- Pelagic muds and oozes
  - Wind-blown dust

  - Distal turbidity currents and sediment plumes
  - Pelagic rain planktonic tests
    * Foraminifera
    * Coccoliths
    * Diatoms
    * Radiolaria

- Turbidity currents
- Tephra (volcanisms)
  - bentonite
- Glacial-marine rain out (IRD)
- Manganese nodules-crust

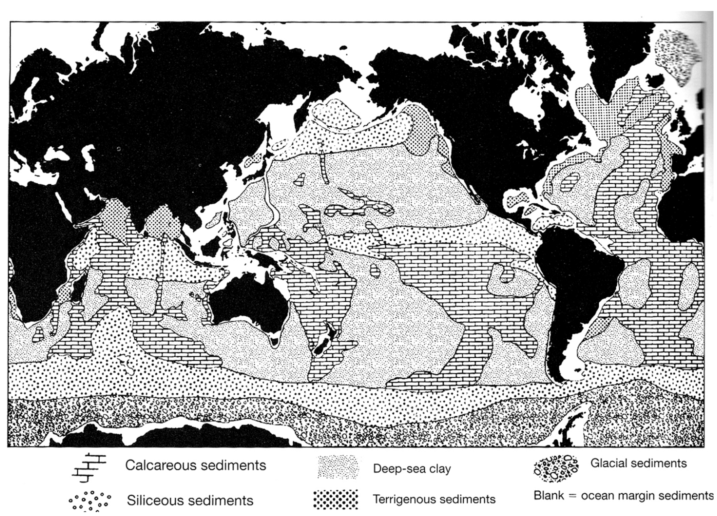



Figure 15.3: Distribution of the main types of deep-sea sediments. From Boggs (2011).

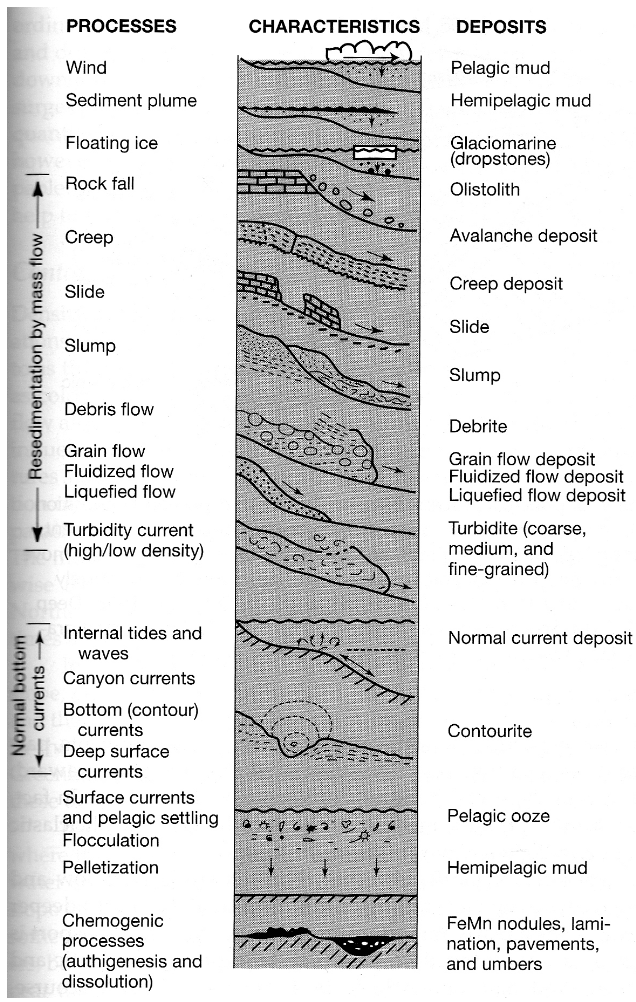



Figure 15.4: The principal processes and their deposits on the slope and in the deep sea. From Boggs (2011).

# Chapter 16

# Carbonate Depositional Environments

Additional Reading: Tucker, Chapter 4 (151-165); Boggs, Chapter 11

## 16.1 Carbonate Systems

Carbonates are deposited in both shallow marine environments and in the deep sea (although not the deepest sea). Most of the thick carbonate successions in the geological record record deposition on continental platforms, although resedimented carbonates in slope settings are not uncommon. Most carbonate shelves occur in clear, shallow tropical seas (although there are also cool water carbonates), where there is little terrigenous input, which tends to dilute carbonate sediments, impede carbonate nucleation, and interfere with biological processes that generate carbonate sediments. These environments are known as *carbonate factories*, and include carbonate sediments that originate as shells and skeletons, ooids, peloids, and carbonate mud that is precipitated directly or indirectly by biochemical processes (e.g. by calcareous algae). Carbonate factories tend to be self-sustaining; once interrupted, for example, by an influx of siliciclastic material or rapid rise or decline in sea level, the carbonate factory may take a long time to recover.

Despite the fact that carbonate sediments are subject to many of the same sedimentary processes as siliciclastic sediments, carbonate systems are also unique as a consequence that most carbonate is produced biochemically (James et al., 2010).

- Grain-size does not always reflect hydrology
- Sediment production in large carbonate systems is produced in-situ and is self-sustaining
- The style of accumulation largely depends on the origin of the sediments

Carbonate systems do not make up a large percentage of modern depositional environments, but this clearly has not been the case throughout Earth history. Carbonates account for about 25% of the sedimentary record; hence, through most of Earth history, carbonate systems have played a larger role in global sedimentation.

Some classic examples of modern carbonate platforms are

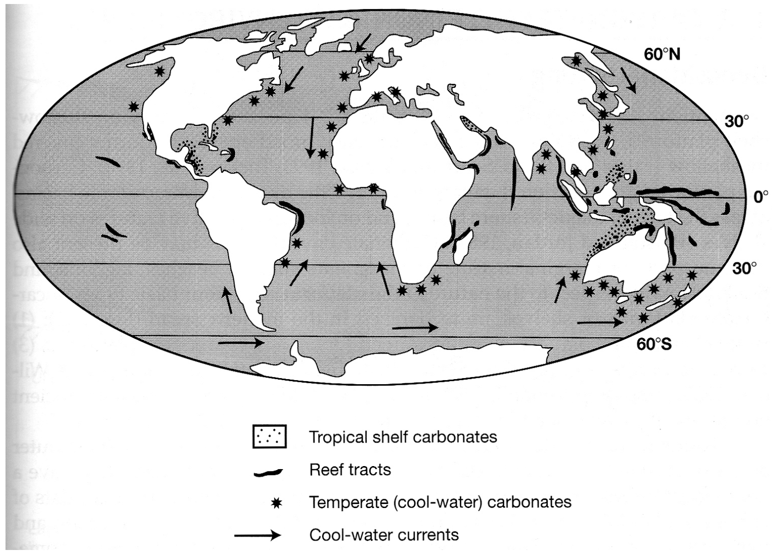



Figure 16.1: Global distribution of modern marine carbonate systems (Boggs, 2011).

- Florida Bay
- western Australia
- Trucial Coast (Persian Gulf)

## 16.2 Carbonate Platforms

Most thick carbonate accumulations in the geological record record deposition on subsiding *carbonate platforms*. Most carbonate platforms (but not all), are in fact *shelves*, meaning they were attached to a landmass. A variety of different platform types occur (Fig. 16.2 with distinctive geometries and facies associations.

### 16.2.1 Rimmed platforms and reefs

Many carbonate shelves are distinct from typical continental shelves in that they are *rimmed* at the edge–that is, at the transition to the slope. Rimmed carbonate platforms are quite common because platform edges in warm environments are well suited to corals. Hence, rims are commonly a reef, but in some cases are sand shoals. In either case, the rim serves the key role of absorbing energy from incoming ocean waves and storms. The result is a calm *back-reef* environment, which is fertile for carbonate sediment production and may become dysoxic to anoxic due to reduced circulation. This may result in deposition of organic-rich sediments and/or evaporites.

On a rimmed platform, many of the sedimentary facies record deposition in relatively shallow water. However, the outer edge of the reef marks an abrupt transition to a deepwater environment and is commonly characterized by resedimented carbonates including talus that is eroded from the the rim. The talus may include very large (car to house sized)

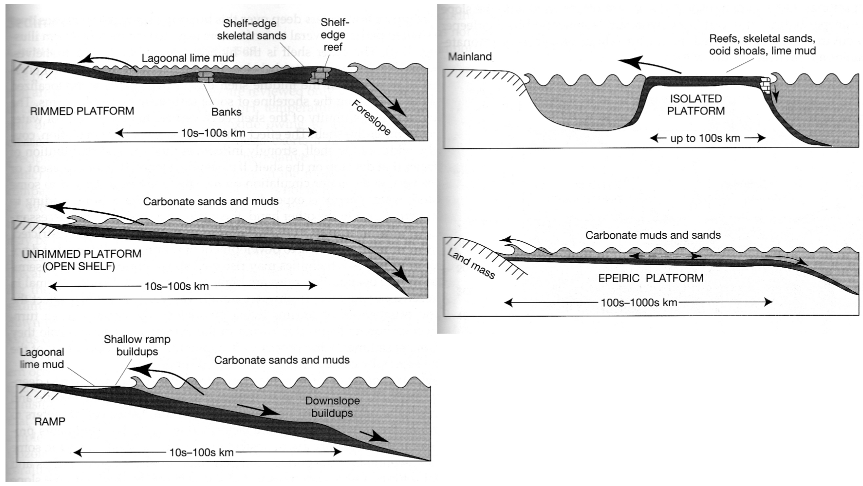



Figure 16.2: Schematic representation of the varies types of carbonate platform. From Boggs (2011).

blocks, known as *olistoliths*.

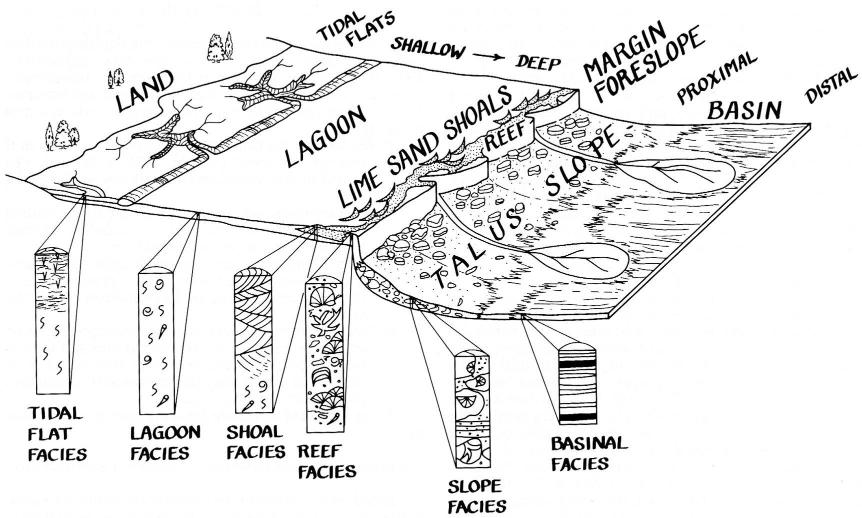



Figure 16.3: Depositional environments on rimmed carbonate platforms. From Demicco and Hardie (1994).

### Reefs

Reefs are shallow, rigid, carbonate build-ups that preferentially form in high-energy environments (i.e. breaking waves).

- Modern reefs are built from a variety of organisms
  - Corals

  - Cyanobacteria
  - Calcifying algae (e.g., Renalcis, Girvanella, Epiphyton)
  - Foraminifera
  - Bryozoa
  - Sponges
  - Molluscs

- Ancient reefs were also built from

  - Stromatolites
  - Archaeocyathids
  - Stromatoporoids

Reef organisms play a variety of roles in the construction of a reef system

- Frame-builders (corals, stromatolites, stramatoporoids)
- Sediment contributers (crinoids, some algae)
- Bafflers (sea grasses)
- Binders (cyanobacteria)
- Precipitators (cyanobacteria)

The best-known reefs are the long, linear reefs, but there a variety of reef morphologies

- Barrier reefs
- Fringing reefs
- Atolls
- Patch reefs
- Pinnacle reefs

### 16.2.2 Unrimmed platforms

*Unrimmed platforms* lack the barrier of rimmed platforms.

- *Ramps* slope gently basinward (typically less than 1°)
- *Open shelves* are platforms that are distally steepend (like siliciclastic dominated-shelves)

In contrast to rimmed platforms, the shallow water environment of unrimmed shelves is high energy and the nearshore environment exhibits complex facies (James et al., 2010), more like siliciclastic shelves. Also, like siliciclastic shelves, sediments are much more easily transported offshore in this environment (e.g. during storms). But unrimmed shelves are distinguished by the fact that carbonate sediment is formed *in situ* (even if much of it is redistributed).

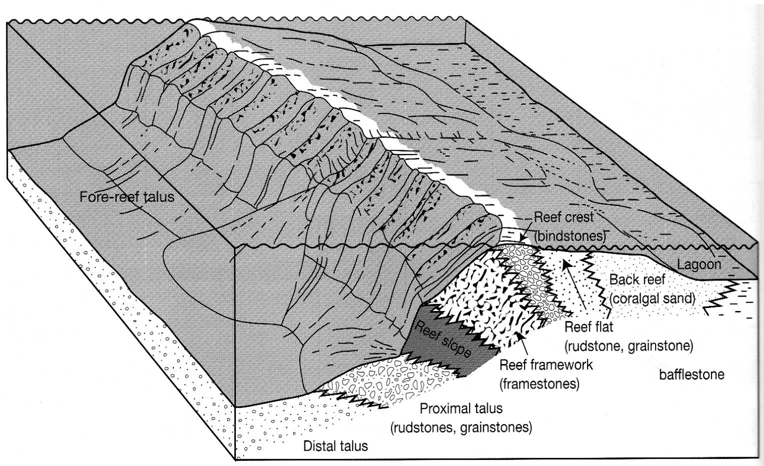



Figure 16.4: Facies associations in the reef environment

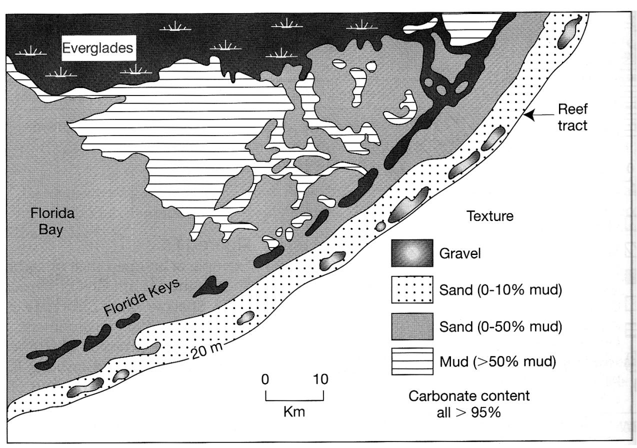



Figure 16.5: Distribution of shallow water facies on a modern carbonate platform (Florida Keys). From Boggs (2011).

### 16.2.3 Carbonate banks

Carbonate banks are isolated carbonate platforms. These include the Bahamas Platform and atolls, which are seamounts that have sunk below the surface of the ocean, but where reefs and related carbonate sedimentation keep pace and maintain a shallow water carbonate environment.

### 16.2.4 Basinal environment

Carbonates do occur in the deep basins, provided that the seafloor is above the carbonate compensation depth. However, deep sea carbonates were rare prior to the middle Mesozoic, a time which saw evolution and proliferation of pelagic calcareous (and siliceous) plankton.

### 16.2.5 Cool water environments

Carbonates do also occur in cooler environments, where the source of sediment is dominantly heterotrophic organisms such as bivalves. The mechanical breakdown and in situ burial of carbonate shells can in places contribute to significant thickness of carbonate sediments in cool water environments. The Australian Bight and South Orkney Plateau are two important settings of modern cool water carbonate deposition.

### 16.2.6 Terrestrial environments

Carbonate form in a wide variety of terrestrial environments, from desert springs to glacial moraines. With the exception of some lake deposits, none of the terrestrial carbonate environments is well represented in the ancient sedimentary record.

- Saline/playa lakes
- Caves (speleothems)
- Tufas (form in highly alkaline lakes)
- Travertines (form around springs due to degassing and/or evaporation)

## 16.3 Carbonate Cycles

Shallow water carbonates are commonly arranged in stacked, meter-scale, shallowing-upward *cycles*. These cycles are a predictable outcome of normal carbonate sedimentation processes, such as progradation of tidal flats or lateral migration tidal channels. A typical cycle consists of a basal, sub-tidal carbonate (commonly limestone), capped by a shallow to emergent carbonate, commonly with tepee structures or other evidence of exposure. However, the actual facies within a cycle can vary, depending on location on a carbonate platform. Grotzinger (1986) traced individual cycles in the Paleoproterozoic Pethei Group (northwest Canada) over 100 kms down depositional dip.

Well developed cycles are intuitively a natural consequence of carbonate sedimentation. A small relative fall in sea level exposes shallow water facies and locally shuts down deposition. A subsequent small rise in sea level produces new accommodation space, which the carbonate factory quickly fills back up to sea level. These m-scale cycles are commonly high order components of larger scale cycles, and it has commonly been argued that they record some *allocyclic* forcing on sea level. This may be either tectonic or eustatic, but eustatic fluctuations are apparently attractive, because the time scale for the cycles commonly approximates Milankovitch time scales. Indeed, these cycles have been related by many carbonate stratigraphers to orbitals cycles (e.g. Goldhammer et al., 1990). Proponents of an allocyclic mechanism for the generation of cycles commonly note that the stacked cycles themselves comprise larger cycles, which together reflect a somewhat complicated, composite fluctuation in sea level.

An opposing school in carbonate stratigraphy recognizes the possible importance of *autocyclic* processes in generating carbonate cycles (e.g. Drummond and Wilkinson, 1993).

This school emphasizes internal controls on carbonate sedimentation and accumulation, and the argument is fairly straightforward. Imagine a carbonate platform that experiences a relatively sharp rise in base level (be it due to tectonics or eustasy). This will generate deepening, and will temporarily slow down the carbonate factory. But once the carbonate factory cranks up again, it will quickly fill this space (say a few or several meters), resulting in a shallowing upward cycle. The top of this cycle need not reflect the cessation of the rise in sea level, however, because as we know, carbonate precipitation tends to keep up with sea level. But it will still look like a complete cycle. At this point, sediment accumulation stops, and presumably the carbonate factory shifts towards off-shore (where there remains some space to precipitate carbonates). With continued increase in sea level, and without that immediate source of carbonate sediments, deeper water conditions will again return, without any cyclicty in base level. Once carbonate precipitation resumes, it will again quickly fill the space, producing a second cycle, and so on, such that a single episode of rising sea level gives rise to multiple cycles.

# Chapter 17

# Glacial Environments and Deposits

Additional Reading: Boggs, Chapter 8 (276-288); ; Bridge and Demicco, Chapter 17

## Introduction

Glacial deposits cover much of the modern-day land surface in the middle to high latitudes. They are relatively less abundant in the geological record, but are hugely significant for interpretation of Earth history. But ancient glacial deposits are both highly controversial (are they truly glaciogenic?) and inherently complex, preserving only a highly incomplete record of the glaciation that they represent.

## 17.1 Ancient Glaciations

Glacial deposits occur throughout episodically Earth history and bear witness to the fact that the ancient climate was not universally hot (and the inverse is true–the absence of glacial deposits through most of Earth history demonstrates that the Earth was not universally cold). The oldest recognized glacial deposits occur in the 2.9 Ga Pongola Supergroup in South Africa (Young et al., 1988). Ancient glacial deposits occur in the

- Palaeoproterozoic (2.4–2.1 Ga)
- Neoproterozoic (750 - 580 Ma)
- Late Ordovician
- Late Devonian
- Carboniferous-Permian
- Mesozoic?
- Cenozoic

## Ice flow

Ice behaves as a *pseudoplastic* and flows under the influence of gravity. It may flow down valleys, as in alpine glaciers, or may simply flow underneath its own weight, as in ice sheets. The relative motion between layers in ice is slow enough that the flow is laminar. Velocity gradients are low in the upper part of a glacier and greatest near the base, due in part to the decrease in viscosity as temperatures increase with depth. Some glaciers are effectively frozen to their substrate, mean there is no slip on the boundary. These are known as *cold-based glaciers*.

The combination of higher temperatures and high pressures mean that water may melt at the base of the glacier (liquid water may also be supplied from *moulins* and other glacial plumbing). Water within the glacier, as well as sediments, reduces viscosity locally (Benn and Evans, 1998). Meltwater at the base of the ice sheet and wet sediments lubricates the flowing glacier, meaning that the velocity of flowing ice at the base is not necessarily zero. Melting may also occur as flowing ice encounters obstacles (like rocks) along its flow path, where it melts on the upstream side, and re-freezes on the downstream side (*regelation*). Where basal sliding occurs, this dominates ice flow in a glacier, and the glacier is said to be *wet-based*. The velocity of basal sliding decreases from the center towards the edges of the glacier.

Ice flows internally as well and where the ice is frozen to the bed beneath, this flow dominates. Flow is driven by accumulation of snow above the *equilibrium line*, where flow lines are downward. Downstream from the equilibrium line, flow is upward. This is why one hears stories (maybe apocryphal) about airplanes that have crash and disappeared on glaciers only to reemerge downstream decades later.

Velocities of ice flow in glaciers range from <1 to >1000 m/year, with the highest flows occurring in *ice streams* and *surging* glaciers. In these cases, ice flow is 10-100 times faster than normal or ambient flow rates and, not surprisingly, involve an increase in sediment erosion and transport. A surging glacier can advance by many kilometers in a matter of years. But because the factors governing not only surging glaciers but more broadly glacial flows are complex and the timescales required for external forcing (e.g. changing climate) to drive changes in glacial flow large, it is difficult to relate change in flow regimes to present conditions.

### Ice erosion and sediment transport

Glaciers are impressive agents of erosion. They erode by

- ice wedging
- plucking
- abrasion

Sediment in glaciers is transported mainly on the edges and at the base, and sediment concentrations can reach up to half of the volume of a glacier. On valley glaciers, rock falls, landslides, and meltwater streams deposit sediments onto the top of the glacier. This sediment can make its way to the interior of the glaciers in moulins and crevasses or by being buried by subsequent snowfall. Because the viscosity of glaciers is so great, it can carry tremendously large blocks. Ice is effectively indiscriminant in the size of sediment it transports, hence it does not sort sediment well. The poor sorting and abundance of fine-grained sediments is one of the distinguishing feature of glacially-derived sediments.

### 17.1.1 Deposition from ice

Deposition occurs at the base of a glacier due to drag on sediments at the interface with the substrate and due to melting of basal ice. Sediments deposited at the base of a glacier are known as *tills*. Those that occur where the glacier is sliding arr prone to deformation by the shear force of the overriding ice and are known as *lodgement tills*. Much sediment is deposited at the edge of the glacier where it melts. *Ablation (or melt) tills* are relatively undeformed, heterogeneous, and porous as compared to lodgement tills.

Deposition also occurs from flowing water at the base of the ice, in outwash plains beyond the ice front, in ice proximal lakes, and from melting ice bergs. These processes will all be discussed in a subsequent chapter on glacial depositional environments.

## 17.2 Types of glaciers

- Valley glaciers
- Piedmont glaciers
- Ice sheets
- Ice domes
- Ice streams

Flow of ice in a glacier varies widely from millimeters per day in cold-based glaciers to many 10s meters per day for surging glaciers and ice streams.

## 17.3 Terrestrial Glacial Environments

- subglacial zone - contact between the glacier and its substrate
  - mainly zone of erosion (plucking)
  - but certain distinct deposits are left
- supraglacial zone - upper surface of the glacier
- ice-contact zone - margin of the glacier
  - moraine accumulation
  - ice-dammed lakes
- englacial zone
- Proglacial environment - downstream from a glacier

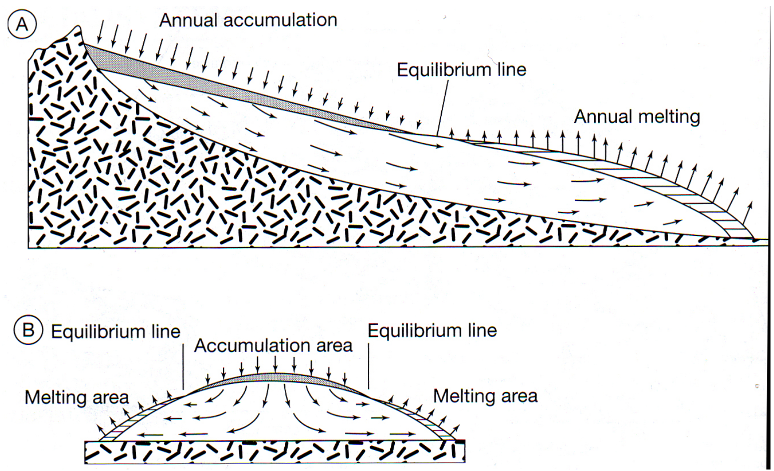



Figure 17.1: Schematic diagram showing directions of ice flow in a glacier. From Leeder (1999).

  - outwash plain
  - proglacial lakes

- Periglacial environment - area extending beyond the glacier, but influenced by the same climate

Erosion, sediment transport, and deposition by land-based glaciers

- Rock flour is a common byproduct of glaciation
- Various types of moraines: end, lateral, medial (*ablation tills*)
  - Typically distinguished by poor-sorting, often varied clast composition
  - Typically unstratified (massive), but sometimes stratified when reworked by meltwaters
- *Lodgement tills* are debris deposited by melting of basal ice
  - deformed by weight and movement of overriding glacier
  - oriented fabrics or deformation features
- Other continental glacial features include
  - *Eskers* - deposited from subglacial streams
  - *Kames* - mounds of sand and gravel accumulated in crevasses
  - *Drumlins* - whale-shaped, waveforms oriented parallel to flow and sculpted into unconsolidated sediments
  - *Ribbed moraines* - similar to drumlins, but transverse to flow
  - *Roche moutonnée* - an erosional, tear-drop-shaped feature pointing downstream

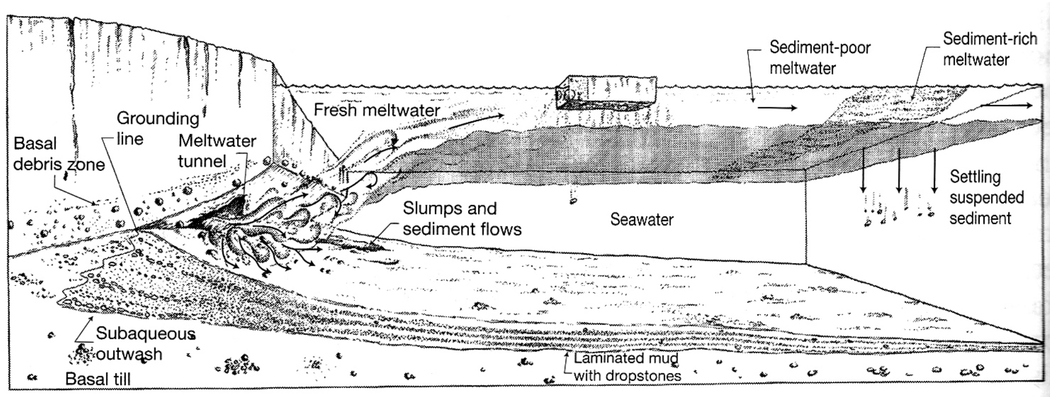



Figure 17.2: The glacial-marine environment. From Edwards (1986).

## 17.4 Glaciomarine Environments

The glacial marine environment is often characterised by very rapid sedimentation rates, resulting in high accumulations of sediments and abundant redeposition by gravity flow processes.

### 17.4.1 The proximal facies

- Unstratified diamictites deposited in the *basal debris zone*, behind and below the *grounding line*
- Transitions into stratified diamictites seaward of the grounding line
- Heavy sediment load from meltwater conduits - fans and lobes of sediment
- Additional sediment from melting icebergs
- Meltwater plumes
- Density currents
- Gravity deposits

### 17.4.2 Distal glaciomarine facies

- Sedimentation heavily influenced by *rain-out* of debris from melting glaciers
- *Ice-rafted debris* (IRD) - very poorly sorted, but somewhat stratified
  - *Dropstones*
  - *Lonestones*
  - *Dump structures*
- Additional sediment input from
  - sediment plumes

  - gravity flows (turbidites and their distal equivalent, rhythmites)
- Plowing of seafloor sediments by icebergs

## 17.5 Evidence of Glacial Influence

- The dominant glacial lithology is diamictite
  - by itself, not sufficient evidence to point to glacial deposition
  - but varied and far-flung source of clasts suggests glacial transport
- Glacial striations
  - Cobbles
  - Pavements
- Faceted clasts
- *Chatter marks*
- *Sand wedges*
- U-shaped valleys

# Chapter 18

# Sequence Stratigraphy

## 18.1 Introduction

It has long been recognized that sedimentary successions are arranged in broadly transgressive-regressive packages. In the 1970's, the oil industry, led by Exxon and work on seismic transects of continental margins, revolutionized how sedimentologists consider and correlate strata bey developing the field of *sequence stratigraphy*. A *sequence* is defined as a "a relatively conformable succession of genetically related strata bounded by unconformities and their correlative conformities." The three controls on the development of sequences are

- rate of crustal subsidence or uplift
- eustatic sea-level change
- rate of supply of sediments

## 18.2 Sequence boundaries and systems tracts

During a given phase of normal regression, forced regression, or transgression, a suite of genetically related sediments are deposited from the onshore to offshore environment. This package of sediments is known as a *systems tract*, and there are four types of systems tract corresponding to the four phases in base level change:

1. highstand systems tract (normal regression)
2. falling stage systems tract (forced regression)
3. lowstand systems tract (normal regression)
4. transgressive systems tract (transgression)

These systems tracts are by definition bound above and below by sequence boundaries. A full sequence will contain up to all four systems tracts, but need to have them all and cannot, by the definition of a sequence, duplicate any of the systems tracts. The important point is that all sediments deposited in a single systems tract are broadly conformable in age.

*Sequence boundaries* are stratigraphic surfaces that develop in response to changing weight of influence of subsidence (or uplift), eustatic sea-level change, and sediment supply rates. These effects are manifested in shifts between transgression and regression of the shoreline, or a shift from forced to normal regression, or vice versa. Sequence stratigraphic surfaces are much better (but in most cases, not perfect) approximations of time lines than lithological boundaries (that is, boundaries between facies), and for the most part, are recognizable in seismic sections (Fig. 18.3). The main sequence boundaries are

- subaerial unconformity
- correlative conformity
- maximum flooding surface (time of highest sea level stand)
- maximum regressive surface
- wave ravinement surface

Different sequence boundaries can be used to define the upper and lower boundaries of an individual sequence. For example, in original definition of *depositional sequences*, the boundaries are the subaerial unconformity and correlative conformity. In the *Transgressive-Regressive (T-R)* designation popularized by Alan Embrey, the bounding surface is the maximum transgressive surface.

## 18.3 Stratal terminations

The accumulation of sequences results in the development of distinct strata terminations (Catuneanu, 2006):

- *truncation*: termination of strata against an overlying erosional surface
- *toplap*: termination of *clinoforms* against an overlying lower angle surface mainly as a result of non-erosion (*sediment bypass*)
- *onlap*: termination of low-angle strata against a steeper surface
  - marine onlap on continental slopes during transgression
  - coastal on lap onto revilement surfaces during transgression
  - fluvial on lap is landward shift of upstream end of coastal plain during transgression or normal regression
- *downlap*: termination of inclined surfaces against a lower-angle surface
- *offlap*: progressive off-shore shift of up-dip terminations of sedimentary surfaces (during forced regressions)

The great power of sequence stratigraphy is that it enables us to assemble a relatively robust chronostratigraphic and spatial framework for sediments and to then interpret the patterns of sedimentation in terms of the competing effects of subsidence/uplift, eustatic

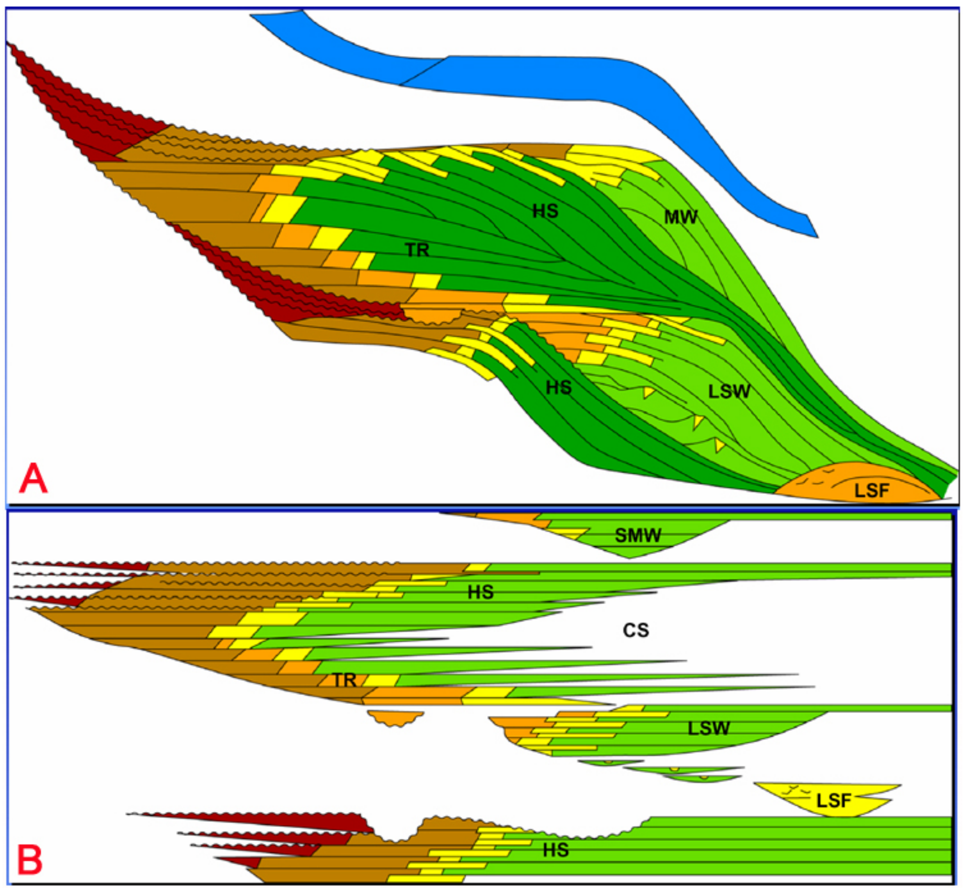



Figure 18.1: The classic Exxon sequence stratigraphic model (from Vail, 1987). A. is a schematic dip profile over a migrating coastline system (showing deposition as function of height and distance from the coastline), which shows key surfaces and depositional sequences, and B. is a *Wheeler diagram*, showing deposition as a function of time and space. That is, the Wheeler diagram demonstrates both where sedimentation is taking place and where non deposition or erosion is occurring through time.

rise in sea level, and sediment supply. Importantly, sequence stratigraphy is formally scale-independent; that is, depositional patterns occur at a variety of temporal and spatial scales (orders). This said, the fundamental units in sequence stratigraphy (those readily resolvable in seismic data) are commonly referred to as *third-order sequences* and are approximately at the scale of a formation.

### 18.3.1 Parasequences

Stratigraphers often simply their sequence stratigraphic analyses to subdividing sections into *parasequences*, which are distinct, coarsening upward cycles, most commonly identified in the coastal marine environment (Catuneanu, 2006). Van Wagoner (1995) defined a parasequences as 'a relatively conformable succession of genetically related beds or besets bounded by flooding surfaces.' A *flooding surface* is an abrupt shift from sediments inferred to have been deposited in relatively shallow-water conditions to sediments deposited in deeper water conditions. Generally, flooding surfaces are identified by a sharp decrease in grain size and are higher order elements of a larger-scale sequence.

Subdivision of successions into parasequences is a popular approach because these coarsening upward packages are common in the sedimentary record, and it provides a relatively robust means of correlating strata that is not lithofacies dependent. However, one criticism of parasequences is that their boundaries, while recognizable in the field, are not strict sequence boundaries. Indeed, a flooding surface may correspond to any of a number of sequence boundaries, and hence it is not always easy to interpret. One way around this problem is to define higher order sequences as transgressive-regressive sequences (T-R sequences), where the boundaries maximum regressive surfaces (or, you could identify them based on the maxim flooding surface).

# Chapter 19

# Sedimentary Basins

Additional Reading: Boggs, Chapter 16, p. 550-567; Bridge and Demicco, Chapter 20, p. 703-731

## 19.1 Concepts in Sedimentary Basin Analysis

What is a sedimentary basin?
Quite simply, we can define a *sedimentary basin* as a depression capable of trapping sediments. That is, a basin is a depression beneath regional *base level*. In the ocean, base level closely approximates sea level, but is located perhaps several meters below sea level since the upper few meters are subject to constant reworking by wave action.

- Basin formation generally requires a method of generating *subsidence*
  - *subsidence* is the downward motion of the crust relative to base level.
- Subsidence generates *accommodation space*, which can be filled by sediments
- *Basin analysis* is the integrative study of the sedimentology, architecture, tectonic setting, and subsidence history of a sedimentary basin

A sedimentary basin need not be a bowl-shaped depression. *Passive margins*, for example, are elongate sedimentary basins that probably hold more sediments than all other types of basins combined. The key to their success is long-term subsidence along the edge of a continent, which enables them to continue to collect sediments for up to a few hundred million years.

The dominant control on basin formation and sedimentation is tectonic. Climate also plays a role, but is subordinate too, and often influenced by tectonics. Importantly, the stratigraphic record in a sedimentary basin is typically the best way to read tectonic history.

### 19.1.1 Stratigraphy

*Stratigraphy*, which literally means "the description of strata" (Bridge and Demicco, 2007), is a subdiscipline of sedimentology and typically involves studying sedimentary patterns at a basinal scale. When sedimentary successions are examined at the basinal scale, it is commonly found that they are arranged in cyclic stratigraphic packages (discussed below).

The end goal of stratigraphy is commonly to reconstruct the 3D geometry of a basin and the distribution of stratal types, which together can be used to determine the type of sedimentary basin in which these rocks were deposited. The individual data sets that can be used to reconstruct basin architecture include outcrop and drill core logs, borehole data (e.g. *gamma ray* and *electrical conductivity logs*), and seismic surveys. Connecting these typically separated data sets requires correlation, which can be accomplished by some combination of radiometric dating, biostratigraphy, chemostratigraphy, and sequence stratigraphy.

### 19.1.2 Sequence stratigraphic contacts and units

*Sequence stratigraphy* is the the analysis of sedimentary basin fill based on the interplay between sediment supply and the generation of accommodation space. It is a powerful and increasingly applied tool in stratigraphy and basin analyses. The fundamentals and application of sequence stratigraphy will be covered in detail in EPSC 425. Here, you are introduced to the basics.

The premise of sequence stratigraphy is that sedimentary successions can be subdivided into surface-bound sequences where these surfaces reflect changes in *base level* in the basin, resulting from either eustatic changes in sea level, or uplift or subsidence of the basin. The implication is that these surfaces can be treated as approximate isochrons, and thus can be used for correlation across the basin, whereas the sediments between the surfaces may change facies laterally (thus being unsuitable for correlation). Sequences of a variety of scales can be recognized. Among the most important sequence boundaries (and the easiest to recognize in the field) are

- *exposure surface*: a surface that has been subaerially exposed and eroded, as a result of a *regressive* event (drop in base level). Thus, it is a disconformity; it may, however, correlate basinward with a conformable surface, marked by a sharp change in facies or the post regression flooding.
- *flooding surface*: a surface that marks a sharp shift to deeper water (usually finer-grained) sediments as a result of a *transgressive* event (rise in sea level).
- *maximum flooding surface*: a surface that records the maximum flooding (peak of transgression), usually in the form of the finest grained interval in a sequence.

## 19.2 Subsidence

Sedimentary basins may be created by a variety of mechanisms in many tectonic settings. Indeed, sedimentary basins are created in all active tectonic environments and simply require the generation of crustal subsidence.

- Crustal thinning: extensional stretching of the crust
- Mantle-lithospheric cooling
- Tectonic loading
- Sedimentary/volcanic loading

- Lithospheric underplating
- Geodynamic (Asthenospheric flow)

Before you can make sense of subsidence, it is necessary first to understand *isostatic compensation.*

- The principal of *isostasy* is that the lithosphere floats atop the asthenosphere in an analogous manner to an ice cube floating in a drink.
- *Isostatic compensaition* occurs when the lithosphere is in buoyant equilibrium with the underlying asthenosphere. That is, we can use Archimede's principle, which states that the mass of an object floating in a fluid is equal to the mass of the fluid it displaces.

### 19.2.1 Extensional subsidence

During rifting (extension), accommodation space is generated mechanically by thinning the lithosphere. The amount of accommodation space generated can be calculated using the principal of isostasy. Assuming uniform extension, the amount of crustal stretching is usually denoted by the stretching factor, $\beta$.

$$\beta = \frac{original - thickness - of - the - crust}{stretched - thickness - of - lithosphere} \tag{19.1}$$

That is, you can calculate the amount of *water-filled* accommodation space in a rift basin, as in Figure 19.2.1, by setting the mass of two imaginary columns, one through the unextended crust, and the other through the extended crustal, equal to each other.

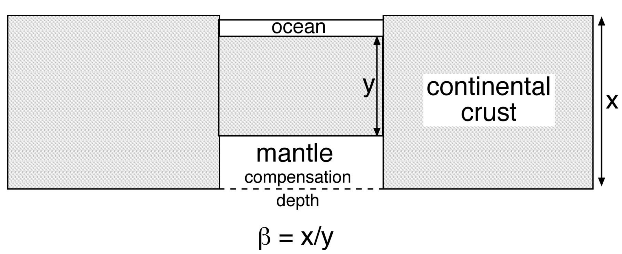



Figure 19.1: The mechanical generation of accommodation space (by extension, or thinning of the crust).

Now consider what happens when this basin begins to fill with sediments. These sediments displace the water, and because they have a great specific gravity, the add additional load to the crust. Isostatic adjustment to this load generates additional accommodation space. This effect is not to be underestimated. A sedimentary basin where only one kilometer of accommodation space was generated by tectonic subsidence will probably end up with 3 or 4 km of sediment.

### 19.2.2 Thermal subsidence

After the mechanical phase of extension, accommodation space continues to be generated, this time as a result of cooling of the mantle beneath the extended crust. This is the *thermal subsidence* phase and is the process by which *passive margins*, such as that on the east coast of North America, are formed. For a clue as to how mantle-lithosphere thickening generates subsidence, considering the bathymetric profile across mid-ocean ridges (Fig. 19.2.2).

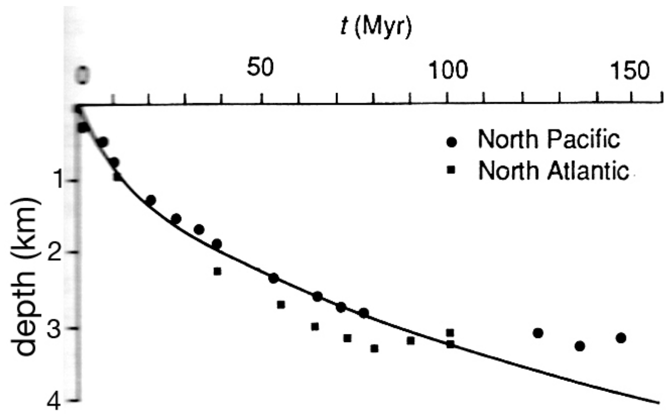



Figure 19.2: The depth to the seafloor across a mid-ocean ridge is proportional to $\sqrt{age}$. From Fowler (1990).

To understand how mantle-lithosphere thickening generates subsidence, consider the McKenzie (1978) 1D thermal subsidence model

- Assume uniform stretching
- Following an extensional event, which thins the lithosphere, the isotherms relax. That is the, the upwelled asthenosphere cools off. The cooling is driven by the heat flow equation:
  - $q = \kappa \frac{dT}{dy}$
    - where $q$ = heat flow (W m$^{-2}$)
    - $\kappa$ = thermal diffusivity
    - $\frac{dT}{dy}$ is the geothermal gradient

### 19.2.3 Flexure-related subsidence

- downward load, $l = D\frac{d^4w}{dx^4} + P\frac{d^2w}{dx^2}$, where
  - $w$ = the downward deflection (subsidence) of the plate
  - $D$ = the flexural rigidity
  - $P$ = a constant horizontal force applied to the plate
- Gives rise to a peripheral bulge that migrates with increased subsidence

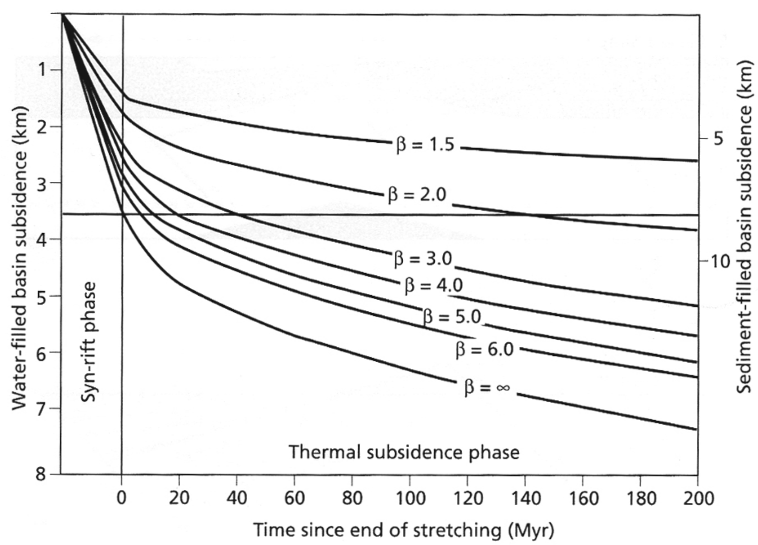



Figure 19.3: Thermal subsidence as a result of cooling of the lithosphere following uniform stretching of the lithosphere. From Dewey (1982).

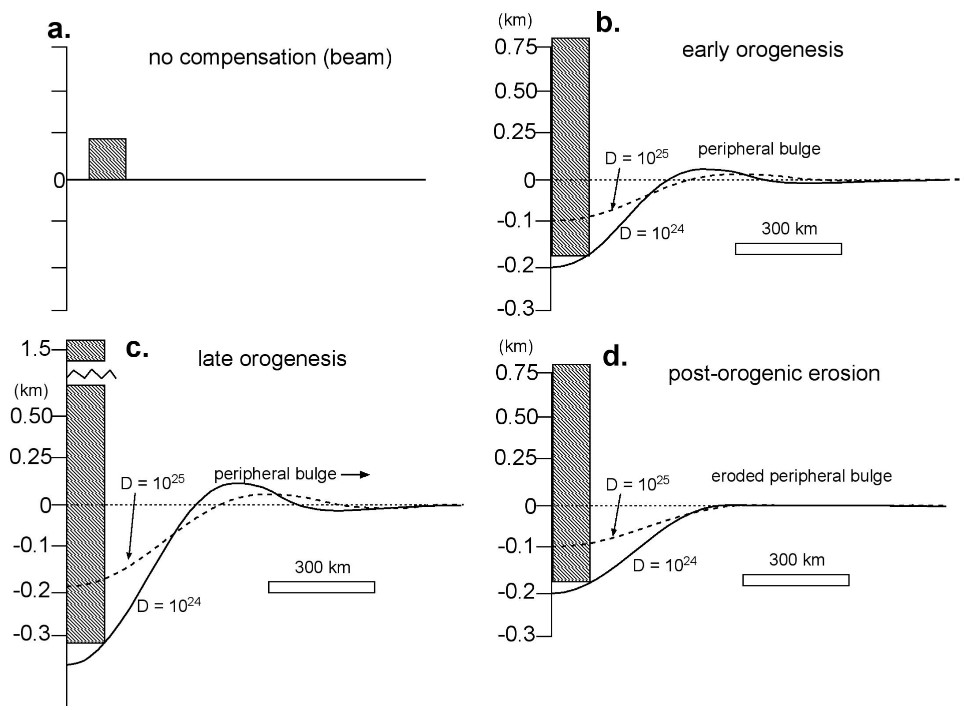



Figure 19.4: Subsidence resulting loading on a plate (e.g. mountain building) can be modeled as the flexure of a rigid beam beneath a line load (modified from (Fowler, 1990) and (Leeder, 1999).

- The amount of subsidence, uplift of the flexural bulge, and location of the flexural bulge all heavily influenced by the flexural rigidity of the plate

- Rate of subsidence limited by viscosity of the mantle
- post-orogenic erosion results in rebound

## 19.3 Tectonic Settings of Sedimentary Basins

Sedimentary basins form in all tectonic (and some geodynamic) environments

- Divergent
- Convergent
- Transform
- Intraplate

### 19.3.1 Divergent settings

Divergent basins essentially are rift basins, although many smaller scale basins form by extension (normal faulting) in a variety of tectonic environments. Rift basins possess various distinct features that may help distinguish them in the stratigraphic record:

- Basins (hanging wall) tend to flank uplands (footwall), generating significant relief and giving rise to very rapid sediment accumulation rates
- Rift basins are long (often 1000 or more km) and narrow (10-50km)
- Associated with bimodal volcanism (that is, rhyolites and basalts)
- Abrupt facies changes, due to steep slopes and rapid changes in topography/bathymetry
- Abundant coarse-grained sediment (i.e. alluvial fans)
- Wedge-shaped (asymmetric) sedimentary packages - onlapping relationships
- Rotation and erosion of units
- May form a network of intersecting sub-basins

Extensional basins can be subdivided into continental rifts, coastal marine rifts, and proto-oceanic basins

- Internal drainage basins characterized by alluvial fans shed off the hanging and foot-walls and grading into large lakes or playas
  - Death Valley
  - East African Rift Valleys, such as Lake Tanganyika
- Axial-through drainage continental rifts are characterized by alluvial fans interacting with meandering river systems
  - Rio Grande Rift and Basin and Range province (U.S.A.)
  - The Rhine Basin (Europe)

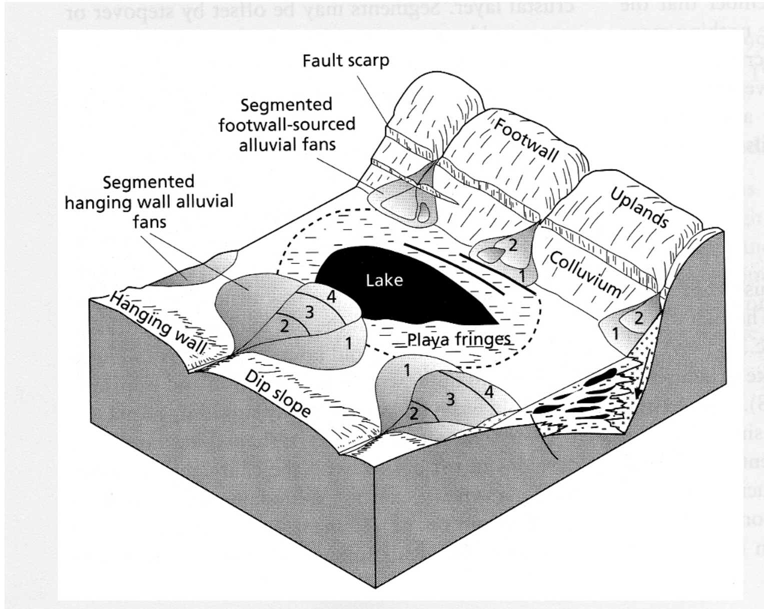



Figure 19.5: Sketch of an internally-drained continental half-graben (Leeder, 1999).

Coastal marine rifts are similar to continental rifts, but characterized by large fan deltas

- May be either deep or shallow
- May be anoxic (black shales)
- Include basins in the Aegean Sea

The Red Sea and Gulf of Aden are quintessential proto-oceanic basins
Proto-oceanic basins precede passive margin basins, often separated by a major regional unconformity

- Contain interbedded basalts
- Dissected by mafic dikes
- Commonly link up to failed rifts, or *aulacogens*

### 19.3.2 Passive margins

After active rifting gives way to seafloor spreading, the continental margin continues to subside under thermal subsidence

- Normal faulting desists
- Older rift-related basins are filled and structures draped
- Broad coastal plain builds out from continental margin
- Commonly 100s of km wide 1000s of km long

- Passive margin sequence takes on a sigmoidal cross section shape as it build out over the continental margin
  - Complete marginal marine to deep basin transition
  - Slope sediments recognized by tabular geometry and reworked nature (debris flow deposits, turbidites, rhythmites)
- seaward expanding wedge
- flexural forebulge in hinterland

Volumetrically, passive margins are the most important of the sedimentary basins preserved in the sedimentary record. Historically referred to as *geosynclines*.

The transition from rift-basin (rift phase) to passive margin (drift phase) generally straightforward to identify in seismic sections/stratigraphic cross-sections.

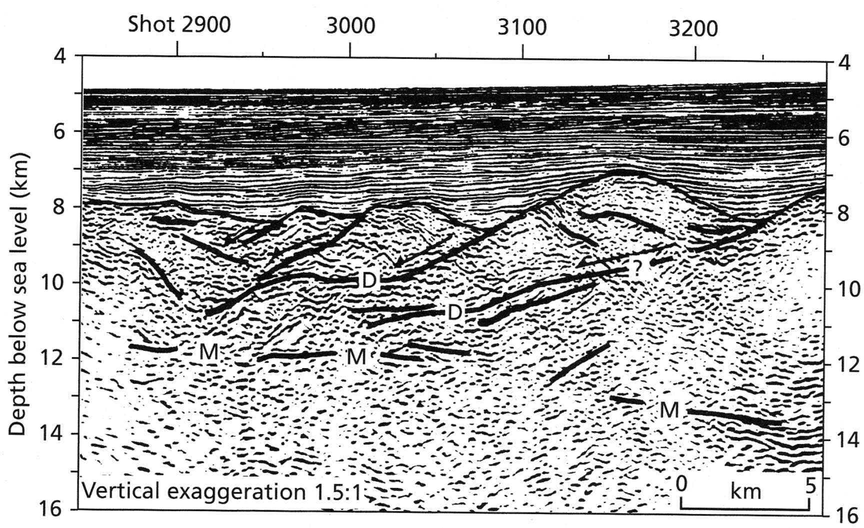



Figure 19.6: Seismic reflection cross section of Iberian margin (Pickup et al., 1996).

### 19.3.3 Intracratonic basins

So-called *Intracratonic* basins occur within a platethat is, far a field from active plate margins. Where complete, they are commonly characterized by their circular geometry (in map view) and lack of obvious connection with faulting.

- Symmetric sag in cross-section
- Hundreds of kilometers wide
- Sedimentation may last for hundreds of millions of years
- Particularly common on North American continent

  - Michigan basin
  - Illinois basin

- Their mechanism of formation has long been a conundrum

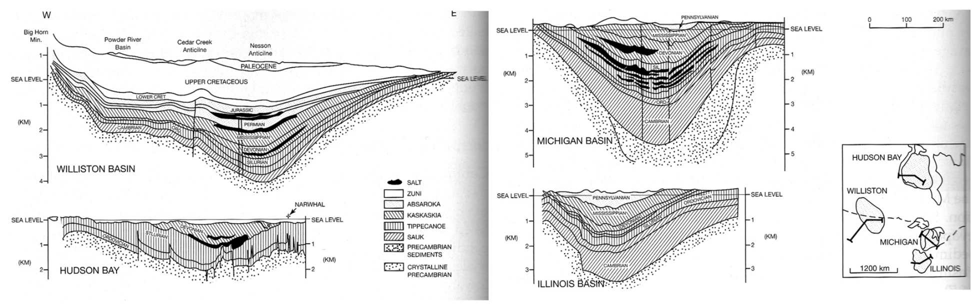



Figure 19.7: Cross sections through several North American intracratonic basins . From Boggs (2011).

Whereas intracratonic basins remain somewhat mysterious, they can be modeled to have formed largely through the effect of thermal subsidence. One idea is that they form where hot mantle plume impinges upon the lithosphere, causing uplift, erosion, and some extension (which generates some accommodation space), and presumably erosion or thinning of the lower lithosphere. When upwelling ceases, the hot mantle within the thinned lithosphere cools and acts as a sort of sinker, generating a circular depression.

Mantle dynamic processes related to subduction zones may also generate intracratonic basins, e.g. the Cretaceous interior seaway (west-central U.S.A.). In this case, it is the downwelling of mantle that deflects a broad swath of overlying lithosphere downward.

- Probably related to mantle flow resulting from phases change in mantle
- Transient by nature (dynamic topography not permanent)

### 19.3.4 Convergent

Convergent tectonic settings are the most complicated of the plate boundaries, and accordingly, a large variety of basins may form in convergent settings. To facilitate the description of these basins, we'll look at the basins that form in subduction settings and collisional settings.

**Basins around subduction zones**

- *Trenches* are the deepest basins on Earth. They are generated by loading of the ocean crust and tend to have long, linear geometries. The sediments deposited in trenches are largely scraped off of the oceanic crust and plastered onto the overriding accretionary prism
- *Fore-arc basins* form on the side of an arc facing the subducting plate. Forearc basins are a type of flexural basin, where accommodation space is generated mainly as the

result of the load of the arc itself. Fore-arc basins lie on the transition between oceanic and continental crust and commonly contain slivers of ophiolites. The are sediment traps for eroding volcanic arcs and typically have high sedimentations rates and are filled with lithic wackestones.

- *Retro-arc foreland basins* occur on the opposite side of the arc and again are largely generated by the load of the volcanic arc. However, they may also have associated fold and thrust belts. The Rocky Mountain foreland and the Andean foreland are both examples of retro-arc foreland basins.

- *Back-arc basins* form where the continental crust on the opposite side of the arc from the subduction zone undergoes extension (i.e., during slab rollback). Extension in back arc basins is commonly sufficient to generate new oceanic crust, and so these basins may accommodate great thickness of typically immature, volcaniclastic sediments.

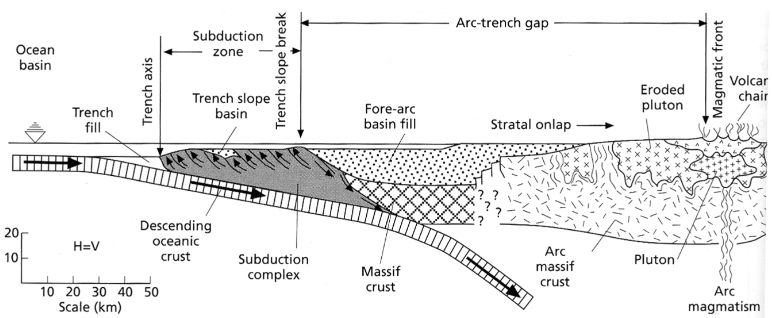



Figure 19.8: Linear *fore-arc basins* occur shoreward of the volcanic arc and landward of the accretionary prism (from Leeder, 1999).

### Basins around continental collision zones

Fore-arc basins lie on transitional basement (between continental and ocean crust)

- Sediment traps for eroding volcanic arcs
- High sedimentation rates
  - volcanic airfalls
  - submarine debris flows and slumps
  - turbidity currents
  - lithick (volcanic) greywacke a common lithology

Back-arc basins are geometrically similar to rift basins

- Abundant volcanics, including pillow basalts and ash fall tuffs
- Lithic arenites and greywackes common
- Following extension, thermal subsidence continues

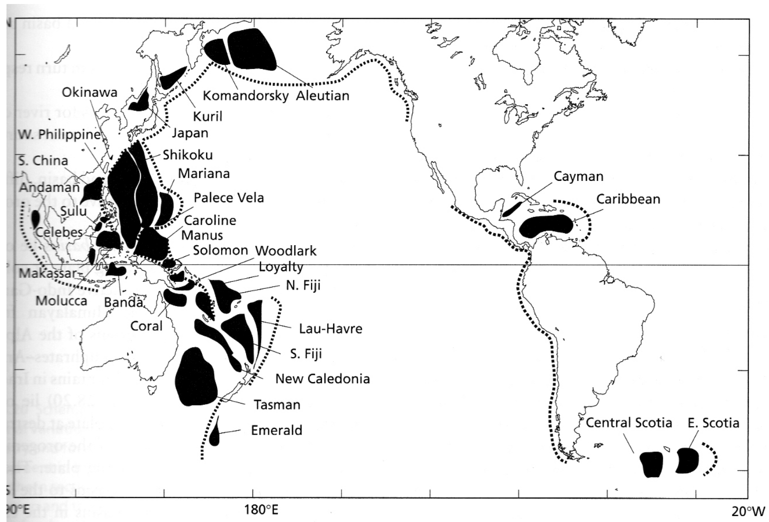



Figure 19.9: *Back-arc basins* are extensional basins that occur on the opposite side of the volcanic arc from the trench and fore-arc basins. From Fowler (1990).

Foreland basins are flexural basins that form on convergent plate margins

- Tend to form long and interconnected series of basins
- Assymetric geometry
- Maybe be marine or non-marine
- Associated with fold-thrust belts
- Basins migrate and sediments are recycled (*cannabilized*).

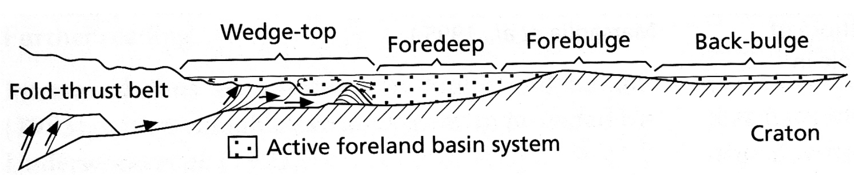



Figure 19.10: Idealized sketch of a foreland basin (Leeder, 1999).

*Foreland basins* form on continental crust along the length of collisional (cont.-cont.) or destructive (subduction zone) plate margins

- Peripheral foreland basins occur on the subducting plate in a collision zone
  - e.g. Indo-Gangetic foreland basins south of the Himalayas
- Retro-arc foreland basins occur on the continental crust of the over-riding plate at destructive margins

– e.g. The Rocky Mountain forelands
– The Andean forelands

Foreland basins are the uppermost component of the classic Wilson cycle trinity: rift-drift-collision

### 19.3.5 Transform

Small, *pull-apart* basins occur along bends in major transform faults.

- Can essentially be thought of as small rift basins
- Have very steep, fault-bound, margins with proximal, coarse-grained sediments
- Include the Sultan Sea, Dead Sea

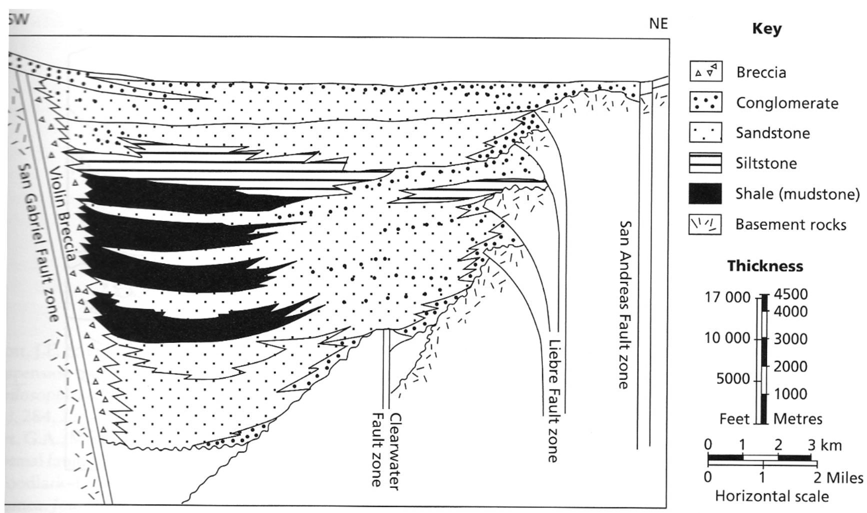



Figure 19.11: Sediment fill in the strike-slip Ridge Basin, southwestern California (from Fowler, 1990).

# Chapter 20

# Resourceful Sedimentary Rocks

## 20.1 Introduction

## 20.2 Coal

## 20.3 Petroleum

What is Petroleum?

*Petroleum* is a mixture of hydrocarbons and other organic molecules containing sulphur, oxygen, nitrogen, and a small amount of metals (mainly Ni and V)

- Petroleum includes tars, oils (heavy - light), and gas
- Hydrocarbons and organic molecules are derived from organic matter, either as a natural organic product or during maturation and alteration of buried, sedimentary organic matter (*Kerogen*).

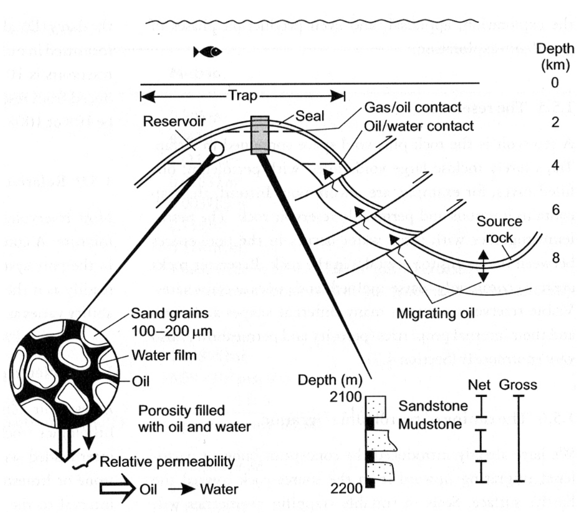



Figure 20.1: Schematic cross-section of a petroleum system (Gluyas and Swarbrick, 2004)

The 5 key ingredients in a petroleum system

- Source
- Timing (migration)
- Seal
- Trap
- Reservoir

### 20.3.1 Source

A *source rock* is one that contains sufficient organic matter such that when it is buried and heated, it will produce petroleum

- organic matter = proteins, carbohydrates, lipids, lignins
- *diagenesis* is the alteration of sediments after deposition and prior to metamorphism. In the case of organic matter, it s the microbrial reprocessing of organic matter that occurs at temperatures of 50°C and less. It involves the remineralization of labile organic matter and the release of gases such as $CO_2$ and (biogenic) $CH_4$.
- *catagenesis* is the thermal alteration of organic matter between 50°C and 200°C. It is during catagenesis that oil and gas-prone source rocks will produce petroleum.

Sedimentary organic matter consists of *kerogen* and *bitumen*

- **Kerogen** is the disseminated, insoluble (in organic solvents or acids) component of sedimentary organic matter residual after diagenesis. It is subdivided into four main types, based on its source:
  - Liptinite
  - Exinite
  - Vitrinite
    * breakdown product of woody tissue
    * Vitrinite Reflectance ($R_0$) a common measure of source rock maturity
    * main phase of oil generation at 0.65-1.30% $R_0$
  - Inertinite
- **Bitumen** is the soluble (in organic solvents) component of organic matter

Kerogen and bitumen are dominantly derived from *lipids*, which are fatty, hydrogen-rich, organic-compounds that are relatively resistant to degradation

Organic carbon generally comprises only a small fraction of the weight percent of sedimentary rocks. 99.5% of all primarily-produced organic matter is *remineralised - that is, not buried*

- The average shale contains about 1 weight% organic carbon
- The average carbonate contains 0.33 weight% organic carbon

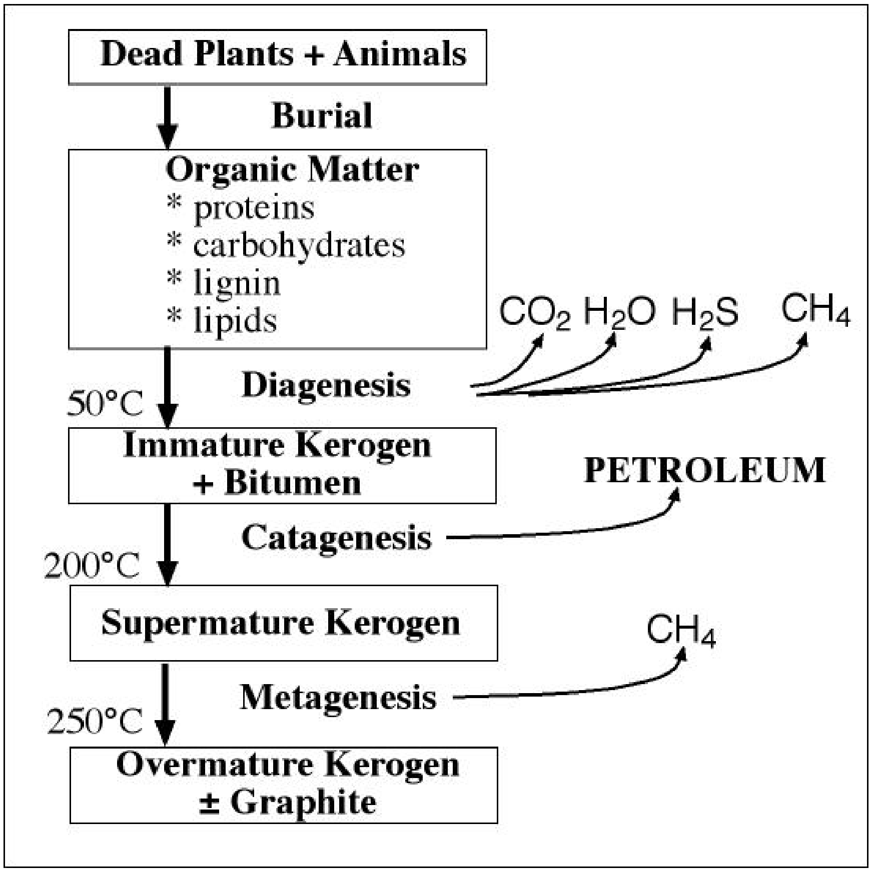



Figure 20.2: Schematic of the evolution of sedimentary organic matter

- The average sandstone contains 0.28 weight% organic carbon

Petroleum source rocks contain 1 - 20% organic carbon.

There are two basic requirements for the preservation of sufficient organic matter to generate a viable source rock:

- high primary productivity
- oxygen deficiency in the water column or sea bed

Four common environments of high primary productivity include

- continental margins
- lagoons and other restricted bodies of shallow seawater
- deltas in low latitudes
- lakes

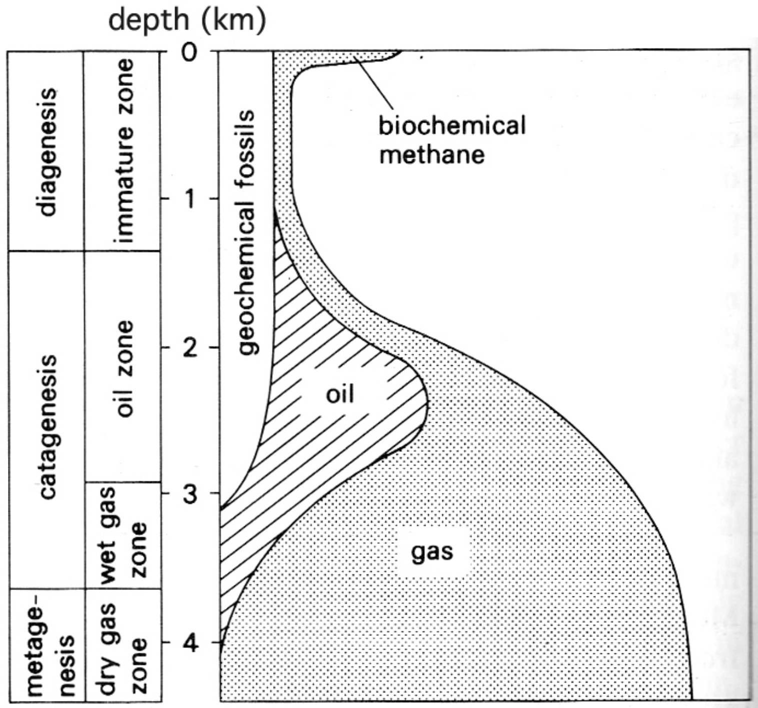



Figure 20.3: The oil-gas window (From Tucker, 1990)

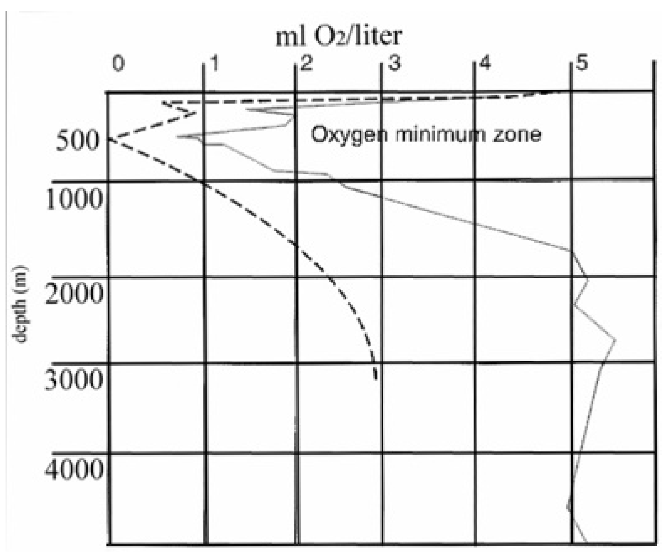



Figure 20.4: The Oxygen Minimum Zone

### 20.3.2 Seal

The *seal* is the component of the petroleum system that prevents petroleum from migrating upwards

- Membrane seals

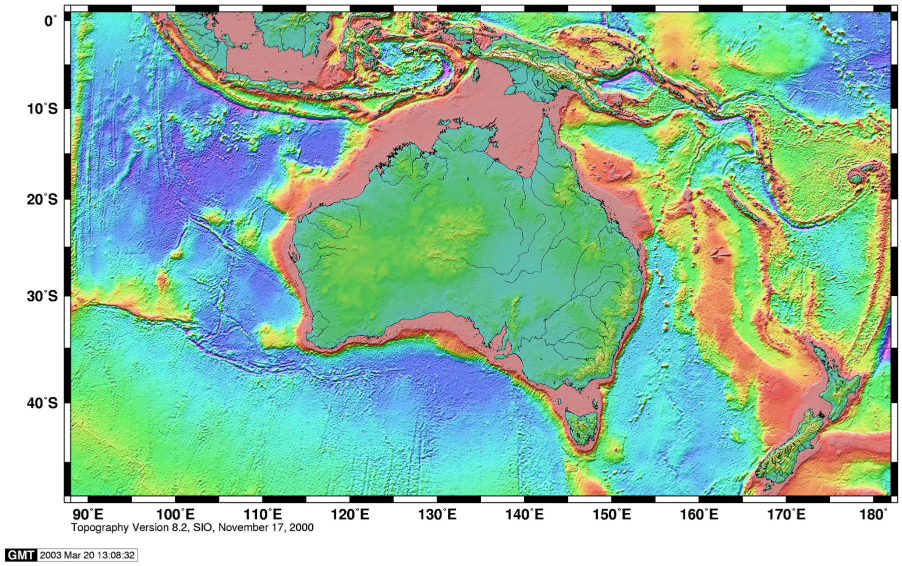



Figure 20.5: The Australian continental shelf

- Hydraulic seals

A *membrane seal* is one in which the capillary pressure in pore spaces exceeds the buoyant pressure of the petroleum

- A membrane seal occurs where $P_d > P_b$
- Buoyant pressure: $P_b = (\rho_w - \rho_p)h$
  - $\rho_w$ = the density of water
  - $\rho_p$ = the density of petroleum
  - $h$ = the height of the petroleum column above the free water level
- Capillary entry pressure: $P_d = \frac{2\gamma cos\theta}{R}$
  - $\gamma$ = the interfacial tension between the water and the petroleum
  - $\theta$ = contact angle between water and petroleum at the pore face
  - $R$ = the radius of the largest pore

Most membrane seals are mudrocks (shales)

- fine-grain and small pore size
- high ductility
- large thickness
- long lateral extent

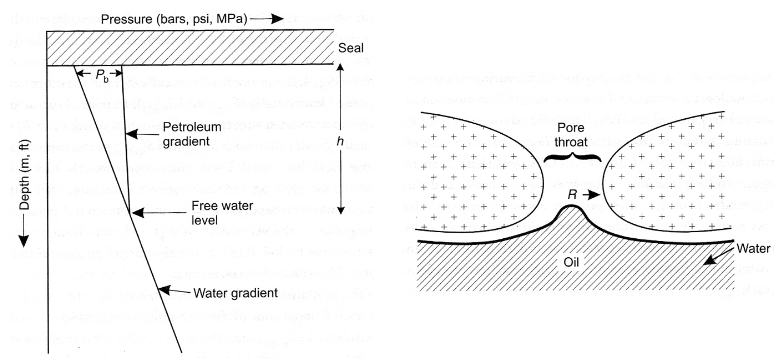



Figure 20.6: Graphic depiction of the Buoyant and Capillary entry pressures governing the effectiveness of a membrane seals

A *hydraulic seal* is one which the cap rock is extremely tight and its sealing properties are controlled by its brittle strength

- Can think of this as a type of membrane seal in which the capillary entry pressure is extremely high
- But when the buoyant pressure is sufficiently high, it will fracture the seal
- Common hydraulic seals include
  - evaporites (salt, anhydrite)
  - very tight shales
  - gas hydrates
  - cherts

Faults can act as both conduits and barriers to fluid flow, depending on the hydraulic conditions and the properties of the rocks juxtaposed across the fault. The processes that reduce fault permeability include

- clay smear
- cataclasis
- cementation

### 20.3.3 Trap

The *trap* is the geometry of the petroleum-bearing container. This geometry can be

- Structural
  - Tectonic
    - Compressional (thrust fault, anticline)
    - Extensional (half graben)
    - Compactional (basement high)

- Diapiric
- Gravitational (rollover anticline)

- Stratigraphic
  - Depositional
    - Pinchouts
    - Unconformities
  - Diagenetic
    - Dissolution
    - Dolomitization
    - Gas hydrates

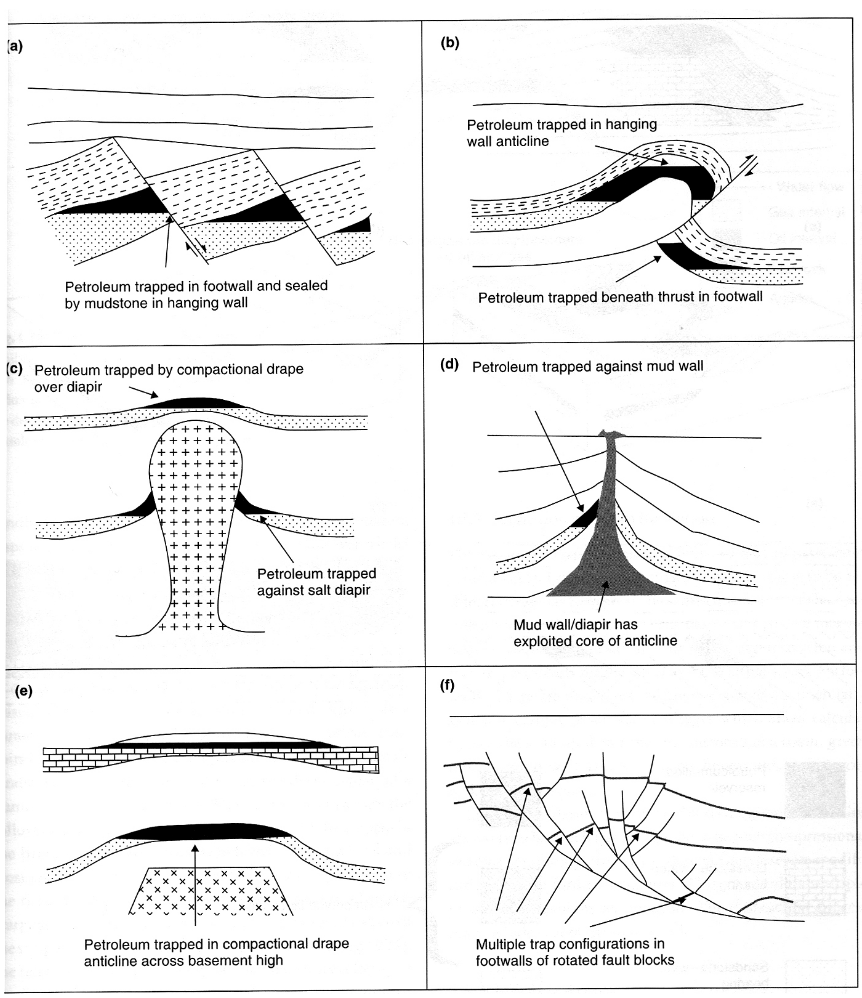



Figure 20.7: Examples of common structural traps (from Gluyas and Swarbrick (2004)

Three classes of traps:

- Class 1: seal strength is high enough that there is no leakage before petroleum fills to spill point

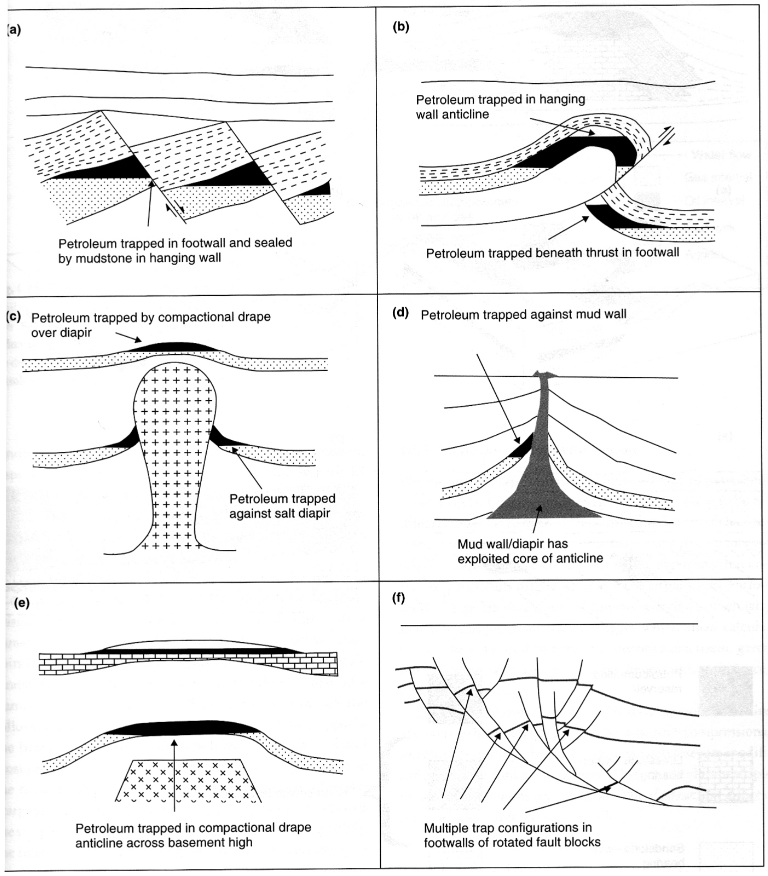



Figure 20.8: Examples of common structural traps (from Gluyas and Swarbrick (2004)

- Class 2: trap fill to spill point, but there is leakage through the seal
- Class 3: leakage through the seal occurs before the spill point is reached

Or you can simply regard traps as 'full' or 'underfull'

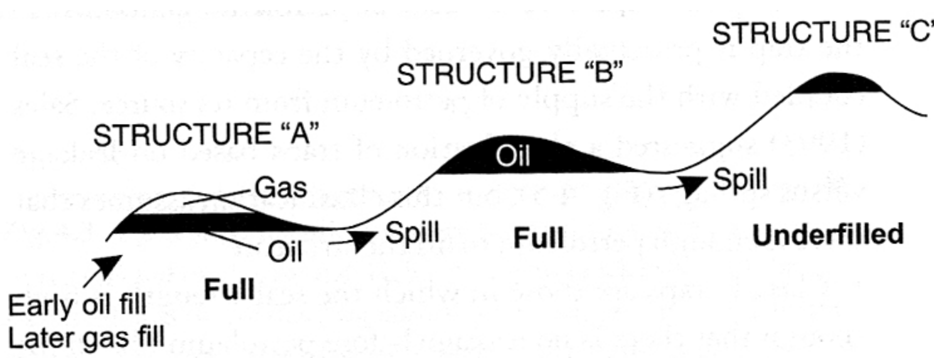



Figure 20.9: Sketch of various state of 'fullness' of traps (Gluyas and Swarbrick, 2004).

### 20.3.4 Reservoir

The principal requirement for a petroleum *reservoir* rock is that it be porous. To be an economically viable reservoir, it must also be permeable. The intrinsic reservoir properties used to establish a reservoir's potential are

- Porosity
  - The void space in a rock
- Permeability
  - Measure of how well a fluid can flow through a material
- Net to gross
  - a measure of the potentially productive part of the reservoir
- Hydrocarbon saturation
  - The proportion of a reservoir that is filled with hydrocarbons

**Porosity** is expressed as a percentage of total void space within a volume of rock. A typical reservoir will have a porosity of 10-30%. In terms of reservoir characteristics, there are three types of porosity

- Intergranular (primary)
  - well sorted sands yield excellent primary porosity
  - some limestone reef rocks (i.e. bafflestones) also have very high primary porosity
- Intragranular (secondary)
  - Fracture porosity
  - Dissolution
  - Karst
- Microporosity

**Permeability** is measured in millidarcy (mD) and is a measure of flow rate for a fluid whose viscosity equals 1 cp under a pressure of $10^{-4}$ atm $m^{-2}$.

- Permeability factors in both porosity and the interconnectedness of pore spaces
- Permeability of reservoir rocks ranges from 0.1-10000 mD.
- Permeability is lower (by approximately an order of magnitude) at typical subsurface pressures.
- The viscocity of petroleum plays a large role in its ability to flow through a reservoir rock
  - a low viscosity petroleum can be recovered from a low permeability reservoir

**Net-to-gross** is measured as a percentage of the potentially productive volume of a reservoir. It is common to establish a threshold permeability for calculating net to gross.
Oil and gas saturation ($S_o$, $S_g$) refers to the amount of pore space occupied by oil and gas, respectively.

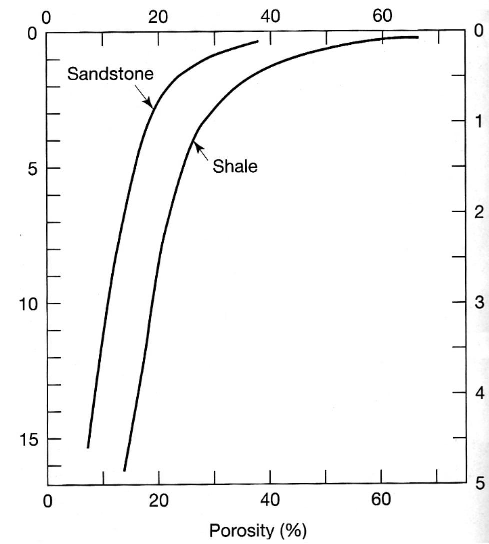



Figure 20.10: Porosity decreases logarithmically with depth

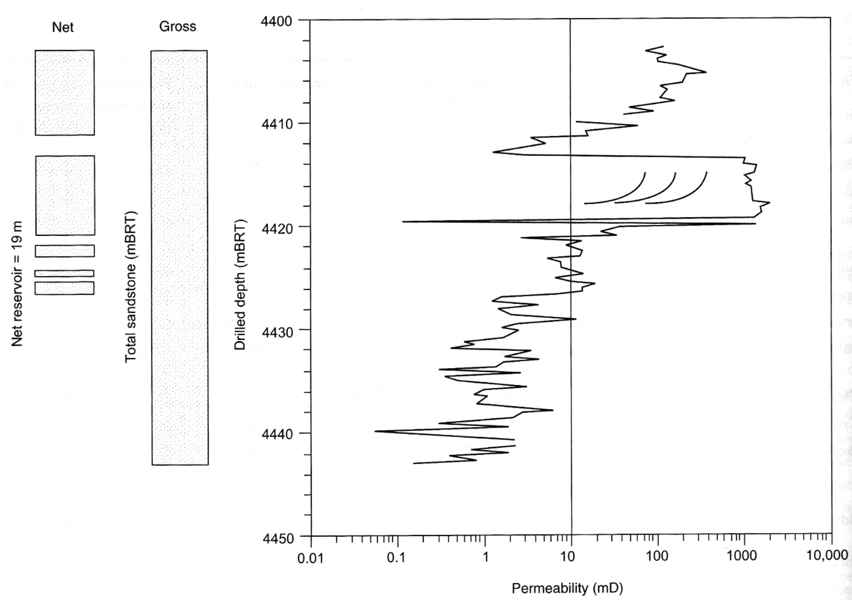



Figure 20.11: Illustration of how net-to-gross is determined on a reservoir (Gluyas and Swarbrick, 2004

### Other factors influencing reservoir potential

In considering the volume of available petroleum in a reservoir (*oil-* or *gas-in-place*), the *formation volume factor* and *recovery factor* must also be considered.

The *formation volume factor* ($B_o$) represents the change in volume of oil that will occur as result of being extracted from the subsurface and placed in a stock tank (at earth surface conditions).

- total oil + gas system is considered
- $B_o = \frac{Volume at reference conditions}{Volume at reservoir conditions}$
- Volume change is a function of
  - partial molar volume of gas in solution
  - thermal expansion of the system
  - liquid compressibility
- cannot be defined by a simple equation of state
- $B_o$ increases with increasing pressure up to the bubble point, where it decreases due to the exsolution of gas.

Some sandstone depositional systems that make suitable reservoir rocks:

- Fluvial systems
  - braided
  - meandering
- Deltas
- Marginal marine systems
  - shoreface sands
  - barrier island
  - tidal systems
- Submarine fans

Carbonate depositional systems that make suitable reservoir rocks are almost entirely those of the shallow marine realm:

- Shelf and ramp carbonates
  - host to the worlds largest petroleum reservoirs (Saudi Arabia)
- Reefs
- Karstified carbonates

### 20.3.5 Migration

Migration is the process by which petroleum moves from its source to the earth's surface

- Primary migration
  - Expulsion of petroleum from the source rock
- Secondary migration
  - The movement of petroleum from the source rock to the trap

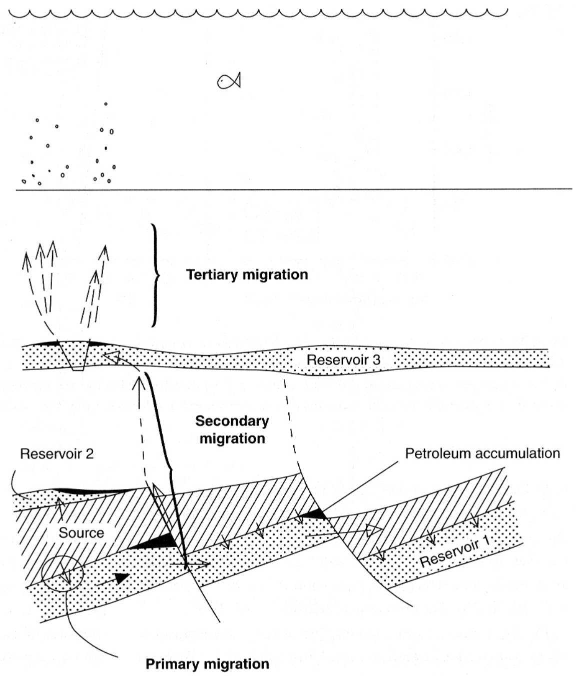



Figure 20.12: Schematic diagram illustrating the three phases of petroleum migration (from Gluyas and Swarbrick)

- Tertiary migration
  - The leakage and dissipation of petroleum at the earth's surface

Primary migration is driven by pressure differences between the source and carrier beds. This migration is generally thought to occur with petroleum as a separate phase (from water).

- Expulsion large controlled by petroleum generation, and thus maturation of the source rock
- Essentially occurs when the volume of generated petroleum exceeds the pore space in the source rock

Subsequent gas generation expels additional hydrocarbons from pore spaces

Methane is commonly transported in solution by water

- highly soluble at higher temperatures and pressures.
- exsolves at shallower depths

Secondary migration channelizes and concentrates dispersed petroleum.

- Driven mainly by buoyancy: $\Delta P = Y_p g(\rho_w - \rho_p)$
  - $\Delta$ P = the difference in pressure (that drives the buoyant force)

  - $Y_p$ = height of petroleum column
  - $g$ = gravitational acceleration
  - $\rho_w$ = the density of water
  - $\rho_p$ = the density of petroleum

- Hydrodynamics can also play a role in transporting petroleum
- Capillary effects cause petroleum to migrate from small to large pores
- Rate of petroleum migration governed by Darcy's Law
  - For typical sandstones: 1-1000 km my$^{-1}$
  - For typical sandstones: 1-10 km my$^{-1}$

A certain percentage of petroleum will be lost to dead-end pore spaces and mini-traps

Secondary migration may causes various chemical and physical changes to petroleum

- Phase changes occur as pressure and temperature decrease
  - exsolution of gases
  - condensation of oils
  - changes viscosity and density of the migrating petroleum
- Bacterial degradation at lower temperatures
- Water-washing
  - increase in viscosity
  - decrease in specific gravity
- *geochromatography* - separation of the petroleum components

Tertiary migration is driven by the same processes as those that drive primary and secondary migration

- Can involve higher flow rates than primary or secondary migration
- Vertical or horizontal motion
- very broad or focused leakage
- Promoted by horizontal bedding and rapid deposition of seals
- An important tool in petroleum exploration
- Some common manifestations
  - gas chimneys
  - mud volcanoemounds
  - gas hydrate layers

Timing of primary and secondary migration is critical: What is the timing of petroleum production and migration relative to development of suitable reservoirs, traps, and seals?

The *API gravity* of an oil is a measure of density

- $°\text{API} = \frac{141.5}{specificgravityat60°F} - 131.5$
- °API increases with decreasing density
- higher °API = larger volume of oil per metric ton

### 20.3.6 Hydrocarbon kitchens and generative basins

A *hydrocarbon kitchen* is the region of

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