SEDIMENTARY GEOLOGY: ROCKS, ENVIRONMENTS AND STRATIGRAPHY

Michael Rygel and Page Quinton
SUNY Potsdam

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Sedimentary Geology: Rocks, Environments and Stratigraphy

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This text was compiled on 11/21/2025

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About this Book

1: Preface

There are numerous books about sedimentary rocks and stratigraphy, but their contents are often locked down by vague or restrictive copyright and not easily used for teaching, coursepacks, etc. This creates a real problem for educators. We found the following resources particularly inspirational and tried our best to emulate them in this single open access resource:

- The comprehensive and organized photographs in Dorrik Stow's 2005 book "Sedimentary Rocks in the Field - A Color Guide", usable and comprehensive photograph collections by **Mark Wilson (Wikimedia Commons), James St. John (Flickr)**, and **Marli Miller (Geology Time Pics)**. We also have to give a nod to photographs at the intersection of art and science by Jessica Martin (@jmgeophoto) and Franco Ricci Lucchi's 1995 book "Sedimentographica: A Photographic Atlas of Sedimentary Structures"
- The elegant hand drafted diagrams in Robert Folk's 1968 coursepack for UT Austin "Petrology of Sedimentary Rocks" and Chelsea Feeney's illustrations in Hyndman and Thomas' 2020 (2nd ed.) book "Roadside Geology of Montana"
- The straightforward explanations and diagrams on **Lynn Fichter's webpages** (much of which was published in Fichter and Pouche's 1993 book "Ancient Environments and the Interpretation of Geologic History", **Brian Rickett's Geological Digressions webpage**, **Alessandro Da Mommio's Alex Strekeisen webpage**, and Samuele Papeschi's Geology is the Way website.
- The paired photomicrographs and descriptions of sedimentary rocks, fossils, and minerals in Scholle and Ulmer-Scholle's 2003 book "A color guide to the petrography of carbonate rocks" and Adam's et al.'s 1984 book "Atlas of sedimentary rocks under the microscope"

BOLD indicates sources that make content available either via licensing or for educational use

2: Use, Adoption, and Suggestions

If you use this book, or some part of it, please drop us an email (rygelmc@potsdam.edu and quintopc@potsdam.edu) to let us know. Please feel free to send us comments or suggestions as well.

3: Book Organization

Chapter 1 provides a brief overview of why sedimentary rocks are important. The rest of the book is organized around three major themes:

Origin, description, and interpretation of sedimentary rocks (Chapters 2-8)

In these chapter we discuss the details of sediment ... where it comes from, how it moves, how to describe it, and how to describe the resulting sedimentary rocks. Ideally, labs to accompany these chapters would focus on hands-on skills including how to use a petrographic microscope, how to identify common sedimentary structures, and how to quantitatively and qualitatively describe both rock and unconsolidated sediment.

Fossils and depositional environments (Chapters 9-10)

As Charles Lyell said, “the present is the key to the past” and we will take a tour of modern environments and then think about what they might look like if preserved in the geologic record. Practical skills in associated labs should include facies analysis, linking processes to environments, and core description and analysis.

Sea level and stratigraphy (Chapters 11-12)

In the final portion of the book we will gain an understanding of the spatial and temporal relationships between units, learn how to lump packages of rock together based on genetic relationships, lithologic similarity, fossil content, and/or age. Ideally, associated lans would focus on correlation exercises and cross sections.

4: About the Authors

Dr. Michael Rygel holds a B.S. in Geology and Planetary Science from the University of Pittsburgh at Johnstown, a Ph.D. from Dalhousie University (Halifax, Nova Scotia), was a post-doctoral fellow at the University of Nebraska-Lincoln, and is a Professional Geologist (licensed in NYS). He is a Professor in the Department of Earth and Environmental Sciences at SUNY Potsdam and teaches courses on Earth history, sedimentary geology, structural geology, GIS, and introductory courses. He worked as a consultant for Devon Energy from 2011-2013 and spends his summers (since 2011) co-teaching Indiana University's "Field Geology in the Northern Rocky Mountains" (X429) course. He is a Past-President of the New York State Council of Professional Geologists and served as an Associate Editor for the Canadian Journal of Earth Sciences and the Journal of Sedimentary Research.

He has deep ties to the steel mills and coal mines of western Pennsylvania and his research focuses on understanding the Paleozoic-aged rocks that these industries were built on. Specifically, he is a sedimentologist and stratigrapher that specializes in understanding ancient environments. Past research projects have focused on outcrops in the coal-bearing rocks in the Appalachian Basin (PA-WV) and Maritimes Basin (Atlantic Canada) as well as coeval ancient glacial deposits in southeastern Australia. His current research focuses on the sedimentology and stratigraphy of Late Paleozoic carbonates in Montana, west Texas, and New Mexico.

Dr. Page Quinton holds a B.S. in Geology from Columbus State University and M.S. and Ph.D degrees from the The University of Missouri. She is an Associate Professor in the Department of Earth and Environmental Sciences at SUNY Potsdam and teaches courses in Earth history, geochemistry, paleontology, scientific writing, and a variety of introductory courses. She serves as the STEM Coordinator at SUNY Potsdam and facilitates hundreds of K-12 students visiting the SUNY Potsdam campus every year. Since 2016 she has co-taught Indiana University's "Field Geology in the Northern Rocky Mountains" (X429) course.

She is a stable isotope geochemist that studies the connection between the carbon cycle, ancient climate change, and major mass extinction events. Recent work has included research focusing on the role of climate change in the dinosaur mass extinction event, understanding the impact of early land plants on shallow seas, and understanding the connection between sea level change and carbon cycling. This work has taken her to Australia, Canada, all over the US (Montana, New Mexico, Texas, and numerous other locations).

5: Artificial Intelligence

All of the written content of this book was written by Rygel and Quinton or adapted from other OER resources. Given the lack of formal copyediting, we ran content through ChatGPT and ask it to look for typos, repeated words, and to make minor editorial suggestions to improve clarity and flow. We individually vetted all AI suggestions. None of the images or diagrams that we created were generated by AI.

6: Acknowledgements

This work was made possible by NSF Grant #2042276 to Quinton and Rygel and sabbaticals granted by SUNY Potsdam.

7: Readings and Resources

- Carbonate World webpage: https://carbonateworld.com/
- SEPM's stratigraphy webpage - http://www.sepmstrata.org/page.aspx?pageid=1
- Sedimentary Rocks entry on the Geology is the Way webpage: https://geologyistheway.com/sedimentary/
- Alek Strekeisen's webpage entries dedicated to sedimentary rocks: https://www.alexstrekeisen.it/english/sedi/index.php

Licensing

A detailed breakdown of this resource's licensing can be found in ***Back Matter/Detailed Licensing****.*

1: Introduction

Sedimentary rocks are composed of material derived from the weathering and erosion of other, older rocks. They cover the vast majority of the terrestrial surface of the Earth and are an important archive of the history of life on Earth, ancient environments and processes, and are of profound economic and environmental significance. This book is a first step in learning how to "read the rocks" and understand the stories they tell us.

Learning Objectives

- Summarize the distribution of sedimentary rocks on the surface of the Earth and the relative abundance of the different sedimentary rock types
- Explain why sedimentary rocks provide important information about the ancient environments and ecosystems
- Summarize the economic, industrial, and practical importance of sedimentary rocks

1.1: Abundance and distribution

Figure 1.1: Sedimentary cover of North America (USGS via Wikimedia Commons; public domain).

Sediment and sedimentary rocks cover the majority of the surface of the Earth; they cover about 70% of the continents and about 90% of the seafloor. Although widespread, they represent a relatively thin veneer and make up approximately 8% of the crust and less than 1% of our planet's total volume.

The geologic record of sedimentary rocks is skewed toward younger rocks and relatively fine grained rocks deposited in ocean basins and rapidly subsiding areas of sedimentary basins. Although conglomerates and breccias are striking, they form in proximal areas that are not likely to be preserved whereas mudrocks, sandstones and limestones are deposited in more distal areas that are more likely to subside and be preserved as part of a sedimentary basin. Numerous workers have provided estimates of the relative abundance of the different types of sedimentary rocks; although the numbers vary widely, the general patterns agree:

Mudrocks: 50-80%

Sandstones: 10-30%

Carbonates: ~10%

Conglomerates and breccias: ~1%

All others together: ~1%

1.2: Historical archive

Sedimentary rocks are deposited at the surface of the Earth and provide an important archive of information about the Earth's past. Their composition and texture contain important clues about tectonics, paleoclimate, and processes that were active at the surface of the Earth when they were deposited. The overwhelming majority of the fossil record is contained in sedimentary rocks and thus they record both the history of the Earth and the history of life on Earth.

1.3: Natural resources

1.3.1: Energy Resources

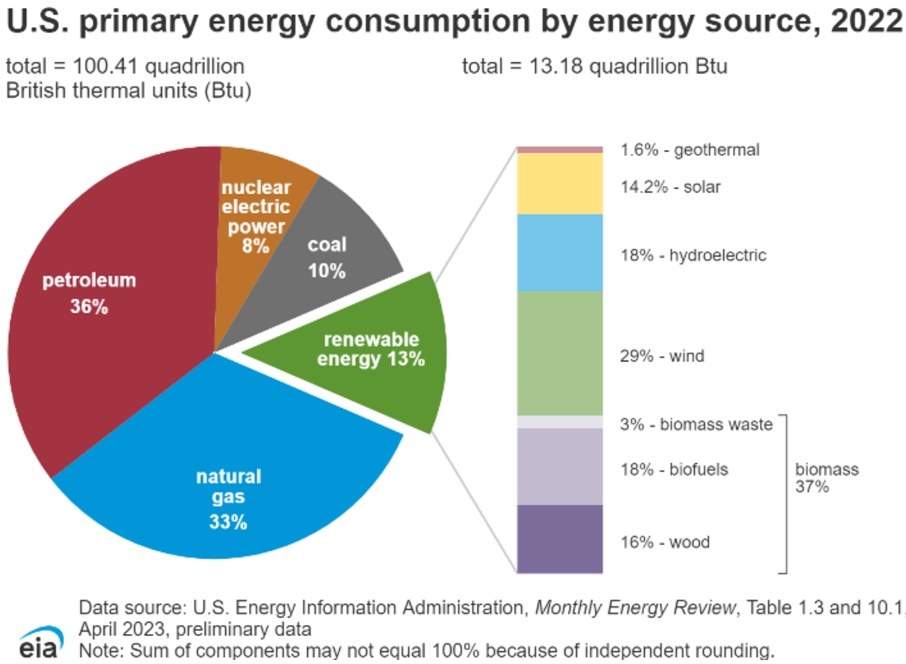


Figure 1.2: Sources of US energy consumption, 2022 (Energy Information Administration via U.S. energy facts explained; public domain)

In terms of the economic significance of sedimentary rocks, most people immediately think of energy resources from fossil fuels including coal, oil, and natural gas. As eloquently described in Richard Alley's book "Earth - The Operator's Manual" these energy resources fueled the industrial revolution and are largely responsible for the convenient, comfortable, and cheap energy intensive world that many of us live in today. Although kicking the carbon habit is a challenging but necessary step to safeguard our future, fossil fuels still represent the majority (~79% in 2022) of the US energy mix and will likely continue to contribute to it for decades to come.

Uranium ore can come from a variety of igneous, metamorphic, and sedimentary sources, but well over half of the world's known sources come from sandstones.

Figure 1.3: A portion of the Signal Hill, CA oilfield circa 1923 (*Library of Congress's Prints and Photographs Division via Wikimedia Commons; public domain)*

1.3.2: Industrial materials

Sediments and sedimentary rocks are important sources of material for the construction of roads and buildings. Some specific examples include:

- Limestone is used to make the binding agent in cement and concrete, building stone and facade, and as a flux in steel making
- Sandstone is used for a building stone and facade
- Gypsum is used for plaster and drywall
- Rock salt is used for road de-icing and a food additive
- Kaolinite and other clays are used for brick and the manufacturing of paper, paint, ceramics, plastics, healthcare products and numerous other applications
- Sand and gravel are used for concrete, filtration, and grading for roads and foundations.
- Pitch is used in asphalt and road construction

Figure 1.4: Limestone quarry near Bellefont, PA (*Dhaluza via Wikimedia Commons; CC BY 3.0)*

1.3.3: Minerals

Sediments and sedimentary rocks are also important hosts for economically important minerals. Banded iron formations are an important source of iron ore, bauxites are an ore for aluminum, and sedimentary phosphorites are an important source of phosphorous.

1.3.4: Water, soil, and geomorphology

Inherently, sedimentary rocks have the potential to be more porous and permeable than crystalline rocks. Consequently, they are very important for groundwater recharge and storage and provide natural filtration as water slowly moves through ore spaces. Given their widespread distribution, the physical characteristics of sediment and sedimentary rocks represent the starting point for soil development; characteristics like mineralogy, grain size, and sorting are important factors that influence the nature of the resulting soils. From a geomorphology perspective, the physical characteristics of sedimentary rocks influences their resistance to erosion which can cause things like karst topography (limestones), susceptibility to mass movement (mudrocks and poorly consolidated materials), and the distribution of ridges, valley, cliffs, etc.

Figure 1.5: Landslides at La Conchita, CA in 1995 and 2005 were caused by unusually heavy rainfall infiltrating poorly consolidated sediments of the Pico and Monterrey Formations (*USGS via Landslide Hazard Program; public domain*).

1.4: Resources and Readings

- Geology of Uranium Deposits by the World Nuclear Association

Chapter thumbnail shows the Chinle Badlands in Grand Staircase-Escalante National Monument, Utah (U.S. Department of the Interior via Wikimedia Commons; public domain).

CHAPTER OVERVIEW

2: Sediment Creation and Transport

Sediment is formed by the weathering of existing rocks and minerals. This sediment can be eroded (transported from one location to another) in a variety of ways by moving fluids or under the direct influence of gravity. Sediment will eventually be deposited and might eventually be transformed into sedimentary rock.

Learning Objectives

- Explain the difference between physical and chemical weathering and provide examples of each.
- Define the most important properties of fluids and sedimentary particles and use this knowledge to make predictions about how these properties will influence fluid motion and sediment transport.
- Describe, identify, and explain the differences between the three types of fluid flow transport
- Describe, identify, and explain the differences between the four types of sediment gravity flows

Chapter thumbnail shows a boulder split by frost wedging (Dominicus Johannes Bergsma via Wikimedia Commons; CC BY-SA 4.0).

2.1: Weathering

2.1.1: Physical Weathering

Physical weathering occurs when existing rocks are mechanically broken into smaller pieces with little or no chemical change. The most important types of physical weathering include freeze-thaw (expansion of cracks via freezing of water), biological activity, changes in volume via wetting and drying (especially important with mudrocks), and stress relief weathering. Locally, physical weathering via thermal expansion and contraction and the growth of salt crystals may be important in deserts and coastal areas, respectively.

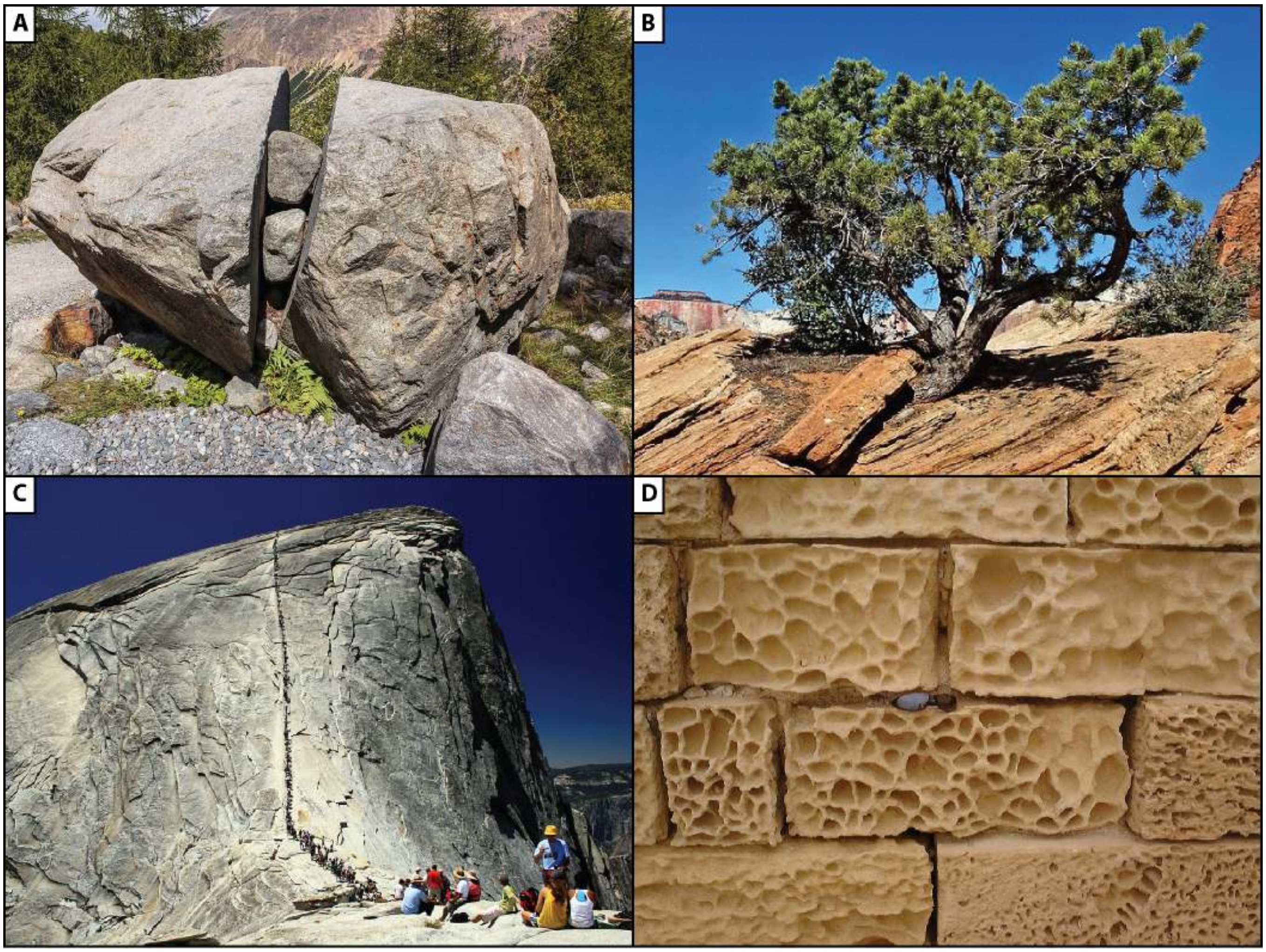


Figure 2.1.1: Examples of physical weathering. A) Boulder split by frost wedging (Dominicus Johannes Bergsma via Wikimedia Commons; CC BY-SA 4.0), B) Tree roots can enhance existing fractures (Don Graham via Wikimedia Commons; CC BY-SA 2.0), C) Half Dome in Yosemite National Park is an exfoliation dome caused by stress relief weathering (HylgeriaK via Wikimedia Commons; CC BY-SA 3.0), D) Salt weathering happens when salt water spray blows on to porous rocks. When the water evaporates, salt crystals grow and break apart connections between grains (Bagamatuta via Wikimedia Commons; public domain).

2.1.2: Chemical Weathering

Chemical weathering is the chemical and/or mineralogical alteration of a rock; it can take place via *in situ* alteration of minerals or by the removal/relocation of dissolved substances.

Solution (aka dissolution) is a type of chemical weathering that takes place when material is dissolved in water, a process that is enhanced if the water is slightly acidic. Common examples of dissolution include halite in water and the dissolution of marble by acid rain.

Hydrolysis is a type of chemical weathering that takes place when acid and water react with a mineral to form clay minerals, silica, and to release ions into solution (commonly K, Na, and Ca). The most common example is the weathering of feldspar to form clay, a process that results in large amount of fine-grained particles and dissolved material.

Oxidation occurs when compounds lose or share electrons with oxygen, the transformation from pyrite to hematite is a common and important example.

Figure 2.1.2: Types of chemical weathering. A) 800 years of solution weathering have made this marble tombstone almost illegible (Palauenc05 via Wikimedia Commons; CC BY-SA 4.0), B) The light colored layers within this bed of coal are bentonites, which form when feldspar-rich volcanic ash undergoes hydrolysis and transforms into clay (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0), C) Acid mine drainage discharging from an exposed shaft in an abandoned coal mine. Oxidation of pyrite (FeS_2) causes the precipitation of orange ferric hydroxide ($FeOH_3$) seen in the foreground (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0)

The processes of physical and chemical weathering commonly act together to speed the breakdown of rocks and minerals. Fractures formed by physical weathering can dramatically increase the surface area of a rock. The increased surface area exposes more of the rock to chemical weathering, which may result in further physical destruction of the rock. This interrelationship of physical and chemical weathering can be seen when fractured blocks develop rounded corners (spheroidal weathering); the increased surface area to volume ratio near the corners results in preferential weathering. Deeply weathered rocks commonly develop a rounded shape because a sphere has the lowest surface area to volume ratio of any shape.

Figure 2.1.3: Notable patterns and processes caused by weathering. A) Spheroidal weathering happens as weathering preferentially acts on the corners and edges of rocks. The longer these fractured rocks are exposed the more round they become (Joachim Himmeröder via Wikimedia Commons; CC BY-SA 2.0 DE), B) Physical and chemical weathering are acting together to degrade this once details sandstone sculpture (Slick via Wikimedia Commons; CC0 1.0).

2.1.3: Some Generalizations

CLimate, Organisms, Relief, Parent material, and Time are the variables that influence the weathering of rock and the creation of the resulting soils; the acronym CLORPT is commonly used to remember these processes (Jenny, 1941, Factors of Soil Formation). Climate, particularly temperature and precipitation, probably exerts the most profound influence by controlling the type, speed, and nature of chemical weathering. Despite the complexities inherent in the system, generalizations can be made:

- Because rocks and minerals are most stable under conditions closest to those under which they formed, Bowen's Reaction Series provides a general framework for predicting igneous rocks' resistance to chemical weathering. Iron-rich ultramafic rocks formed at high temperatures are particularly prone to chemical weathering; silica-rich felsic rocks and minerals formed at relatively low temperatures tend to be the most resistant.
- Quartz, rutile, anatase, magnetite, and zircon are some of the most common stable minerals formed from melts.
- With the notable exception of chert, minerals formed by precipitation (halite, gypsum, etc.) tend to be easily dissolved and prone to weathering.
- Feldspars are prone to hydrolysis and are rapidly destroyed in humid climates.
- Sandstones are common ridge-formers in humid climates; Carbonates are common ridge-formers in arid and semi-arid climates but are valley-formers in humid climates
- Generally speaking, increasing the temperature and the amount water tends to increase chemical weathering rates.

2.2: Fluid Mechanics

Water, air, ice, and gravity are agents of sediment transport (erosion). Although they have very different physical properties, the first three are all fluids because they flow when shear stress is applied to them. As these materials flow, generally in response to gravity, they can transport sediment. Sediment can also move in direct response to gravity without the aid of a fluid – If you don't believe that, just roll a rock down a hill or push it off the edge of a table! Although the fluid is not needed, its presence can facilitate movement by decreasing friction along existing planes of weakness.

2.2.1: Density and Viscosity

Density and viscosity are the properties of a fluid that most profoundly influence the way in which it flows and its capacity for sediment transport. Density is a measure of mass per unit volume (ex: g/cm^3). Liquids (water) have a much higher density than gasses (air) and most naturally occurring solids (minerals) have a higher density than liquid water. One noteworthy exception is water ice, this solid form of water floats because it has a lower density than the liquid form.

Viscosity is the measure of a substance's internal resistance to flow (commonly measured as kg/(m·s) or m^2/s). Fluids with a high viscosity have a high resistance to flow; those with a low viscosity have a low resistance to flow. Honey has a much higher viscosity than liquid water, which in turn has a much higher viscosity than air. The viscosity of a given material can also change with temperature; syrup stored in a refrigerator has a higher viscosity than syrup that has been warmed.

2.2.2: Laminar vs. Turbulent Flow

Fluid flow occurs in two very different modes (laminar and turbulent flow) depending primarily on the flow velocity, fluid viscosity, and flow depth. The differences in flow conditions are best considered in terms of flow lines, which represent the paths of individual particles and can be visualized by adding dye tracer from a point source within the water column. Laminar flow occurs where flow lines are broadly parallel to one another, curve smoothly around obstructions and do not form eddies behind them, and stay at the same orientation through time. Laminar flow can be created when flow smoothly streams out of a faucet or where a smooth body moves through shallow water. Turbulent flow exists where flow lines cross and become indistinct, form eddies behind obstructions, or show random changes in direction at a given location through time. Turbulent flow can be generated by increasing flow out of a faucet until it becomes chaotic and contains air.

Reynolds Number is a dimensionless number that quantifies the balance between viscous forces (tendency for particles to smoothly shear past one another) and inertial forces (tendency for moving particles to resist changes in velocity and direction). Laminar flow exists where viscous forces dominate and the fluid moves in an organized fashion; turbulent flow exists where inertial forces and chaotic motion dominate. Reynolds Number can be calculated by the following equation:

$$Reynold's\ Number = \frac{\mathrm{VD}\rho}{\mu}$$

where:

V = flow velocity

D = flow depth

ρ = density

μ = viscosity

In nature, the transition between laminar and turbulent flow happens at Reynolds Number values of 500 to 2000. Values of less than 500 are typically laminar and represent organized flow with relatively little capacity for sediment transport. Values above 2000 are typically turbulent and are much more effective at transporting sediment. This equation shows that increasing variables in the numerator (flow velocity, depth, or density) and/or decreasing viscosity would tend to push the flow toward turbulent conditions and increase its potential for sediment transport.

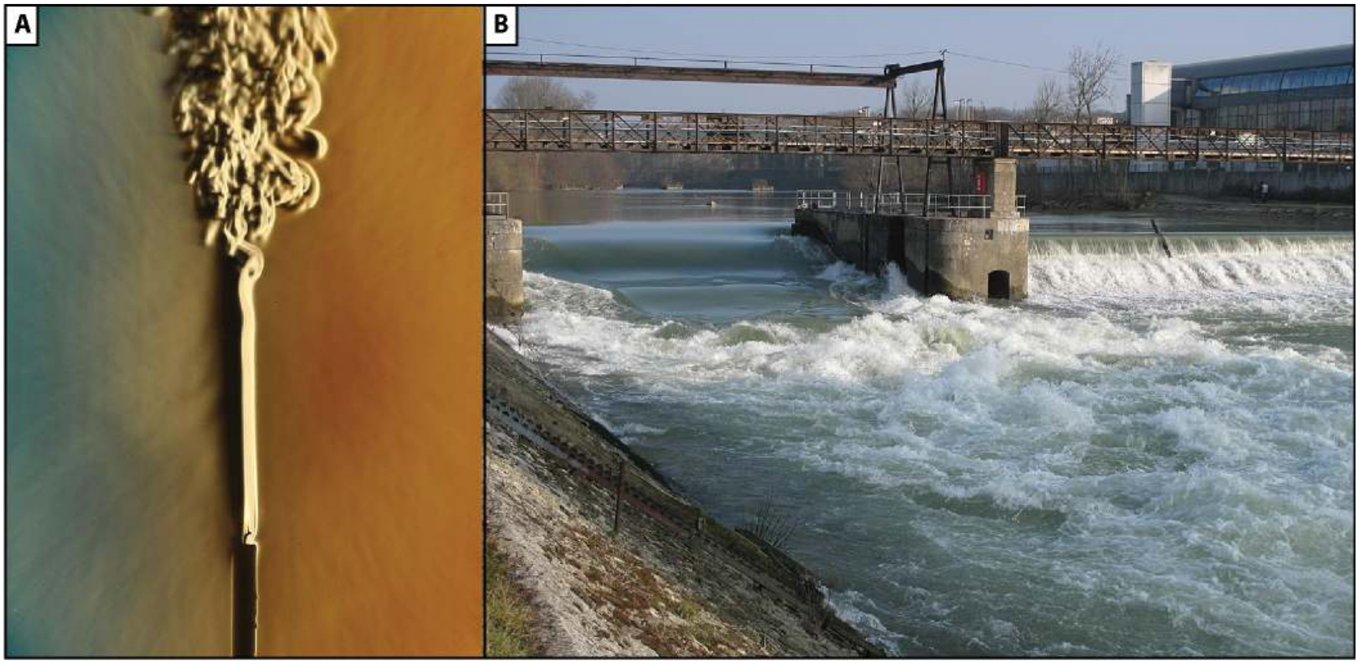


Figure 2.2.1: Examples of laminar and turbulent flow. A) Photograph showing the transition from laminar to turbulent flow in a plume of air rising from a candle. The plume starts of as laminar but transitions to turbulent in the top 1/3 of the frame (Gary Settles via Wikimedia Commons, CC BY-SA 3.0). B) Water flowing over the left side of the dam starts of as laminar and transitions to turbulent flow downstream. Water flowing over the right side of the dam almost immediately transitions to turbulent flow (Tangopaso via Wikimedia Commons; public domain).

Video 2.2.1: Why laminar flow is awesome! Great flume- and field-based examples of laminar and turbulent flow.

2.2.3: Settling Velocity

Stoke’s Law is an equation that shows the relationship between settling velocity, fluid density, particle density, and particle size. This equation is used for small particles (<0.1 mm diameter) where it is assumed that the particles are spherical and that the water moving around them is experiencing laminar flow. Given these assumptions, settling velocity can be calculated as follows:

$$V = \frac{g\left(\frac{\rho s}{\rho f} - 1\right)D^2}{18\mu}$$

where:

V is the settling velocity

ρ_s is the density of the particle

ρ_f is the density of the liquid

g is the gravitational constant

D is the diameter of the sphere

μ is the viscosity of the liquid

Effectively, this equation tells us that, for small particles, the most important factors are the diameter of the particle and the difference in density between the particle and the fluid. It also tells us that increasing the viscosity of the fluid decreases the settling velocity.

Video $2.2.2$: A plain language explanation of Stoke's Law.

For particles larger than very fine sand, turbulent flow around the particle is more likely and the viscosity of the fluid becomes less important. In this case, its best to use a modified version of the equation:

$$V = \frac{g\,RD^2}{18\mu + \sqrt{0.75gRD^3}}$$

where:

V is the settling velocity

R is the specific gravity of the particle

D is the diameter of the particle

μ is the viscosity of the liquid

g is the gravitational constant

This equation is very similar to the one for smaller particles. The big take away from it is that the extra variables in the denominator factor in the drag that comes with larger particles (by increasing the value of the denominator you decrease the settling velocity). Taken together, these equations tell us several basic, but important, things about how particles behave when settling out of the water column:

2.2.3.1: Size

Bigger particles settle out faster than smaller ones (if all other variables are held constant). Some representative settling velocities for different sizes of spherical quartz particles are:

- 0.001 mm (clay) = 0.0001 cm/s
- 0.01 mm (silt) = 0.01 cm/s
- 1 mm (sand) = 15 cm/s
- 10 mm (pebble) = 71 cm/s

The relationship between particle size and settling velocity explains why the largest clasts are found at the bottom of graded beds and why many sedimentary rocks are at least moderately sorted.

2.2.3.2: Density

Particles with a high density settle out faster than those with a low density. The relationship between density and settling velocity explains why, in beach sands, relatively small grains of dark magnetite (density = 5.21 g/cm^3) commonly occur with much larger quartz grains (density = 2.65 g/cm^3).

2.2.3.3: Shape

It is also worth noting that particle shape can profoundly influence settling velocity and that these differences in shape are not factored in to the simplified equations provided above. Many sandstones contain broadly spherical sand grains of a certain size fraction mixed with considerably larger sheet-like grains of mica. The flat shape of the mica grains greatly increases the drag force which slows it's descent. The overall result is that, from a hydrodynamic perspective, large platy grains settle out at a velocity similar to smaller, spherical grains.

2.2.4: Froude Number: Supercritical vs. Subcritical Flow

Perhaps the most abstract fluid flow concept is that of supercritical and subcritical flow – conditions that can be quantified using the Froude Number:

$$F_r = \frac{V}{\sqrt{gL}}$$

Where: F_r is Froude Number (dimensionless)

V is flow velocity

g is the gravitational constant

L is flow depth

This number quantifies the relationship between inertial forces (tendency for moving particles to resist changes in velocity and direction) and gravitational forces (downward attraction toward the Earth). In terms of what this means to flowing water, calculation of Froude Number reveals whether a wave (generated by a disturbance or obstruction) is capable of moving upstream. A Froude number of less than one indicates tranquil, subcritical flow where the wave could move upstream and influence material upstream of it (wave velocity > current velocity). A Froude Number of greater than one indicates supercritical flow where wave energy cannot move upstream and cannot exert an upstream influence (current velocity>wave velocity).

A disturbance in supercritical flow (F_r>1) is analogous to sounding an air horn out the window of a supersonic jet; the sound waves will immediately fall behind and will never get ahead of the jet. A more sedimentological example is to think about throwing a pebble into a river; if the ripples can move upstream the flow is tranquil (F_r<1), if they cannot, the flow is supercritical (F_r>1). Standing waves in a rapidly moving current form when the flow is transitional between these two states (F_r=1).

Video 2.2.3: Flume examples of supercritical, critical, and subcritical flow as well as examples of hydraulic jumps and how obstructions interact with different types of flow.

2.2.5: Sediment Entrainment

Fluid dynamics matter because fluids (air, ice, and water) are important to sediment transport. Whether or not one of these flowing fluids gets things moving is determined by the balance between forces that are preventing movement (gravity, friction, cohesion) and the forces that are contributing to movement (drag, lift forces c/o Bernoulli's Principle).

Video 2.2.4: An overview of Bernoulli's Principle with numerous examples.

If we were engineers, or just really interested, we could sit down and calculate these forces if we knew the properties of the liquid (shear stress, fluid viscosity) and the particle (shape, size, and density). But, we are not going to bother with that because many of them are difficult to predict or impractical to predict in nature.

Instead, we will rely on the Hjulstrom diagram to give us a sense of whether or not things are being eroded or deposited. This plot of grain size versus velocity is useful because it allows us to make some basic predictions based on particle size and flow velocity. Overall, what you can see is that it takes faster flow to move bigger particles … except for the case of mud where sediment cohesion makes particles stick together.

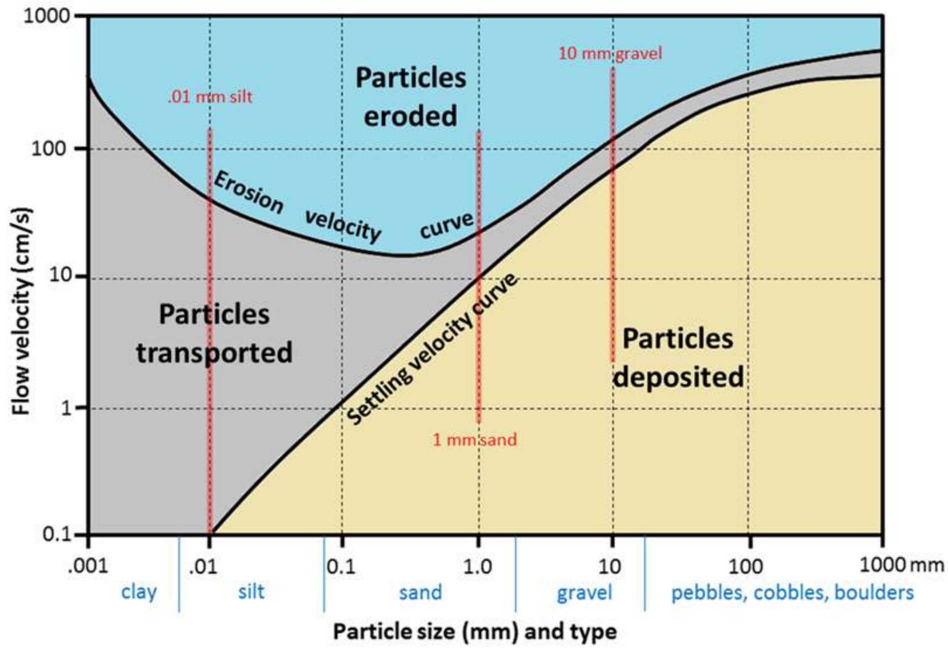


Figure 2.2.2:Hjulstrom diagram the relationship between particle size and the tendency to be eroded, transported, or deposited at different current velocities. Figure from Steven Earle via Physical Geology (2nd); CC BY 4.0)

2.2.6: Readings and Resources

- Froude and Reynolds Number: https://www.geological-digressions.com/fluid-flow-froude-and-reynolds-numbers/
- Settling Velocity: https://www.geological-digressions.com/fluid-flow-stokes-law-and-particle-settling/
- Calculations using Stoke's Law: https://stormwaterbook.safl.umn.edu/sedimentation-practices
- Explanation of the Hjulstrom Diagram: https://www.geological-digressions.com/fluid-flow-shields-and-hjulstrom-diagrams/

2.3: Fluid-Flow Transport

Fluid flow transport happens when sediment is being carried along with fluids that are responding to gravity. If there is no motion in the fluid then there is no movement of sediment. Sediment can get carried along with these moving fluids in the three main ways described below.

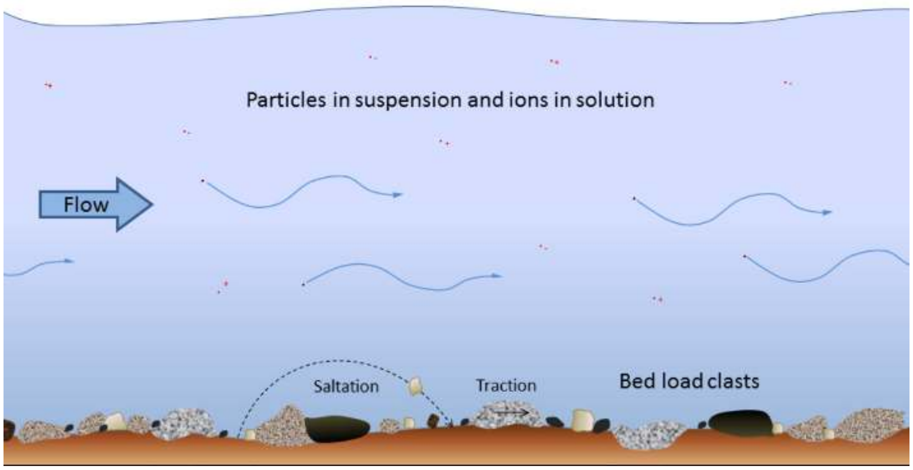


Figure 2.3.1: Sediment transport in a stream via bedload, suspended load, and dissolved load. Image from Steven Earle via Physical Geology (2nd); CC BY 4.0.

2.3.1: Dissolved Load

The products of chemical weathering result in ions that are dissolved in water. These dissolved materials remain aloft in the water even if it is completely still, the only way to get them out is by a change in concentration and/or biological activity. If the water moves, then dissolved materials are transported along with it.

2.3.2: Suspended Load

Suspended load happens when (usually) silt- and clay-sized particles are held aloft in the water column by the turbulence of flow conditions. Unlike dissolved load, suspended load will settle out if the water stops moving.

2.3.3: Bedload

Bedload transport happens when larger particles (usually sand and gravel) move along with the fluid through some combination of rolling, sliding, creeping, impacts, saltation, eddies, etc.

Video 2.3.1: Bedload sediment transport in an Alaskan stream.

Figure 2.3.2: Sediment-laden waters from the Mississippi River dumping into the Gulf of Mexico. Although all three forms of fluid flow transport are happening in this image, suspended load (the brown plumes) is the most striking (Mississippi River Sediment Plume by NASA Earth Observatory; public domain).

2.4: Gravity Mass Movements

It does not always take moving water to get sediment transport to happen. Sometimes materials move under the direct influence of gravity … this is the process of gravity mass movement. Fluid may play a role by reducing internal friction and supporting grains, but its not what's driving movement. We will consider the three main types of gravity mass movements described below.

2.4.1: Rockfall

Rockfalls happen when gravity causes rocks to free fall or roll down steep slopes. Fluid is not needed for this motion to occur, but it's presence can lubricate clasts and thus encourage movement.

Video 2.4.1: Rockfall in the Swiss Alps

Figure 2.4.1: Rockfall closed a portion of Oregon 138E in March, 2017 (Rock blocks OR 138E by Oregon Department of Transportation, CC BY 2.0).

2.4.2: Slides, Glides, and Slumps

Although it might be important to tease them apart if you were doing surficial mapping, we will consider slides, glides, and slumps together. All of them represent movement of material that happens when there is shear deformation along discrete planes with little internal deformation.

Figure 2.4.2: Slide and debris flow at LaConchita, CA (La Conchita by R.L. Schuster (U.S. Geological Survey), public domain).

2.4.3: Sediment Gravity Flows

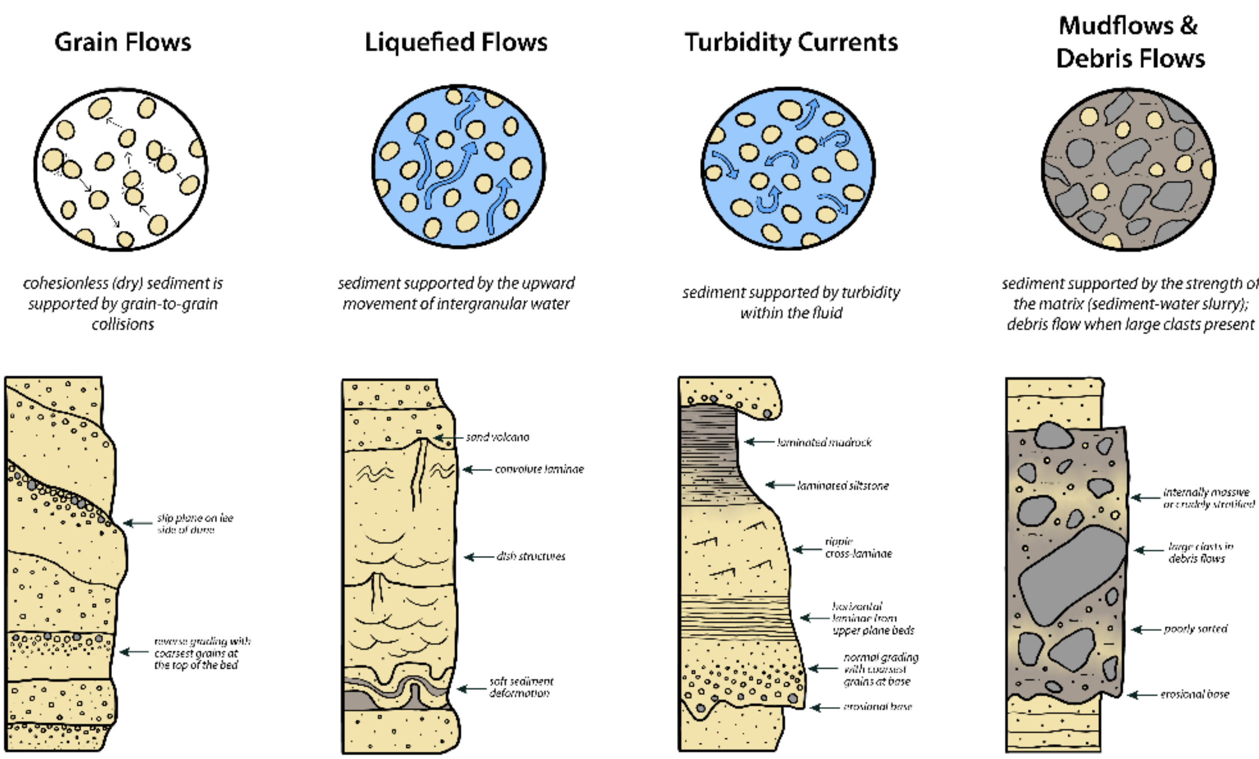


Figure 2.4.3: Types of sediment gravity flows, sediment support mechanisms, and resulting deposits (Page Quinton via Wikimedia Commons; CC BY-SA 4.0). Diagram after Middleton and Hampton (1976).

Sediment gravity flows are the final category of gravity mass movements and because they are often preserved in the sedimentary record we will explore them in more detail and talk about the different types. In all cases, they requires massive internal deformation and separation of the parent mass into discrete particles to form a mixture of particles and the fluid. Generally speaking, this is a very fluid type of motion where grains supported by collision with each other and by the fluid. The main distinguishing factor between the different types is the mechanism by which the particles are supported.

2.4.3.1: Mud flows and debris flows

Mudflows and debris flows (we will treat them as synonymous) happen when a dense muddy matrix (a slurry of sediment and water) is capable of moving and supporting large clasts. They can be subaerial or subaqueous. Although we cannot embed it in this page, there is a great video of an active debris flow available on Youtube via Storyful at https://youtu.be/Fsh5E9m3PrM?si=DSFONNDTEzY305d7.

Debris flow deposits are typically poorly sorted, have a sandy/muddy matrix, show little internal organization, and often have large “floating” clasts.

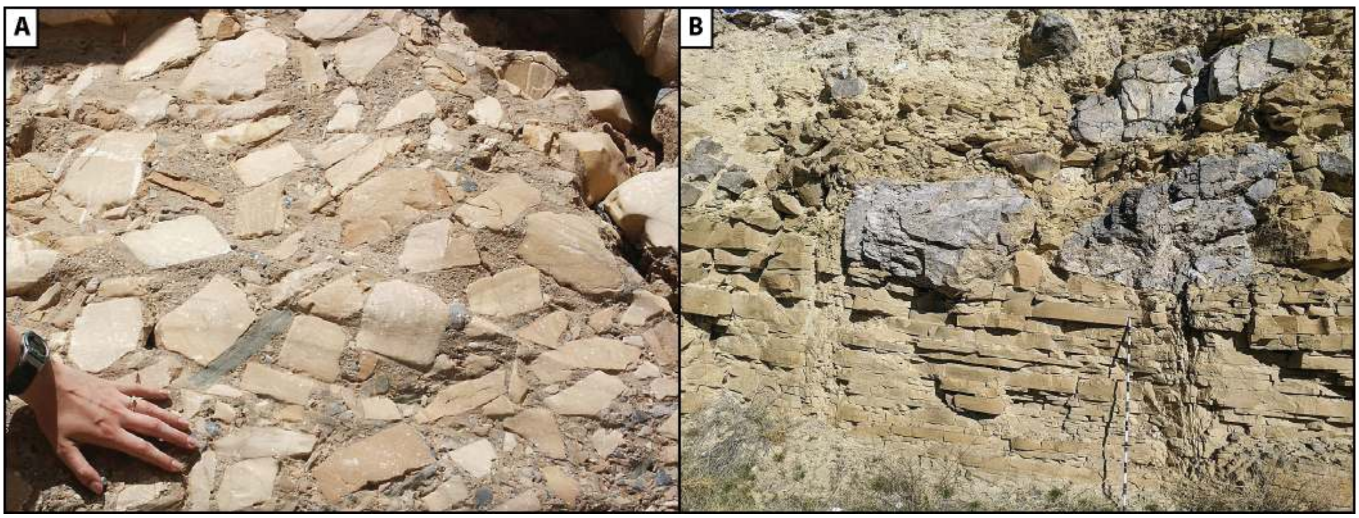
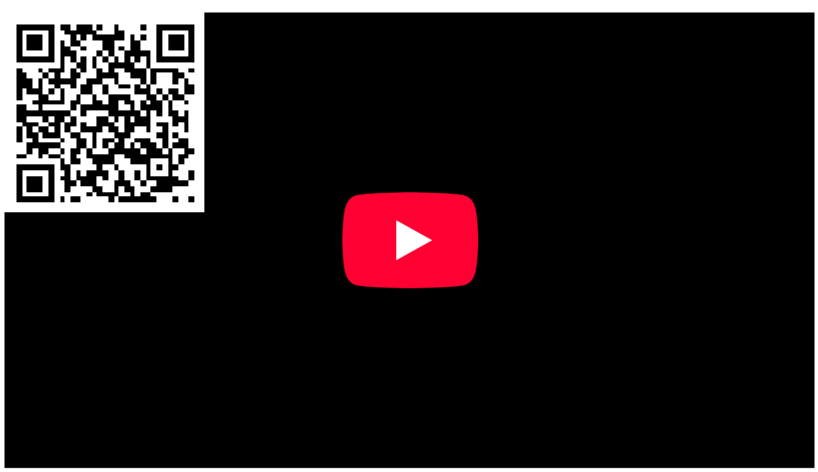


Figure 2.4.4: Debris flow deposits. A) Breccia formed in an alluvial fan with floating angular clasts encased within a fine-grained muddy matrix (Michael C. Rygel via Wikimedia Commons, CC BY-SA 3.0). B) Deep water debris flow deposits with blocks of carbonate floating within siltstone matrix (Michael C. Rygel via Wikimedia Commons, CC BY-SA 3.0).

2.4.3.2: Grain flows

Grain flows happen when cohesionless (dry) sediment is supported in the air by grain-to-grain collisions. Grain flow deposits are generally thin (a few cm thick), lobe-shaped in 3D, and are typically sandy with massive or inverse grading.



Video 2.4.3: Grainflow in dry desert sand.

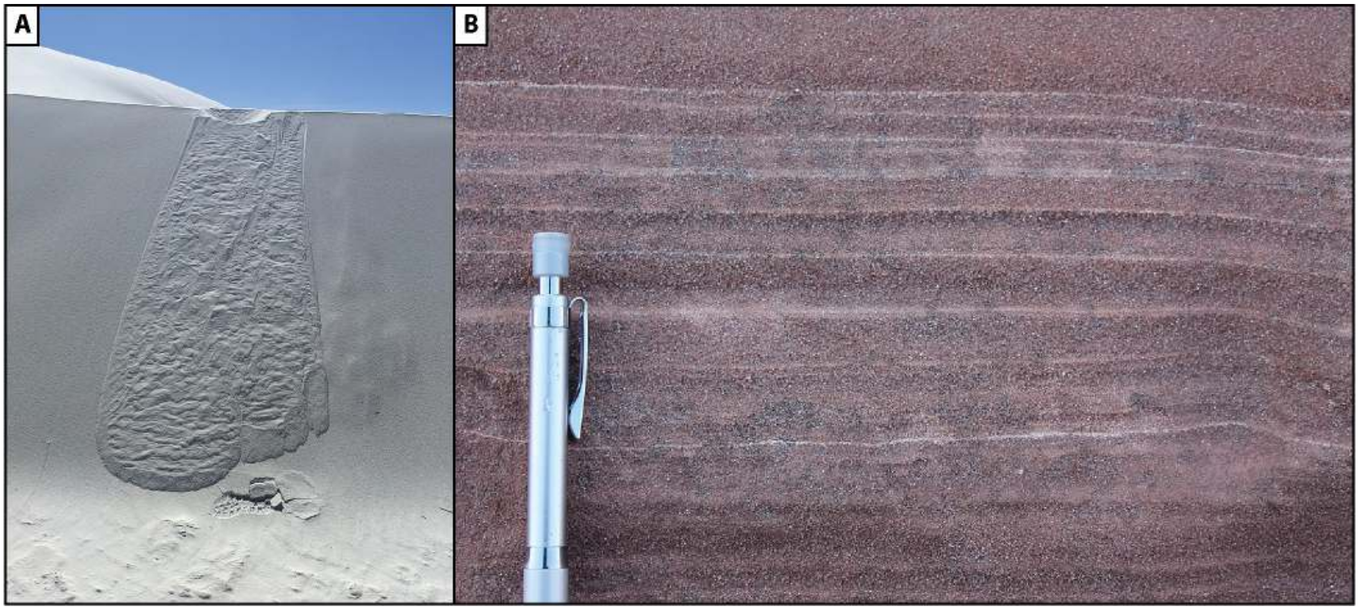


Figure 2.4.5: A) Lobe-shaped grainflow deposits on the lee side of an eolian dune, White Sands National Park, NM (Michael C. Rygel via Wikimedia Commons, CC BY-SA 3.0). B) Reverse grading in eolian strata of the Jurassic Carmel Formation. Up is to the top of the image; a typical package of reverse graded sediment consists of a light color dusty zone that passes upward into increasing coarse sand (dark; Michael C. Rygel via Wikimedia Commons, CC BY-SA 4.0).

2.4.3.3: Liquefied Flows

Disturbances like earthquakes or crashing storm waves can cause saturated sediments to compact in the subsurface. When this happens, the upward movement of intergranular water can temporarily supports sediment grains to create liquefied flows. The resulting deposits typically have massive soft-sediment deformation (flame structures, load structures, convolute laminae) and possibly sand volcanoes at the surface.

Video 2.4.4: Liquidized/fluidized flows interacting with large particles and layered sediment.

Video 2.4.5: Eruption of sand volcanoes after an earthquake in New Zealand.

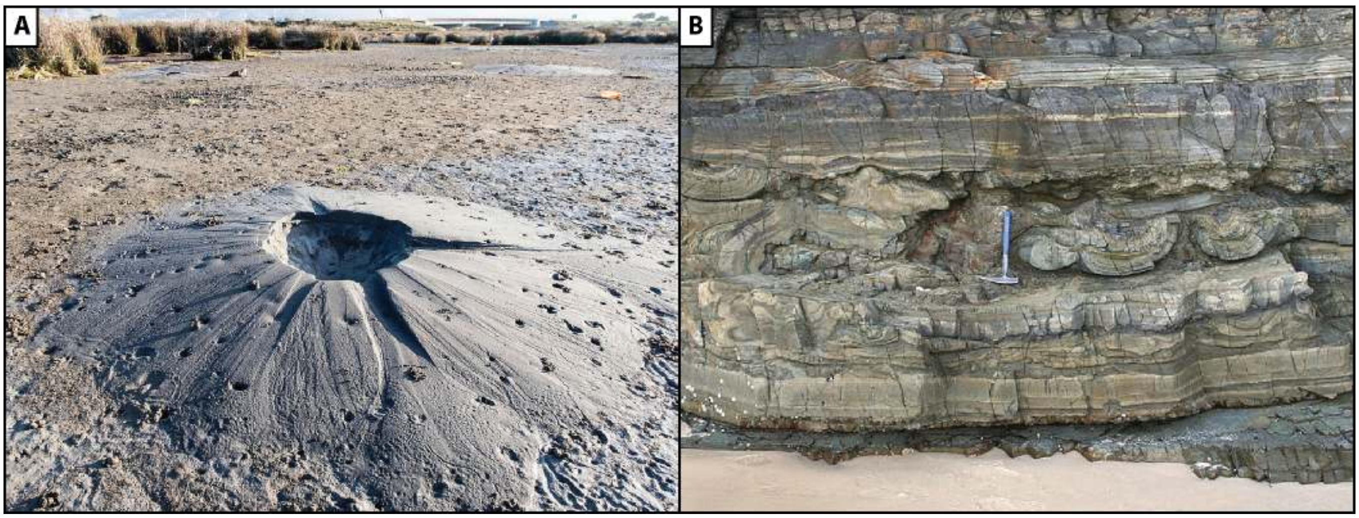


Figure 2.4.6: Liquefied flows. A) Modern sand volcano caused by soil liquefaction following an earthquake in New Zealand (Martin Luffvia Wikimedia Commons, CC BY-SA 3.0). B) Soft sediment deformation caused by dewatering in the Permian Booti Booti Sandstone (Michael C. Rygel via Wikimedia Commons, CC BY-SA 3.0).

2.4.3.4: Turbidity currents

Turbidity currents happen where sediment is held aloft by turbidity within the fluid. The result is that you have a relatively dense fluid with sediment and water that flows in response to gravity, typically downslope through less dense clean water. The resulting deposits show normal grading (coarsest grains at the bottom) and an overall fining upwards trend. These coarse-fine successions are commonly referred to as Bouma sequences. Remember from earlier in this chapter that Stoke's Law tells us that the biggest particles settle out first.

Video 2.4.6: Turbidity current in a flume.

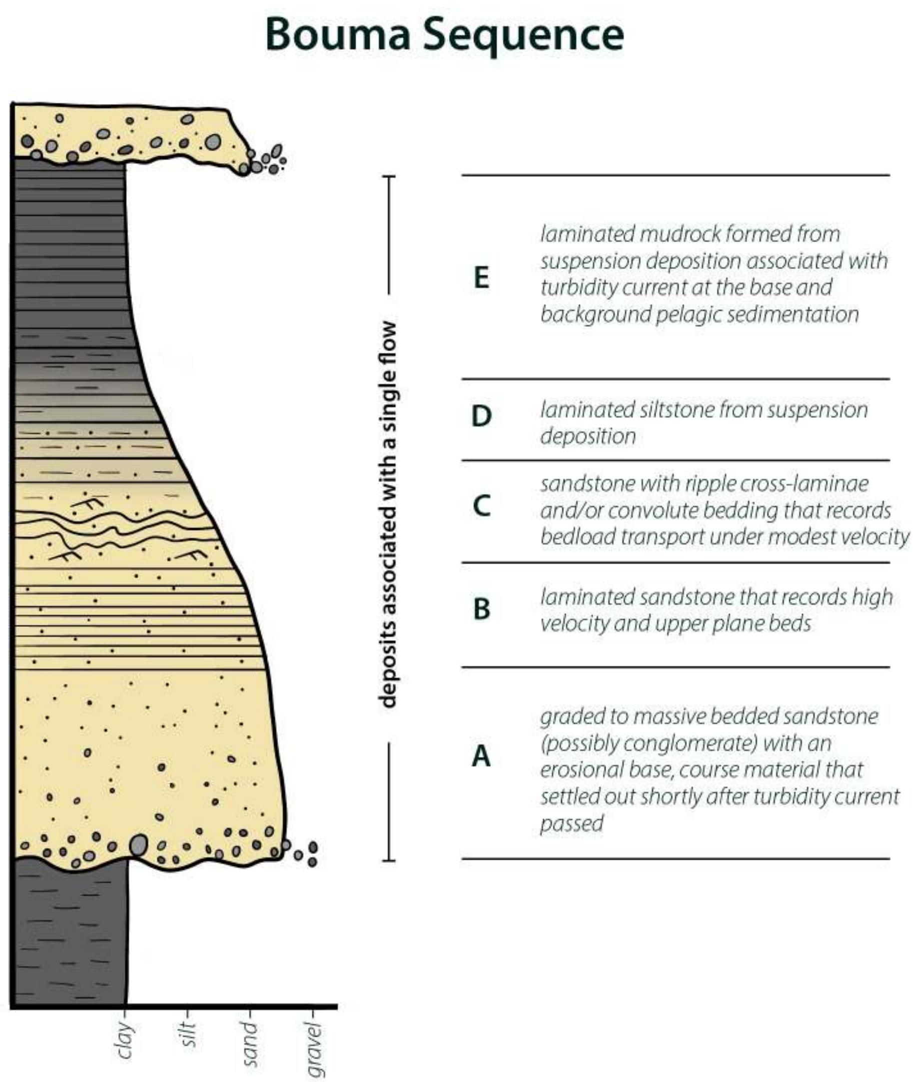


Figure 2.4.7: Diagram showing a complete Bouma sequence formed by the passage of a turbidity current (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

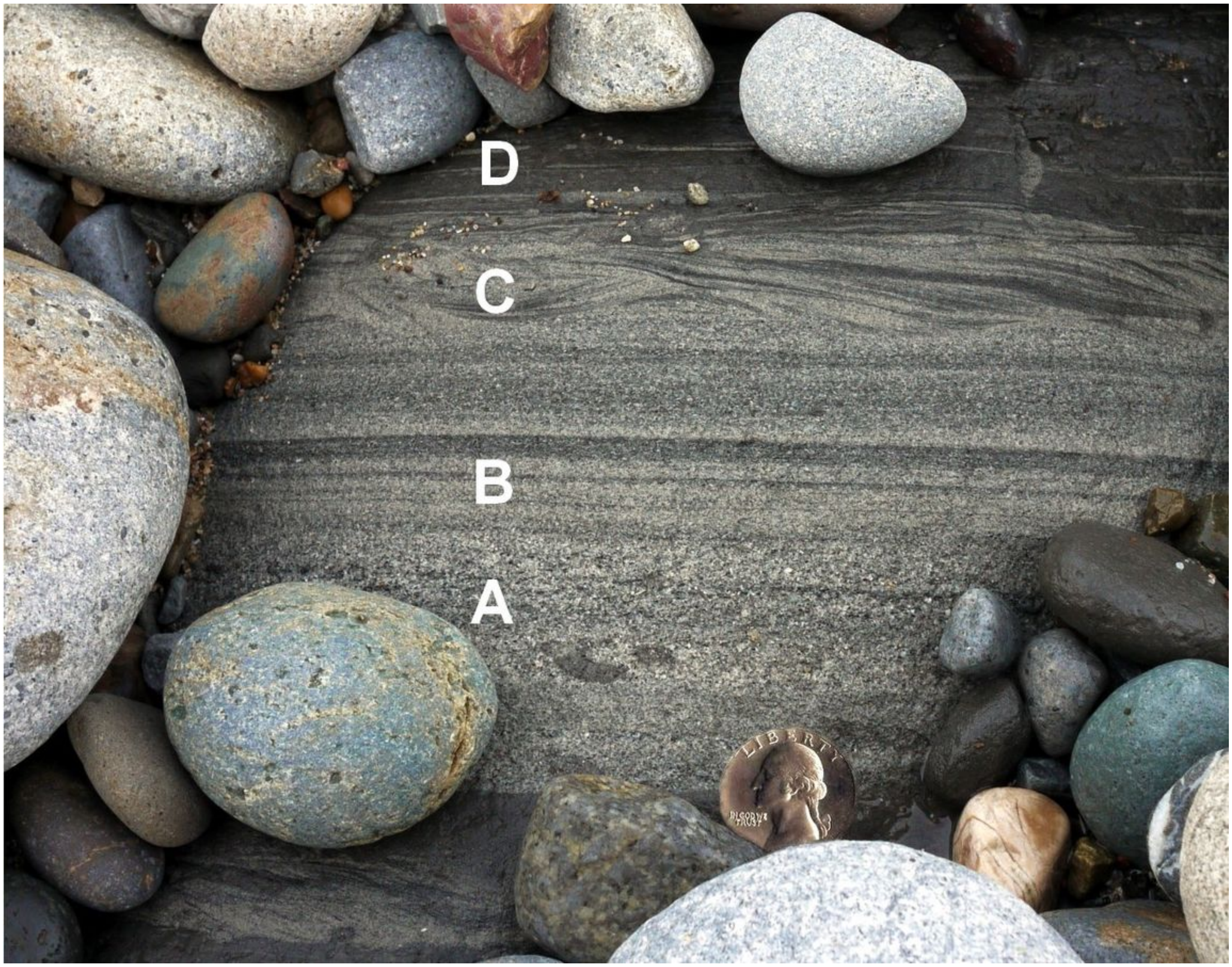


Figure 2.4.8: Outcrop example a Bouma sequence formed by a turbidity current (Mikesclark via Wikimedia Commons, CC BY-SA 3.0).

2.4.4: Readings and Resources

- Dasgupta, P. and Manna, P., 2011, Geometrical mechanism of inverse grading in grain-flow deposits: An experimental revelation, Earth-Science Reviews, v. 104, no. 1-3, p. 186-198. https://www.sciencedirect.com/science/article/pii/S0012825210001376#f0025
- Middleton, G.V. and Hampton, M.A., 1976, Subaqueous sediment transport and deposition by sediment gravity flows in Stanley, D.H. and Swift, D.J.P. (eds.), Marine sediment transport and environmental management, Wiley, New York, p. 197-218.
- Geological Digressions, Sedimentary structures: Turbidites - https://www.geological-digressions.com/2020/04/09/sedimentary-structures-turbidites/
- USGS, Landslide Types and Processes - https://pubs.usgs.gov/fs/2004/3072/fs-2004-3072.html(opens in new window)

CHAPTER OVERVIEW

3: Describing Sediment and Sedimentary Rocks

The size, shape, and composition of particles are the most important characteristics for the description and classification of sediment and sedimentary rock. Specific characteristics include particle size and size distribution, grain shape and the nature of any edges, and the maturity of a sample as recorded by the composition of framework grains and interstitial material.

Learning Objectives

- Classify sediment and sedimentary rocks using an Udden-Wentworth grain size chart
- Convert between millimeters and phi units
- Describe grains using sorting and rounding terminology
- Discuss how composition and/or grain size is used to classify sedimentary rocks and provide examples
- Describe specimens using the concepts of compositional and textural maturity and make general interpretations about what these properties record in terms of environments and processes.

Topic hierarchy

Cover diagram shows a stylized portrayal of an immature sandstone (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

3.1: Grain Size

3.1.1: Udden-Wentworth Grain Size Scale

The Udden-Wentworth grain size scale (or some derivative thereof) is the most common one used by geologists and forms the basis for subdividing clastic sedimentary rocks based on clast size. We tend to make the most basic subdivisions based on size because the maximum clast size is a function of the amount of energy in the system. We've used this classification scheme as the basis for Table 4.1.1, which combines grain size, phi units, and more detailed naming of clastic sediments and clastic sedimentary rocks.

Diameter (mm)	Phi units	Particle Size		Sedimentary Rocks	Grain/Clast Composition: Quartz >90%	Feldspar > 10%, Feldspar > Lithics	Lithics[1] > 10%, Lithics > Feldspar	Settling velocity (cm/s)	Entrainment velocity (cm/s)
256	-8	Boulder	Gravel (rounded) or Rubble (angular)	Conglomerate (rounded) or Breccia (angular)	X	X	______[2] Boulder Conglomerate (Breccia)		500
64	-6	Cobble			Quartz Cobble Conglomerate (Breccia)	X	______[2] Cobble Conglomerate (Breccia)	163	175
4	-2	Pebble			Quartz Pebble Conglomerate (Breccia)	X	______[2] Pebble Conglomerate (Breccia)	43	60
2	-1	Granule			Quartz Granule Conglomerate (Breccia)	Feldspathic Granule Conglomerate (Breccia)	______[2] Granule Conglomerate (Breccia)	26	50
1	0	Very coarse	Sand	Sandstone	Quartz Arenite (<10% matrix) or Quartz Wacke[4] (>10% matrix)	Feldspathic Arenite[3] (<10% matrix) or Feldspathic Wacke[3,4] (>10% matrix)	Lithic Arenite (<10% matrix) or Lithic Wacke[4] (>10% matrix)	15	38
1/2	1	Coarse						7	28
1/4	2	Medium						3	23
1/8	3	Fine						1	20
1/16	4	Very fine						0.3	Cohesion may become important and entrainment velocities may increase
1/256	8	Silt	Mud	Mudrock (massive) or Shale (fissile)	Siltstone (silt > clay) Mudstone (silt ≈ mud) Claystone (silt < clay)			0.08	
		Clay							

X = clasts of this size and composition are unlikely.
[1]Lithic fragments could be composed of chert, limestone, igneous, or metamorphic rock fragments (and many others).
[2]Lithologic descriptor of most abundant clast type should be used in the blank space (ex: basalt cobble conglomerate); use the term polymictic if no one lithology is dominant (ex: polymictic pebble conglomerate).
[3]Feldspar-rich sandstones are informally termed arkose.
[4]Sandstones with abundant muddy matric are informally called graywackes.

Table 3.1.1: Grain size classification from Wentworth (1922) and phi scale from Krumbein (1934). Settling velocities from http://www.filtration-and-separation...g/settling.htm and entrainment (erosion) velocities from http://en.Wikipedia.org/wiki/File:We...Size-Chart.pdf; both assume spherical particles of quartz.

3.1.2: Phi Units

In hydrogeology, we commonly describe sediment size in terms of phi units (Φ), where the conversion to real world units is:

Diameter (mm) = $1/2^n$
where n = phi (Φ) value

They key thing to remember about this is that the bigger the phi value the smaller the diameter of the particle.

3.1.3: Sorting

Sorting is a measure of the uniformity of grain size in a specimen. A well sorted sample will have relatively uniform grain size whereas a poorly-sorted sample has a wide range of grain sizes. Geologists generally apply positive-sounding terms to uniform grain size because those samples have the most porosity and thus have the best potential for fluid flow and storage.You can estimate sorting visually (Figure 3.1.1) or measure it quantitatively using sieve analysis and plotting up the data on a cumulative distribution plot and histogram (Figure 3.1.2). Once plotted, you can use the equations and techniques in Figure 3.1.2 to determine mean (average), median (middle value when data is sorted smallest to largest), and mode (value that occurs most frequently) values for grain size, as well as calculate a numeric value for sorting and skewness (a measure of the symmetry of grain size distribution).

Sorting terminology (describes grain size distribution)	
Descriptive terminology	***Sorting value (Φ)***
Very well sorted	<0.35Φ
Well sorted	0.35Φ to 0.50Φ
Moderately well sorted	0.50Φ to 0.71Φ
Moderately sorted	0.71Φ to 1.00Φ
Poorly sorted	1.00Φ to 2.00Φ
Very poorly sorted	2.00Φ to 4.00Φ
Extremely poorly sorted	>4.00Φ

Skewness terminology (describes symmetry around ϕ50)		
Mathematical terminology	***Values***	***What it means***
Strongly positive skewed	+1.00 to +0.30	Very skewed toward coarse (-ϕ) size fraction
Positive skewed	+0.30 to +0.10	Slightly skewed toward coarse (-ϕ) size fraction
Near symmetrical	+0.10 to -0.10	Nearly symmetrical distribution
Negative skewed	-0.10 to -0.30	Slightly skewed toward fine (+ϕ) size fraction
Strongly negtive skewed	-0.30 to -1.00	Very skewed toward fine (+ϕ) size fraction

Table 3.1.1: Sorting and skewness terminology and values from Folk (1966).

Engineering Terminology

Engineers and geologists live in opposite terminology worlds when it comes to grain size distribution. They use the term "grading" to describe the distribution of grain size. For them, compaction is what is important and a well-graded specimen has a wide range of grain sizes (can be densely compacted) and a poorly graded specimen has uniform grain size and does not compact as well.

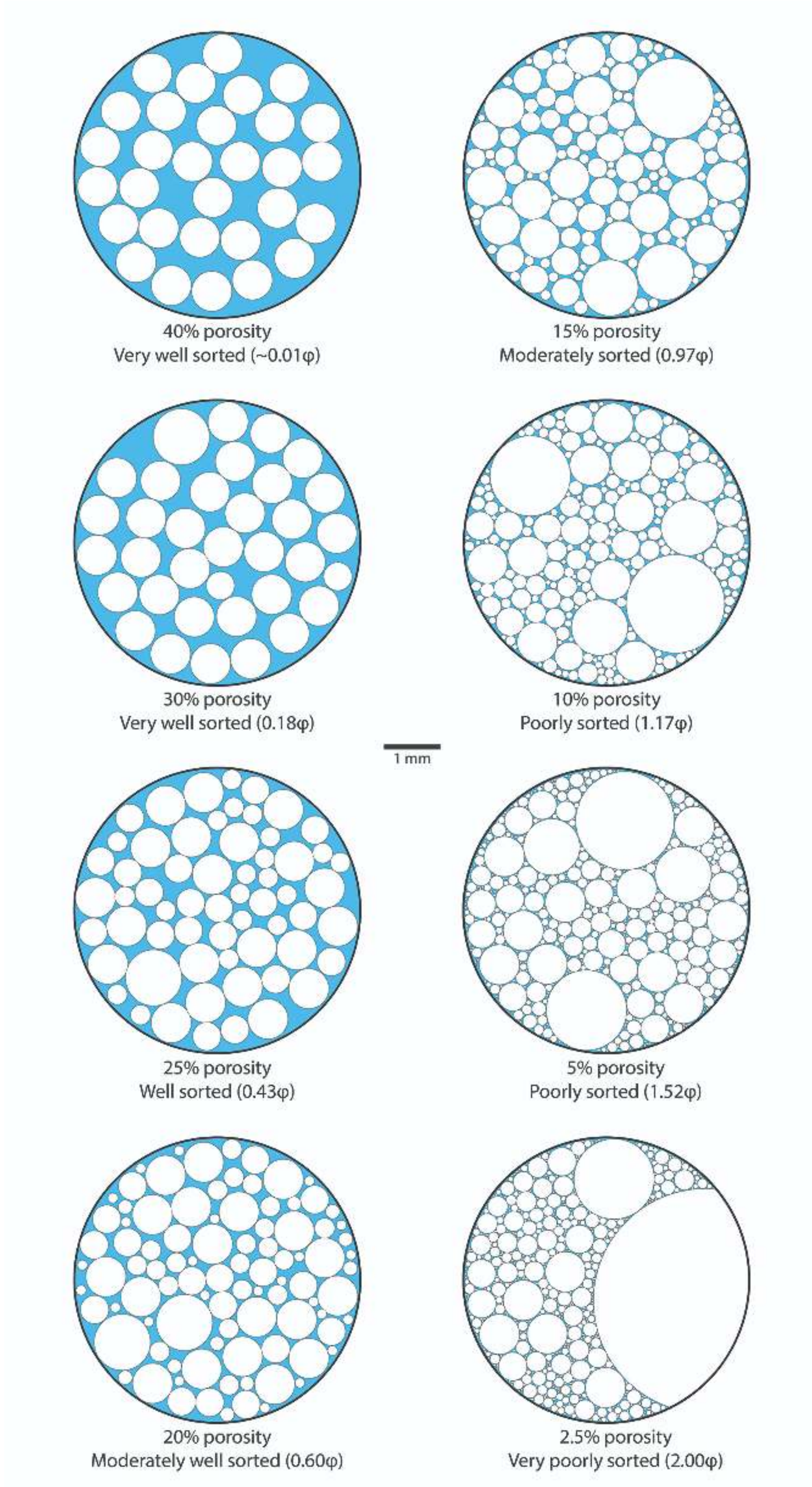
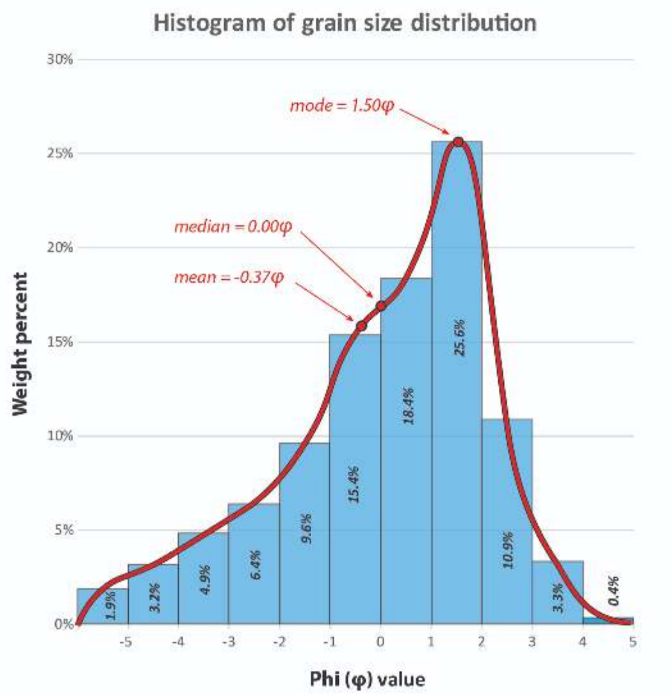


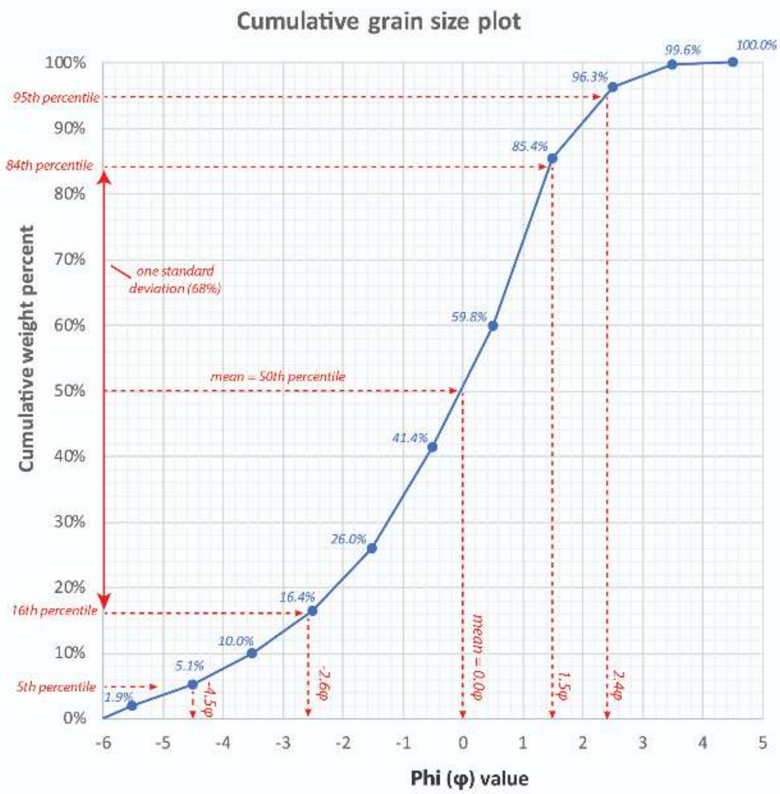
Figure 3.1.1: Illustration of spherical sand grains (white) and pore spaces (blue) showing changes in sorting and porosity (Michael C. Rygel via Wikimedia Commons; *CC BY-SA 4.0*).

Grain Size Data Table

Screen size in φ units	*Average Diameter (φ) retained*	*Diameter (mm)*	*Weight (g)*	*Individual weight (%)*	*Cumulative weight (%)*
-5		32	15	1.9%	1.9%
-4	-4.5	16	25.0	3.2%	5.1%
-3	-3.5	8	38.0	4.9%	10.0%
-2	-2.5	4	50.0	6.4%	16.4%
-1	-1.5	2	75.0	9.6%	26.0%
0	-0.5	1	120.0	15.4%	41.4%
1	0.5	0.5	143.5	18.4%	59.8%
2	1.5	0.25	200.0	25.6%	85.4%
3	2.5	0.125	85.0	10.9%	96.3%
4	3.5	0.0625	26.0	3.3%	99.6%
5	4.5	0.01	3.0	0.4%	100.0%
		End weight (g)	780.5	100.0%	
		Start weight (g)	795		
		Difference (%)	-1.8%		

Characteristic	Value	Source
Median grain size (φ units)	0.00	*50th percentile on cumulative chart (φ50)*
Median grain size (mm)	1.00	*converted from above*
Mean grain size (φ units)	-0.37	*calculated from mean equation*
Median grain size (mm)	0.77	*converted from above*
Mode (ϕ units)	1.50	*midpoint of most abundant class on histogram*
Mode (mm)	2.83	*converted from above*
Sorting (φ) *inclusive graphic standard deviation*	2.07	*calculated from inclusive graphic standard deviation*
Skewness (symmetry) *inclusive graphic skewness*	-0.29	*calculated from inclusive graphic skewness equation*





Equations

$$\text{Mean} = \phi 50$$

$$\text{Median} = \frac{\phi 16 + \phi 50 + \phi 84}{3}$$

$$\text{Sorting} = \frac{\phi 84 - \phi 16}{4} + \frac{\phi 95 - \phi 5}{6.6}$$

$$\text{Skewness} = \frac{\phi 84 + \phi 16 - (2 * \phi 50)}{2(\phi 84 - \phi 16)} + \frac{\phi 95 + \phi 5 - (2 * \phi 50)}{2(\phi 95 - \phi 5)}$$

Figure 3.1.2: Example grain size data shown in tabular, histogram, and cumulative distribution curve form. Median, mean, mode, sorting, and skewness can be calculated from this data using the equations on the bottom right of the diagram. Techniques and equations from Folk (1966).

3.1.4: Readings and Resources

- Folk, R. L., 1966, A review of grain-size parameters. Sedimentology, v. 6, no. 2, p. 73-93.
- Krumbein, W. C. M., 1938, Size frequency distributions of sediments and the normal phi curve. Journal of Sedimentary Research, v. 8, no. 3, p. 84-90.
- Wentworth, C. K., 1922, A scale of grade and class terms for clastic sediments. The Journal of geology, v. 30, no. 5, p. 377-392

3.2: Particle Morphology

3.2.1: Form

The overall form or shape of a particle can be important because it influences the hydrodynamic properties of the particle and because it might be influenced by mineralogy which in turn can impact how it responds to diagenesis and/or fluid flow. We commonly use the term equant to describe a particle that has broadly comparable dimensions in all directions (broadly spherical) and platy to describe a particle that is elongate in one direction (think elongate rectangle). Keep in mind that this refers to the overall shape not to the number or nature of the corners.

3.2.2: Roundness and Angularity

Roundness and angularity are end members that described the nature of the edges/corners of a particle. Keep in mind that this is different from the description of grain morphology.

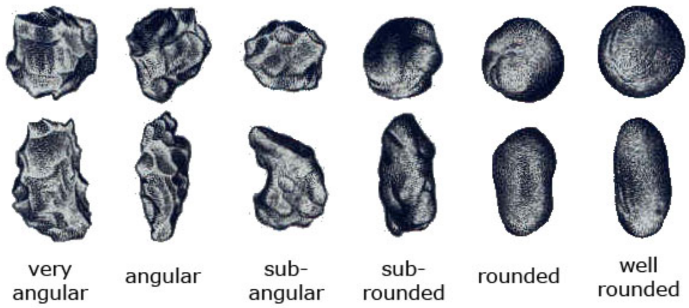


Figure 3.2.1: Visual reference for describing rounding of equant (above) and platy (below) particles (Oceanography 101 (Miracosta) via source content; Public Domain).

3.2.3: Surface texture

One can also characterize the surficial texture of a particle. Common descriptors include things like bright, frosted, polished, pitted, scratched, etc.

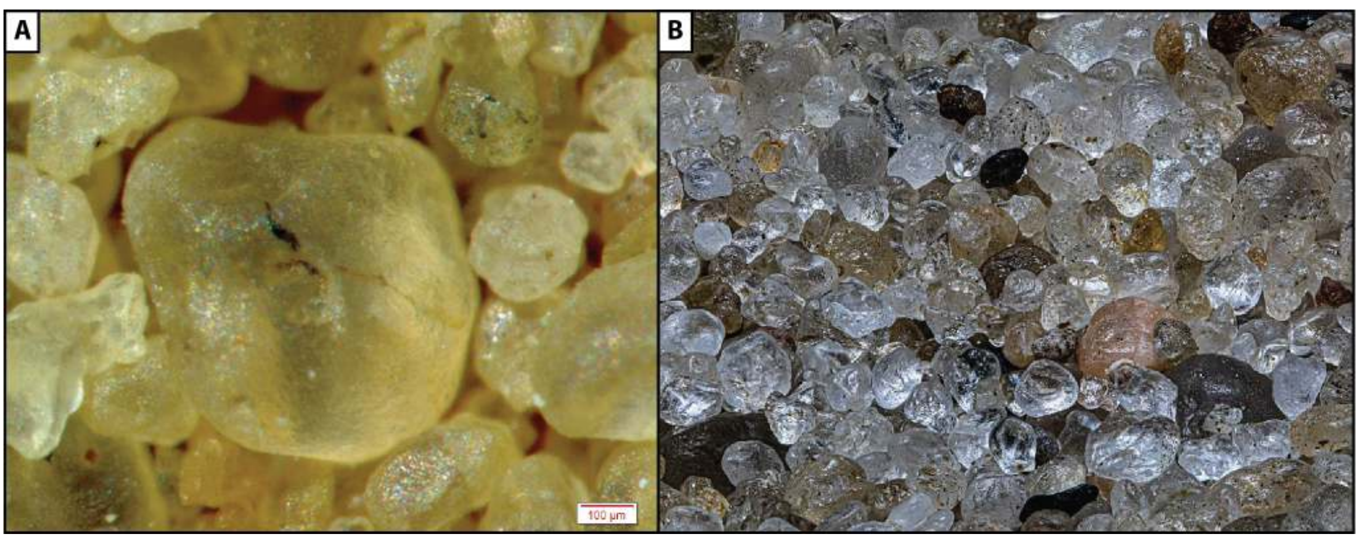


Figure 3.2.2: Surface textures of sand grains. A) Frosted and pitted sand grains typical of aeolian environments (Mark Wilson via Wikimedia Commons; CC0 1.0) and B) bright, glassy sand grains more typical of fluvial transport (Krzem Anonim via Wikimedia Commons; CC BY-SA 4.0).

3.2.4: Packing

Packing refers to the arrangement of grains/particles in a rock. Although its not something that we can meaningfully describe in our work, its worth being aware of because of the potential to influence porosity and compaction of sediment. Have a look at the diagram below which shows various arrangements of spherical grains of the same size. By moving the grains around, you can go from a theoretical maximum of 47.64% porosity to a minimum of 25.95%.

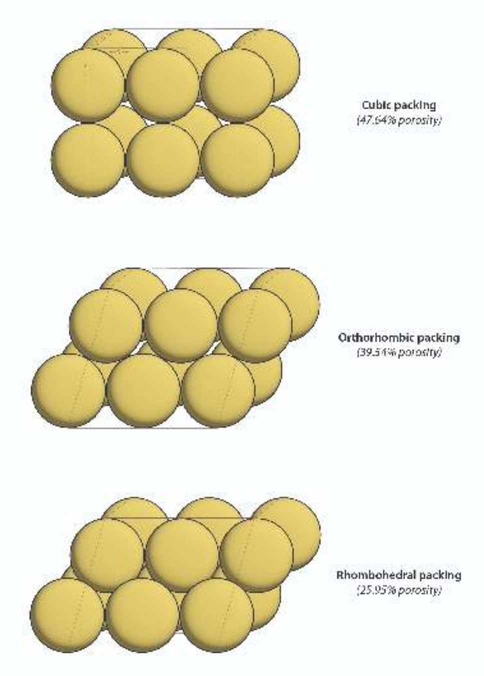


Figure 3.2.3: Three arrangements of spherical, identically size particles showing how the packing of grains influences porosity. Diagram is after Graton and Fraser (1935) and from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0.

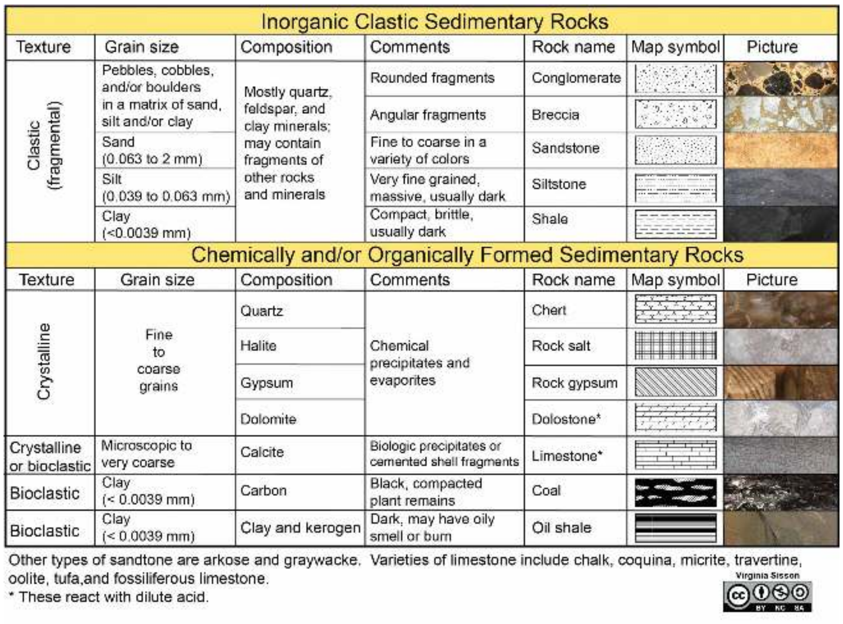
3.2.5: Readings and Resources

- Graton, L.C. and Fraser, H.J, 1935, Systematic Packing of Spheres: With Particular Relation to Porosity and Permeability, Journal of Geology, v. 43, no. 8, pt.1, p. 785-909.

3.3: Composition

We will explore composition in more detail when we discuss different types of sedimentary rocks. But for now, its worth mentioning that:

1. No matter the scheme you use, the major subdivisions of sedimentary rocks are based on their composition. This includes sub-types like carbonates (made of carbonate minerals), siliclastic sedimentary rocks (made of transported particles that came to rest when flow ceased), biochemical sedimentary rocks (made of things derived from- or associated with living organisms), etc.



Inorganic Clastic Sedimentary Rocks						
Texture	Grain size	Composition	Comments	Rock name	Map symbol	Picture
Clastic (fragmental)	Pebbles, cobbles, and/or boulders in a matrix of sand, silt and/or clay	Mostly quartz, feldspar, and clay minerals; may contain fragments of other rocks and minerals	Rounded fragments	Conglomerate		
			Angular fragments	Breccia		
	Sand (0.063 to 2 mm)		Fine to coarse in a variety of colors	Sandstone		
	Silt (0.039 to 0.063 mm)		Very fine grained, massive, usually dark	Siltstone		
	Clay (<0.0039 mm)		Compact, brittle, usually dark	Shale		
Chemically and/or Organically Formed Sedimentary Rocks						
Texture	Grain size	Composition	Comments	Rock name	Map symbol	Picture
Crystalline	Fine to coarse grains	Quartz	Chemical precipitates and evaporites	Chert		
		Halite		Rock salt		
		Gypsum		Rock gypsum		
		Dolomite		Dolostone*		
Crystalline or bioclastic	Microscopic to very coarse	Calcite	Biologic precipitates or cemented shell fragments	Limestone*		
Bioclastic	Clay (< 0.0039 mm)	Carbon	Black, compacted plant remains	Coal		
Bioclastic	Clay (< 0.0039 mm)	Clay and kerogen	Dark, may have oily smell or burn	Oil shale		

Other types of sandtone are arkose and graywacke. Varieties of limestone include chalk, coquina, micrite, travertine, oolite, tufa,and fossiliferous limestone.
* These react with dilute acid.

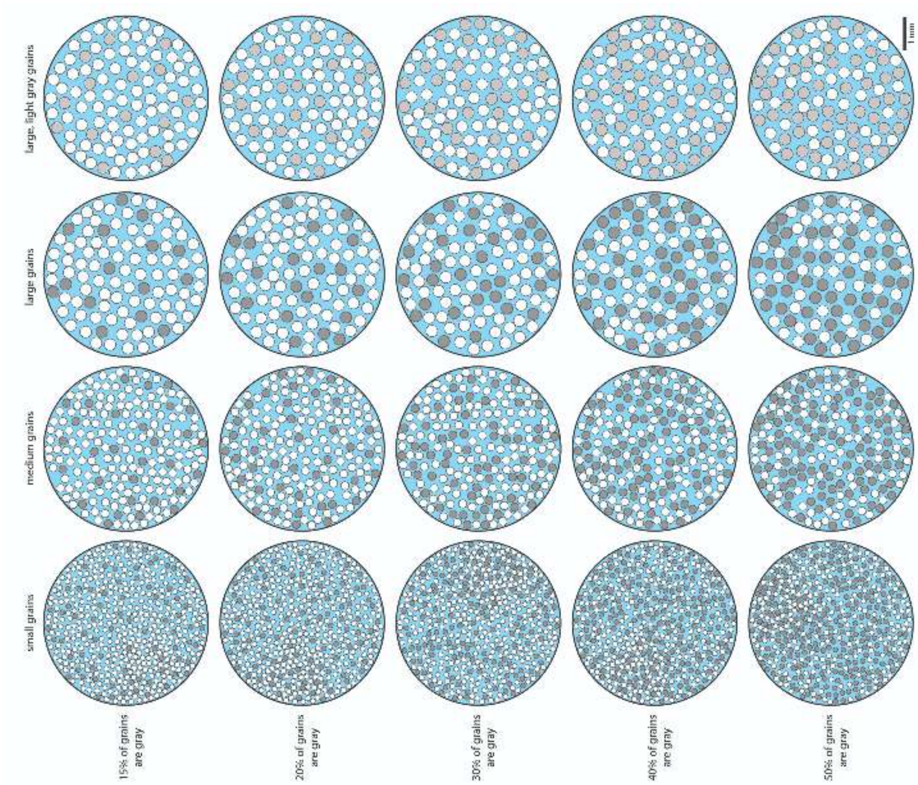
Virginia Sisson

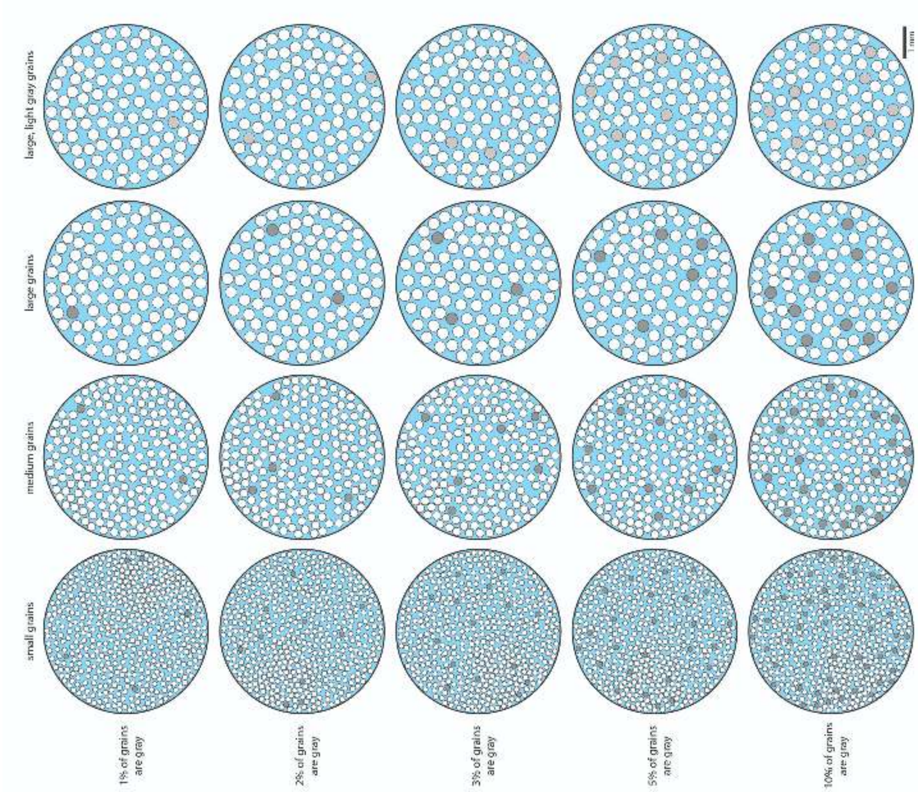
Figure 3.3.1: The primary subtypes of sedimentary rocks are defined based on their grain size and their composition (Virginia Sisson via Pressbooks; CC BY-NC-SA 4.0)

2. The composition of sedimentary rocks influences everything from chemical reactions with subsurface fluids to nutrient values in soil.
3. Composition is influenced by a variety of factors including the type of source rock as well as the amount/type/duration of weathering the sediment experienced, and the cumulative amount of energy that the sediment was exposed to during transport.
4. Geologists tend to classify sandstones based on the relative abundance of quartz, feldspar, and lithic (rock) fragments and keep an eye out for important accessory minerals like mica, magnetite, clay minerals, etc. Although the petrographic microscope is the ideal tool for looking at the mineralogy of sandstones, you can make some very educated guesses using 10x magnification via a hand lens or binocular microscope. Point counting hundreds of grains can provide valuable quantitative data, but quick visual estimates of abundance can also be of value (especially with practice).

Next page →

Figure 3.3.2: Visual guides for estimating grain abundance (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).


large, light gray grains
large grains
medium grains
small grains
15% of grains are gray
20% of grains are gray
30% of grains are gray
40% of grains are gray
50% of grains are gray
1 mm


large, light gray grains
large grains
medium grains
small grains
1% of grains are gray
2% of grains are gray
3% of grains are gray
5% of grains are gray
10% of grains are gray
1 mm

3.4: Maturity

Together, many of the parameters described in the preceding sections can be combined to talk about the textural and compositional maturity of a sample. Textural maturity refers to the nature of interstitial materials (especially the abundance of mud) as well as the sorting and rounding of the framework grains. Compositional maturity refers to the relative abundance of stable framework grains like quartz to unstable grains such as feldspars, micas, and mafic minerals. Maturity is traditionally considered to be a function of cumulative input of energy into the system, transport distance from source area, and exposure to chemical and physical weathering. The more mature the sample the more of these things it underwent.

However this is an oversimplification because things like composition of source rocks and diagenetic processes can have an even more pronounced influence on maturity. Despite these complications, some generalizations about environments and processes can be made: environments like beaches tend to have mature or supermature sediments and proximal environments like alluvial fans tend to have immature sediment.

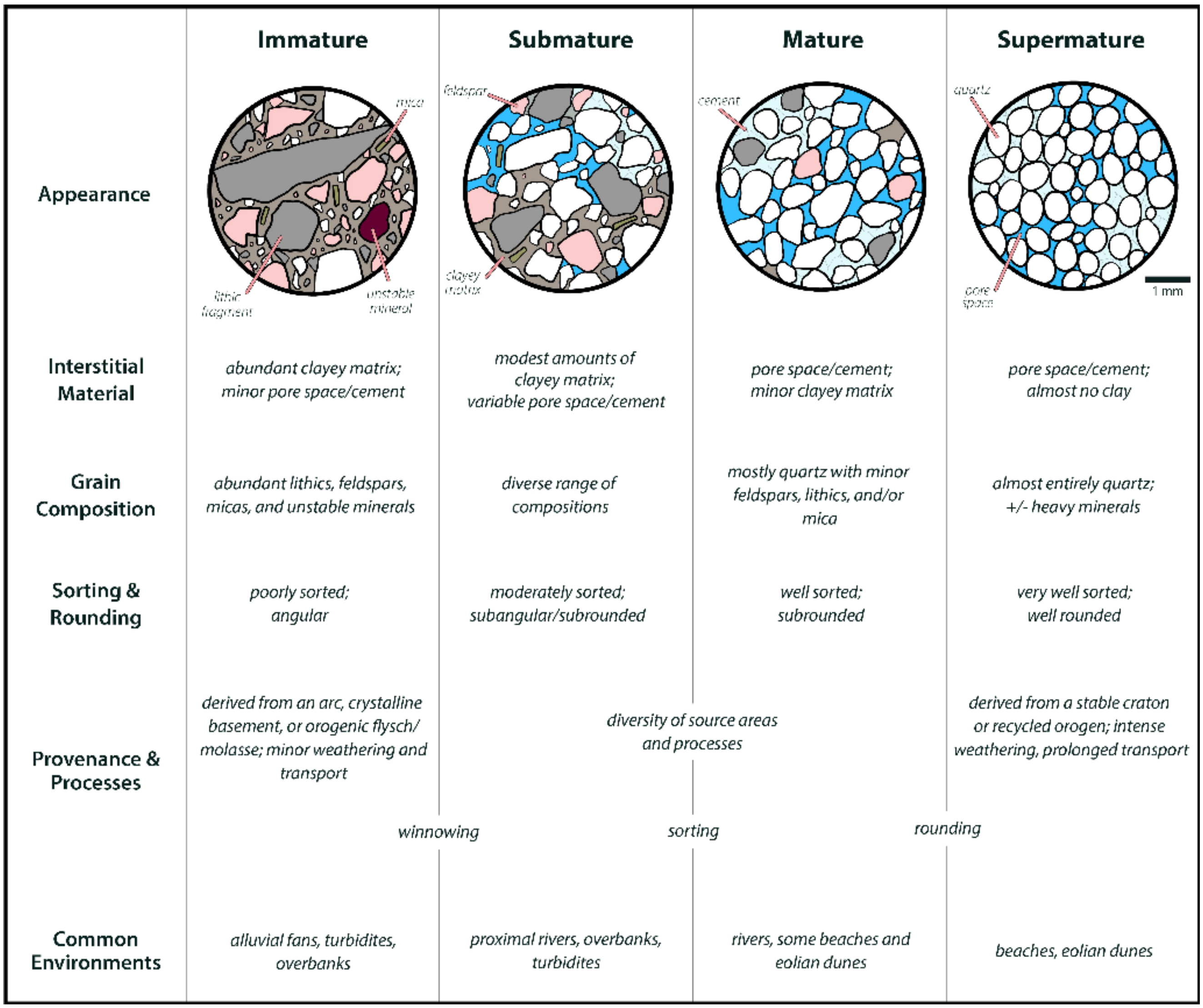


Figure 3.4.1: Textural and compositional maturity of sandstones (Page Quinton via Wikimedia Commons; CC BY-SA 4.0). Figure is after after Folk (1951), Dott (1964), Folk (1968), and Garzanti (2017).

3.4.1: Readings and Resources

- Folk, R.L., 1951, Stages of textural maturity in sedimentary rocks: Journal of Sedimentary Petrology, v. 21, p. 127–130
- Dott, R.H., 1964, Wacke, graywacke and matrix: What approach to immature sandstone classification?, Journal of Sedimentary Petrology, v. 34, p. 625–632
- Folk, R.L., 1968, Petrology of Sedimentary Rocks, Hemphill's, Austin, TX, 170 p.
- Garzanti, E., 2017, The maturity myth In sedimentology and provenance analysis, Journal of Sedimentary Research, v. 87, p. 353–365.

3.5: Color

3.5.1: Munsell Color System

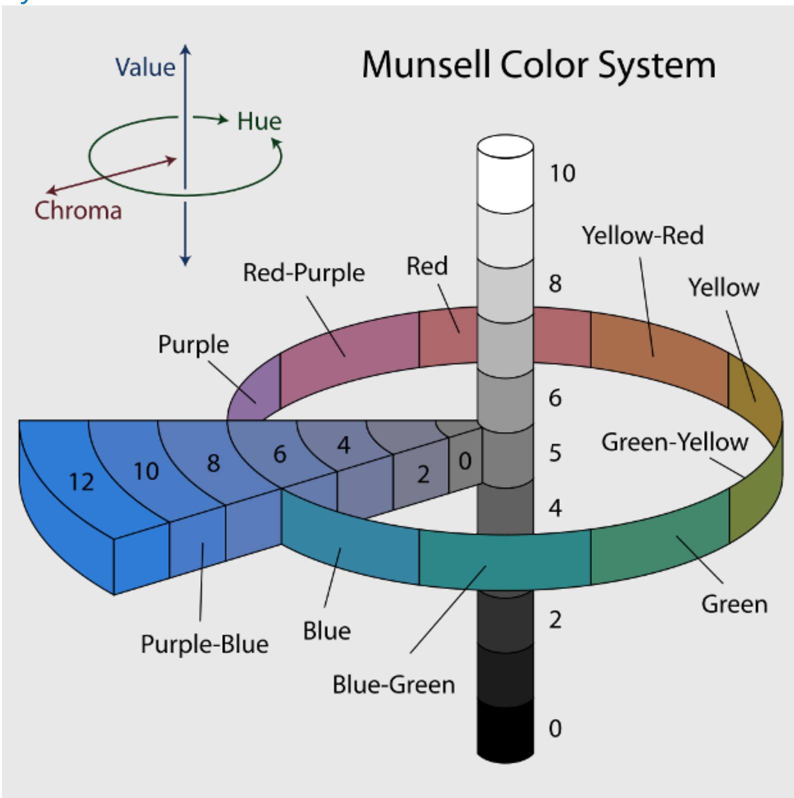


Figure 3.5.1: Diagram showing hue, chroma, and value - the three principal parts of the Munsell Color System (Jacob Rus via Wikimedia Commons, CC BY SA 3.0).

Although first order assessment of color are useful for many purposes, the Munsell Color System is widely used to give specific names to the colors that naturally occur in rocks and soils. Specifically, you can order small, rugged booklets with color plates suitable for use in the field from numerous sources. The Munsell books allow you to give a specific name and code to each observed color (ex: Pale Yellowish Orange, 10YR 8/6).

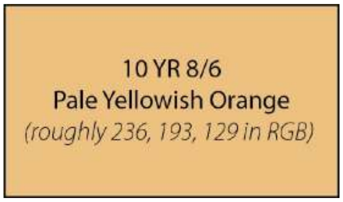


Figure 3.5.2: Rectangle infilled with a RGB fill and the (approximate) corresponding Munsell Color System values.

The coded part of the system refers to three components of color:

1. Hue refers to the actual color family, such as red, yellow, green, blue, etc. Hues are arranged in a circular format around a color wheel, with five principal hues (red, yellow, green, blue, and purple) along with five intermediate hues (e.g., YR) halfway between adjacent principal hues. In the case of pale yellowish orange - 10YR 8/6, the hue is 10 YR (10 parts yellow and one part red).
2. Value represents the lightness or darkness of a color. In the Munsell system, this attribute is depicted vertically, with lighter colors at the top and darker colors at the bottom. The scale ranges from 0 (pure black) to 10 (pure white), with various

gradations in between to show different levels of lightness. A value of 5 represents the true color. Value is given as the number before the slash. In the case of pale yellowish orange - 10YR 8/6, the value is 8.

3. Chroma describes the vividness of the color; it is also referred to as saturation or intensity. High chroma colors are far away from the central axis and are vibrant and intense. Low chroma colors are close to the central axis and tend to be more muted or grayish. Chroma is designated with the number after the slash. In the case of pale yellowish orange - 10YR 8/6, the chroma is 6.

Color can be profoundly influenced by diagenetic processes and is typically something that is only described for mudrocks - we will discuss this more in detail in Chapter 5.

3.5.2: Readings and Resources

- Link to digital Munsell Rock Color chart
- Link to digital Munsell Soil Color Book
- Link to a website that converts from RGB to Munsell colors - https://pteromys.melonisland.net/munsell/

CHAPTER OVERVIEW

4: Sedimentary Structures

Sedimentary structures are important because they tell us about processes that were operating during the deposition of that sediment. Things like flow velocity, depth, and direction can be determined from some sorts of structures. From there, we can start to make some educated guesses about what sorts of depositional environments these processes were taking place in. But, the first step is having the vocabulary to describe what you are seeing.

Learning Objectives

- Identify common sedimentary structures
- Interpret appropriate sedimentary structures to determine paleoflow direction
- Interpret and describe the processes responsible for and recorded by sedimentary structures
- Use appropriate sedimentary structures to determine way up and top/bottom of beds

Topic hierarchy

Chapter thumbnail shows eolian sand dunes in Death Valley National Park (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

4.1: Stratification

Stratification is an intentionally vague term used to describe layering in a sedimentary rock without implying any specific scale or thickness. Layering can be obvious or subtle ... you need to look for some sort of difference in grain size, mineralogy, cements, color, texture, texture, etc.

4.1.1: Beds and laminae

If you want to start talking about the scale of layering, we can break it down into two broad groups: layers that are less than 1 cm thick are called laminae and layers that are more than 1 cm thick are called beds. If you need more precise terminology, a table of more specific terms is provided below.

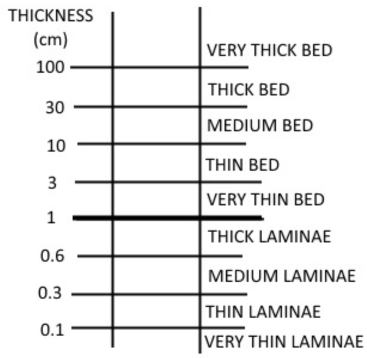


Figure 4.1.1: Descriptive terms for the thickness of beds and laminae (Tenzinsonam995 via Wikimedia Commons; CC BY-SA 4.0).

4.1.2: Internal changes in grain size

4.1.2.1: Normal grading

Within some strata, it is possible that there might be a systematic change in grain size. Normal grading is when a bed shows an upward decrease in grain size. This is usually caused by a decrease in flow velocity or particles falling out of suspension.

4.1.2.2: Reverse (inverse) grading

Reverse (inverse) grading is exactly the opposite … the largest grains are at the top of the bed and the smallest are at the bottom. Reverse grading typically occurs in response to shaking or winnowing in cohesionless sediment (grainflows) which causes the smallest grains to work their way downward through gaps between the larger grains.

4.1.2.3: Massive beds

And although relatively rare, internally massive beds that lack any sort of internal stratification can form when sediment is rapidly dumped or when previously stratified sediments are reworked via bioturbation.

Figure 4.1.2: Grain size distribution with beds. A) Normally graded bed that fines upward (James St. John via Wikimedia Commons; CC BY 2.0). B) Inverse grading that coarsens upward (John Waldron via University of Alberta; CC BY-NC-SA 4.0). C) Crudely stratified to internally massive conglomeratic sandstone (Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

4.1.3: Bed geometry and contacts

Once you've identified bedding, you might find it useful to describe the overall form of the bed by thinking about whether the beds are continuous or not, whether the contacts are planar or wavy, and whether the contacts are are parallel or truncate each other. Another aspect of bedding that you can describe is the nature of contacts between beds. Specifically, we find that phrases like sharp, gradational, interbedded are very useful especially when paired with a semi-quantitative estimate of scale (ex: gradational over 5 cm).

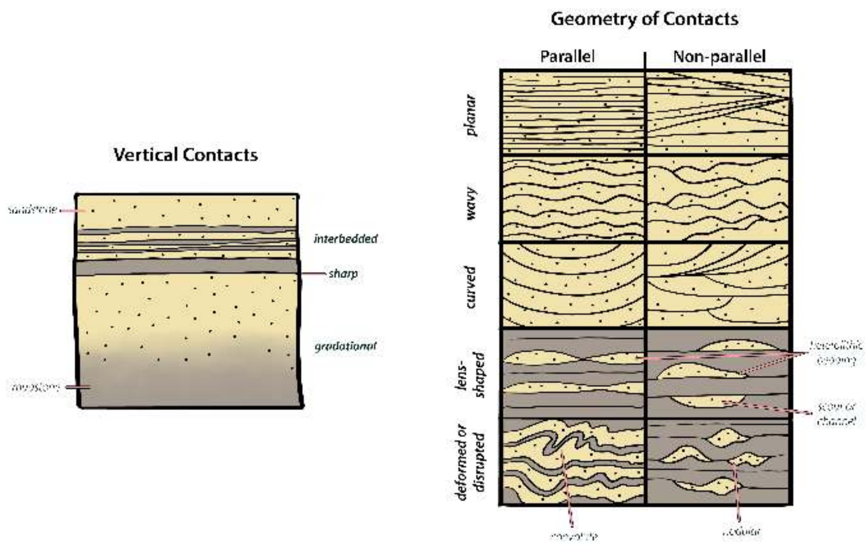


Figure 4.1.3: Types and geometries of bedding contacts after North American Commission on Stratigraphic Nomenclature (2005) and Campbell (1967). Diagram from Page Quinton via Wikimedia Commons; CC BY SA 4.0.

Figure 4.1.4: Different types and scales of bedding contacts are present in this exposure of a redbed paleosol complex. *(Michael C. Rygel via Wikimedia Commons; CC BY SA 4.0).*

4.1.4: Resources and Readings

- North American Commission on Stratigraphic Nomenclature, 2005, North American Stratigraphic Code, AAPG Bulletin, v. 89, no. 11, p. 1547-1591
- Campbell, C.V., 1967, Lamina, laminaset, bed and bedset, Sedimentology, v. 8, p. 7-26.

4.2: Structures Formed by Unidirectional Currents

We moved from sediment transport directly into sedimentary structures because these structures are formed in the sediment during the transport process and tell us about what was going on at the time of deposition. While transport is happening, the sediment may be organized into a three-dimensional bedform. If preserved (or partially preserved), that bedform becomes a sedimentary structure in the rock. Sometimes the two go by the same name and sometimes they don't; when we first introduce the terms we will try to make it clear what we are referring to.

4.2.1: Bedform Stability Diagram

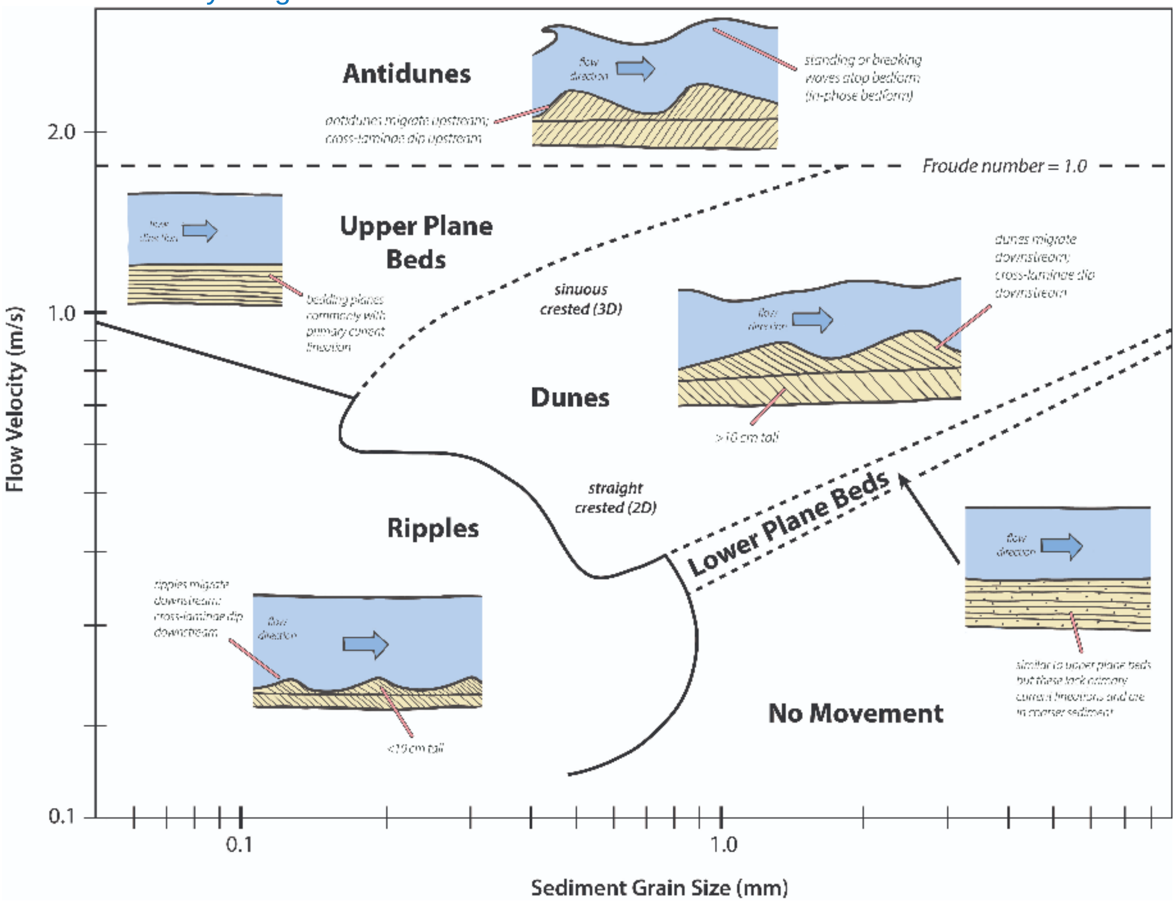


Figure 4.2.1: Bedform stability diagram showing grain size versus flow velocity for unidirectional currents in water at 10 °C and 0.25-0.40 m deep (Page Quinton via Wikimedia Commons; CC BY-SA 4.0). Diagram after Southard and Boguchwal (1990) and Potter and Pettijohn (1977).

Before we launch into a more detailed description about the different types of sedimentary structures that form in response to unidirectional flows, it is worth taking a moment to point out that:

- A systematic relationship exists between bedforms and velocity.
- Flume studies reveal that you can predict what bedform will be present/stable given a certain grain size and flow velocity (provided that you hold all other variables constant)

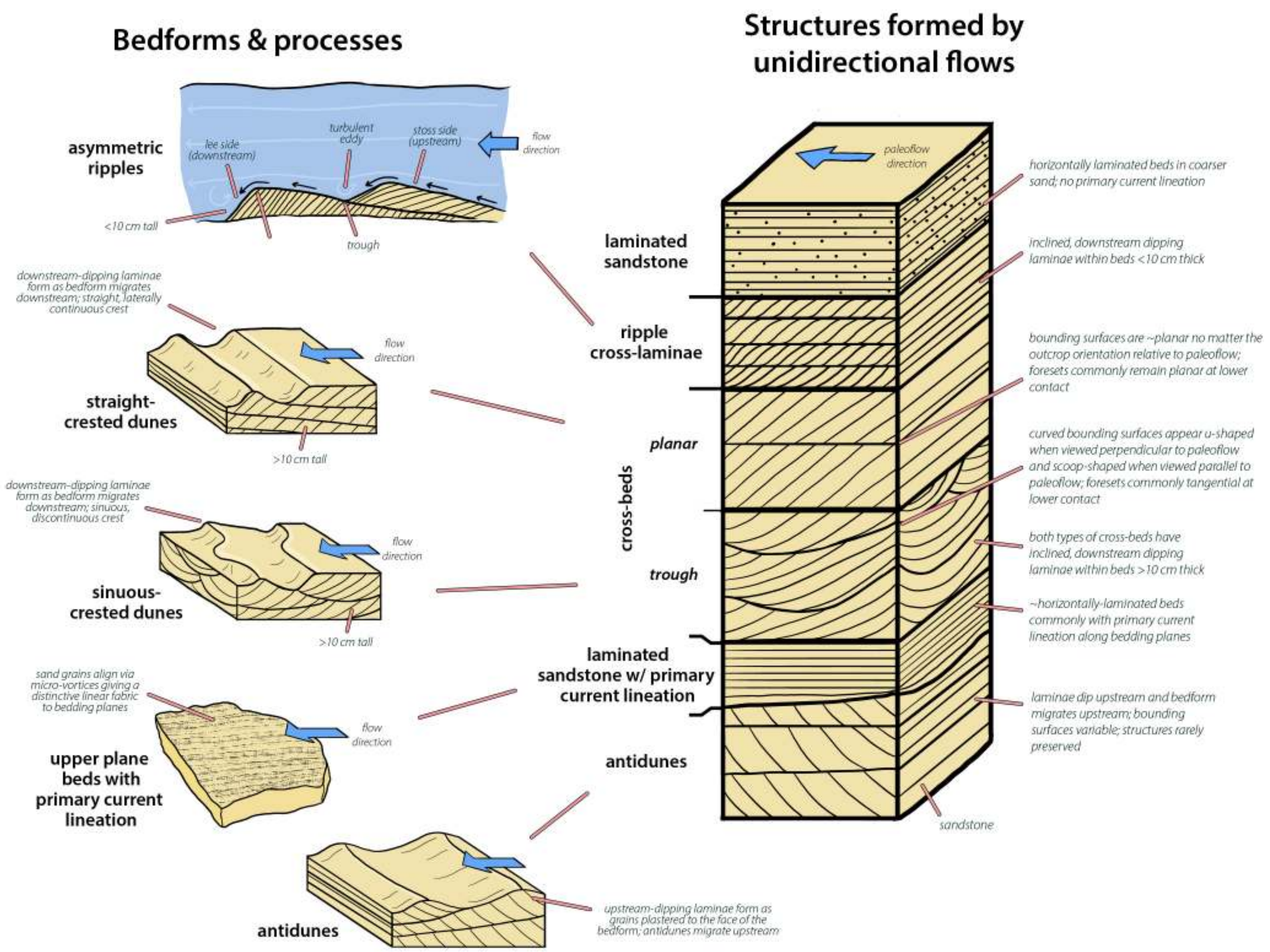


Figure 4.2.2: Bedforms and sedimentary structures formed in response to unidirectional flows. In all cases the flow direction was/is from right to left. (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

4.2.2: Lower Plane Beds

Relatively course-grained sand (above about 0.6 mm diameter) will not form ripples under low flow velocities. Instead, rolling and tumbling of grains will form simple horizontal layers known as lower plane beds (term refers to the bedform and resulting sedimentary structure).

4.2.3: Asymmetric Ripples & Ripple Cross-Lamination

Asymmetric ripples (aka current ripples) are bedforms that are less than 10 cm tall; they have a pronounced asymmetric shape with a gently dipping upstream (stoss) side and a relatively steeply dipping downstream (lee) side. They form as sediment, and the bedform itself) migrates in a downstream direction. Silt- and sand-sized particles move up the stoss side through bedload transport and then accumulate as downstream-dipping layers on the sheltered lee side.

Asymmetric ripples are only rarely preserved in their three-dimensional form, instead the downstream dipping laminae deposited on the stoss side are preserved as ripple cross-laminae. Ripple cross-laminae are just smaller versions of cross-beds and consist of subhorizontal layers with internally dipping laminae that are tangential with the bottom of the bed and truncated at a higher angle at the top of the bed. If the sedimentation rates is very high, climbing ripples may form, these features preserve both the ripple crest and the downstream-dipping laminae.

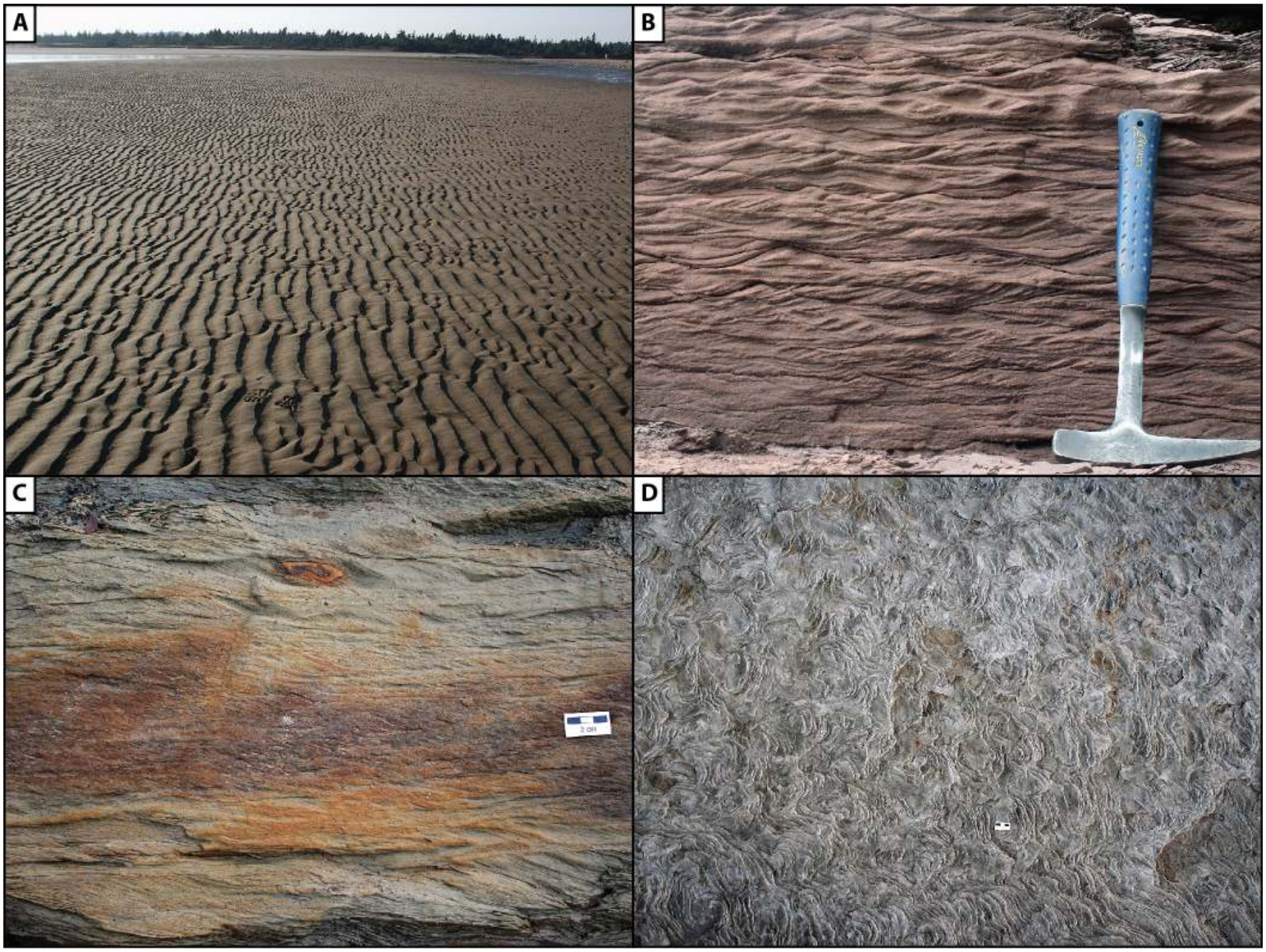


Figure 4.2.3: Ripples and ripple cross-laminae (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0 or CC BY-SA 4.0). A) Modern asymmetric ripples exposed in the intertidal zone. Flow was from left to right; footprint near bottom center of the image for scale. B) Cross-sectional exposure of a ripple cross-laminated sandstone; Although not fully preserved in three dimensions, these structures are strongly aggradational and could be considered climbing ripples. C) Cross-section view of a ripple cross-laminated sandstone. Paleoflow was from left to right. D) Bedding plane exposure of ripple cross-laminated sandstone; individual cross-laminae are often covered with fine fragments of plant materials (black) and they dip into the bed toward the left indicating that paleoflow was from right to left.

4.2.4: Dunes and Cross-Beds

Given a constant grain size, an increase in velocity will cause ripples/lower plane beds to transition into dunes (a bedform). The main difference between ripples and dunes is size; by definition, dunes are greater than 10 cm tall. Other than that, the processes and terminology are largely the same … they migrate in a downstream direction by deposition of inclined layers on the steeper downstream side.

Based on the extent and morphology, we can subdivide these bedforms into sinuous-crested dunes which have curved and often laterally discontinuous crests and straight-crested dunes with have a linear crest and may be continuous for tens of meters.

Dunes are preserved as cross-beds, which have broadly horizontal bedding planes separating beds that have inclined laminae internally. Trough cross-beds are the product of sinuous-crested dunes. We are using the terminology of Potter and Pettijohn (1977) who differentiate trough and planar cros-beds based on the morphology of the bounding surfaces between individual foreset beds (individual cross-bedded layers). Trough cross-beds have curved bounding surfaces that appear u-shaped when viewed in an up- or down-flow direction and scoop-shaped when viewed perpendicular to flow direction. When viewed from above, foresets in trough cross beds appear u-shaped and open in a downstream direction. Tabular cross-beds are the product of straight-crested dunes. They too have downstream-dipping foresets but the bounding surfaces are much more planar and continuous than in trough-cross beds.

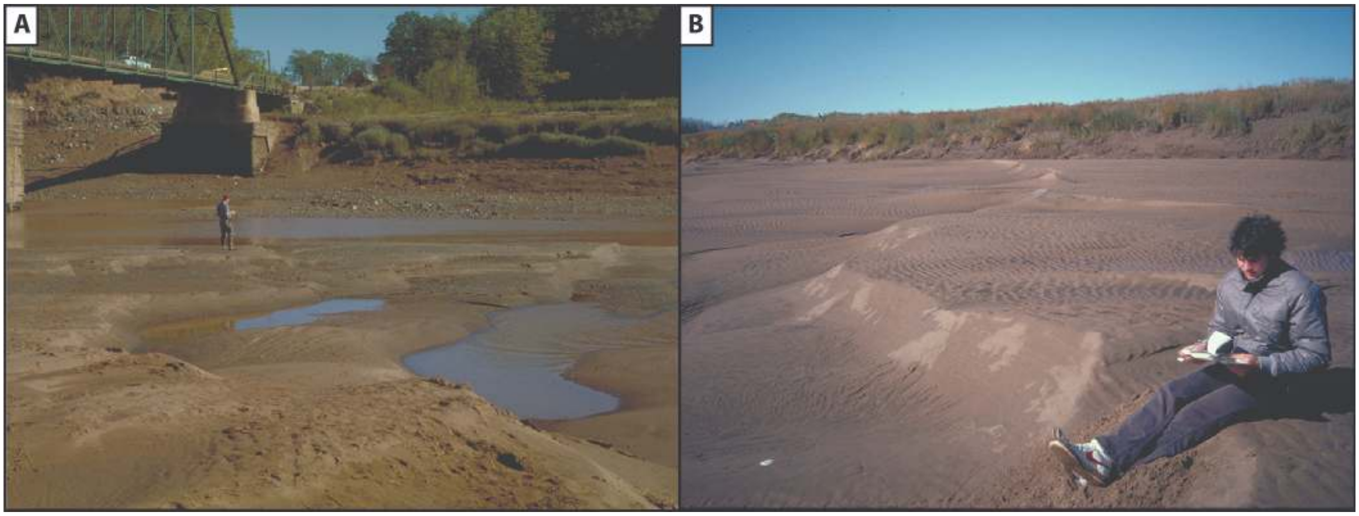


Figure 4.2.4: Dune crest shape. A) Sinuous-crested dunes from the Kennetcook River. B) Straight-crested dunes from the Kennetcook River estuary (both images courtesy John Waldron via University of Alberta; CC BY-NC-SA 4.0).

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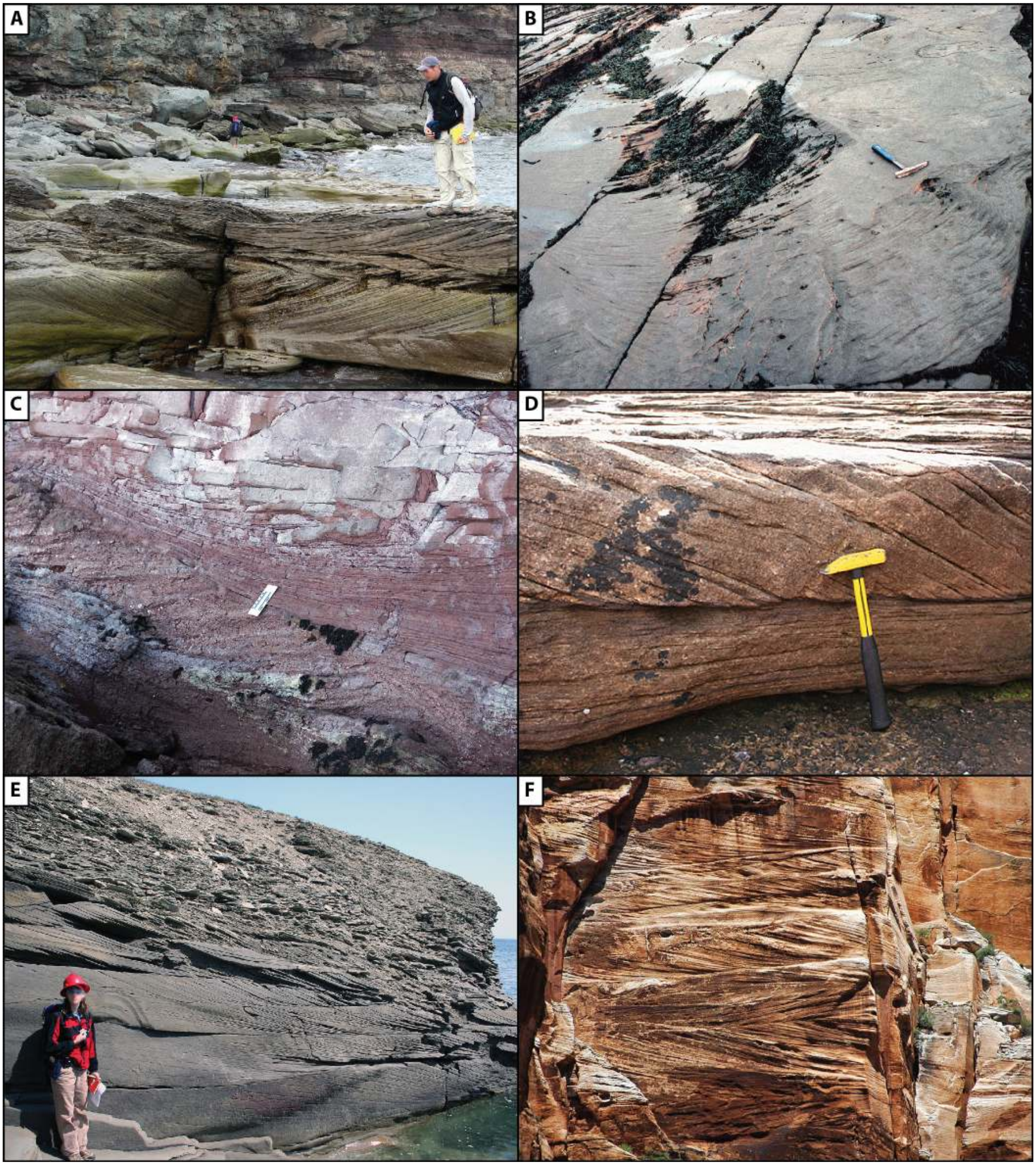
Figure 4.2.5: Photographs of cross-beds. A) Cross-sectional view of trough cross-beds; paleoflow is almost directly into the face. B) Slightly oblique bedding plane view of trough cross-beds; paleoflow is into the image toward about 11 o'clock. The u-shaped foreset laminae open in a downstream direction and dip downstream. C) Cross-sectional view of trough cross-beds; paleoflow is from right to left. Note that cross-laminae become tangential with the lower bounding surface and that there is modest relief along the bounding surface. D) Planar cross-beds with foreset laminae that intersect the lower bounding surface at a steep (close to the angle of repose) and that the lower bounding surface is nearly planar. Paleoflow direction is from left to right. E) Planar cross beds with tangential foreset laminae but planar bounding surfaces. Paleoflow direction is from left to right. Although there is some difference of opinion about whether to use foreset or bounding surface morphology to classify cross-beds, the descriptive terminology doesn't ultimately matter all that much. F) Planar cross-beds with wedge-shaped bounding surfaces. Paleoflow direction is generally from left to right. Parts A, B, C, and E from Michael C. Rygel via Wikimedia Commons, CC BY-SA 3.0 or CC BY-SA 4.0; D from Anne Burgess via Wikimedia Commons, CC BY SA 2.0; E from James St. John via Wikimedia Commons, CC BY 2.0.


A
B
C
D
E
F

4.2.5: Upper Plane Beds and Laminated Sandstone with Primary Current Lineation

A continued increase in velocity will cause dunes to wash out and the formation of a flat sediment surface where individual grains are rolling or streaming along the bed. These horizontally laminated upper plane beds (bedform) are nearly identical to lower plane beds except they can occur in much finer grained sand and that bedding plane surfaces are ornamented with primary current lineation. This distinctive linear fabric forms parallel to paleoflow direction as sand grains align behind one another via micro-vortices (just how bicyclists draft behind one another when racing).

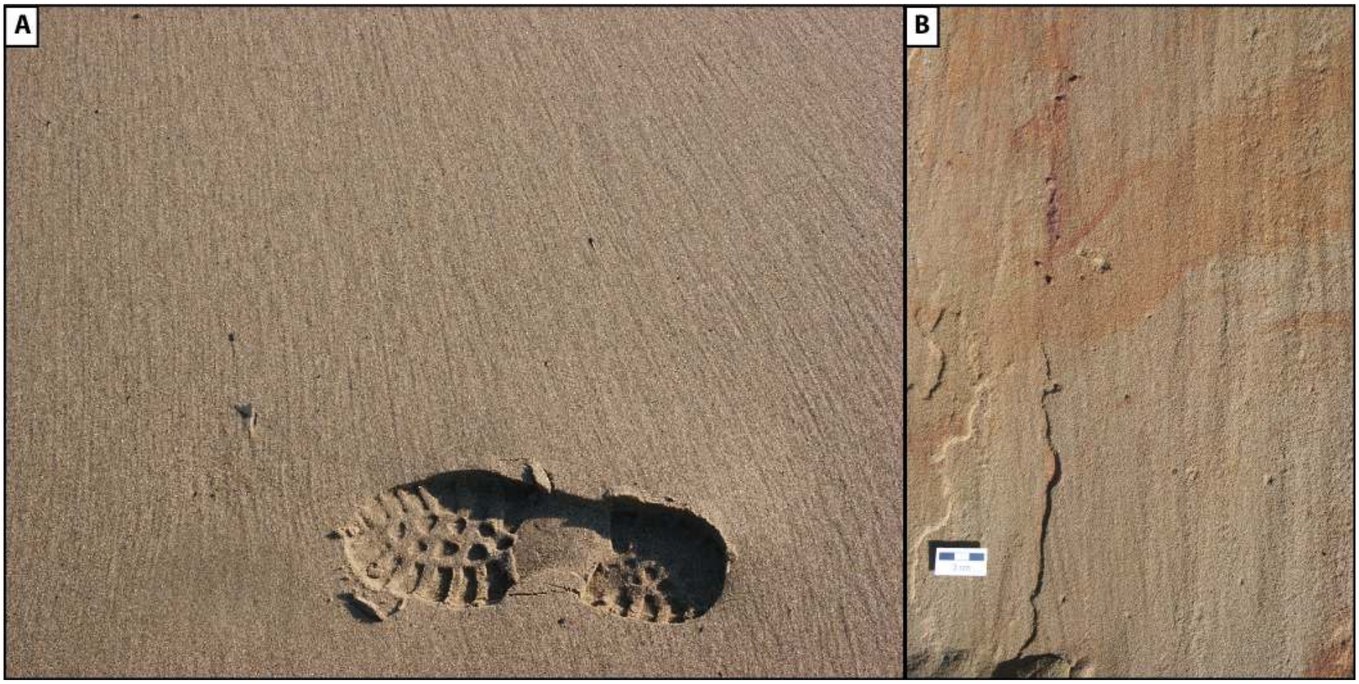


Figure 4.2.6: A) Modern primary current lineation in a modern beach sand. The linear fabric is caused by the alignment of sand grains. The accumulation of sand behind larger pebbles indicates that paleoflow was from top to bottom. B) Ancient primary current lineation on a bedding plane in a laminated sandstone. Paleoflow was either from top to bottom or bottom to top. (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0)

4.2.6: Antidunes

As the name implies, antidunes (name for bedform and sedimentary structure) are, in many ways, the opposite of regular dunes. They form under fast shallow flow conditions (Fr > 1) and the bedform migrates in an upstream direction along with the standing wave that sits atop it (see video below). They do this as sediment is plastered onto the steeply-dipping upstream face of the bedform; the result is that laminae dip in an upstream direction. They'd be easy to confuse with regular cross-beds, but don't fret because they are only rarely preserved because as flow velocity wanes the sediment is commonly reworked into upper plane beds and dunes. When preserved, they are closely associated with wavy- or undulatory laminae formed with the transition to upper plane beds.

Figure 4.2.7: Video of active antidunes and standing waves.

Figure 4.2.8: Possible antidunes and undulatory laminae formed in association with upper flow regime conditions (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

4.2.7: Readings and Resources

- Potter, P.E. and Pettijohn, F.J., 1977, Paleocurrents and Basin Analysis (2nd); Springer-Verlag, NY, 425 p. - https://link.springer.com/book/10.1007/978-3-642-61887-1
- Southard, J.B., and Boguchwal, L.A., 1990, Bed configurations in steady unidirectional water flows. Part 2. Synthesis of flume data, Journal of Sedimentary Petrology v. 60, no. 5, p. 658-679

4.3: Structures Formed by Bidirectional, Oscillatory, and/or Fluctuating Flows

4.3.1: Wave Motion and Wave Ripples

In deep bodies of water (where depth>½ wavelength) energy is transferred through water causing particles to move in a circular pattern.

When water depth is less than ½ wavelength, the wave "feels" bottom and friction causes the circular motion to change into back-and-forth motion which in turn causes symmetric wave ripples to form. Unlike asymmetric ripples, wave ripples are most commonly preserved in three dimensions on bedding planes. The crests of symmetric ripples form parallel to the local shoreline direction.

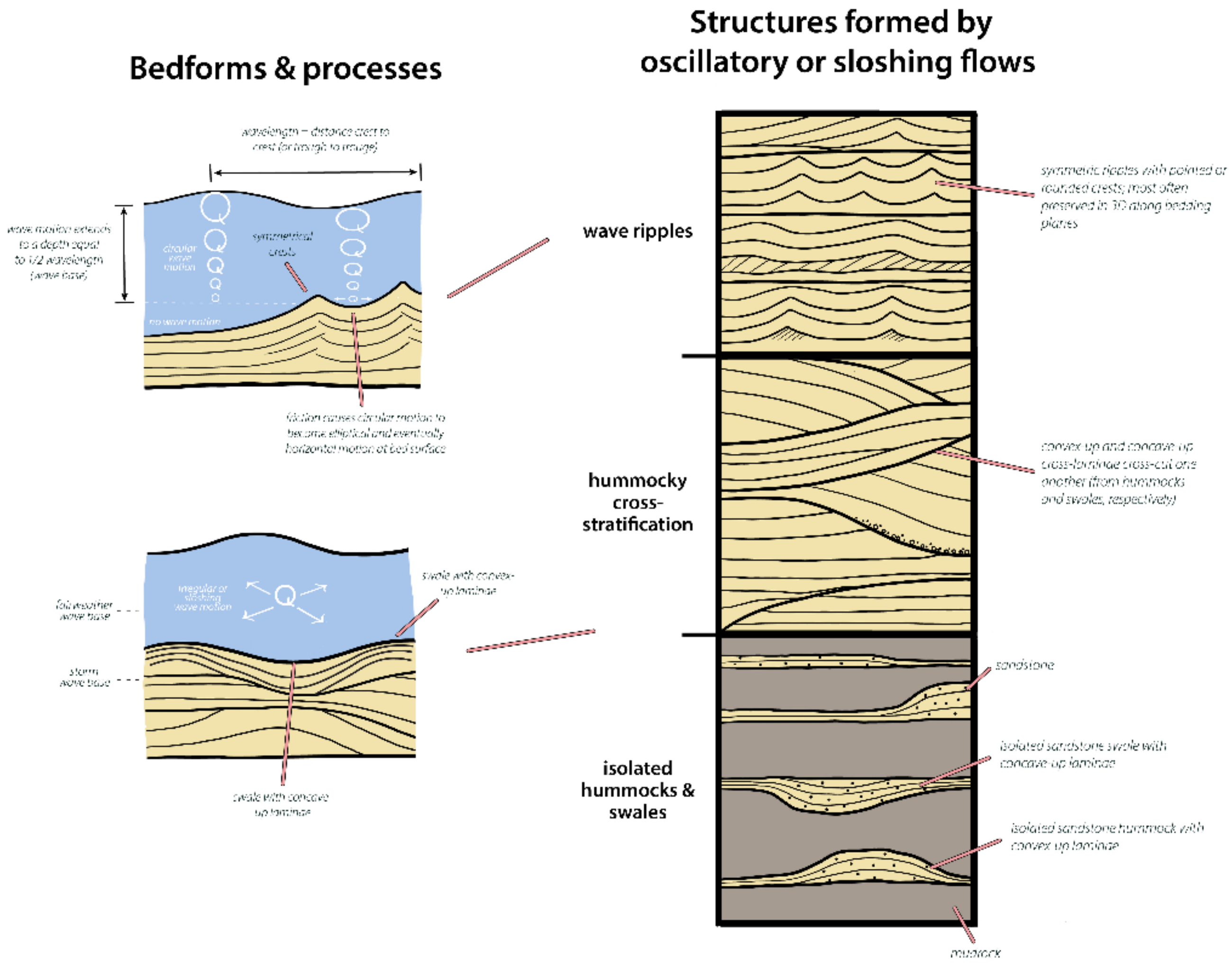


Figure 4.3.1: Processes, bedforms and structures formed by waves, oscillatory or sloshing flows (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

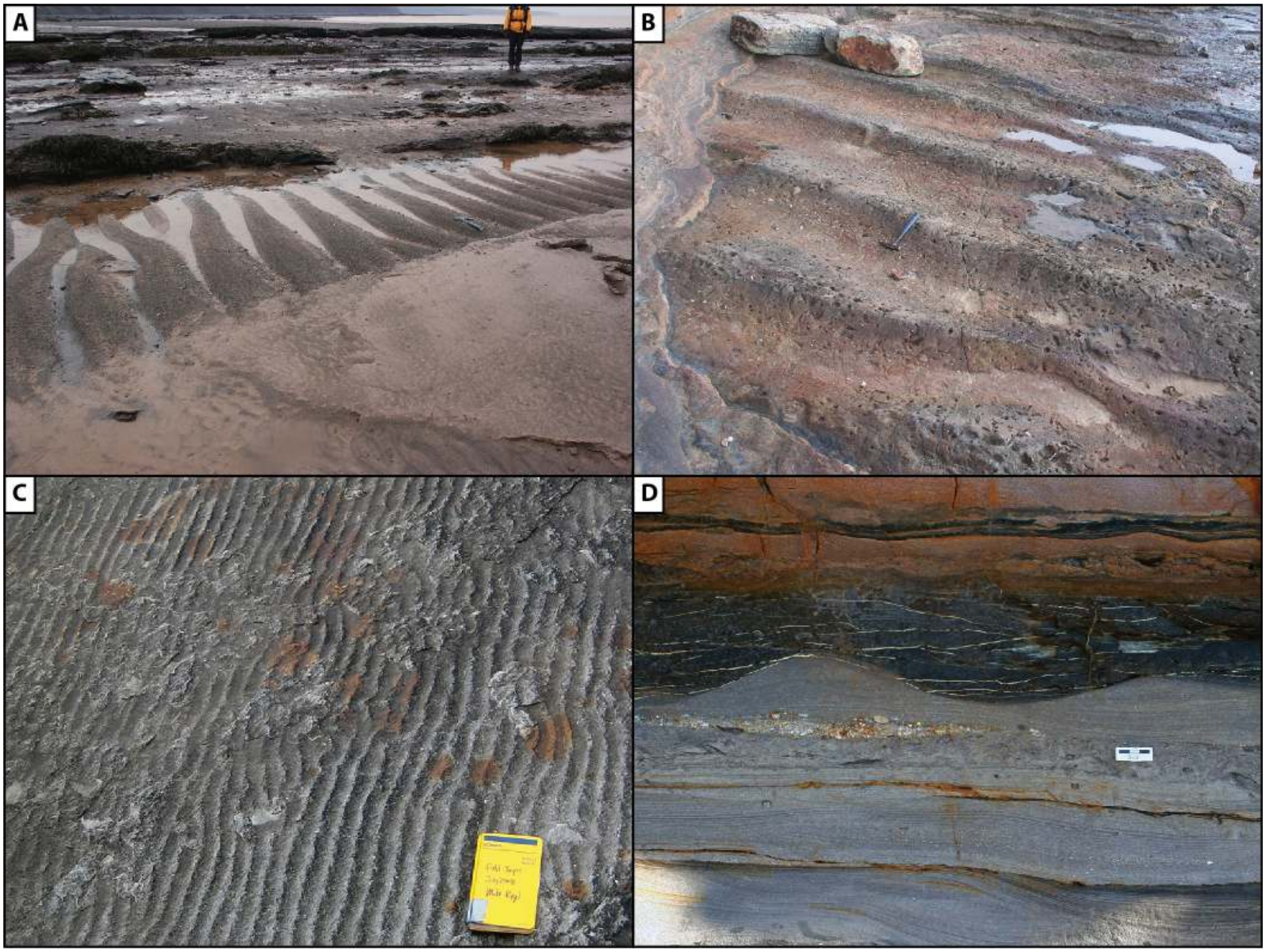


Figure 4.3.2: Modern and ancient symmetrical wave ripples (all from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0 or CC BY-SA 4.0). A) Modern wave ripples exposed at low tide on the Bay of Fundy are similar in size and shape to those shown in B) from the Permian Snapper Point Formation, New South Wales. C) A more typical bedding plane exposure of symmetrical wave ripples from the Cambrian Potsdam Sandstone. D) Although much less common, wave ripples are also occasionally preserved in cross section as with this example from the Permian Wasp Head Formation, New South Wales.

4.3.2: Hummocky Cross-Stratification

Hummocky cross-stratification (sedimentary structure) forms when storms cause chaotic wave motion and irregular "sloshing" of water at depth. The resulting hummocky cross-stratification consists of convex-up (hummocks; bedform) and concave-up (swales; bedform) cross-laminae that cross-cut one another.

In areas with muddy bottoms, storms may transport sand from shallow water and form them into isolated hummocks.

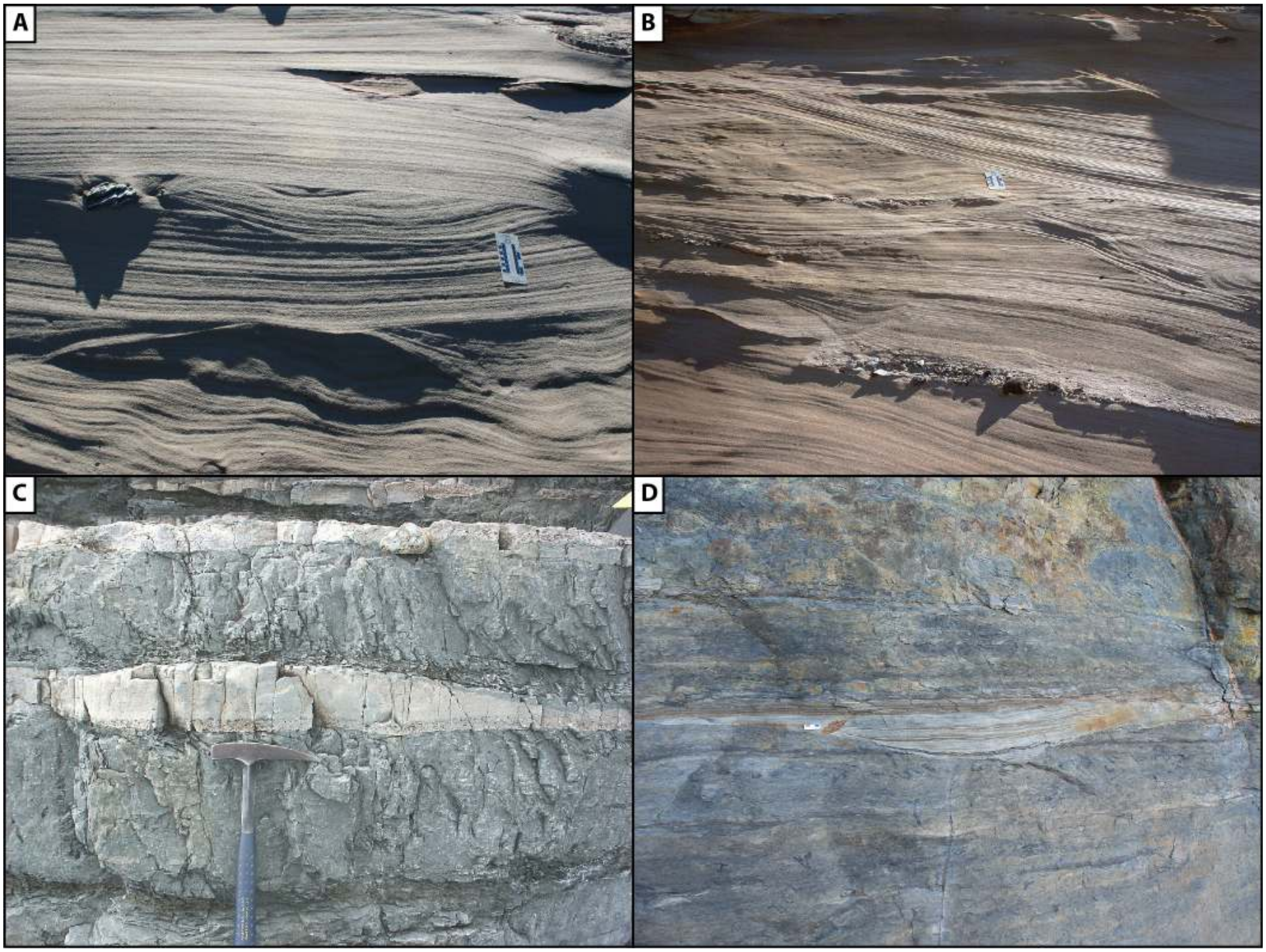


Figure 4.3.3: Examples of hummocky cross-stratification (all from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0 or CC BY-SA 4.0). A & B) Hummocky cross-stratified sandstone from the Permian Wasp Head Formation, New South Wales. C) Isolated sandstone hummock in mudrock of the Silurian Arisaig Formation, Nova Scotia. D) Isolated sandstone swale in offshore transition mudrocks of the Permian Pebbley Beach Formation, New South Wales.

4.3.3: Heterolithic Bedding

As the name implies, heterolithic bedding contains alternations of sand and mud. This feature indicates fluctuations in flow velocity and is most commonly formed in tidal regimes where water is moving rapidly during rising and falling tide and calm at high and low tide. In addition to fluctuations in flow velocity, many examples have ripple cross-laminae that indicate reversing flow - a trait that, if present, supports the tidal interpretation.

Four main types of heterolithic bedding are recognized:

- Flaser bedding is dominantly sand with isolated lenses of mud
- Wavy bedding has subequal amounts of sand and mud
- Lenticular bedding is mostly mud with elongate lenticular sandy zones
- Starved ripples are isolated ripple forms preserved within mud.

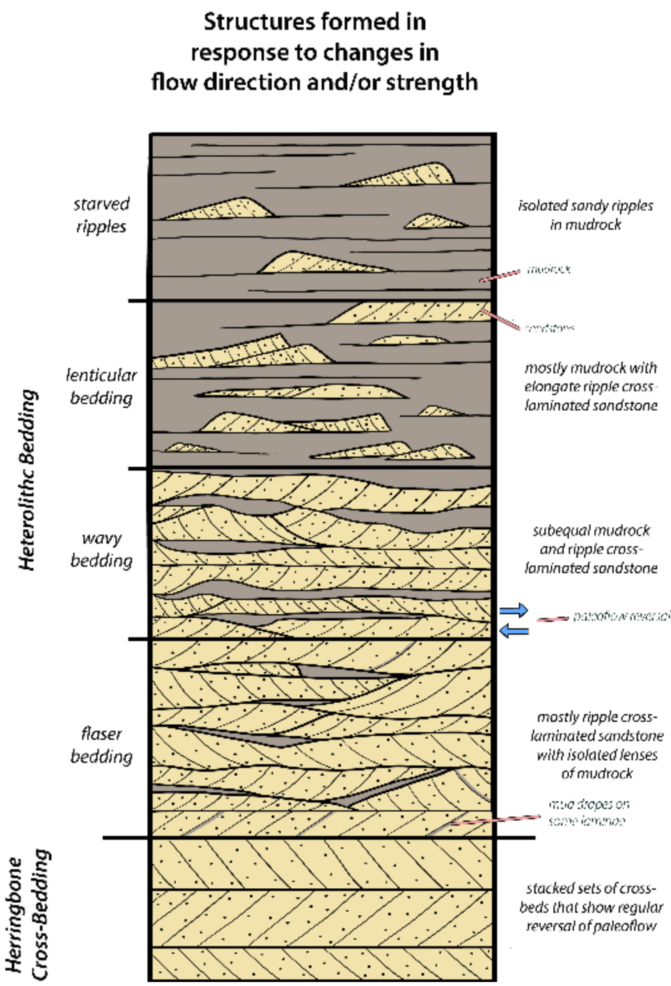


Figure 4.3.4: Structures formed in response to changes in flow direction and/or strength (Page Quinton via Wikimedia Commons; CC BY-SA 4.0). Diagram after Reineck and Wunderlich (1968).

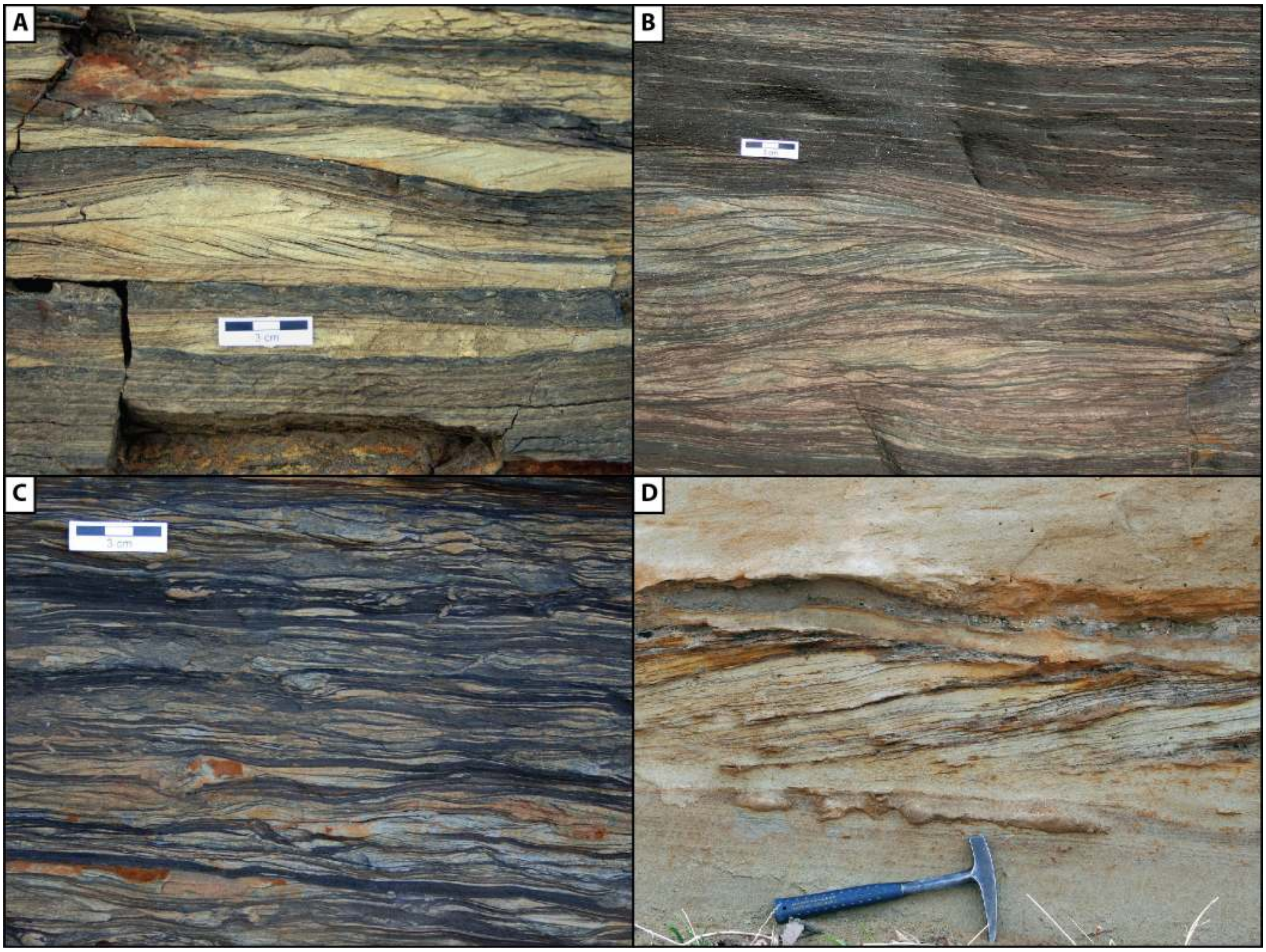


Figure 4.3.5: Examples of heterolithic bedding (all from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0 or CC BY-SA 4.0). A) Wavy bedding with ripple cross-laminae that record reversal of flow direction B) Starved ripples atop a sandy zone interval that could be described either as flaser bedding or ripple cross-laminae with mud drapes. C) Wavy bedding. A-C all from the from the upper Pebbley Beach Formation (Permian). D) Mud drapes on foreset laminae in the Indian Cave Sandstone (Pennsylvanian), Nebraska.

4.3.4: Herringbone Cross-Stratification

Herringbone cross-stratification is a name applied to successions where cross-beds overlying one another record paleoflow in opposite directions. Because this requires fast flow that rapidly reverses direction, most occurrences record tidal conditions.

Figure 4.3.6: Herringbone cross-bedding in a tidal channel, Eocene Delmar Formation, Torrey Pines State Park, California (Geozz86 via Wikimedia Commons; CC BY-SA 4.0).

4.3.5: Readings and Resources

- Reineck, H.-E., and Wunderlich, F., 1968, Classification and origin of flaser and lenticular bedding, Sedimentology, v. 11, p. 99-104.

4.4: Erosional and Post-Depositional Structures

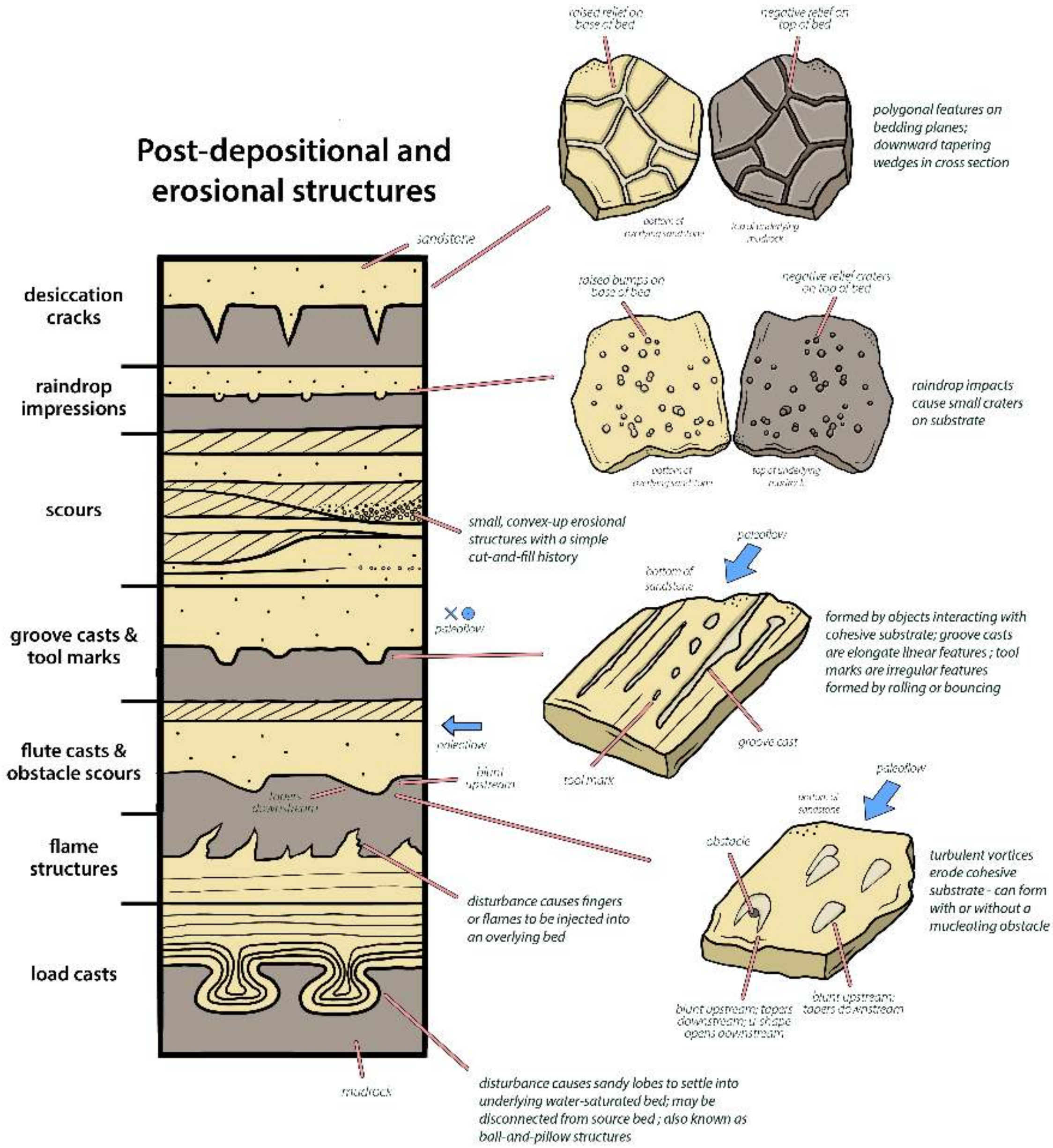


Figure 4.4.1: Diagram showing erosional and post-depositional structures (Page Quinton via Wikimedia Commons; CC BY-SA 4.0). Note that labels on some of the pop-outs to the right show whether the top or bottom of a bed is pictured.

4.4.1: Erosional Structures

4.4.1.1: Scours and Channels

Channels form where flowing water concentrates and erodes the underlying sediment. Eventually the sediment being carried by the river fills the channel and the river avulses to another location on the floodplain. Although we commonly think of channel deposits as being simple u-shaped features, their morphology is a function of the type of river, the orientation of the outcrop relative to paleoflow direction, and whether or not the older channel deposit was modified by younger ones. We will get into some of these differences when we start talking about depositional environments later in the course.

We are not aware of a strict size range applied to channels. But generally speaking, many people use the term scour to refer to small, nondescript erosional features within floodplain or beach sediments indicating a simple cut-and-fill history.

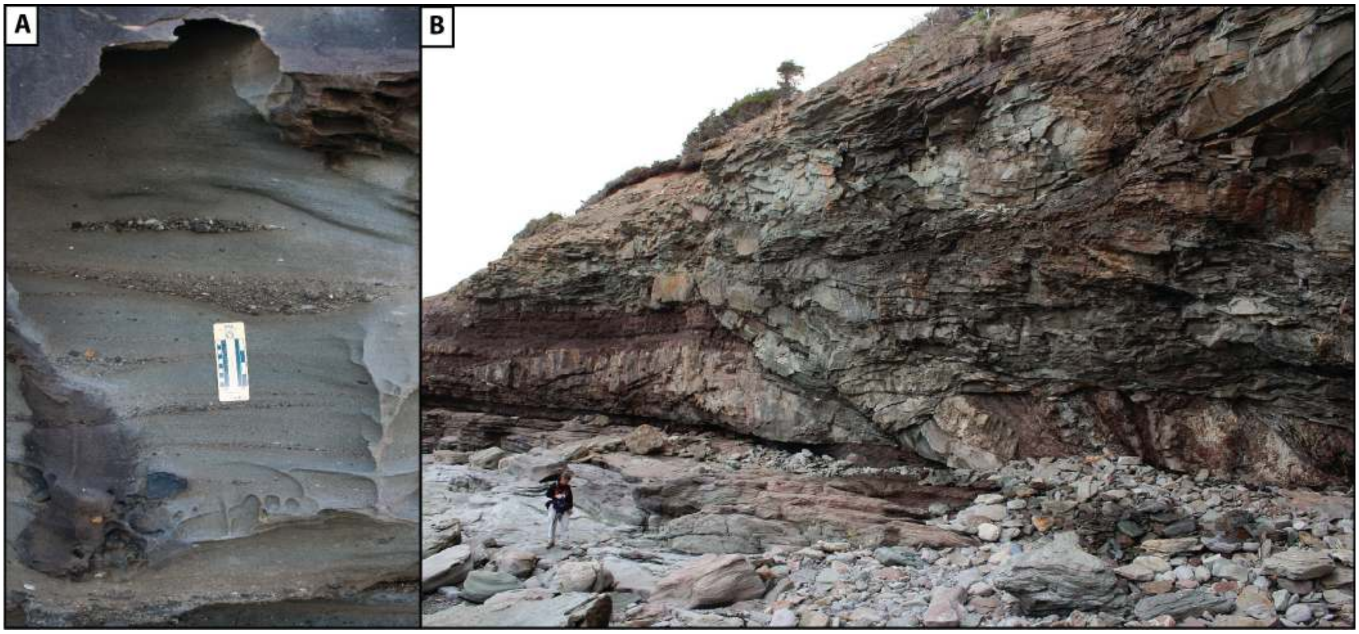


Figure 4.4.2: A) Small scour in shoreface sandstones of the Permian Wasp Head Formation, New South Wales. B) Erosive margin of a paleovalley in the Pennsylvanian Waddens Cove Formation, Nova Scotia (both images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

4.4.1.2: Groove Casts and Tool Marks

Objects drug by a moving current can make a variety of indentations into cohesive substrate. Groove casts refer to elongate linear features caused by dragging and tool marks refer to less linear features formed by rolling or bouncing along the sediment surface. If sand later blankets the sediment surface these marks may be infilled and preserved on the base of the overlying sandstone.

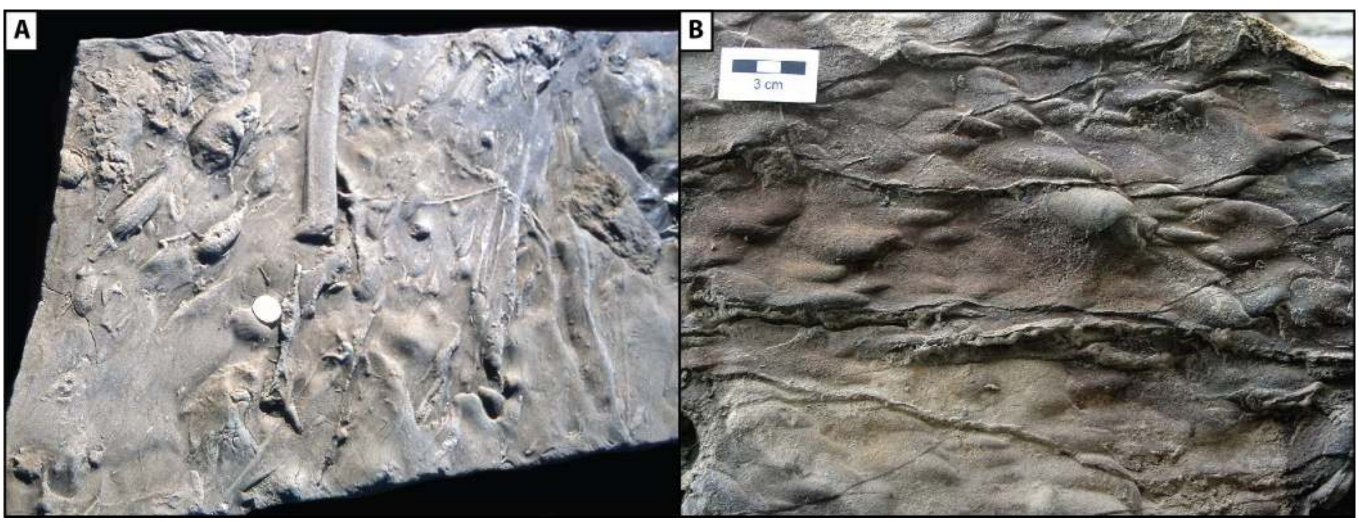


Figure 4.4.3: A) Groove casts and tool marks on the base of a bed of sandstone. Given that many of these marks taper toward the top of the image, the paleoflow direction was probably bottom to top. B) Flute casts on the base of a sandstone from the Inverness Formation (Pennsylvanian), western Cape Breton, Nova Scotia. Flute casts are bulbous on the upstream side and taper on the downstream side; in this case recording paleoflow from right to left (both images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0 or CC BY-SA 4.0).

4.4.1.3: Flute Casts

Flutes form where a turbulent vortex erodes into muddy sediment. It forms a depression that is blunt upstream and tapers downstream. If later infilled with sand, they can be preserved as flute casts that stand in relief on the base of a bed of sandstone.

They are useful for determining paleoflow direction because the blunt end faces upstream.

4.4.1.4: Obstacle Scours

Where you have a fixed obstruction (rock, tree, etc) a moving current will commonly scour a deep hole around the side and upstream edge of the feature forming a u-shaped depression that opens downstream. Although not terribly common, it is possible to preserve these obstacle scours in the geologic record.

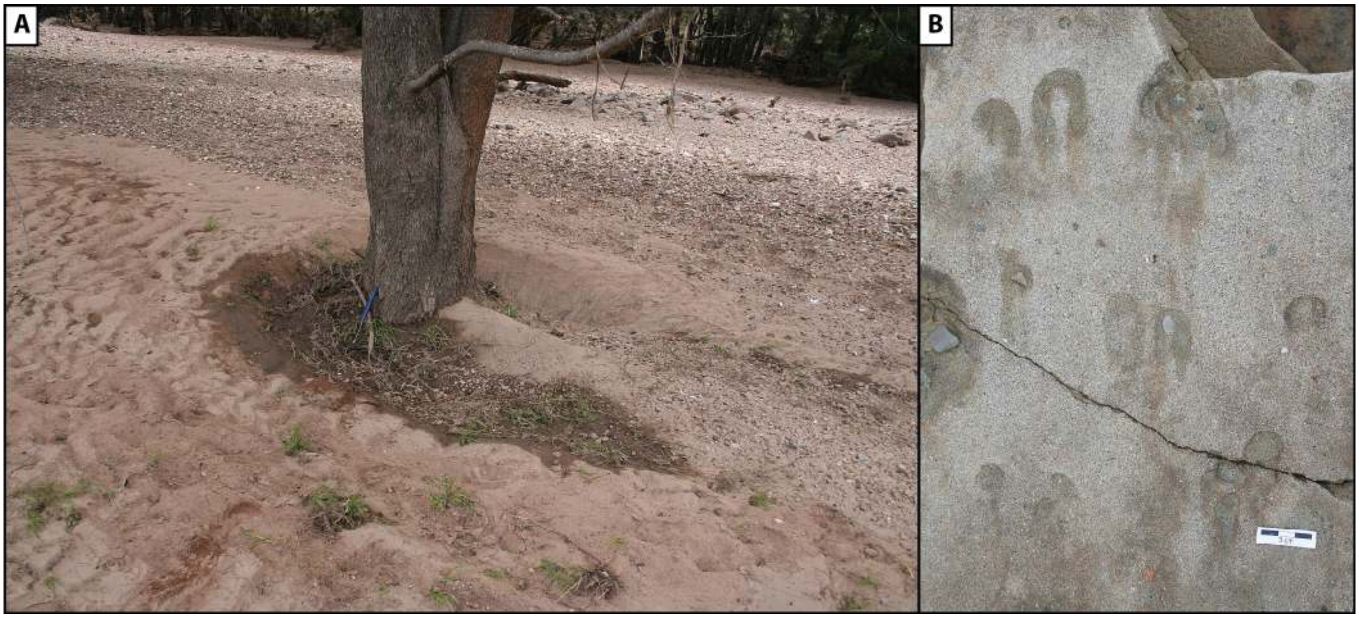


Figure 4.4.4: A) Obstacle scour around a tree. The deepest part of the scour is on the upstream side of the obstacle; turbulent vortices are shed off the sides of the obstacle and create a u-shaped scour pit that opens in a downstream direction; in this case flow was from left to right. B) Obstacle scours (aka current crescents) preserved on a bedding plane in a fluvial sandstone of the Pennsylvanian Ragged Reef Formation. Flow direction was from top to bottom (both images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0).

4.4.2: Soft-Sediment Deformation Structures

The sedimentary structures described above were all “primary” sedimentary structures in that they record conditions at or near the time that the sediment was deposited. The structures described below are all "secondary" structures in the sense that they show modification of the sediment after it was deposited but before it underwent diagenesis (the process that transforms sediment to rock).

4.4.2.1: Flame Structures

When water-saturated mud is overlain by more dense sand, compaction and/or a slight disturbance can cause flame structures which are tongues of mud to inject upward into the overlying sand

4.4.2.2: Load Casts

Load casts (aka ball-and-pillow structures) are sandy lobes of mud that can form when dense mud settles down into water-saturated mud as a result of compaction or disturbance. They may be be connected to source bed or may be isolated from it. Superficially they can appear similar to flute casts but they lack the pronounced blunted edge.

A wide variety of other types of nameless synsedimentary deformation features can form when shaking or loading of water-laden muds can cause overall chaotic bedding in rocks. The key to distinguishing these features from those formed as a result of tectonic forces is that over and underlying beds are flat-lying and relatively undeformed.

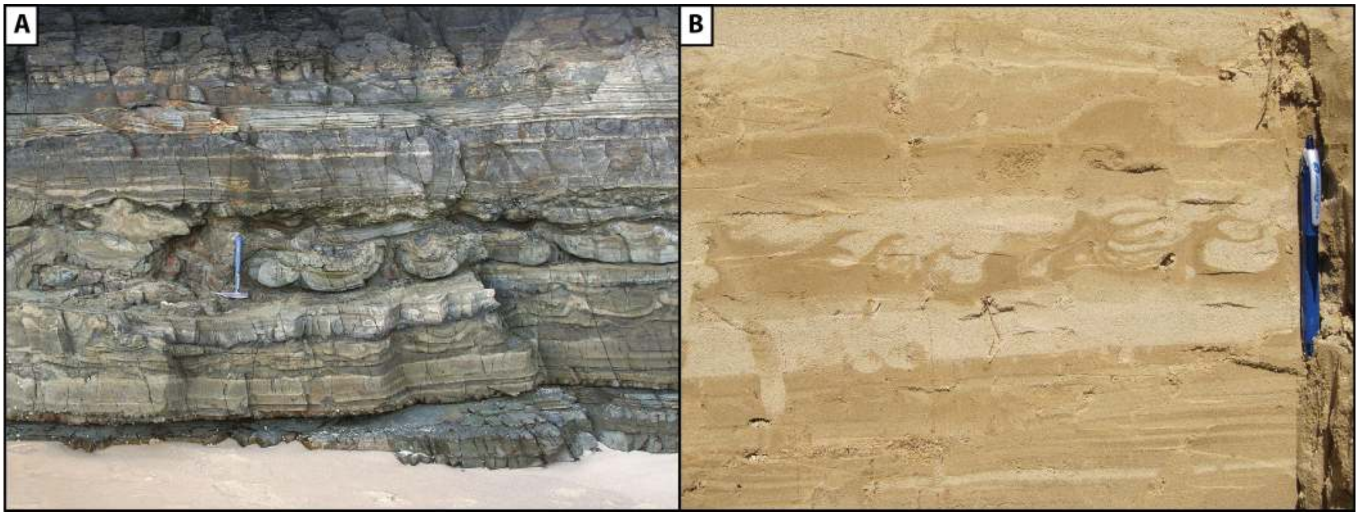


Figure 4.4.5: A) Cross-sectional view of load casts (aka ball-and-pillow structures) in the Booti Booti Sandstone (Mississippian), New South Wales (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0). B) Flame structures in unconsolidated sand, (Catrin1000via Wikimedia Commons; public domain).

4.4.2.3: Desiccation cracks

As mentioned above, mud can be 70% water. When it dries out it contracts forming polygonal desiccation cracks (aka mudcracks). These features can be preserved as polygonal features on bedding planes or as downward tapering wedges if viewed in cross section.

4.4.2.4: Raindrop Impressions

When raindrops fall onto fine-grained sediment they make small depressions or craters on the top of that sediment. These features can be preserved either as depressions on the top of the fine-grained sediment that they fell on or as bumpy casts on the base of the overlying bed.

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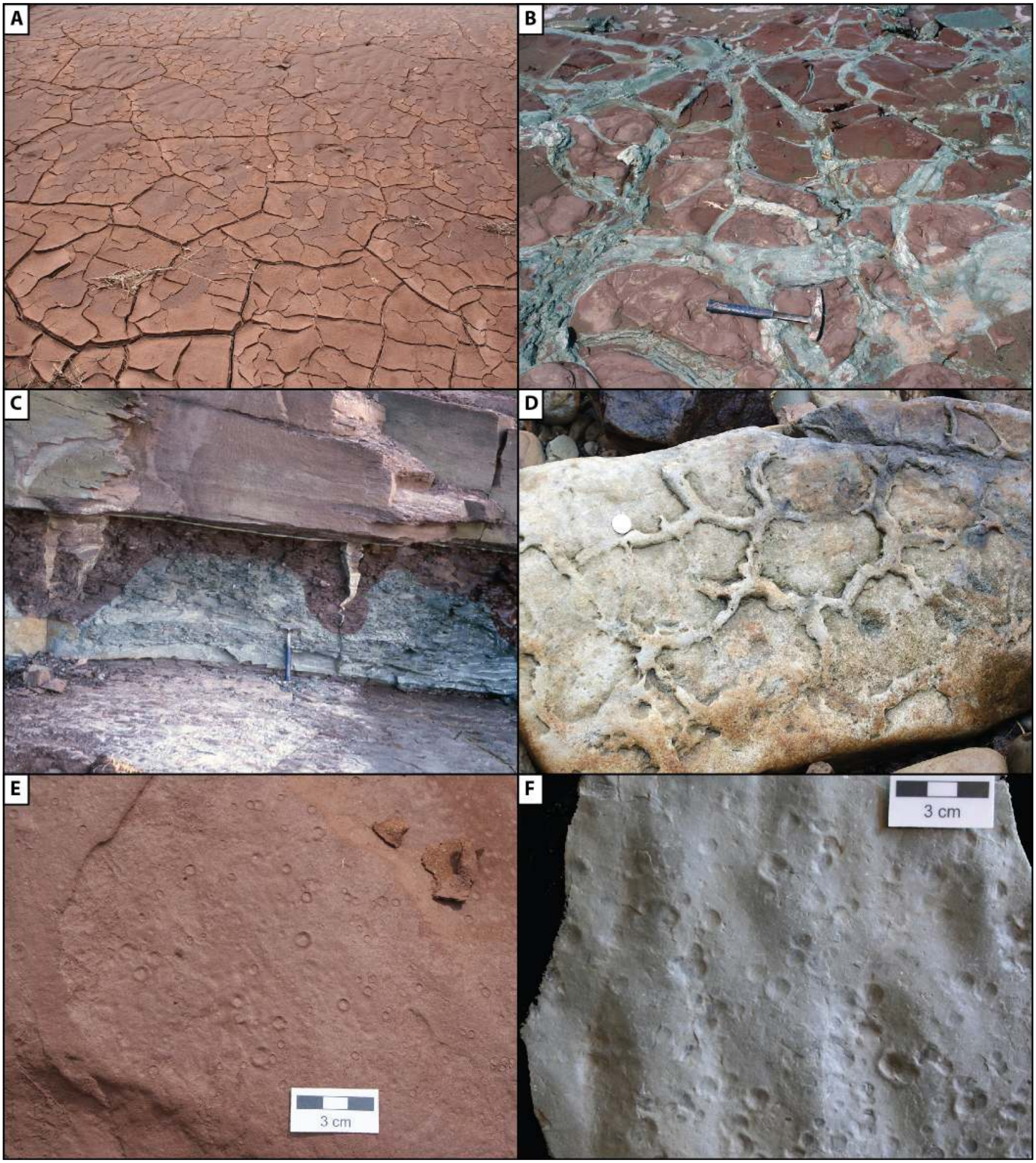
Figure 4.4.6: A) Oblique view of modern desiccation cracks atop a tidally-influenced point bar in the Salmon River near Truro, Nova Scotia. B) Slightly oblique bedding plane view of green reduced zones associated with desiccation cracks in the Pennsylvanian Clifton Formation, New Brunswick. C) Cross-sectional view of mudcracks in the Mississippian Mabou Group, Nova Scotia. D) Dessication crack preserved as a cast on the base of a sandstone, Inverness Formation (Pennsylvanian), Cape Breton Island, Nova Scotia. E) Modern raindrop impressions on a tidally-influenced point bar in the Salmon River near Truro, Nova Scotia. F) Fossilized raindrop impressions on the top of a wave-rippled sandstone from the Horton Bluff Formation (Mississippian), near Avonport, Nova Scotia. All images by Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0 or CC BY-SA 4.0).


A
B
C
D
E
3 cm
F
3 cm

CHAPTER OVERVIEW

5: Siliciclastic Sedimentary Rocks

Siliciclastic sedimentary rocks are made of solid particles that were transported as suspended load, bedload, or under the direct influence of gravity. In all cases, the clastic particles will settle and come to rest when motion ceases.

Grain size is the most important characteristic in the classification of clastic sediments and sedimentary rocks. As discussed in Chapter 3, unconsolidated sediment can be classified as gravel, sand, or mud; these size fractions make up conglomerates, sandstones, and mudrocks, respectively. More specific names for the different size fractions and sedimentary rocks are provided in the table below.

Although dissolved materials are the primary building blocks for carbonates and other types of chemical and biochemical sedimentary rocks, dissolved materials may form mineral cements that bind clastic grains during the transformation from sediment to sedimentary rock.

Diameter (mm)	Phi units	Particle Size		Sedimentary Rocks				Settling velocity (cm/s)	Entrainment velocity (cm/s)
					Grain/Clast Composition				
					Quartz >90%	Feldspar > 10% Feldspar > Lithics	Lithics[1] > 10% Lithics > Feldspar		
		Boulder	Gravel (rounded) or Rubble (angular)	Conglomerate (rounded) or Breccia (angular)	X	X	______[2] Boulder Conglomerate (Breccia)		
256	-8								500
		Cobble	Gravel (rounded) or Rubble (angular)	Conglomerate (rounded) or Breccia (angular)	Quartz Cobble Conglomerate (Breccia)	X	_____[2] Cobble Conglomerate (Breccia)		
64	-6							163	175
		Pebble	Gravel (rounded) or Rubble (angular)	Conglomerate (rounded) or Breccia (angular)	Quartz Pebble Conglomerate (Breccia)	X	_____[2] Pebble Conglomerate (Breccia)		
4	-2							43	60
		Granule	Gravel (rounded) or Rubble (angular)	Conglomerate (rounded) or Breccia (angular)	Quartz Granule Conglomerate (Breccia)	Feldspathic Granule Conglomerate (Breccia)	______[2] Granule Conglomerate (Breccia)		
2	-1							26	50
		Very coarse	Sand	Sandstone	Quartz Arenite (<10% matrix) or Quartz Wacke[4] (>10% matrix)	Feldspathic Arenite[3] (<10% matrix) or Feldspathic Wacke[3,4] (>10% matrix)	Lithic Arenite (<10% matrix) or Lithic Wacke[4] (>10% matrix)		
1	0							15	38
		Coarse	Sand	Sandstone					
1/2	1							7	28
		Medium	Sand	Sandstone					
1/4	2							3	23
		Fine	Sand	Sandstone					
1/8	3							1	20
		Very fine	Sand	Sandstone					
1/16	4							0.3	Cohesion may become important and entrainment velocities may increase
		Silt	Mud	Mudrock (massive) or Shale (fissile)	Siltstone (silt > clay) Mudstone (silt ≈ mud) Claystone (silt < clay)				
1/256	8							0.08	
		Clay	Mud	Mudrock (massive) or Shale (fissile)					

X = clasts of this size and composition are unlikely.
[1]Lithic fragments could be composed of chert, limestone, igneous, or metamorphic rock fragments (and many others).
[2]Lithologic descriptor of most abundant clast type should be used in the blank space (ex: basalt cobble conglomerate); use the term polymictic if no one lithology is dominant (ex: polymictic pebble conglomerate).
[3]Feldspar-rich sandstones are informally termed arkose.
[4]Sandstones with abundant muddy matric are informally called graywackes.

Table 5.1: Grain size classification from Wentworth (1922) and phi scale from Krumbein (1934). Settling velocities from http://www.filtration-and-separation...g/settling.htm and entrainment (erosion) velocities from http://en.Wikipedia.org/wiki/File:We...Size-Chart.pdf; both assume spherical particles of quartz.

Learning Objectives

- Identify the main components of sandstones in thin section and use this information to name them.
- Use appropriate terminology to described conglomerates and breccias
- Use appropriate terminology to described mudrocks and make interpretations based on mudrock color.
- Explain the four main types of clay minerals and list some economically important examples.
- Explain the possible origins of diamictites and sandstones/mudrocks with outsized clasts.

Topic hierarchy

Chapter thumbnail shows a sandy, matrix-supported conglomerate (Michael Rygel via Wikimedia Commons; CC BY-SA 3.0).

5.1: Sandstones

Sandstones are clastic sedimentary rocks that are composed largely of sand-sized grains (>50% particles between 1/16 and 2 mm diameter). Sandstones make up approximately 14% of sedimentary rocks and can be found in a wide variety of depositional environments including alluvial fans, rivers and floodplains, beaches, and submarine fans. Porous sandstones can form important reservoirs for water and hydrocarbons; homogenous, well-cemented sandstones are commonly used as building stones.

5.1.1: Methods

Although much information can be gained with a hand lens and an acid bottle, sandstones are best studied in thin section with a petrographic microscope that has polarizing filters. The major components of sandstones include sand-sized framework grains and interstitial areas that originally contained either muddy interstitial matrix and/or open pore space. During diagenesis, mineral cements may precipitate into pore space.

5.1.1.1: Petrographic Microscopes

If you want a quick refresher on use of the petrographic microscope, a helpful YouTube video is embedded below and several other resources are provided at the end of this chapter.

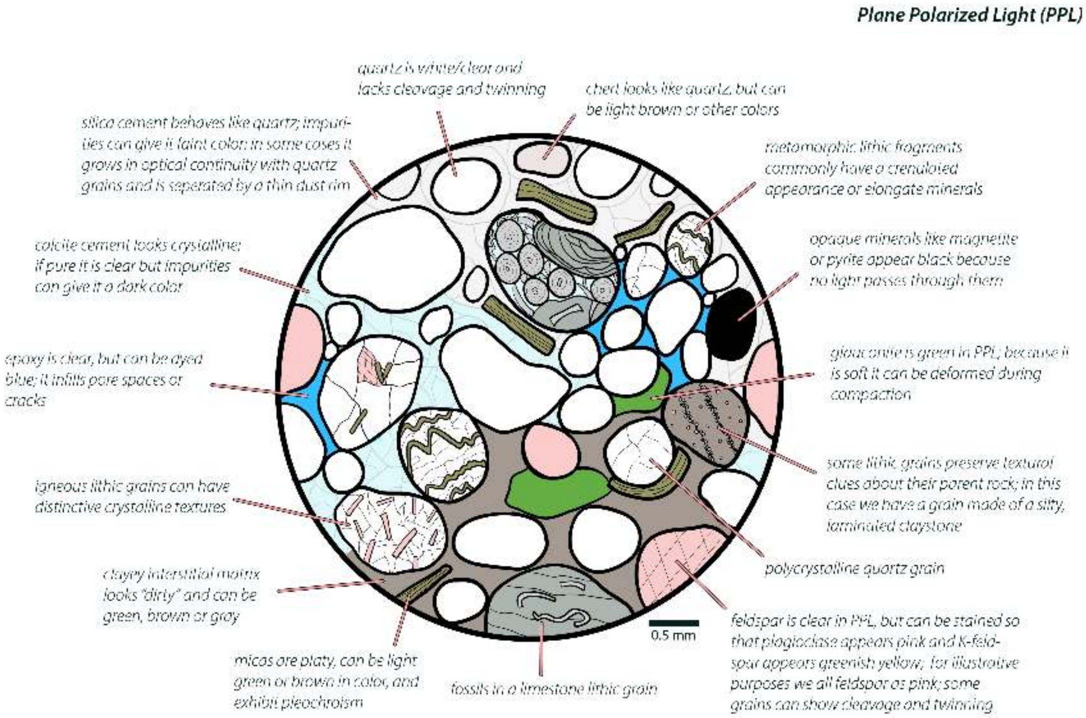
Video 5.1.1: Review of petrography fundamentals.

5.1.1.2: Rocks and Minerals in Thin Section

As shown in the video above, the combination of thin sections (30 micron thick slivers of rock glued to a glass microscope slide) and petrographic microscope with polarizing filters can provide significant information and insight into rocks and minerals. In thin section, we can use the following properties to identify minerals:

- Color (in plane polarized light) - the hue observed in a mineral. It results from the selective absorption of certain wavelengths of light due to the mineral's composition and crystal structure. Color can vary due to impurities or alteration.
- Pleochroism (in plane polarized light) - the change in color observed in an anisotropic mineral when the stage is rotated. This occurs because the mineral absorbs light differently along different crystallographic directions. Pleochroism can range from weak to strong and can vary to include two or three colors depending on the mineral's optical properties.
- Interference colors (crossed polars) - colors that result from the splitting of light into two rays that travel at different speeds through a mineral, causing a phase difference when they recombine. The resulting colors correspond to a specific order on Michel-Lévy's interference color chart (see link below).
- Relief - the degree to which a mineral stands out relative to its surrounding minerals or the mounting medium. It is a function of the difference in refractive index (RI) between the mineral and its surroundings. High-relief minerals appear strongly outlined with pronounced shadows at grain boundaries, while low-relief minerals blend more seamlessly with their surroundings.
- Twinning - refers to the presence of intergrown crystal domains within a mineral that share a specific crystallographic relationship but have different optical orientations. It is observed under both plane-polarized light (PPL) and cross-polarized

light (XPL) in a petrographic microscope. Twinning can appear as parallel, crosshatched, or wedge-shaped patterns.



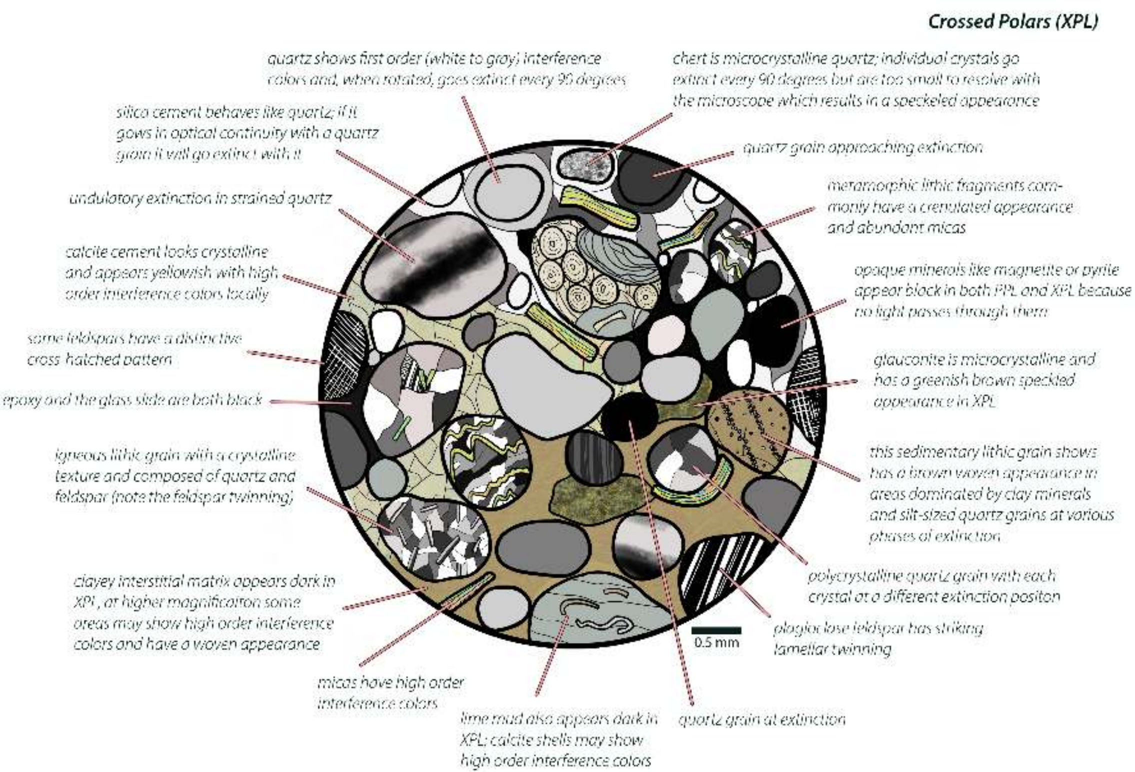


Figure 5.1.1: Summary diagram showing the most common elements of clastic sedimentary rocks in thin section (after Page Quinton via Wikimedia Commons; CC BY-SA 4.0). Note that the same field of view is shown in both, but that the top image shows what the components look like in plane polarized light and the bottom one shows what they look like with crossed polars.

5.1.2: Framework Grains

Framework grains are sand-sized particles that make up the majority of a sandstone. These particles were transported by a moving current and came to rest when turbulence and bedload transport ceased. Grains composed of quartz, feldspar, or lithic fragments (rock fragments) are the most common type of sand grains and their relative abundance is used to classify the sandstone; all others are considered accessory minerals.

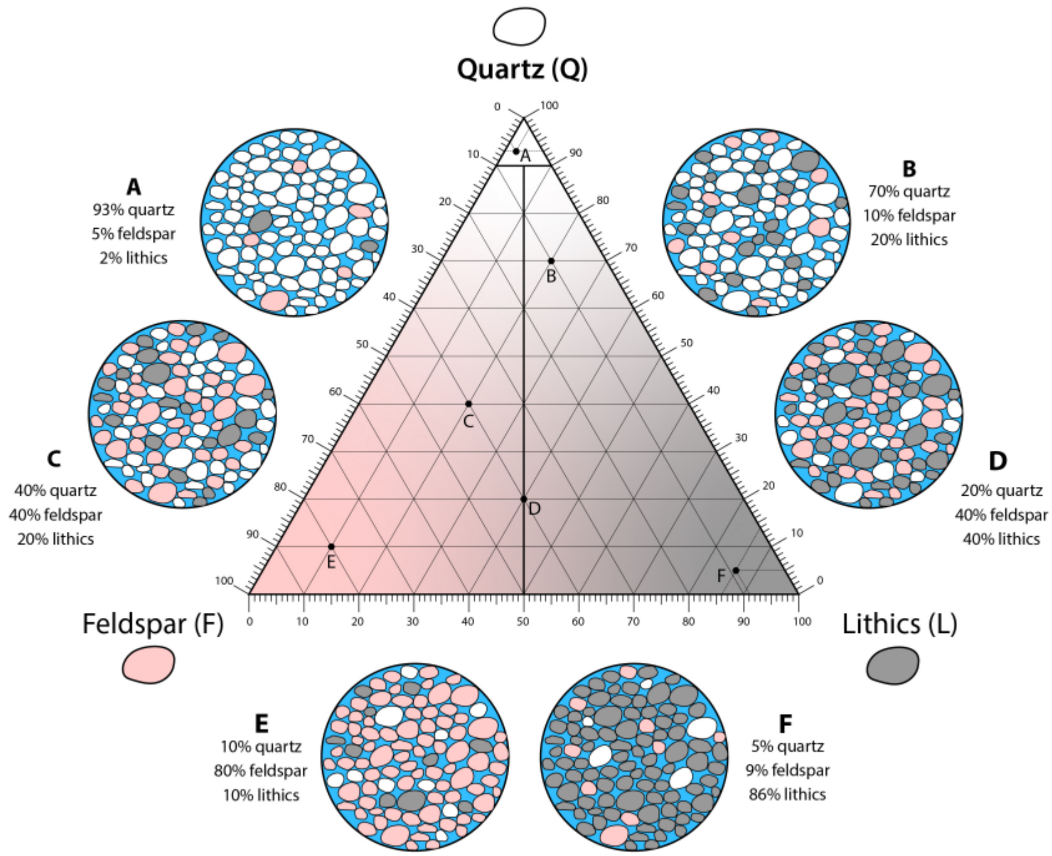


Figure 5.1.2: Ternary diagram showing the relative abundance of quartz, feldspar, and lithics in a sandstone with a few points showing what those abundances would look like (from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

5.1.2.1: Quartz

Given its resistance to chemical weathering and abundance in continental crust, quartz is the most common framework grain in most sandstones. In hand sample, quartz grains are easily identified by its glassy appearance and lack of cleavage. In thin section, if the quartz grain is made from a single crystal (monocrystalline) it will appear clear/white in plane polarized light (PPL) and uniformly go extinct every 90° in cross polarized light (XPL). Grains derived from an igneous source may actually be composed of several interlocking crystals (polycrystalline quartz) and those from a metamorphic source may show undulatory extinction. Note that polycrystalline quartz grains (and chert, see lithic grains below) can be counted as quartz if the goal is to focus on compositional maturity. Alternately, polycrystalline quartz and chert can be counted as lithics if the goal is to focus on the composition of the source rocks.

5.1.2.2: Feldspar

Feldspar grains can be particularly abundant in sandstones derived from granitic source areas and/or in arid climates with limited chemical weathering. Feldspar grains appear clear in PPL and commonly exhibit twinning in XPL. In both cases, cleavage may give the external form and internal planes of weakness a blocky appearance. When making thin sections, it is possible to stain them to help distinguish feldspars from other broadly similar minerals. If stained with cobaltinitrite, potassium feldspar will take on a yellow color; if stained with barium rhodizonate, plagioclase feldspar will turn red.

5.1.2.3: Lithics

Lithic grains are sand-sized particles of preexisting metamorphic, sedimentary, or igneous rocks. These grains have a variety of appearances, generally they can be composed of multiple minerals, have a complex fabric, and/or have a distinctive composition (ex: carbonate). As discussed above, one could count polycrystalline quartz and chert as lithic fragments if the goal of the analysis was to emphasize source (rather than compositional maturity).

5.1.2.4: Accessory Minerals

All other sand-sized particles are classified as accessory minerals. Muscovite and biotite micas are common accessory minerals in sediment found adjacent to crystalline source areas. The micas are easily identified in hand sample as particularly large, platy grains concentrated along bedding planes; in thin section they appear as large play grains with obvious cleavage and high order birefringence colors in XPL. Magnetite, rutile, tourmaline, zircon, and other resistant minerals may be abundant in some areas.

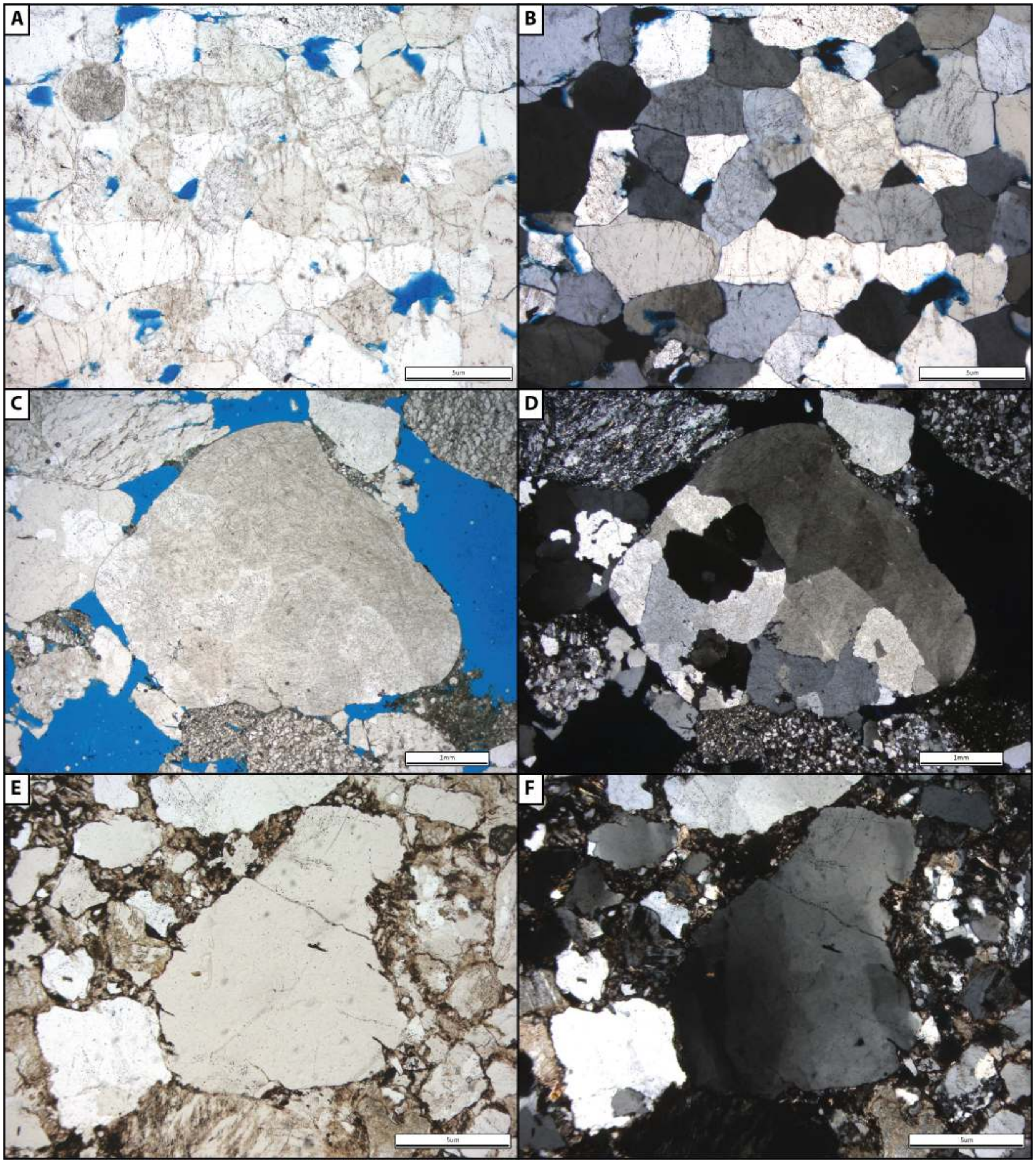


Figure 5.1.3: Photomicrographs of quartz grains in sandstones; images in plane polarized light are on the left and cross polarized light on the right (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A & B) Monocrystalline quartz grains. Note that the apparent tight fit of the grains is actually caused by silica cement growing in optical continuity with the grains; in many cases a dust rim shows the grain boundary. C & D) A large polycrystalline quartz granule. E & F) A monocrystalline quartz grain showing undulatory extinction.

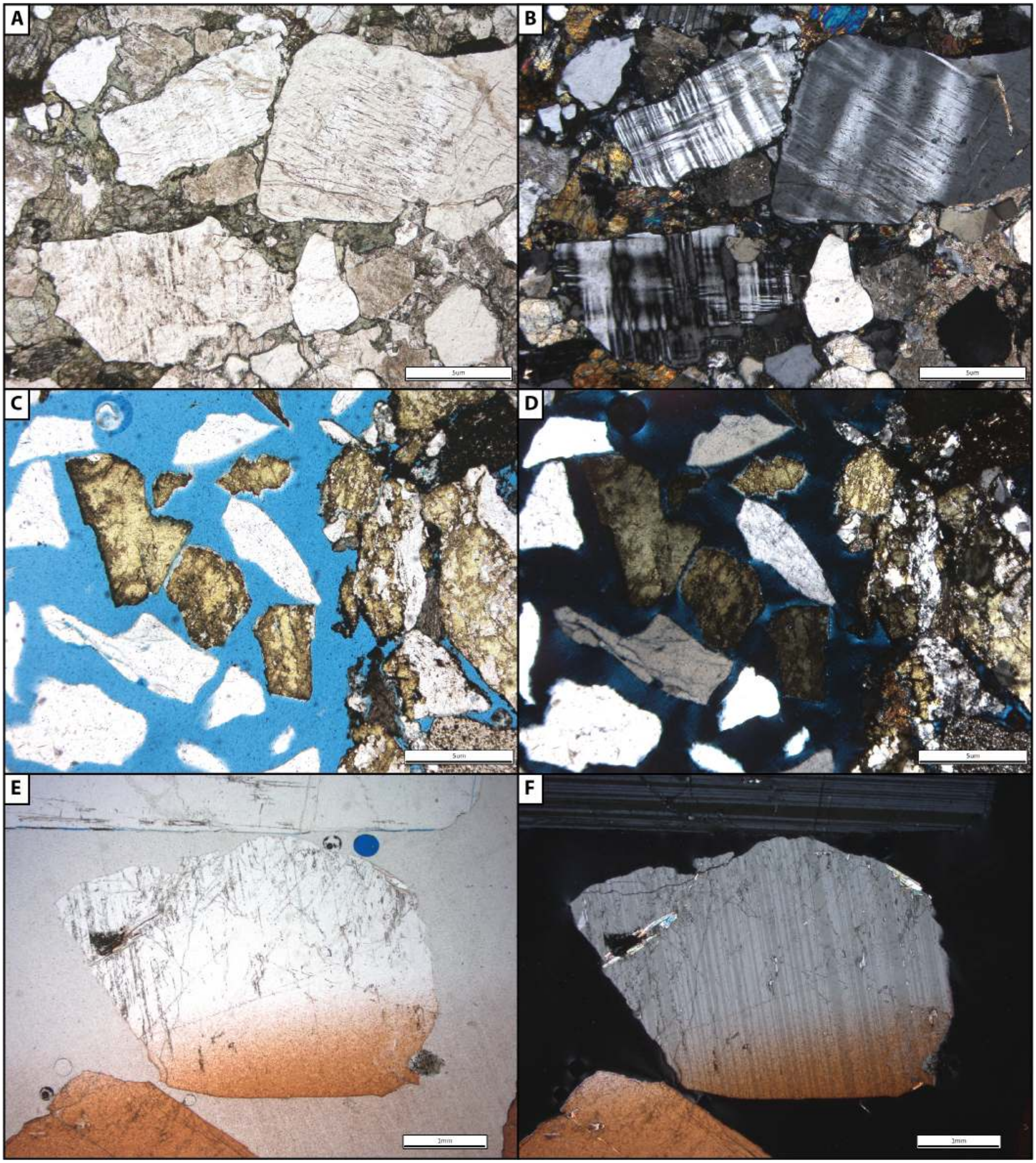


Figure 5.1.4: Photomicrographs of feldspar grains; images in plane polarized light are on the left and cross polarized light on the right (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A & B) Large unstained feldspar grains showing twinning. Scattered quartz grains and abundant interstitial chlorite C & D) Feldspar grains stained yellow for potassium feldspar. The simple grains on the left are single feldspar crystals; the more complex grains on the right are lithic fragments with feldspar in them. E & F) Plagioclase feldspar grains; the bottom half of the image is stained red for plagioclase; the top half of the image is unstained.

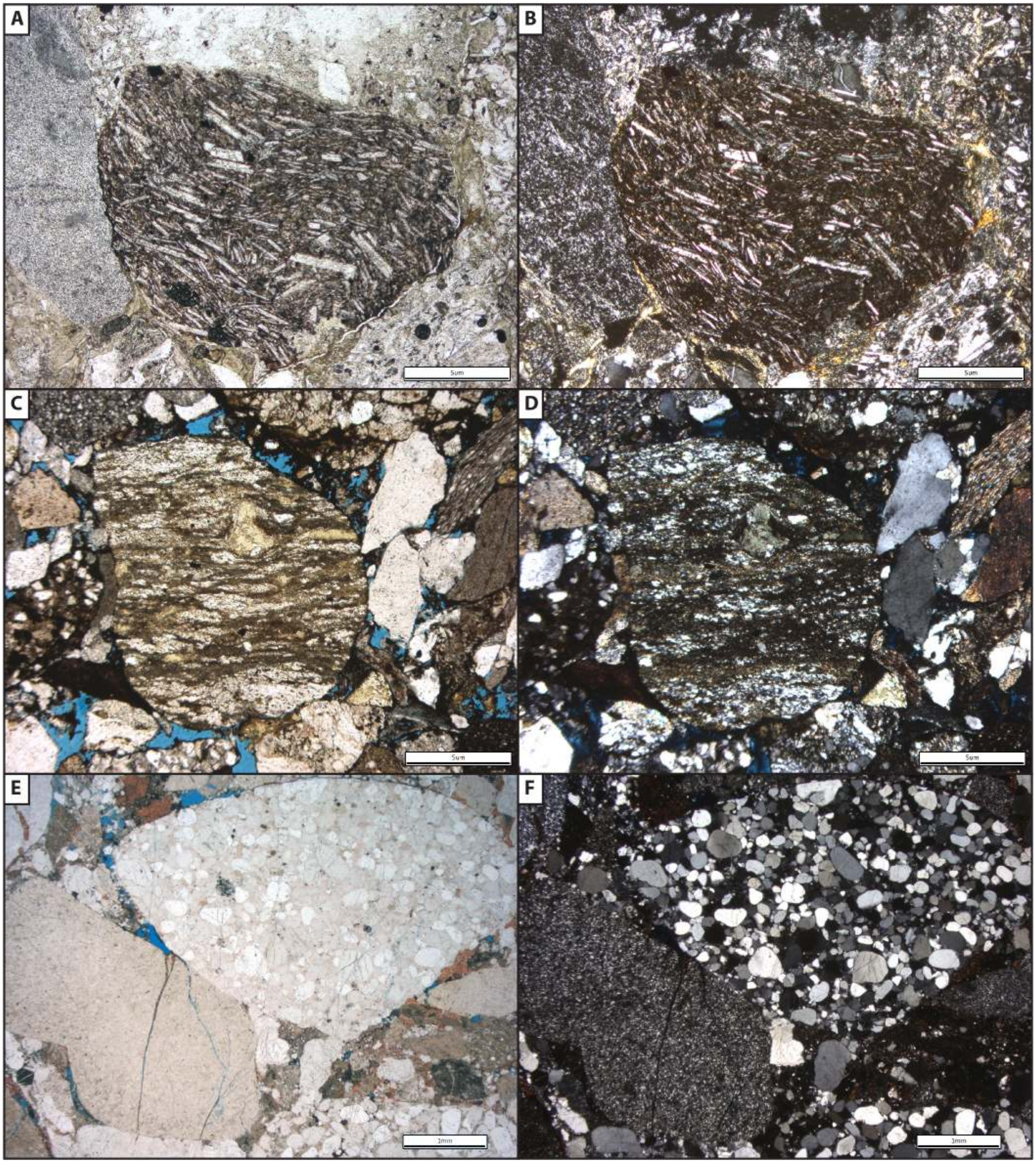


Figure 5.1.5: Photomicrographs of lithic grains; images in plane polarized light are on the left and cross polarized light on the right (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A & B) Lithic grains derived from an igneous rock. C & D) Lithic grains derived from a metamorphic rock - likely a schist or a gneiss. E & F) Lithic granules derived from sedimentary rocks. The large grain near the top right is a fine-grained sandstone; the large grain in the left half of the image is chert.

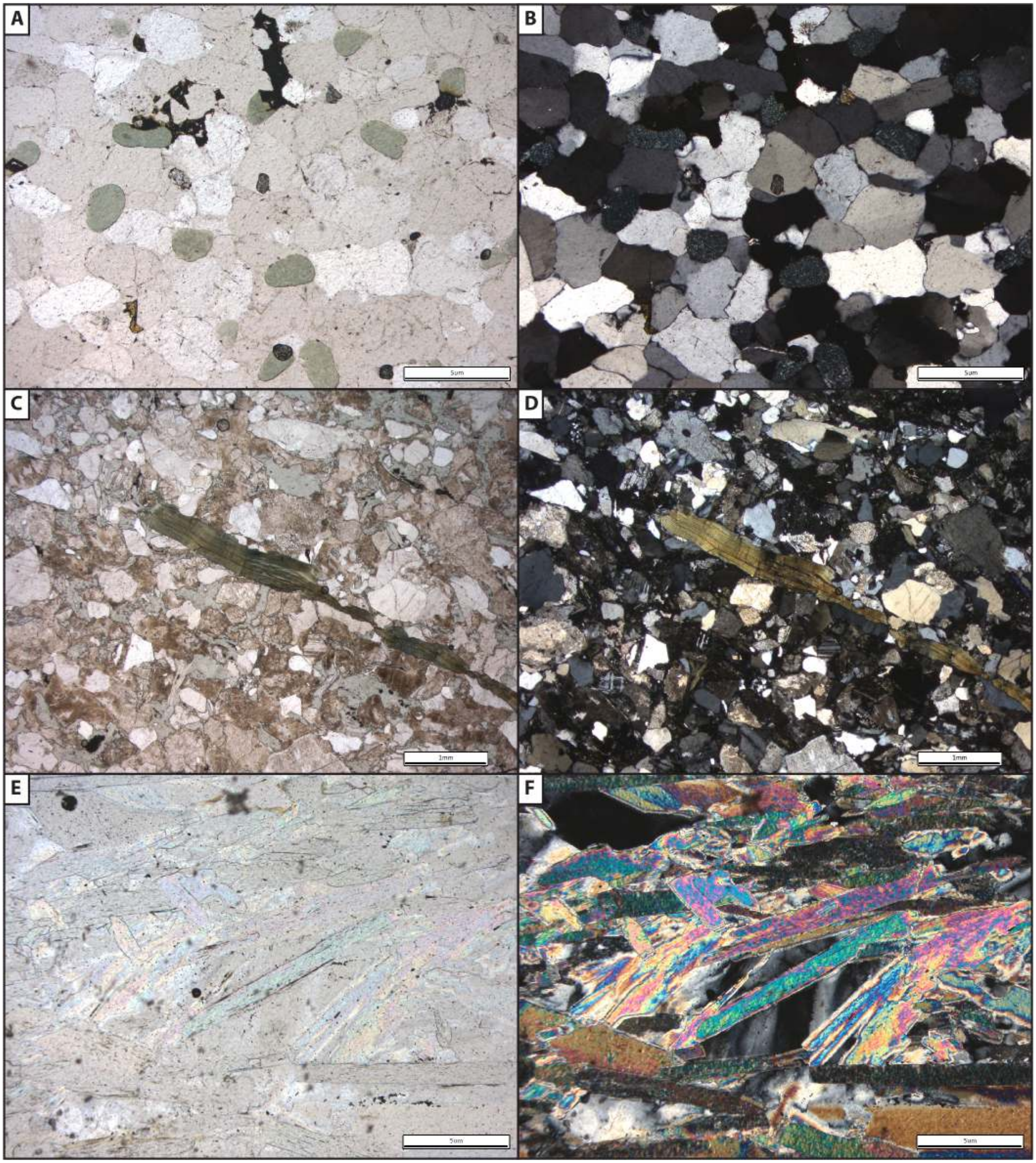


Figure 5.1.6a: Photomicrographs of accessory minerals, part 1; images in plane polarized light are on the left and cross polarized light on the right (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A & B) Green glauconite grains and quartz grains from the Flathead Formation (Cambrian), SW Montana. C & D) Large biotite flake surrounded by quartz grains, feldspar grains, and matrix that has been metamorphosed to chlorite. E & F) Muscovite mica (long bladed crystals) with small amounts of quartz.

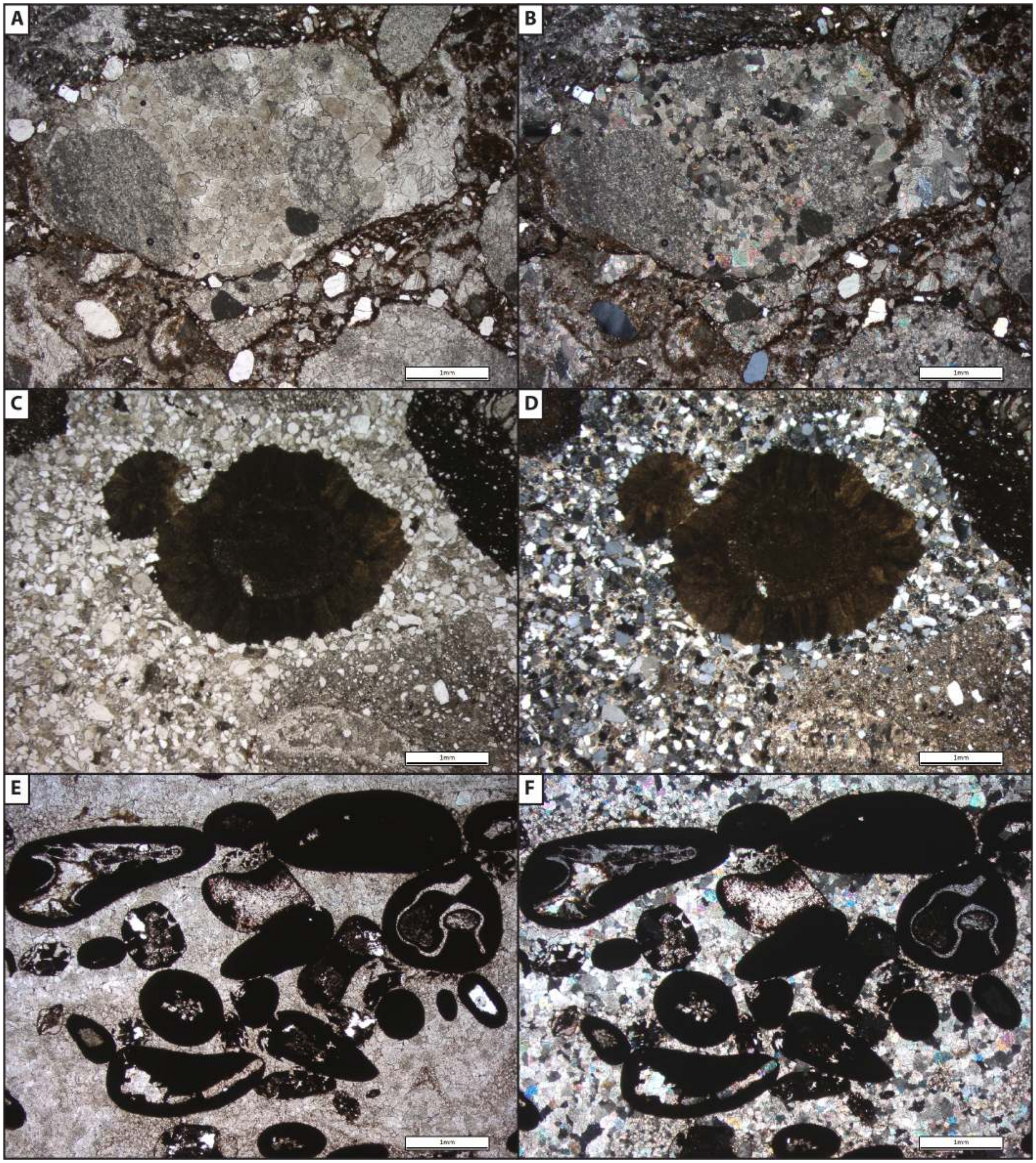


Figure 5.1.6b: Photomicrographs of accessory minerals, part 2; images in plane polarized light are on the left and cross polarized light on the right (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A & B) Large dolomite grains and granules in a muddy breccia. C & D) Siderite (probably sphaerosiderite) grains/granules and lithic grains/granules in a lithic sandstone. E & F) Ooids replaced with opaque hematite and surrounded by recrystallized calcite.

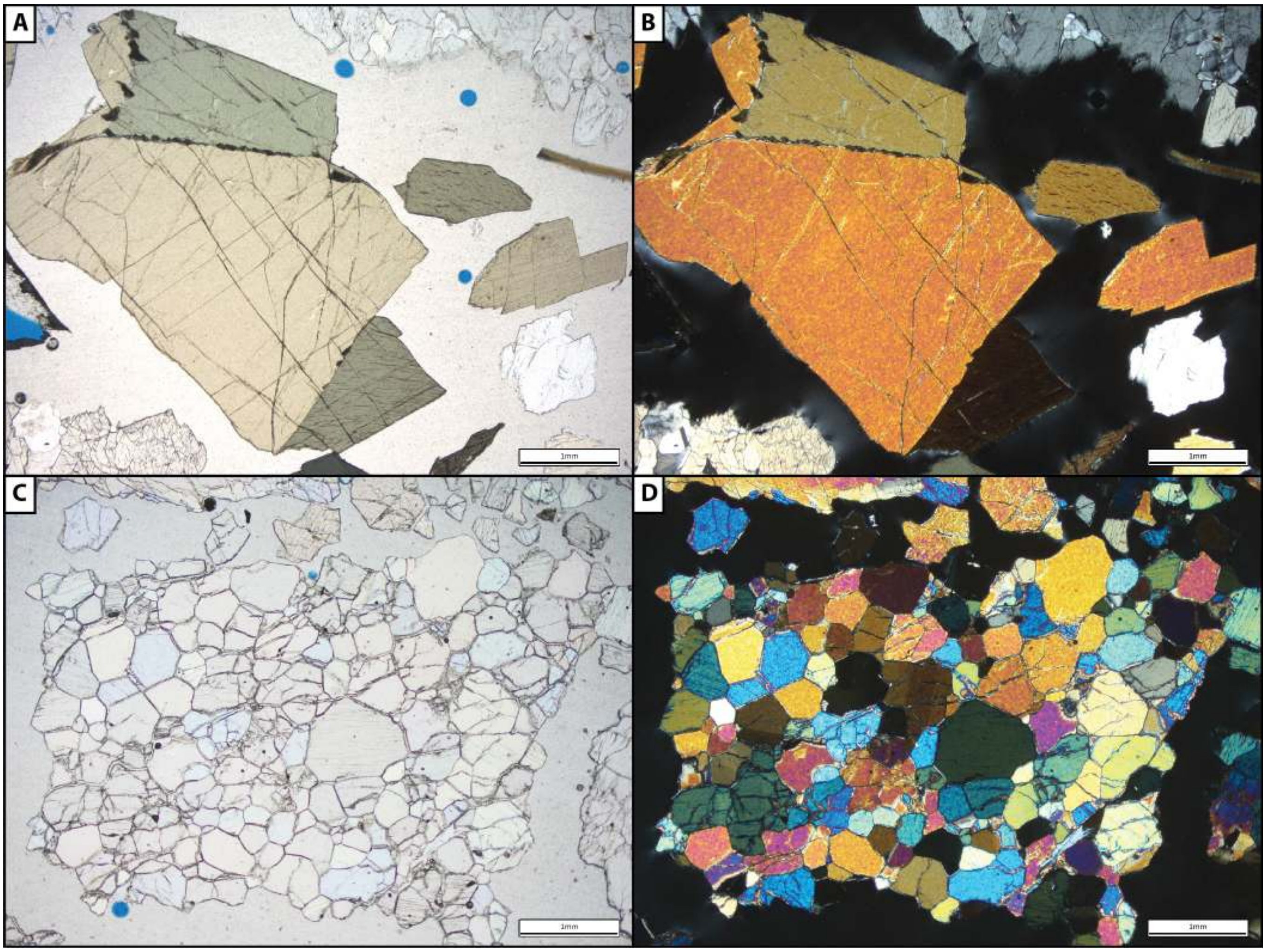


Figure 5.1.6c: Photomicrographs of accessory minerals, part 3 - mafic mineral in igneous rocks; images in plane polarized light are on the left and cross polarized light on the right (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A & B) In plane-polarized light, hornblende is green to brown, shows pleochrosim, and commonly cleavage. Its shows 2nd to 4th order orange colors under crossed polars C & D) In plane-polarized light, olivine is clear to light green, shows high relief, a lack of cleavage, and fracture. Under crossed polars, it shows a variety of bright 2nd to 3rd order interference colors.

5.1.3: Interstitial Material

5.1.3.1: Matrix

Matrix is fine-grained, clayey material between the framework grains in some sandstones. The origin of matrix is not well understood; possibilities for its formation include deposition of mud in protected areas between grains, deposition as thin layers that are later homogenized, larger particles of flocculated clay, or alteration products derived from unstable grains (pseudomatrix). Regardless of its exact origin, matrix (or its precursor) forms at the time of deposition. Matrix-rich sandstones are commonly dark colored in hand sample; matrix appears as dark, semi-translucent, amorphous material between sand grains in PPL and as dark material with a fine “woven” texture that may have high-order birefringence colors in XPL.

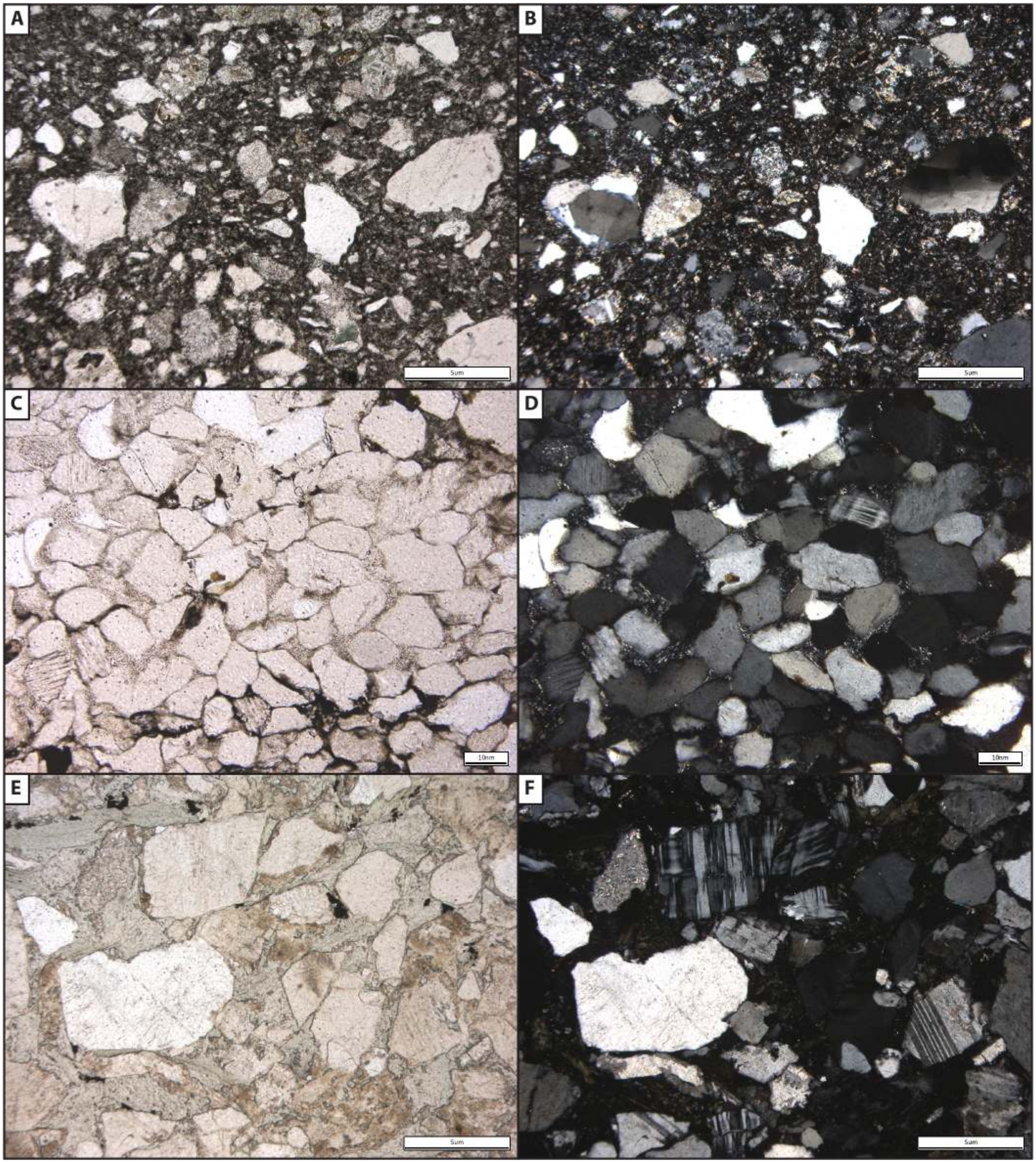


Figure 5.1.7: Photomicrographs of matrix in sandstones; images in plane polarized light are on the left and cross polarized light on the right (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A-D) Matrix appears dark in PPL and can have a woven appearance with crossed polars. E & F) Matrix has transformed to chlorite in a sandstone that has experienced low-grade metamorphism.

5.1.3.2: Pore space

If not clogged with muddy matrix, the interstitial areas between sand grains (pore spaces) are voids filled with water or air. Although sand spherical grains organized into a cubic stacking pattern has a theoretical maximum of 47.6% pore space, actual porosity values of 5-25% are more typical. During the manufacture of thin sections, pore space is commonly filled with epoxy which appears clear or blue (if dyed) in PPL and black in XPL.

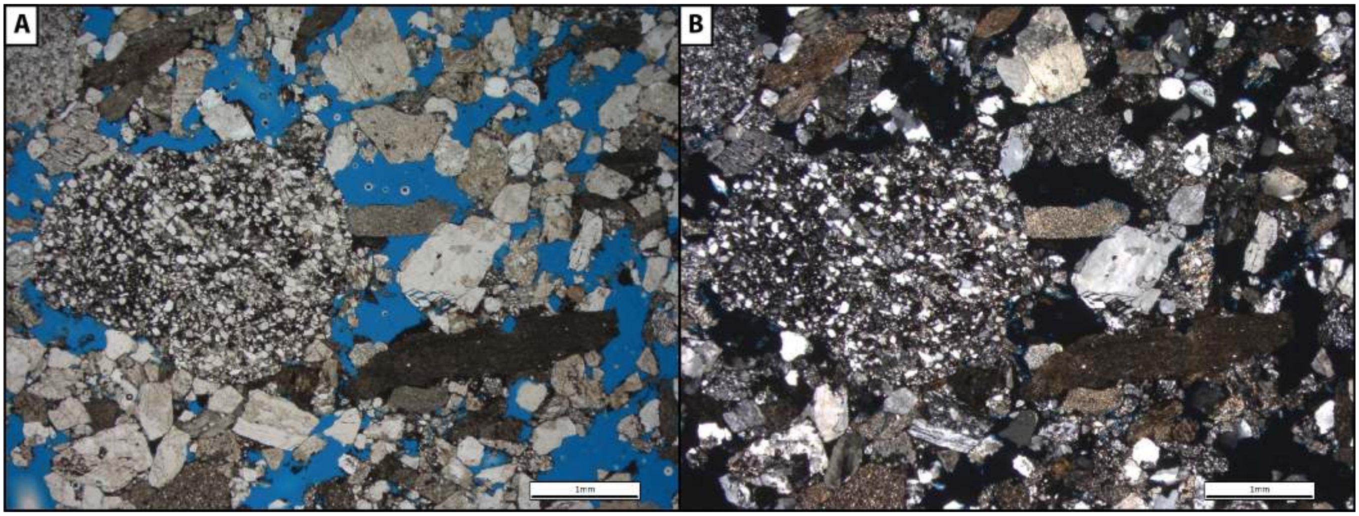


Figure 5.1.8: Photomicrographs of pore space that was filled with blue epoxy via vacuum impregnation when the thin section was made; images in plane polarized light are on the left and cross polarized light on the right (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

5.1.3.3: Cement

During diagenesis, material dissolved in groundwater commonly precipitate into pore spaces forming mineral cements that bind the sandstone together. The two most common cements in sandstones are silica and calcite. Silica cement commonly nucleates on the surface of quartz grains; in PPL dust rims may delineate the edge of the grains; the silica commonly grows in optical continuity with the grain and may go extinct with it as the microscope stage is rotated in XPL. Although calcite cement appears similar to silica in PPL, it is easily distinguished by high order interference colors in XPL. Beyond these common cements, some sandstones may be cemented with pyrite, hematite, gypsum, kaolinite, illite, or a variety of other minerals.

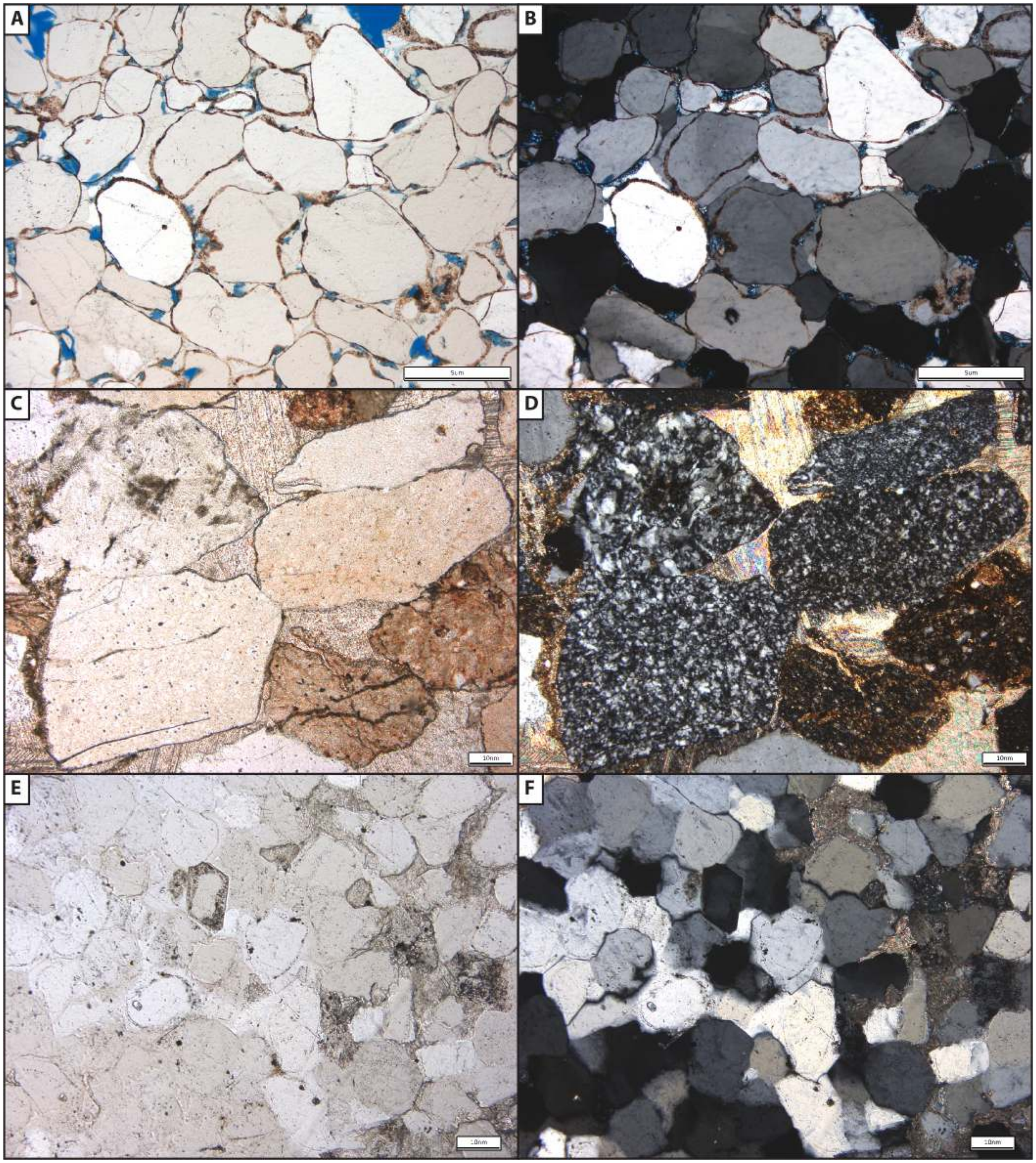


Figure 5.1.9: Photomicrographs of silica and calcite cement; images in plane polarized light are on the left and cross polarized light on the right (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A & B) Silica cement that grew in optical continuity with quartz grains; grain boundaries are highlighted by hematite dust rims. C & D) Coarse calcite cement between chert grains. Note the high order birefringence colors. E & F) A quartz arenite sandstone with silica cement predominantly filling interstitial areas on the left and calcite cement largely restricted to the right.

5.1.4: Classification

Older literature and many introductory textbooks recognize three main types of sandstone: quartz sandstone (quartz-rich), arkose (feldspar-rich) sandstone, and greywacke (muddy) sandstone. Although useful when applied to the type of endmember sandstones included in rock and mineral kits, these terms are ambiguous and do not capture much of the variability present in the field.

Dott (1964) proposed a simple two-word naming scheme where the first word (quartz, feldspathic, or lithic) represents the relative abundance of quartz, feldspar, and lithic framework grains and the second word represents the abundance of interstitial matrix (arenite if <10% matrix or wacke if >10% matrix). This scheme allows for a name that speaks to the compositional maturity (framework grain composition) and textural maturity (amount of mud) of the rock.

A few things to consider when classifying sandstones using this scheme include:

- Only quartz, feldspar, and lithic framework grains are used for naming; plotting composition on the ternary diagram requires that their relative abundances be normalized to 100%.
- Cement is a diagenetic feature and not used to name sandstones.
- The amount of matrix is estimated for the entire volume of the rock. This represents the entire field of view in thin section. In hand sample, wackes are distinguished by the presence of dark, fine-grained material rather than glassy or shiny cement.
- Although proper classification requires point counting of ~300 sand grains in thin section, quick visual estimation using a hand lens or thin section usually produces satisfactory results.

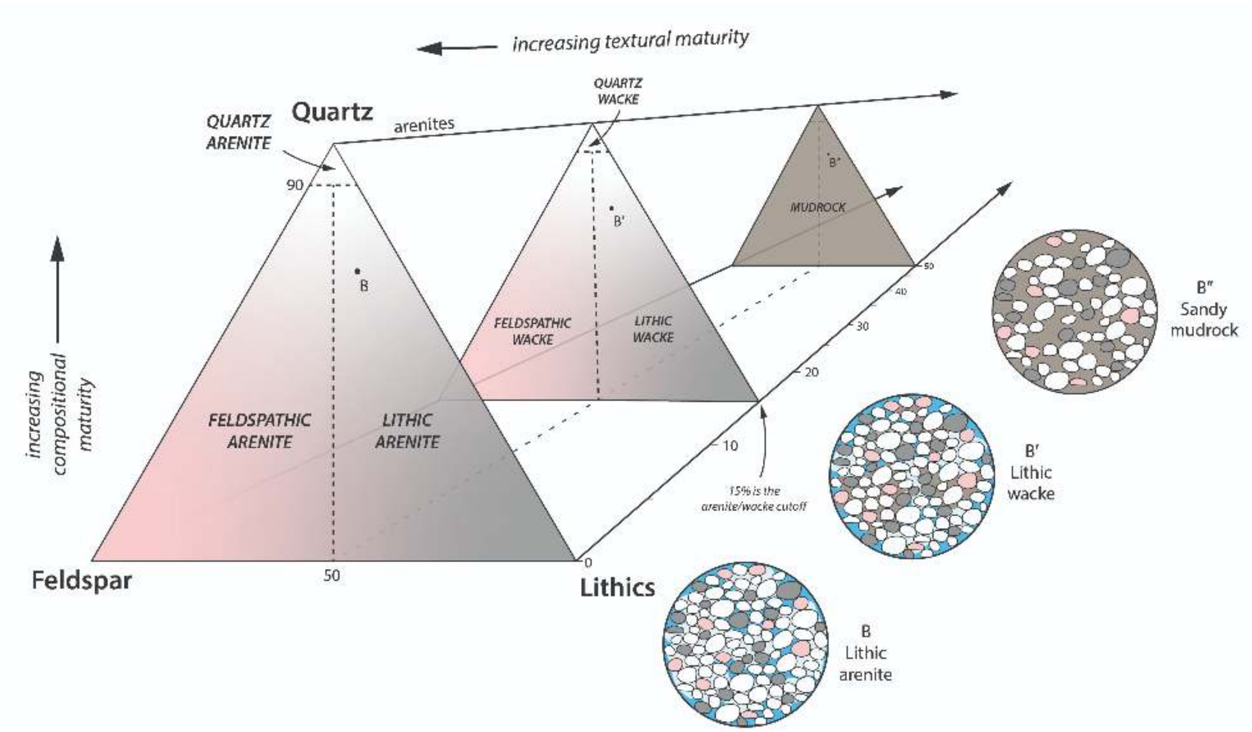


Figure 5.1.10: A three dimensional diagram for the classification of sandstones. The triangle is a ternary diagram showing the relative abundance of quartz, feldspar, and lithics. The third dimensions shows the relative amount of interstitial matrix (from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). Modified from Dott (1964).

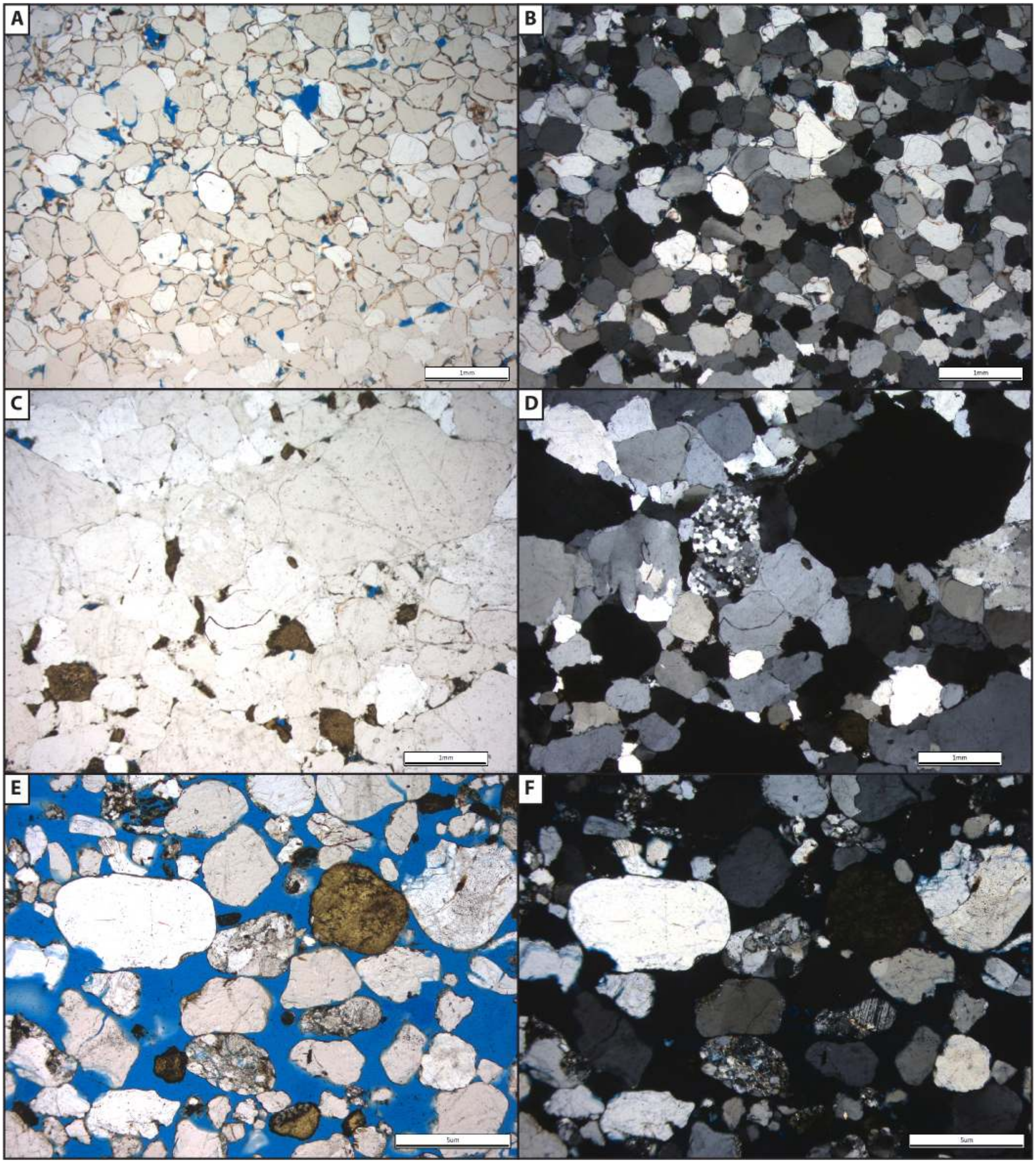


Figure 5.1.11: Photomicrographs of arenite sandstones with little/no interstitial matrix; images in plane polarized light are on the left and cross polarized light on the right (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A & B) show a quartz arenite with silica cement and modest amounts of pore space. C & D) show a silica-cemented feldspathic arenite; note that this slide was stained for potassium feldspar. E & F) show a poorly cemented and friable lithic arenite.

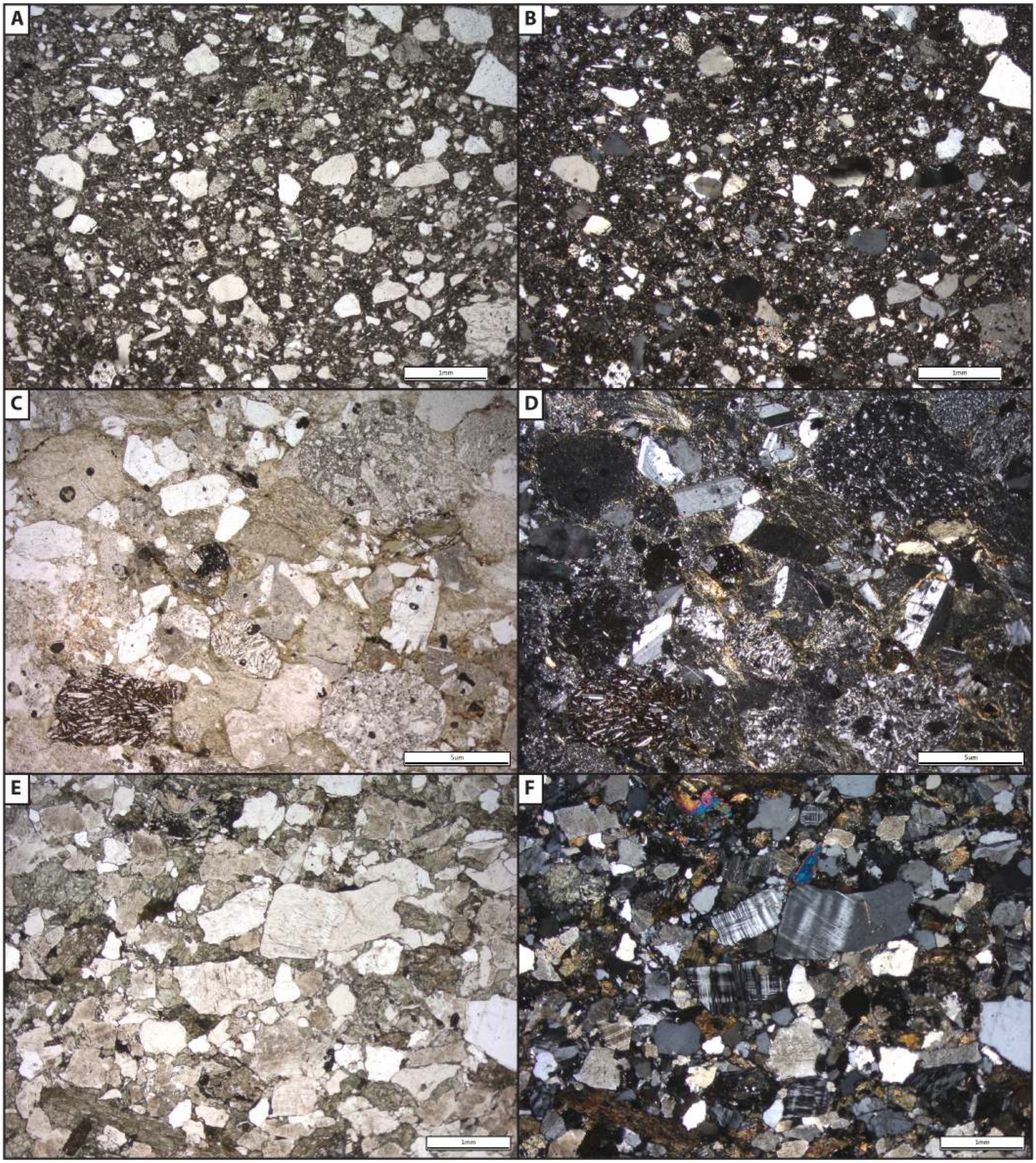


Figure 5.1.12: Photomicrographs of wacke sandstones with significant amounts of matrix; images in plane polarized light are on the left and cross polarized light on the right (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A & B) show a quartz wacke with a modest number of lithic grains. C & D) show a lithic wacke with mainly igneous lithic grains, E & F) show a feldspathic wacke with matrix that has been transformed to chlorite.

5.1.5: Readings and Resources

- Short video about stains for feldspars and carbonate minerals: https://www.youtube.com/watch?v=qIGWga1ElKE
- Discussion about provenance, plate tectonics, and what to do with polycrystalline quartz and chert: https://www.geological-digressions.com/provenance-and-plate-tectonics
- Michel-Lévy interference colour chart issued by Zeiss Microscopy: https://en.wikipedia.org/wiki/Interference_colour_chart#/media/File:Michel-L%C3%A9vy_interference_colour_chart_(21257606712).jpg
- Introduction to Petrology OER textbook with great explanations and videos about thin sections and the petrographic microscope - https://viva.pressbooks.pub/petrology/front-matter/table-of-content/
- Dott, R.H., 1964, Wacke, Graywacke and Matrix - What Approach to Immature Sandstone Classification?, Journal of Sedimentary Petrology, v. 34, p. 625-632
- Thin section analysis of oil reservoirs and source rocks: https://wiki.aapg.org/Thin_section_analysis

5.2: Conglomerates and Breccias

Conglomerates (rounded clasts) and breccias (angular clasts) are composed of >50% gravel-sized particles (>2 mm). Although they are the most coarse-grained clastic rocks, they are relatively rare in the rock record because they are derived from tectonically active areas and have relatively little preservation potential unless transported into the subsiding part of a basin.

5.2.1: Composition

For our purposes, we will consider conglomerates and breccias to have more than 50% clasts. When describing these rocks, clasts are the larger particles that make up the rock; they are >2 mm diameter and are analogous to the framework grains in a sandstone. The smaller particles between the clasts are called matrix; they are <2 mm diameter and are analogous to the muddy matrix in a sandstone. In terms of describing these rocks, we can consider the following characteristics:

5.2.1.1: Clast composition

Extraformational clasts are composed of fragments of sedimentary, igneous, and/or metamorphic rocks that formed outside of the sedimentary basin. We use the term oligomict to describe a conglomerate/breccia that is composed of one or a very few types of lithic clasts and the term polymict to refer to one that is composed of many types of lithic clasts.

Intraformational clasts are composed of partially lithified clasts that formed within the sedimentary basin. Common examples of intraformational clasts include caliche nodules, siderite nodules, mud chip rip-up clasts, bones, scales, or a variety of other materials.

5.2.1.2: Matrix composition

The matrix of a conglomerate/breccia can essentially be described in the way that you would describe a sandstone (they will be overwhelmingly composed of sand and gravel). Properties like composition, sorting, cement, etc. Because mud beyond what one would expect as clayey matrix is difficult to explain from a hydrodynamic perspective, rocks with a mixture of sand, mud, and gravel are termed diamictites (discussed later on in this chapter).

5.2.2: Support and Fabric

Clast-supported conglomerates are exactly that - the clasts generally touch one another, they typically contain less than 15% matrix, and if the matrix was removed the clasts would not collapse or significantly shift. Matrix-supported conglomerates typically contain 15-50% matrix, which means that the clasts appear to be "floating" within the matrix and would collapse if the matrix was removed. Some conglomerates and breccias have clasts that are imbricated; they were aligned by flowing water and have "shingled" appearance with the long axis dipping in an upstream direction.

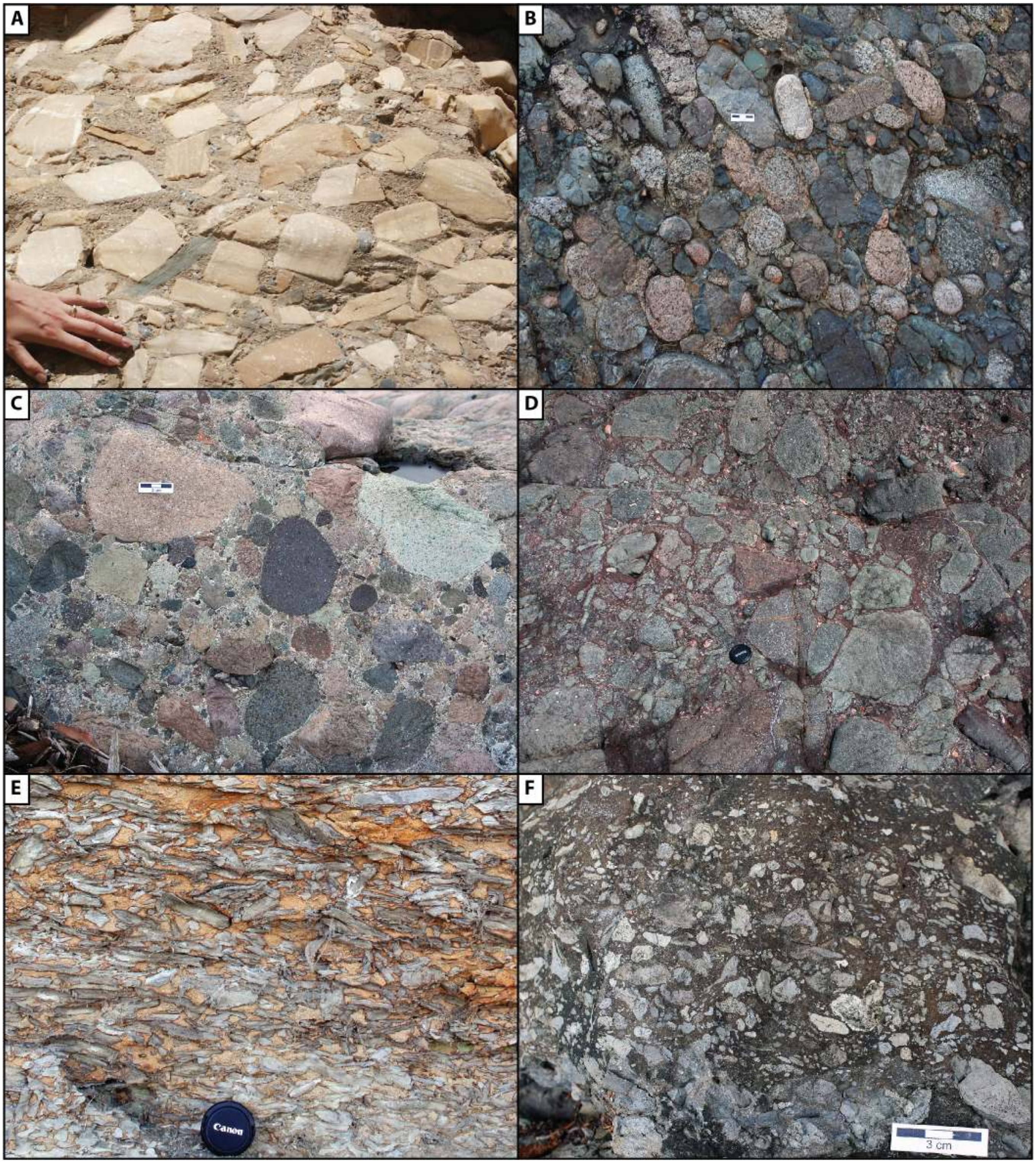


Figure 5.2.1a: Characteristics of conglomerates and breccias, part 1 (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A) Breccia with angular clasts and B) conglomerate with rounded clasts. C) Extraformational polymict conglomerate with a variety of clast lithologies. D) Oligomict breccia with clasts composed almost entirely of basalt clasts derived from the nearby (~25 m) basin margin. E & F) Intraformational breccias composed of mud-chip rip clasts and poorly lithified carbonate, respectively.

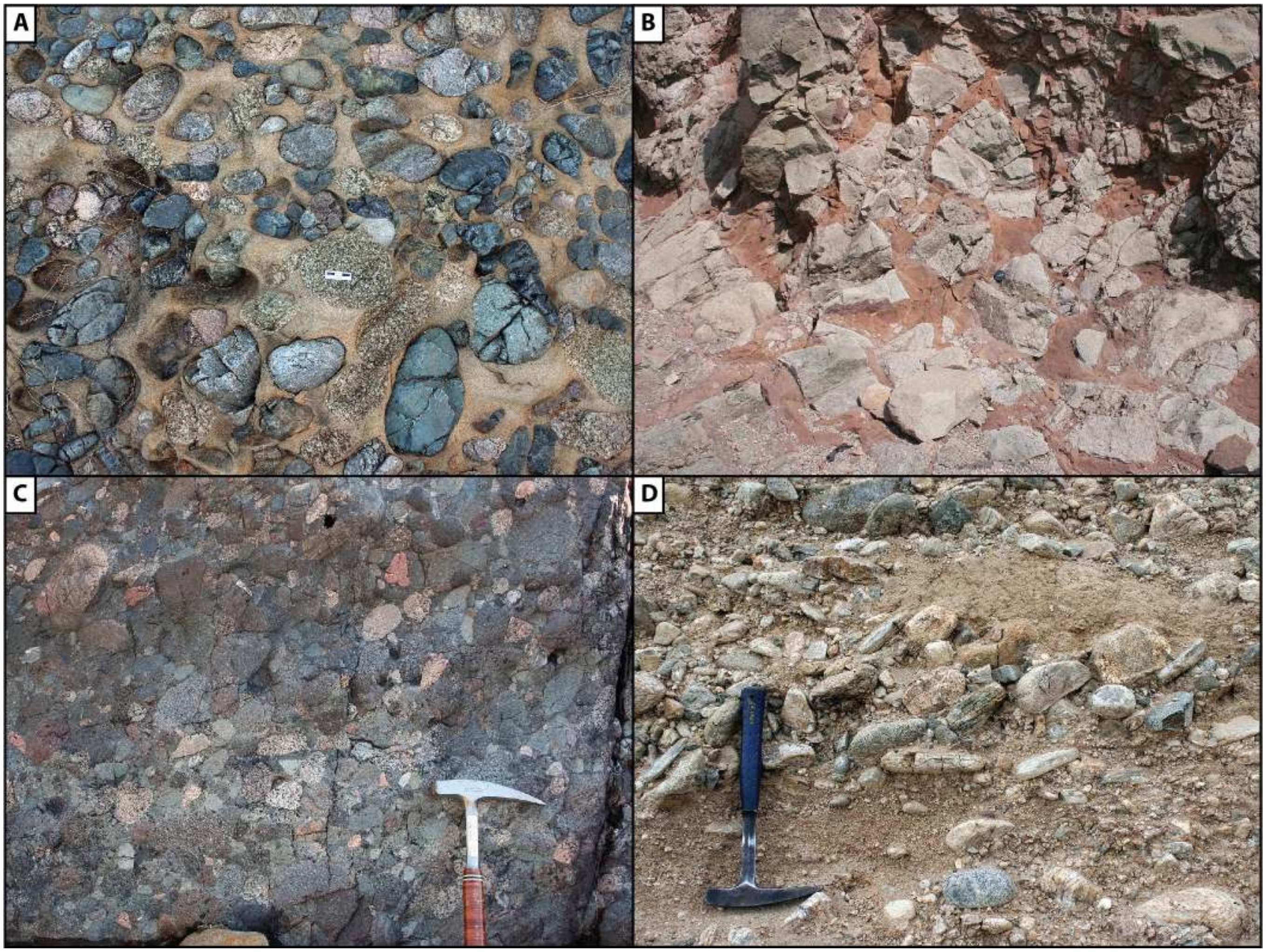


Figure 5.2.1b: Characteristics of conglomerates and breccias, part 2 (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A) Conglomerate with a sandstone matrix that could be described using the descriptive terminology and techniques discussed in the previous chapter. B) Matrix-supported breccia. These basalt clasts tumbled from the basin margin into eolian sand dunes. C) Clast-supported polymict conglomerate. D) Conglomerate with imbricated clasts that record paleoflow from left to right.

5.2.3:

5.2.4: Readings and Resources

5.3: Mudrocks

By definition, mudrocks contain more than 50% mud-sized particles (silt and clay). They are the most abundant sedimentary rocks because the sediment typically is deposited in calm environments relatively far from tectonically-active areas. Although they commonly contain abundant organic material and are important source rocks in petroleum systems, they are understudied because they weather quickly and are generally not well exposed at the surface and because specialized techniques and equipment are needed to fully understand the fine-grained particles that make up these rocks.

5.3.1: Mudrock Classification

Unfortunately, there is no one universally accepted classification scheme for mudrocks. We use the term mudrock to refer to all rocks containing >50% clay and silt. If further subdivision is required, we use the terms siltstone (<33% clay), mudstone (33-67% clay), and claystone (>67% clay).

We use the term shale to describe a mudrock that has pronounced fissility (breaks into sheets); this takes the place of terms like mudrock, siltstone, mudstone, and claystone. Modifiers like sandy, pebbly, etc. are useful in some situations.

5.3.2: Important Properties

5.3.2.1: Stratification and Weathering

The outcrop appearance of a mudrock reflects some combination of resistance to weathering and nature of primary stratification. Adjectives like shale, platy (flattened), blocky (equant), hackly (equant with sharp edges), etc. are very useful when describing mudrocks in outcrop.

5.3.2.2: Color

If its primary and reflects the characteristics of the sediment at the time of deposition, the color of a mudrock can tell your something about the organic content and amount of oxygen in the environment where the sediment was deposited. Red mudrocks reflect very little organic material and abundant oxygen in the system at the time of deposition; many are deposited in terrestrial environments. Green/gray color reflects reducing conditions with a modest amount of organic material; these conditions exist in a variety of environments. Black mudrocks have abundant organic material and record anoxic conditions at the time of deposition; these sorts of conditions can occur in a variety of environments from oxbow lakes on a floodplain to deep marine sediments.

Mudrocks can form anywhere there are calm waters. Understanding how and where they formed is facilitated by comparing the physical characteristics described above with characteristics like context, fossil content, color, and detailed analysis of mineralogy and geochemistry (especially of organic components).

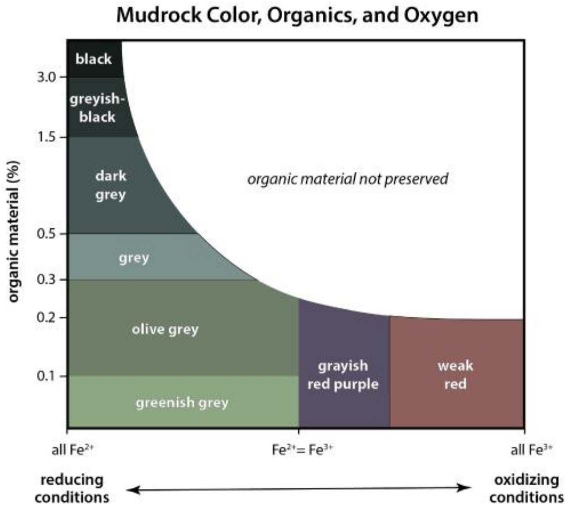


Figure 5.3.1: Diagram showing the relationship between mudrock color, organic content, and oxidation state of iron. Colors and names are consistent with the Munsell Rock Color Chart. Diagram is after Potter et al. (1980). From Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0.

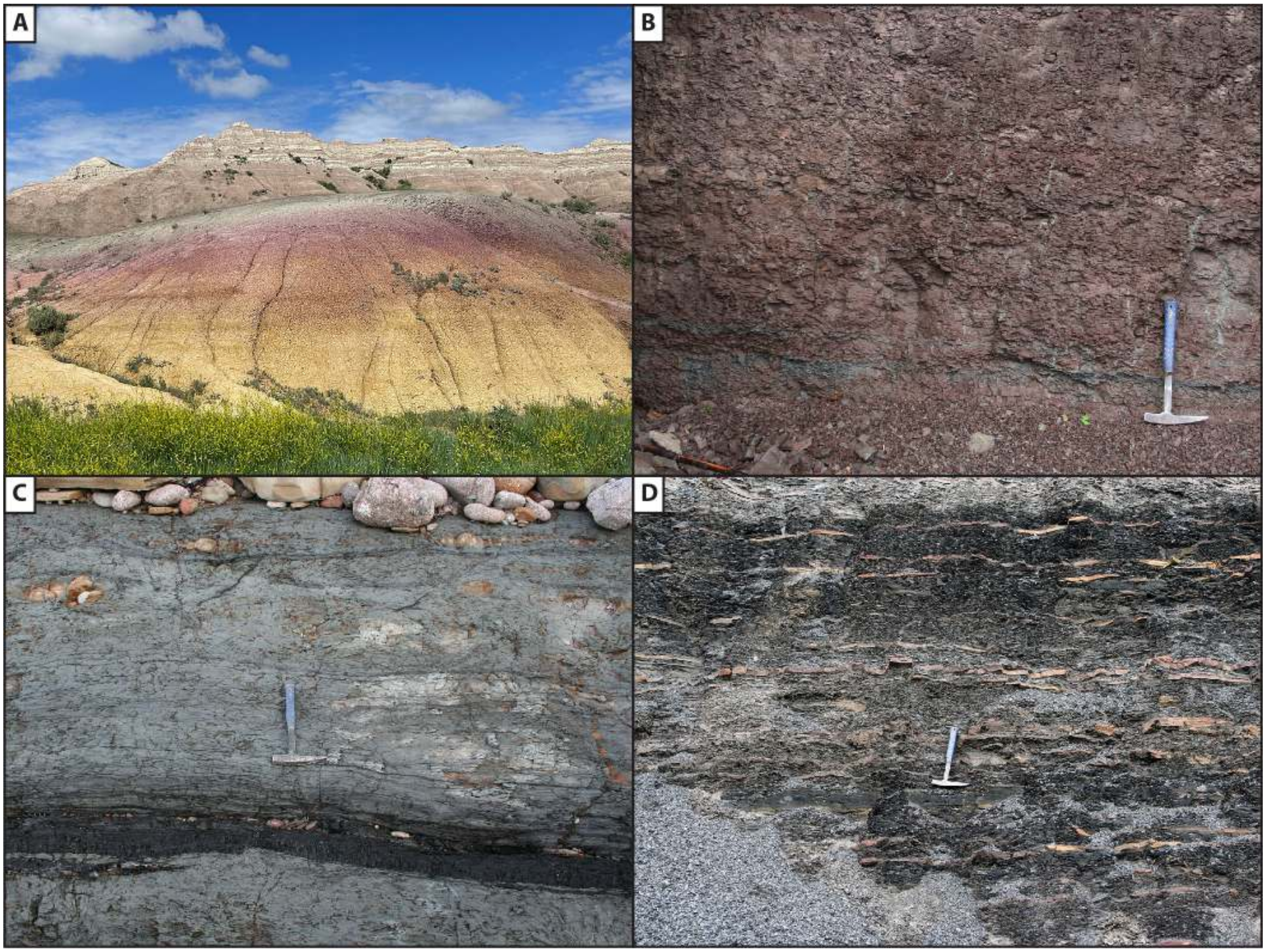


Figure 5.3.2: Mudrock colors and textures. A) Bright yellow, red, and/or purple colors are commonly associated with paleosols (James St. John via Wikimedia Commons; CC BY 2.0). B) The blocky texture and red/green colors of this mudrock suggest that it is a poorly developed paleosol that formed in mostly oxidizing conditions. C) Internally massive greenish gray mudrock with siderite nodules and associated coal; together these features suggest deposition in a reducing swampy environment. D) Platy dark gray to black shales with minor siltstone and sandstones. These rocks likely formed in an organic-rich standing body of water. B-D from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0.

5.3.3: Components

Mudrocks are composted of silt-sized particles that are typically composed of quartz, feldspar, or other common minerals … some might even be large enough to classify using the petrographic microscope.

Although the terminology is confusing, many of the clay-sized particles (<1/256 mm) in a mudrock happen to be "clay minerals" which is a name applied to a specific group of aluminosilicate and phyllosilicates minerals. These clay minerals are derived from weathering (hydrolysis) of feldspar and mica.

5.3.3.1: Mineralogy

Clay minerals are made of alternating layers of silicon-oxygen tetrahedra (tetrahedral sheets) and alumina-oxygen octahedra (octahedral sheets). These sheets can be layered in a variety of ways and may even alternate with interlayer cations or water. Clay minerals can be grouped into two main families: 1:1 clay minerals and 2:1 clay minerals.

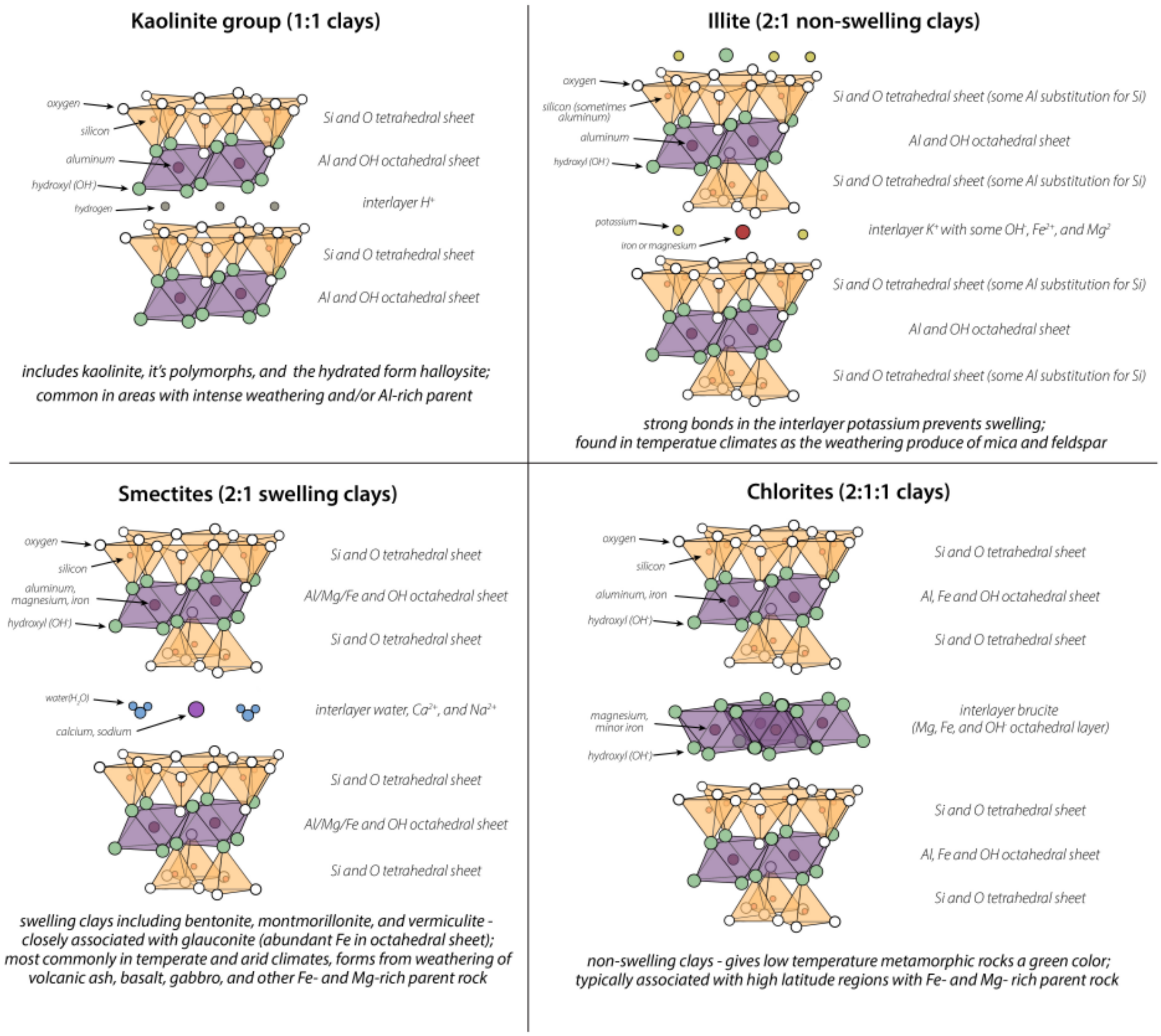


Figure 5.3.3: Overview of the major types of clay minerals in sedimentary rocks (Page Quinton via Wikimedia Commons; *CC BY-SA 4.0*). Diagram after Kumari and Mohan (2021).

1:1 clay minerals are the product of more intense/prolonged weathering that progressively strips interlayer cations from 2:1 clay minerals which concentrates Si, Fe, and Al in a simpler structure. Their fundamental building block is a single octahedral layer bonded to a single tetrahedral layer. Kaolinite is probably the best known and most important 1:1 clay mineral because of its close association with alumina- and/or iron-rich soils (bauxites and laterites, respectively)

The fundamental building block of 2:1 clay minerals is an octahedral (O) sheet sandwiched between two tetrahedral (T) sheets. Between these T-O-T packages, variable amounts of Al, Mg, Fe, water, or other interlayer cations determine the mineralogy of the specimen. These relatively complex clay minerals tend to form in temperate areas. Noteworthy 2:1 clay minerals include:

- "Swelling clays" like montmorillonite and vermiculite are 2:1 clays that can retain significant amounts of water
- Glauconite is a green, iron-rich clay mineral that forms in reducing seafloor conditions where smectitic clays in existing materials (fecal pellets, fossils, volcanic debris, and various detrital minerals) uptake potassium and iron from the seawater. Glauconization is enhanced in environments with abundant organics and a slow sedimentation rate.
- Bentonite is a 2:1 clay formed from weathered volcanic ash beds by alkaline & acidic water, respectively. Given their volcanic origin, dateable minerals like zircon are commonly found with bentonites.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B
C
D
No. 200
VOLCLAY
HIGHEST QUALITY
WESTERN
BENTONITE
PRODUCED SI CE 1927
AMERICAN COLLOID COMPANY
E
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Figure 5.3.4: Noteworthy clay minerals. A) Granule-sized Silurian glauconite pellets from Ohio (James Cheshire via Wikimedia Commons; public domain) B) Black shales and bentonites from the Cretaceous Benton Shale (James St. John via Wikimedia Commons; CC BY 2.0). C) White pellets of vermiculite are commonly added to potting soil to help retain water (M Tullottes via Wikimedia Commons; public domain). D) Bag of bentonite - a swelling clay that is commonly used to seal well casings (Corey872 via Wikimedia Commons; public domain). E) Kaolinite and, to a lesser degree, montmorillonite and bentonite, are commonly used in pottery clay (Knecht03 via Wikimedia Commons, CC BY SA 3.0). Laterites are soils that are enriched in kaolinite and iron via intense tropical weathering; they are commonly used as brick and road bed material (Vinayaraj via Wikimedia Commons; CC BY-SA 4.0).

5.3.4: Readings and Resources

- Kumari and Mohan, 2021, Basics of Clay Minerals and Their Characteristic Properties
- Potter et al., 1980, Sedimentology of Shale, Springer-Verlag, New York, 306 p., ISBN 0-387-90430-1

5.4: Diamictites, Pebbly Sandstones, and Outsized Clasts

5.4.1: Diamictites

Diamictites are poorly-sorted, generally matrix-supported conglomerates/breccias with 15-50% matrix that is a mixture of sand and mud. The presence of mud is important because if purely a product of bedload transport, the silt and clay grains should have been swept downstream by a current that was moving fast enough to transport gravel. Diamictites are important because they can only form in one of three ways: debris flows (slurry of sediment and water), glacial till (transport by viscous glacial ice), or bioturbation (biological mixing of discrete beds of mud/sand/gravel).

5.4.2: Pebbly Sandstones and Sandstones/Mudrocks with Outsized Clasts

Although really a subset of mudrocks, we will discuss pebbly sandstones and mudstones here because our eyes are typically drawn to the coarse-grained fraction of the rock.

Pebbly sandstones typically occur in situations where flows are occasionally fast enough to transport gravel and winnow away sand but then slow to allow for the deposition of sand atop and between the lags of coarser-grained materials. These deposits are common along the base of channels or along scours or other erosional surfaces within sandstones. If the clasts are intraformational, its likely that they were locally derived and that erosion incorporated robust clasts derived from adjacent un- or partially-lithified sediments (ex: a river eroding its banks).

Video 5.4.1: Video of coarse sand transportation and the transport/deposition of pebbles within sandy sediment.

Mudstones and sandstones with exceptionally large clasts (outsized clasts) require a bit more explanation because, once again, there is a problem with hydrodynamics. Although bioturbation is possible, large, outsized clasts actually deposited within mudrocks or sandstones can typically only be attributed to ice rafting, rafting by tree roots, gastroliths derived from certain types of vertebrates or volcanic bombs. Context and close examination of the clasts might reveal which of the four options is most likely.

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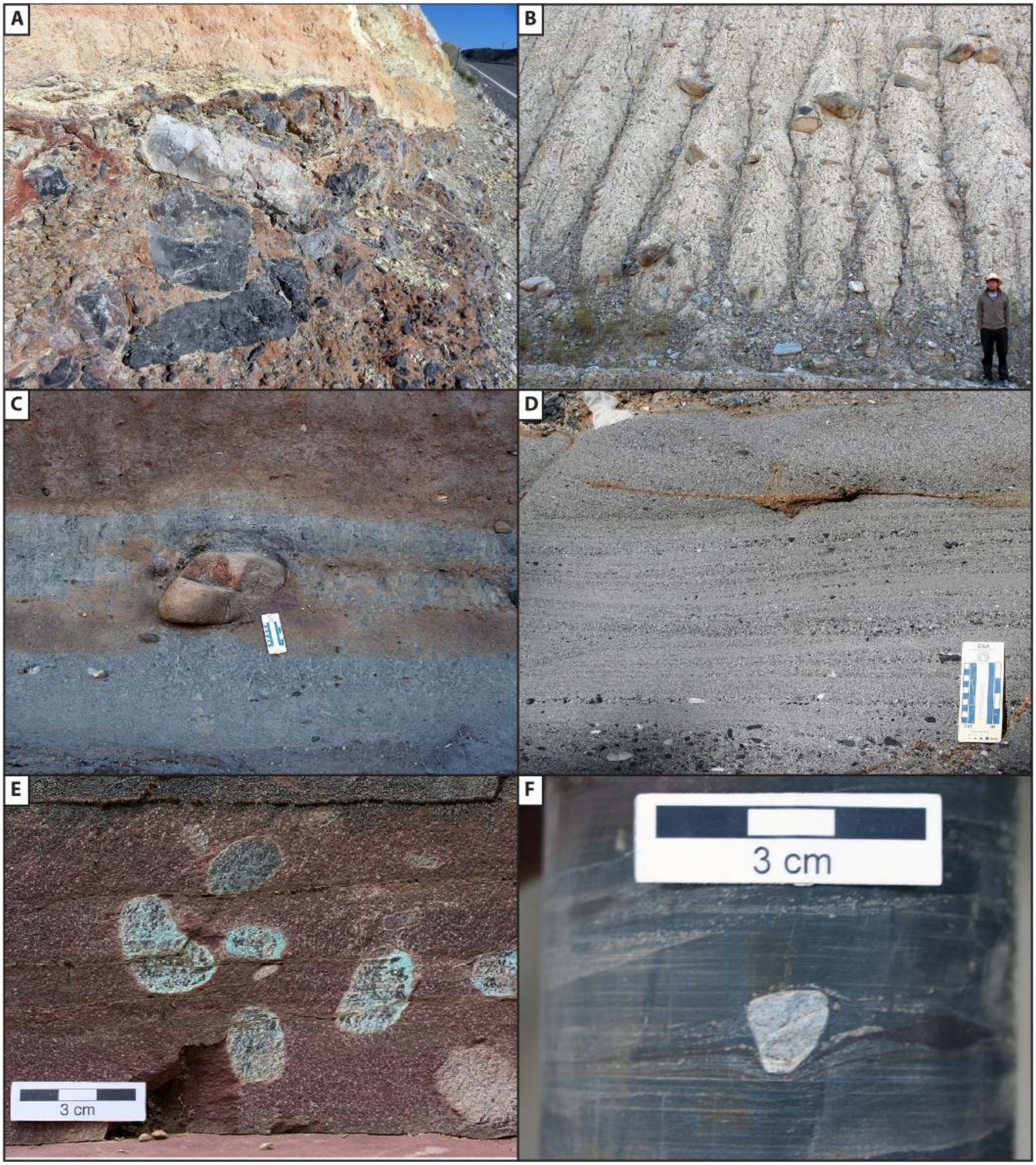
Figure 5.4.1: Photographs of diamictites, pebbly sandstones and mudstones. A) Tertiary debris flow deposit, California (Wilson44691 via Wikimedia Commons; public domain). B) Poorly-sorted glacial till outside of Whitefish, Montana. C) Former layers of mud, sand, and gravel have been completely homogenized to form this Permian diamictite. D) A coarse-grained pebbly sandstone where gravel is hydrodynamically concentrated along bedding planes. E) Although these mineralized mud-chip rip-up clasts in a sandstone are much bigger than any other clasts in this formation, they were locally derived from underlying mudrocks and don't require any special explanations. F) Glacial dropstone in a deep-marine black shale. Images B-F from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0 or CC BY-SA 4.0.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CHAPTER OVERVIEW

6: Carbonate Sedimentary Rocks

Carbonate sedimentary rocks make up approximately 22% of the sedimentary record and are second in abundance to mudrocks. They are of great economic importance because they serve as both source and reservoir rocks in petroleum systems and are used in everything from metallurgy to cement. They are also of environmental importance because of their ability to buffer acidic water, association with karst topography, and influence on soil chemistry.

Learning Objectives

- List and identify the most important carbonate minerals.
- Explain the chemical reactions responsible for carbonate precipitation and how the amount of carbon dioxide in water influences carbonate precipitation.
- Explain how dolomite forms.
- Identify the main components of carbonate rocks and name the rocks using common classification schemes.

Chapter thumbnail shows details of ooids in the Silurian Mifflintown Formation, PA (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

6.1: Composition

6.1.1: Mineralogy

Carbonate sedimentary rocks are dominated by the Ca^{2+}, Mg^{2+} and CO_3^{2-} ions. Although there are many carbonate minerals, the four most important ones are:

Calcite ($CaCO_3$) is the more stable polymorph of calcium carbonate; its atoms are arranged into a rhombohedral architecture. Mg can substitute for up to 16% of Ca in calcite; specimens with <4% are referred to as low-Mg calcite and those with 4-16% are referred to as high-Mg calcite.

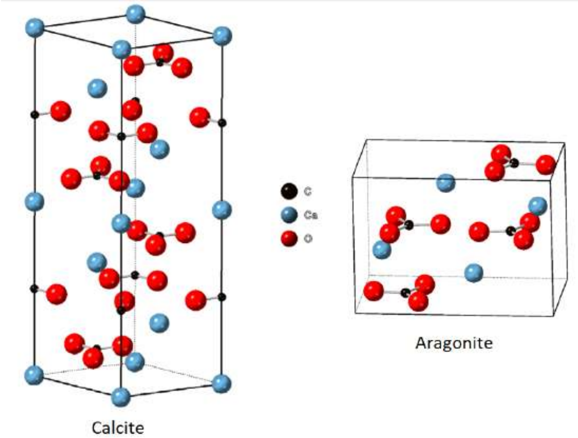

Figure 6.1.1: Atomic structure of calcite (rhombohedral) and aragonite (orthorhombic; Wikimedia Commons; CC BY-SA 4.0).

Under atmospheric temperatures and pressures, aragonite ($CaCO_3$) is a less stable polymorph of calcium carbonate and given time it will often turn into calcite. The atoms are arranged into an orthorhombic configuration. Many invertebrate shells that have the “mother of pearl” luster are composed of aragonite.

Dolomite ($CaMg(CO_3)_2$) has equal parts Ca and Mg and is thus a different mineral than high-Mg calcite which has a maximum of 16% Mg substituting in for Ca. Although extremely common in the geologic record, direct precipitation at the Earth’s surface is unlikely because it can only form at high temperature and/or under conditions where the Mg/Ca ratio greatly exceeds typical ocean chemistry (see Figure 6.3.1).

Siderite ($FeCO_3$) is common in reducing, freshwater conditions and is often nucleated around organic material that causes localized reduction. It is uncommon as an early formed mineral in marine environments because the iron will preferentially bond with the sulfur (and thus form pyrite) that is abundant in marine settings.

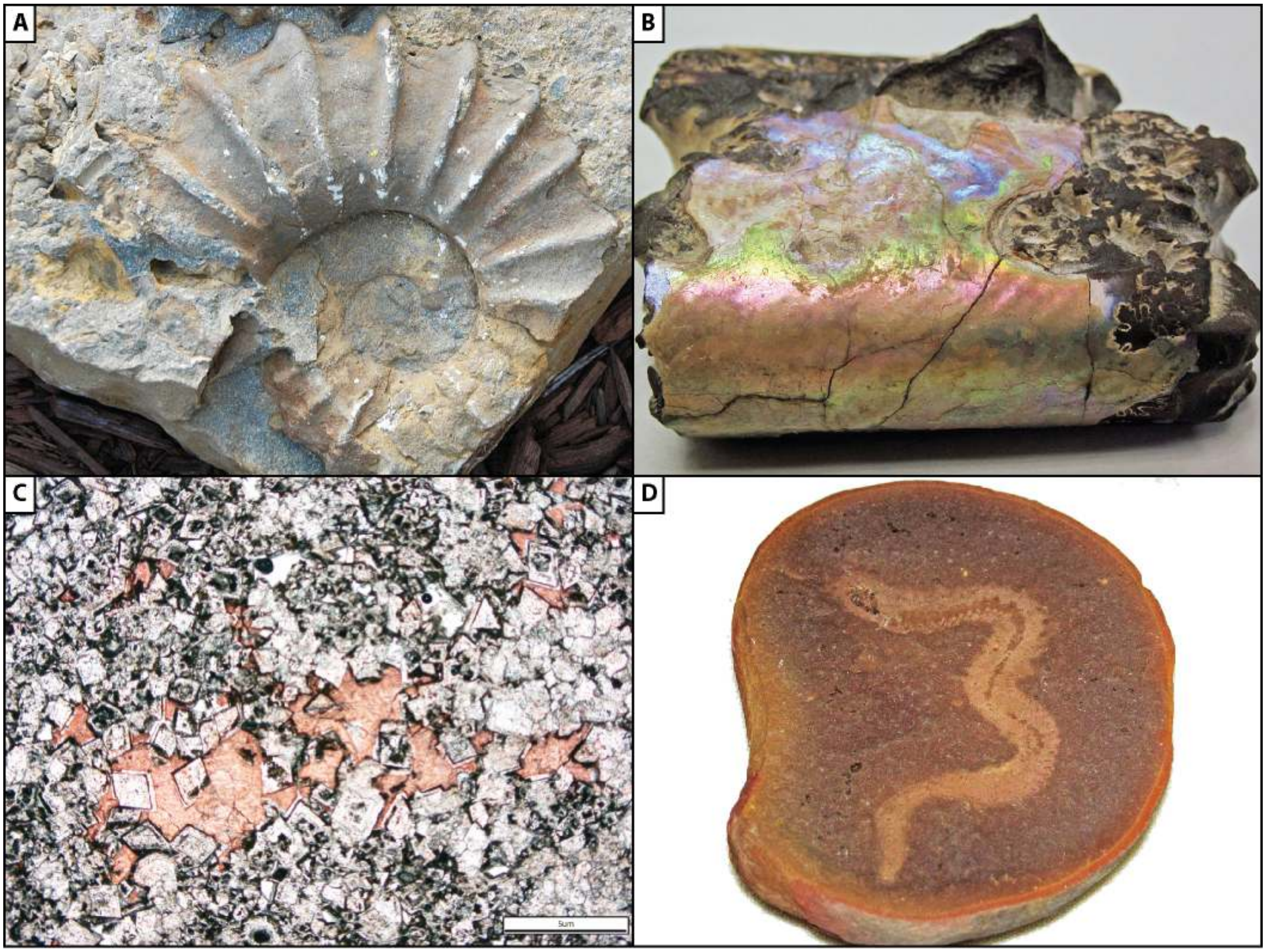

Figure 6.1.2: Carbonate minerals. A) This ammonite shell was probably originally composed of aragonite, but during diagenesis it was transformed into calcite which is more stable (James St. John via Wikimedia Commons; CC BY 2.0). B) Exceptional preservation allowed for pristine preservation of this nacreous aragonite ammonite shell (James St. John via Wikimedia Commons; CC BY 2.0). C) Photomicrograph of a carbonate (stained for calcite). showing rhombohedral dolomite crystals and stained (red) calcite crystals (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0) D) Polychaete worm *preserved in a siderite nodule (James St. John via Wikimedia Commons; CC BY 2.0)*

6.1.2: Calcite and aragonite seas

The crystal structure of aragonite does not allow Mg to substitute for Ca easily. Conversely, calcite allows Mg to readily substitute, but increased Mg concentrations make the calcite unstable. The end result is that when the ocean is relatively enriched in Mg (as is the case when there is rapid seafloor spreading), aragonite and high-Mg calcite become the most abundant types of inorganic carbonate in the oceans ("aragonite seas"). Conversely, "calcite seas" exist at times of low Mg concentration. However, most ancient aragonite transforms into calcite during diagenesis.

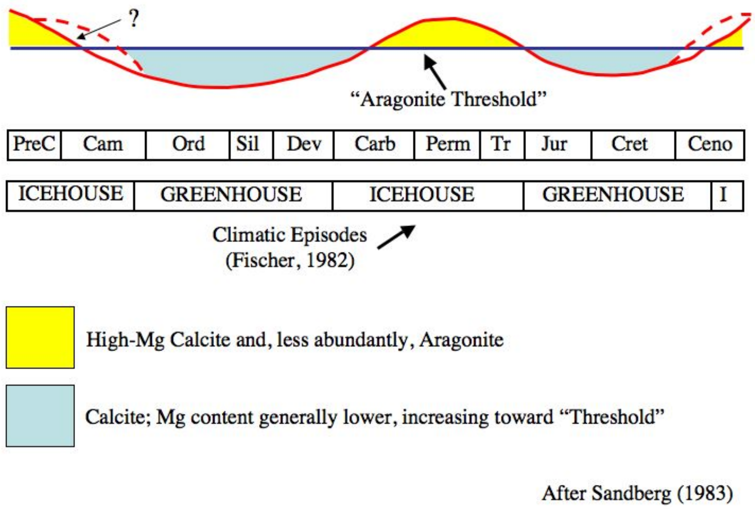

Figure 6.1.3: Alternation of calcite versus aragonite seas through geologic time (Wilson44691 *via Wikimedia Commons; public domain)*

6.1.3: Carbonate Rocks

Rocks that are made up of >50% calcite/aragonite are classified as limestones and those with >50% dolomite are called dolomite (or dolostones). Although there are numerous other carbonate minerals, calcite, aragonite, and dolomite make up approximately 98% of them and are the most important ones for rocks names. Please also keep in mind that other minerals can occur in accessory amounts in carbonate rocks; quartz grains, chert, and glauconite are particularly common.

6.2: Carbonate Precipitation

6.2.1: Chemistry of Carbonate Precipitation

The chemistry of carbonate precipitation can be summarized by the following, somewhat counterintuitive, statements:

- Anything that adds CO_2 to water increases the acidity of the water which discourages carbonate precipitation and increases it's solubility.
- Anything that removes CO_2 from the water increases carbonate precipitation and decreases its solubility

We can back those statements up with a bit of chemistry. In the four following equations, the arrow points in the direction that the reaction proceeds and the k values are equilibrium constants (smaller values are slower reactions; note the negative exponent):

1. $CO_2 + H_2O \rightarrow H_2CO_3$ (fast, $k=10^{-1.43}$)

(Carbon dioxide plus water creates carbonic acid)

2. $H_2CO_3 \rightarrow H^+ + HCO_3^-$ (med., $k=10^{-6.40}$)

(Carbonic acid dissociates into hydrogen and bicarbonate)

3. $HCO_3^- \rightarrow H^+ + CO_3^{2-}$ (slow, $k=10^{-10.33}$)

(Bicarbonate dissociates into hydrogen and carbonate)

4. $Ca^{2+} + CO_3^{2-} \rightarrow CaCO_3$ (med., $k=10^{-8.33}$)

(Dissolved calcium combines with carbonate to form calcium carbonate)

If you just looked at these four equations without paying attention to the k value, you'd think that CO_2 would be a good thing for carbonate precipitation. But the k value in that third equation is slow, so much slower that the H^+ produced in Reaction 2 overwhelms the system (it makes the water acidic) and causes Reaction 3 to reverse. Its because of this fact that having lots of CO_2 in the system is bad for carbonate precipitation.

6.2.2: Controls on Carbonate Deposition

There are seven major factors that have an important influence on whether significant amounts of carbonate are deposited. Several of them are derived from the chemistry described above and might be intuitive to those of you that drink carbonated beverages.

Water temperature: All other factors being equal, warm water holds less CO_2 than cold water. If you've got a soda stream, you might know that you can make cold water much fizzier than warm water. Thus, warm water encourages carbonate precipitation.

Agitation: Increasing agitation allows excess CO_2 to go from the water to the atmosphere - just think about what happens if you shake a can of soda. Thus, increased agitation encourages carbonate precipitation.

Water depth: You can keep more CO_2 in solution at higher pressures (deeper water) and much less in solution at lower pressures. Think about what happens when you open a bottle of soda. Thus, shallow water is friendlier to carbonate deposition.

This variable becomes very apparent in the deep ocean. Shallow waters are frequently supersaturated in calcite and many microorganisms build calcareous skeletons. When the organisms dies, their skeletons settle and move into the cold, calm, waters of the deep ocean. If the water is deep enough, they will eventually reach the Carbonate Compensation Depth (CCD) which is the point beyond which carbonate does not accumulate because dissolution overcomes supply.

Life: The activity of organisms can encourage the accumulation of carbonate by building calcareous skeletons, photosynthesis (removal of CO_2), bacterial activity, and the decay of organic material. Respiration adds CO_2 to the water and this has a negative impact on carbonate precipitation.

Salinity: Increasing salinity increases the number of dissolved cations in the water which has a negative impact on the precipitation of carbonates because the introduction of foreign cations tends to make calcium carbonate less stable.

Light: Many calcite-producing organisms are photosynthetic. That, combined with some of the other variables described above means that the majority of carbonate is produced at shallow depths.

Clastic sediment input: Clastic sediment input can overwhelm carbonate production and thus dilute the amount of carbonate. Additionally, clastic sediment can interfere with many shell-building organisms and decrease the amount of light available for photosynthesis.

6.2.3: Dolomite Formation

6.2.3.1: Primary Dolomite

The chemical formula for the creation of true stoichiometric dolomite is:

$$Ca^{2+}(aq) + Mg^{2+}(aq) + 2CO_3^{2-}(aq) \rightarrow CaMg(CO_3)_2$$

But, chemical conditions are suitable for the precipitation of primary dolomite only at temperatures above 60°C or in cases where the ratio of Mg to Ca is much higher than it is in the open ocean which makes primary dolomite formation unlikely to be responsible for much of the dolomite present in the geologic record.

6.2.3.2: Dolomitization

Dolomitization is the process by which $CaCO_3$ is transformed to $CaMg(CO_3)_2$. This reaction can occur at normal atmospheric temperature and pressures and the reaction is:

$$Mg^{2+}(aq) + 2CaCO_3(solid) \rightarrow CaMg(CO_3)_2 + Ca^{2+}(aq)$$

In order to make this reaction happen, existing carbonates must be flushed with Mg-rich brine. The “hypersaline model” of dolomitization postulates that these brines can form in areas where evaporation of seawater causes the precipitation of gypsum ($CaSO_4$) and aragonite ($CaCO_3$) and that the remaining brine has an Mg:Ca ratio >8:1, which is much greater than that of seawater (~1:3). Although this model has its problems, its be best explanation we have for this poorly understood process.

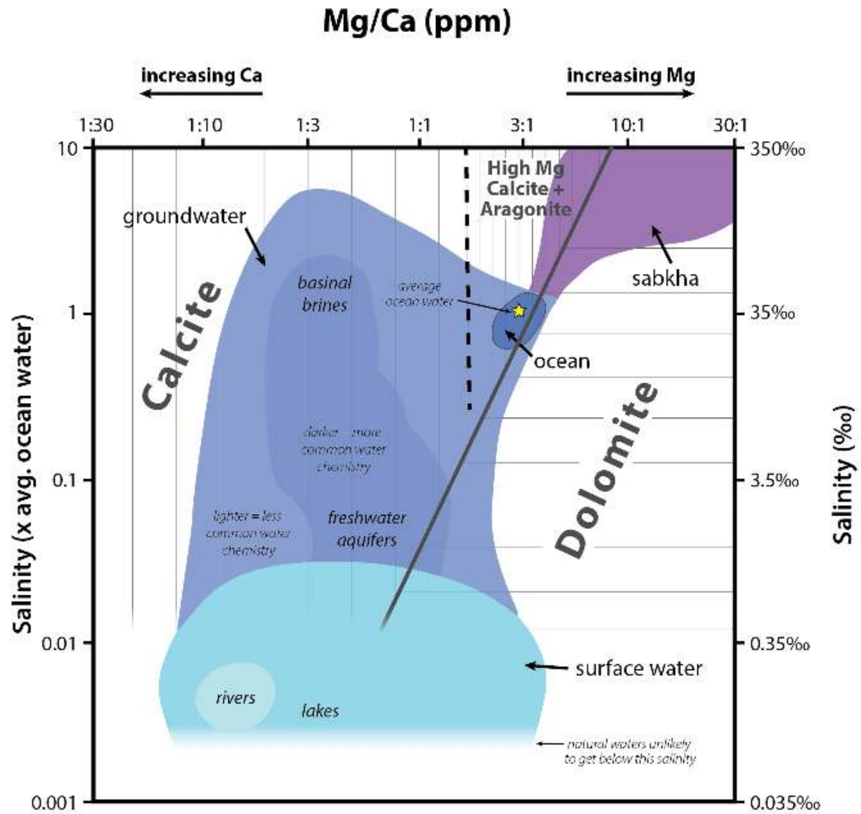

Figure 6.2.1: Plot of magnesium to calcium ratio versus salinity with overlays showing the chemistry of natural waters (colors), calcite and dolomite stability (dark lines and labels). After Figure 1 in Folk and Land (1975) available from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0.

6.2.4: Readings and Resources

- Folk, R. L., & Land, L. S., 1975, Mg/Ca ratio and salinity: two controls over crystallization of dolomite; AAPG Bulletin, v. 59 (1), p. 60-68.
- García-Ruiz, J. M., 2023, A fluctuating solution to the dolomite problem. *Science*, *382* (6673), 883-884.
- Kim, J., Kimura, Y., Puchala, B., Yamazaki, T., Becker, U., & Sun, W., 2023, Dissolution enables dolomite crystal growth near ambient conditions. *Science*, *382* (6673), 915-920.

6.3: Carbonate Components and Classification

The most common carbonate description schemes are comparable to those used for sandstones in that coarse- and fine-grained components are present at the time of deposition and that they can later be cemented together when minerals precipitate in pore spaces. Although you can gain a significant amount of information from carbonates in hand sample, thin sections are very useful for more nuanced descriptions and interpretations. Brief descriptions of the various components are provided in the sections that follow.

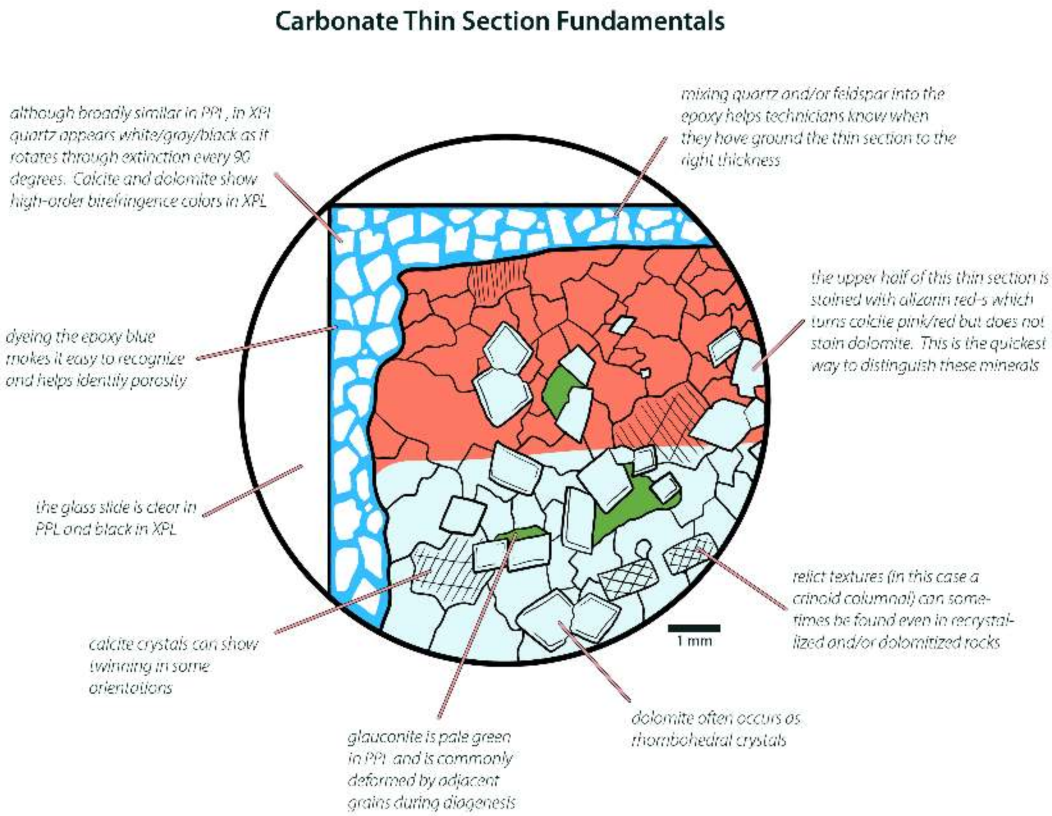

Figure 6.3.1: Illustration showing some fundamentals of carbonate rocks in thin section as they would appear in plane polarized light (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

6.3.1: Components

6.3.1.1: Carbonate Grains

Carbonate grains are analogous to the framework grains in a sandstone and consist of the following types:

6.3.1.2: *Skeletal Particles (Fossils)*

Skeletal particles are composed of calcareous shells or skeletal material. Identification keys for fossils in thin section are provided in Chapter 9.6

6.3.1.2.1: Coated Grains

Coated grains form as carbonate is progressively added to a grain. Ooids are the most common type, they form when wave activity suspends a particle and allows for concentric layers of inorganic carbonate precipitation around a nucleus. Oncoids are a less common type formed where algae causes calcite to precipitate, but storms periodically reorient the grain causing precipitation to occur on alternating sides of the grain. Pisoids are superficially similar to ooids, but they are larger (>2 mm diameter) and grow because of *in situ* accumulation of calcite and aragonite in the vadose zone.

6.3.1.2.2: Peloids and Pellets

Peloids are silt-to sand-sized particles composed of fine-grained lime mud (micrite); most are invertebrate fecal pellets.

6.3.1.2.3: Intraclasts

Intraclasts are fragments of lithified carbonate that were incorporated into younger sediment (effectively rip-up clasts).

6.3.1.3: Micrite

Micrite is fine-grained lime mud made of clay-sized crystals of microcrystalline calcium carbonate. Its analogous to the matrix in a sandstone in that its present at the moment of deposition. Micrite is most commonly formed from particles derived from algae or microfossils. Its also possible that it can from direct precipitation of these crystals in whiting events.

6.3.1.4: Spar

Spar is relatively coarse-grained calcite cement that grows in the pore spaces of carbonates. The large, clear crystals are relatively easy to distinguish from the other carbonate components. Spar in carbonates is analogous to cement in siliciclastic sedimentary rocks.

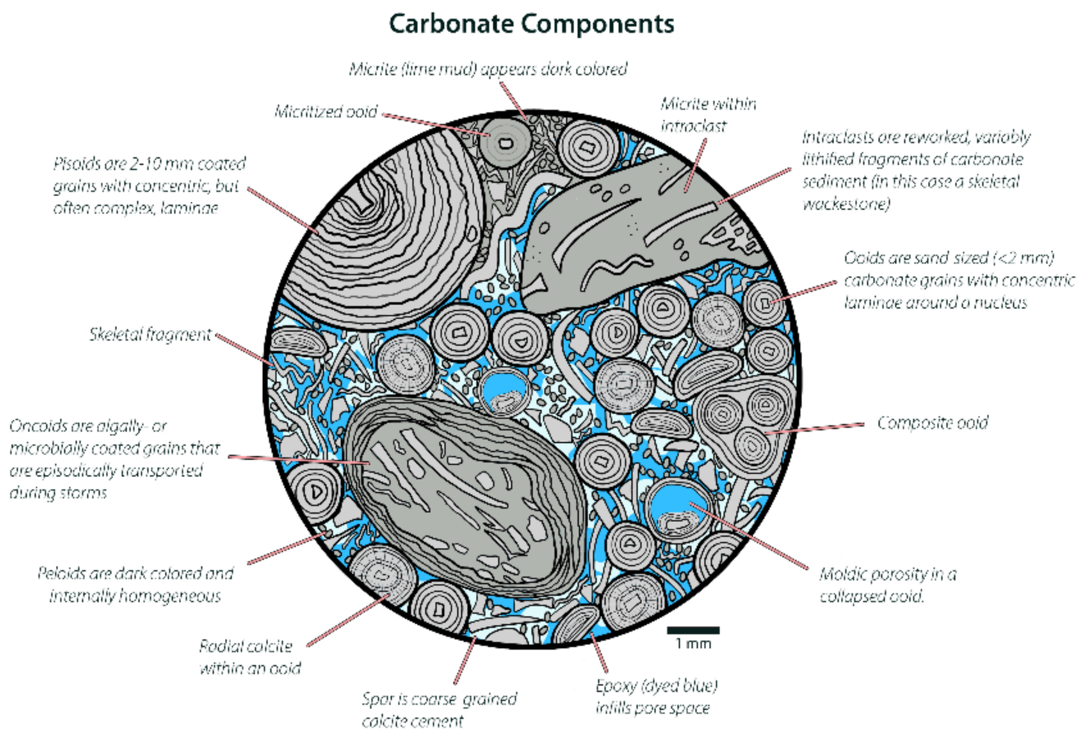

Figure 6.3.2: Schematic thin section (PPL) diagram showing the important components of carbonate rocks in thin section (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

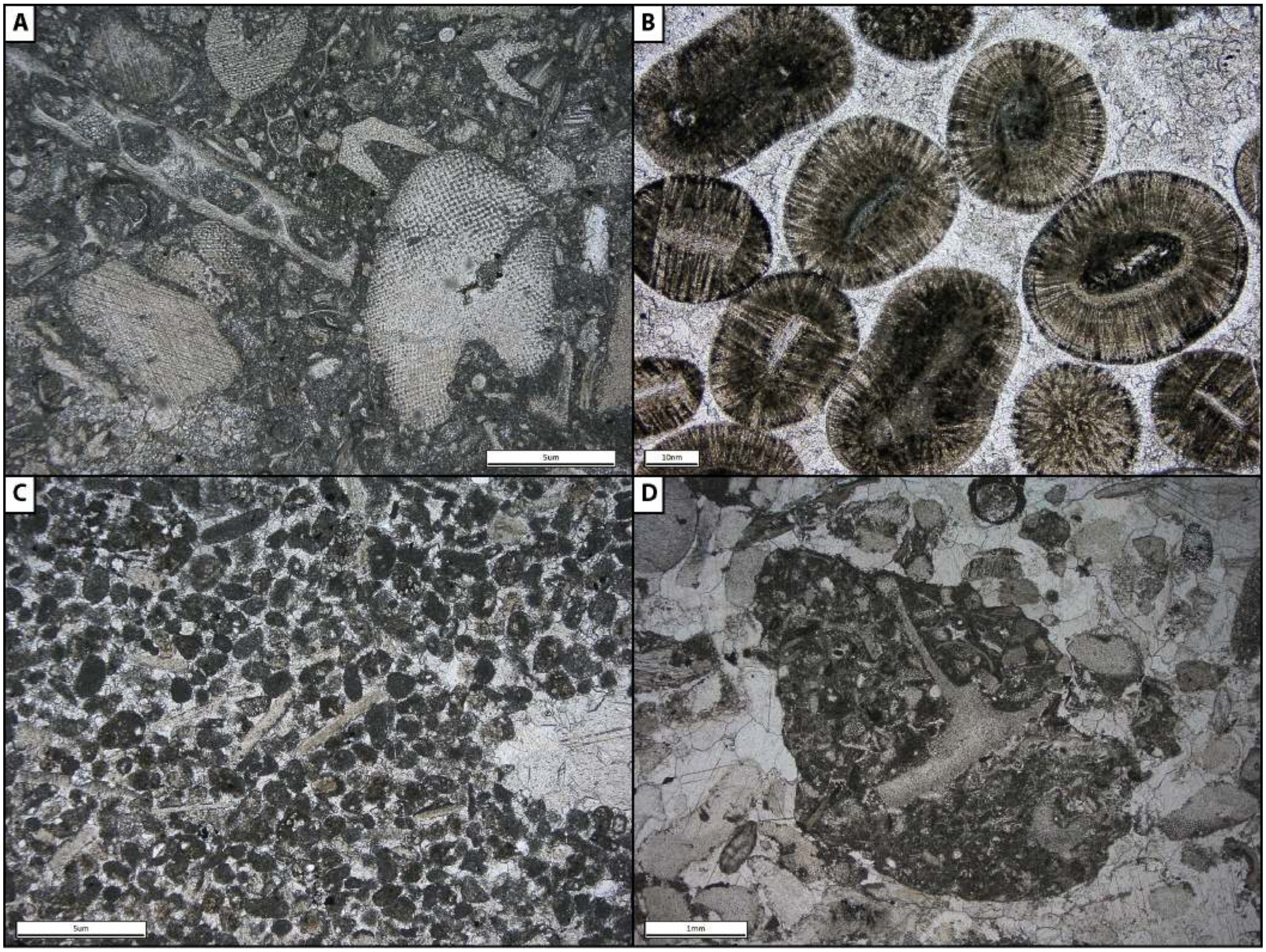

Figure 6.3.3: The four main types of grains in carbonates (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A) Photomicrograph (PPL) of bryozoan fragments in the Mississippian Lodgepole Formation, SW Montana. Skeletal fragments can come in an impressive variety of shapes and sizes ... this diversity of form is even more pronounced by the 2D slice of the thin section. B) Photomicrograph (PPL) of ooids in the the Cambrian Mifflintown Formation, PA. C) Photomicrograph (PPL) of peloids in the bryozoan fragments in the Ordovician Benner Formation, PA. D) Photomicrograph (PPL) of lithoclasts in Cambrian carbonate from upstate NY.

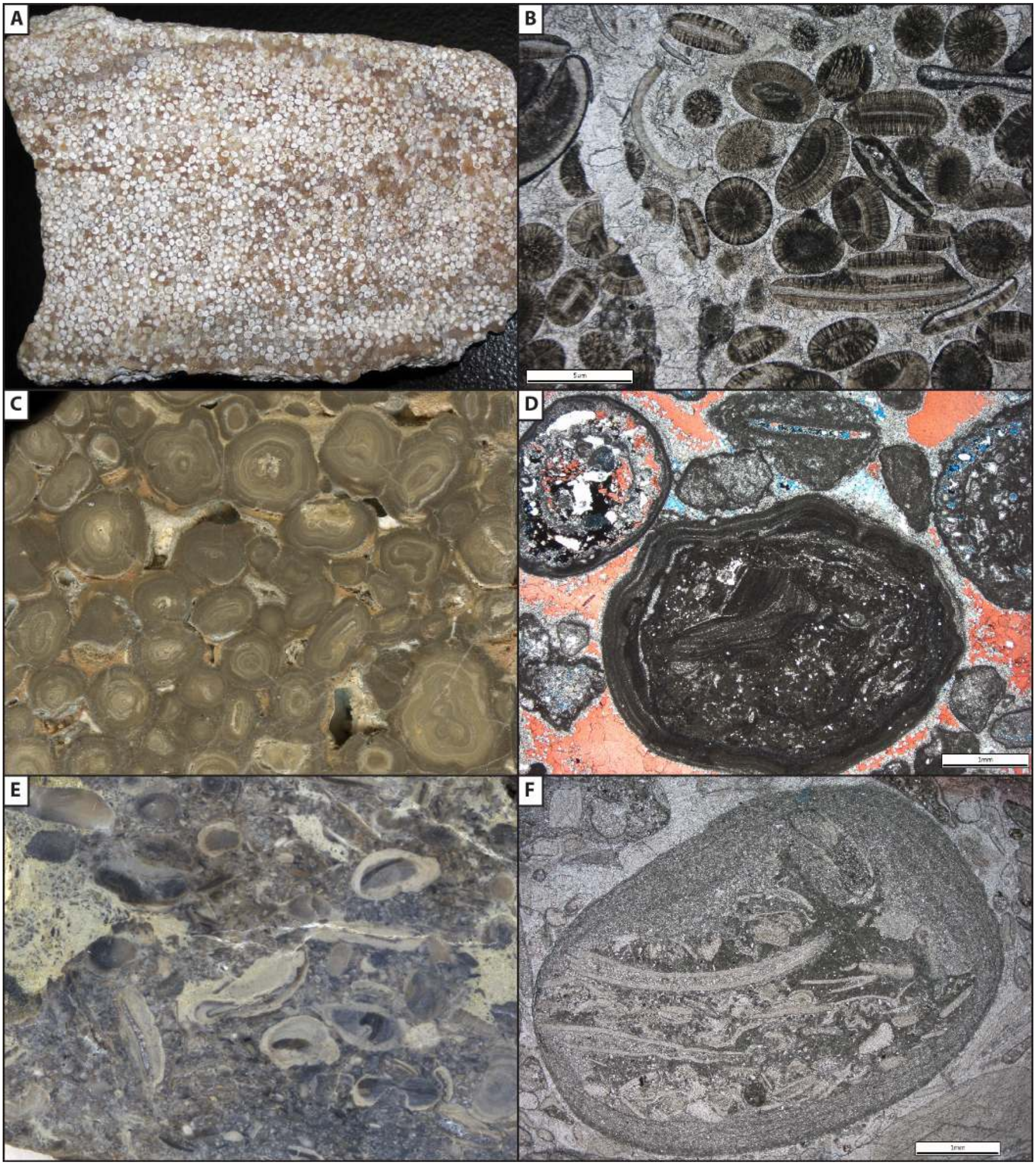

Figure 6.3.3: Hand sample and thin section photographs showing coated grains. A and B are ooids (field of view in A is ~20 cm), C and D are pisoids (field of view in C is ~12 cm), and E and F are oncoids (field of view in E is ~10 cm). A and E are from James St. John via Wikimedia Commons and Flickr; CC BY-SA 4.0. All other images are from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0.

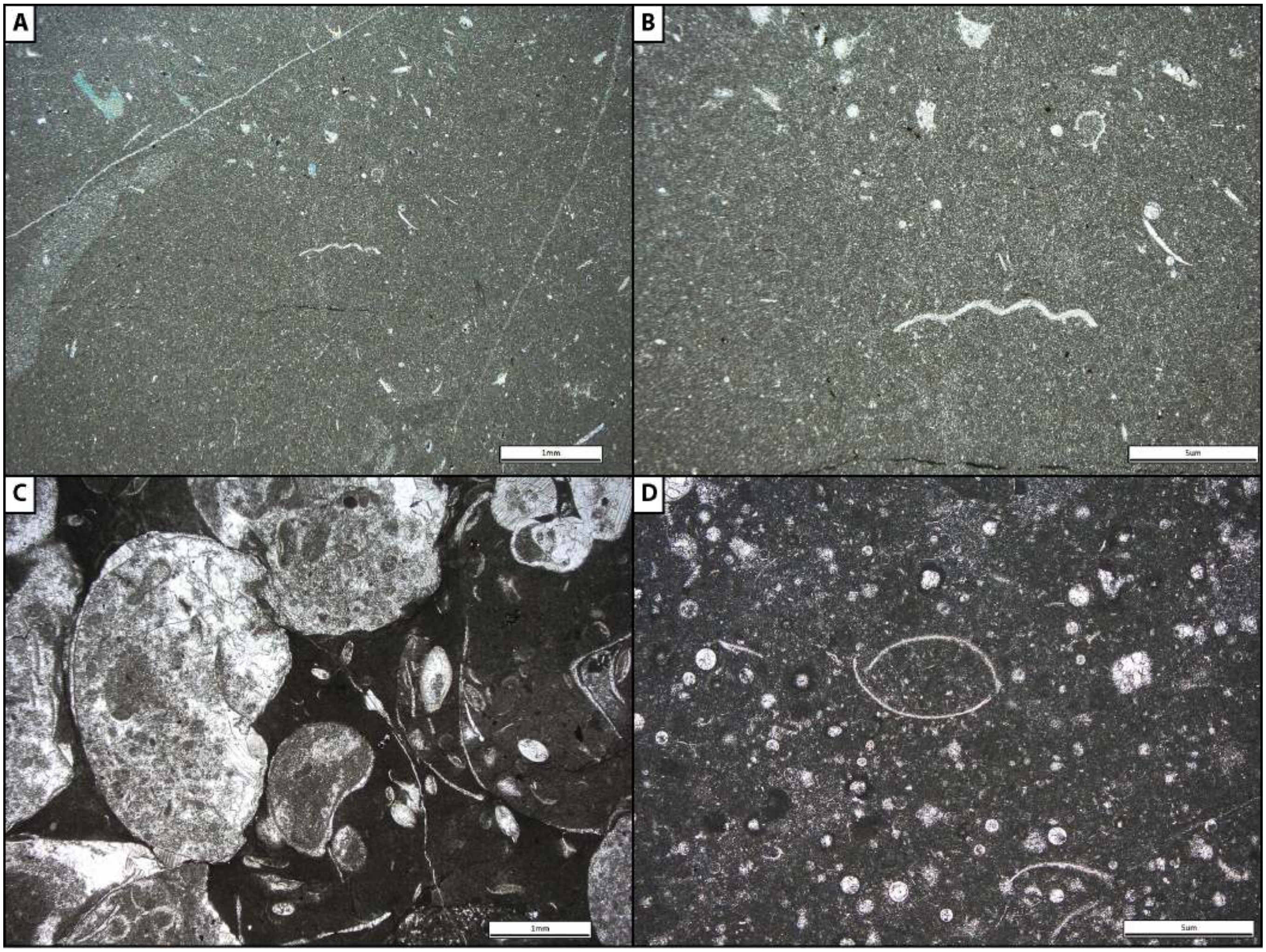

Figure 6.3.4: Examples of micrite (lime mud); all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0. A) Photomicrograph (PPL) of micrite in a lime mudstone in the Mississippian Lodgepole Fm., SW Montana. B) Photomicrograph (PPL) showing a more magnified view of the central part of the image in Fig. 2A. Note that even at this scale, the micrite appears homogeneous. C) Photomicrograph (PPL) of micrite with coarse spar infilling gastropods and ostracods. Specimen is from the Gastropod Limestone Member of the Cretaceous Kootenai Fm., SW Montana. D) Photomicrograph (PPL) of micrite with scattered ostracods and calcispheres. Specimen is from the Mississippian Lodgepole Fm., SW Montana.

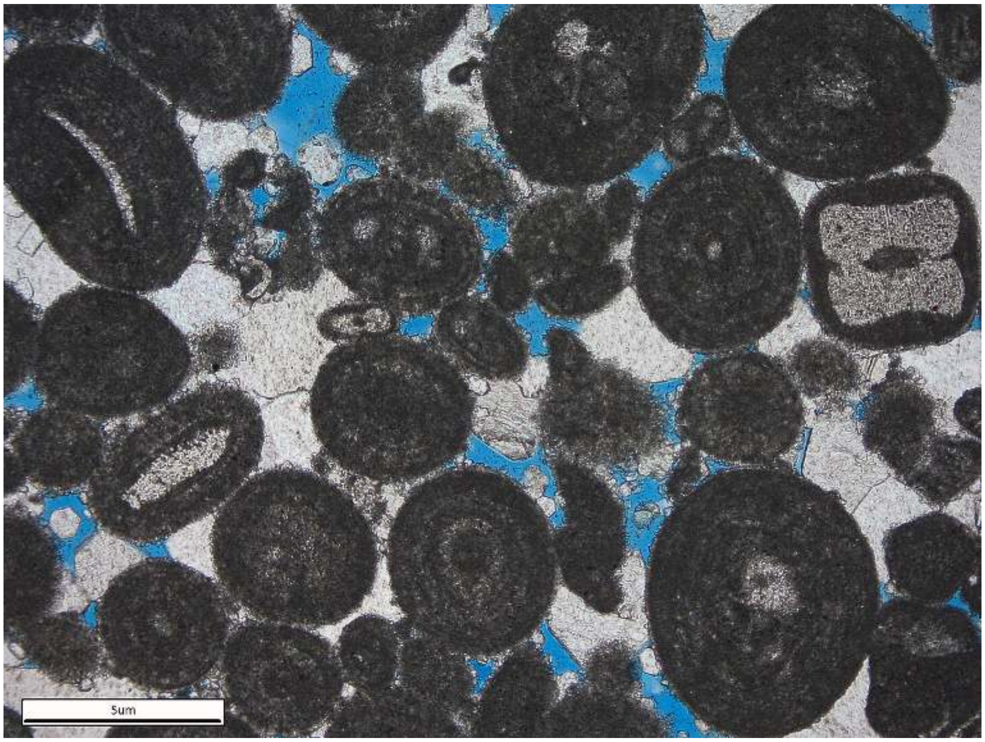

Figure 6.3.5: Photomicrograph (PPL) showing an oolitic grainstone with interstitial areas filled with spar (white) and/or blue epoxy (which infilled formerly open spore spaces during preparation of the slide; Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

6.3.2: Classification

6.3.2.1: Folk classification scheme

The Folk classification scheme is most useful for petrographic work. Samples are given a two part name; the first part is the most abundant grain type (bio-, pel-, oo-, or litho-) and the second indicates the nature of the interstitial material (-sparite or -micrite). Thus, a pelsparite would be a limestone composed mostly of peloids with with coarse spar between the shells and a biosparite would be most shell fragments with sparry calcite cement.

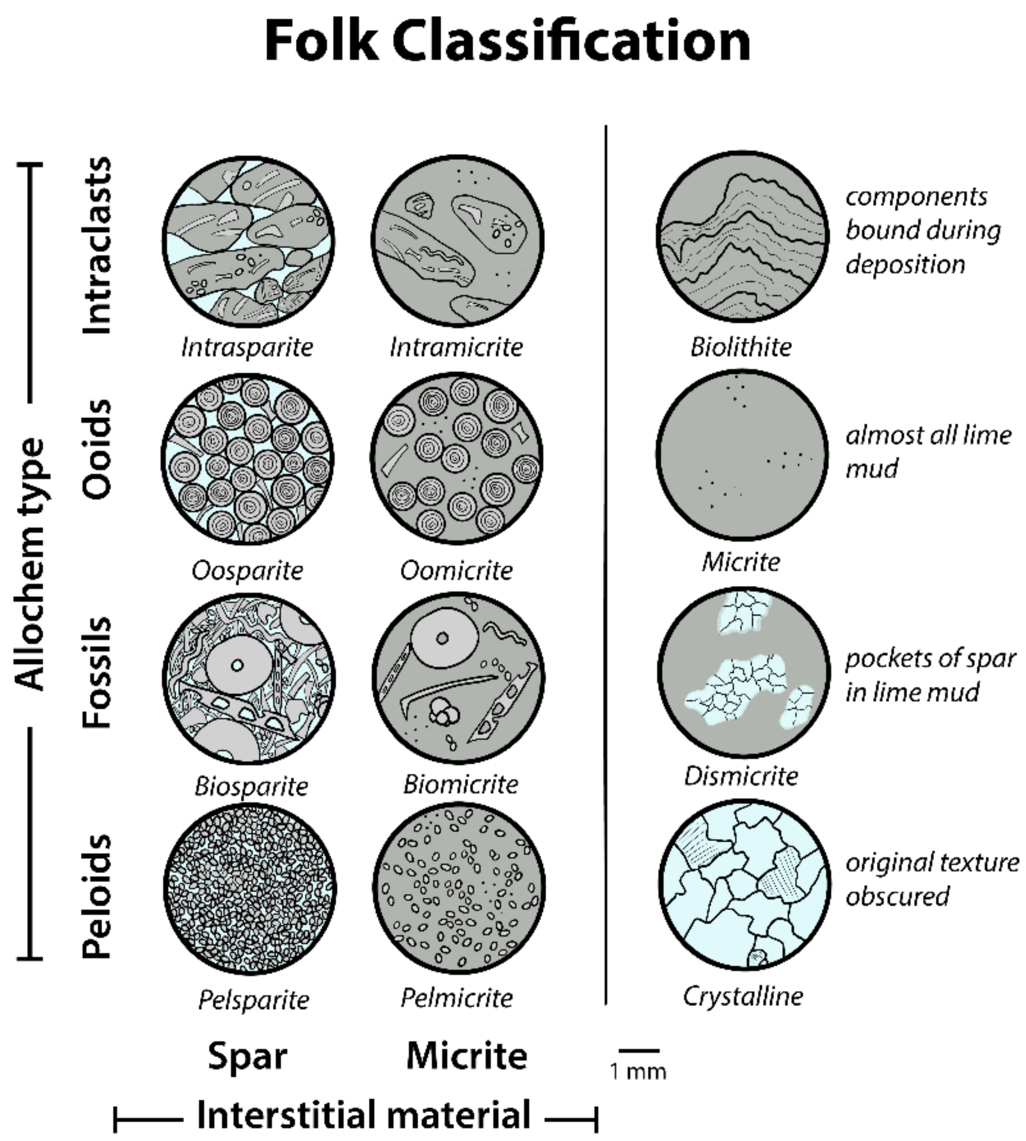

Figure 6.3.6: The Folk classification scheme assigns names to limestones based on the most abundant grain(s) and the nature of the interstitial material (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

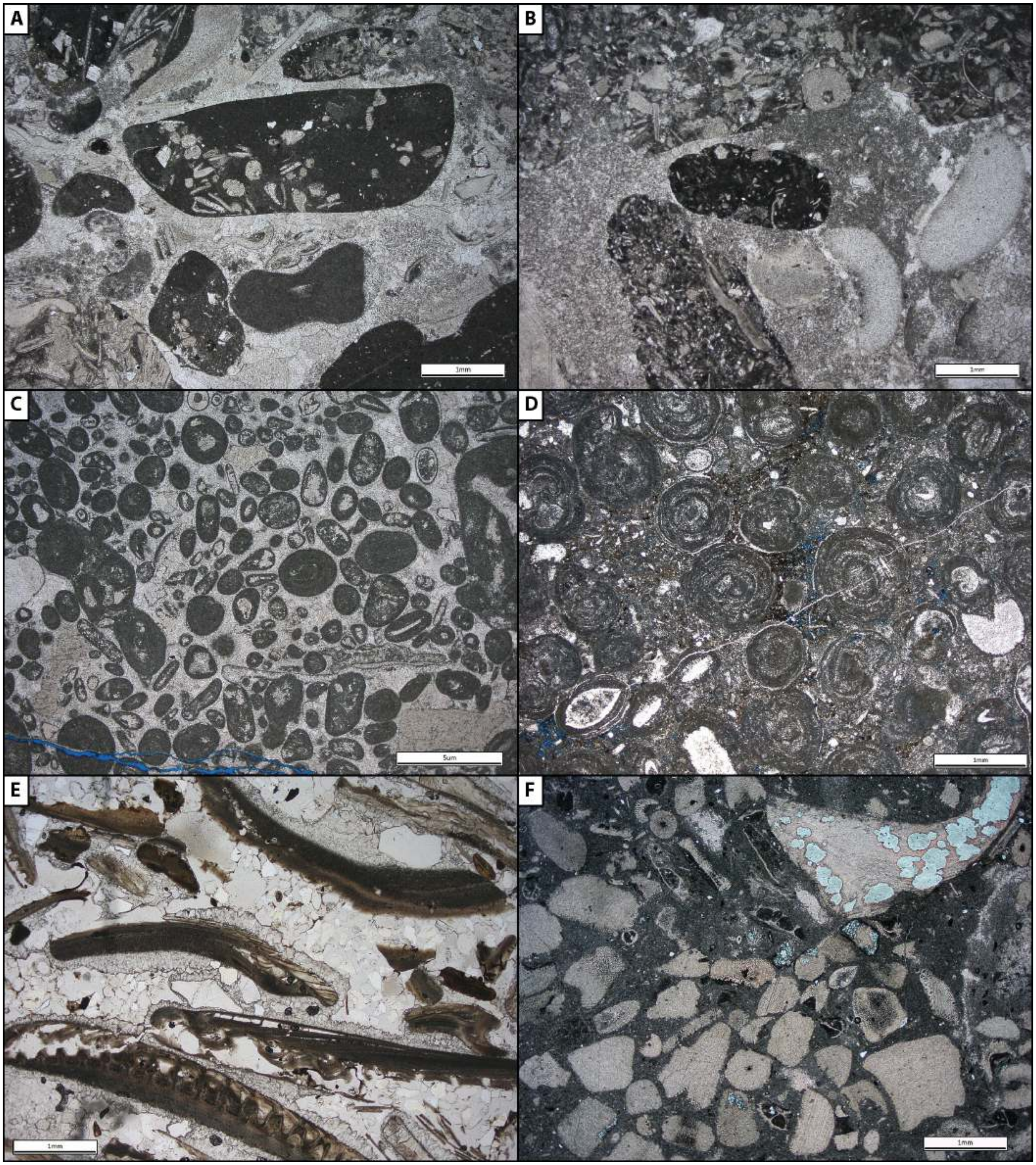

Figure 6.3.7: Photomicrographs (PPL) of different types of limestones in the Folk classification - part 1 (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A) Lithoclasts and skeletal material in an intrasparite from upstate NY. B) Lithoclasts and skeletal material in an intramicrite from upstate NY. Note that some of the dark micrite has started to recrystallize in the left half of the image. C) Oosparite from the Mississippian Lodgepole Formation, SW Montana. D) Oomicrite from the Silurian Keel Formation, Oklahoma. E) Biosparite. F) Biomicrite from the Mississippian Lodgepole Formation, SW Montana.

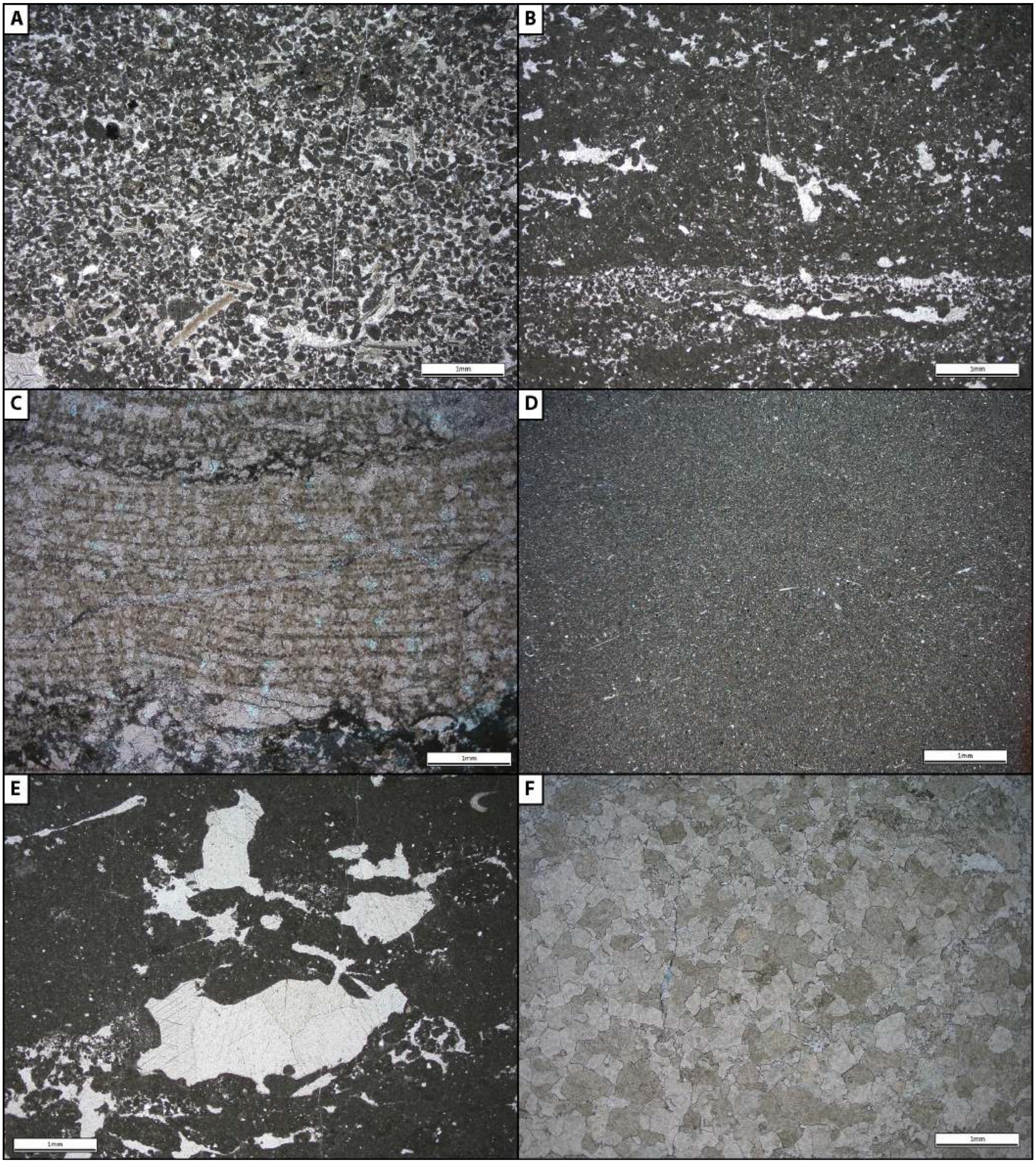

Figure 6.3.8: Photomicrographs (PPL) of different types of limestones in the Folk classification - part 2 (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A) Pelsparite from the Mississippian Lodgepole Formation, SW Montana. B) Possible pelmicrite. In a situation where you have muddy peloids surrounded by micrite, the odds of being able to distinguish individual peloids after diagenesis is slim. However, the presence of pelsparite zones makes the "blobby" areas of micrite likely candidates for pelmicrite. C) Stromatoporoid boundstone from the Devonian Jefferson Formation, SW Montana. D) Mudstone Mississippian Lodgepole Formaton, SW Montana. E) Dismicrite. F) Recrystallized limestone from the Cambrian Pilgrim Formation, SW Montana.

6.3.2.2: Dunham Classification System

The Dunham classification scheme is most useful for hand sample and outcrop work. It is a four-fold scheme based on the presence/absence and relative abundance of mud:

- (Lime) mudstones have <10% grains
- Wackestones have >10% grains and are supported by mud
- Packstones are grain supported with interstitial mud.
- Grainstones are grain supported with little or no interstitial mud.

It is worth noting that with carbonates, shells are often the largest grains and thus maximum grain size could ultimately be controlled by biology. Because of this, the amount of mud is the best proxy for the amount of energy in the system.

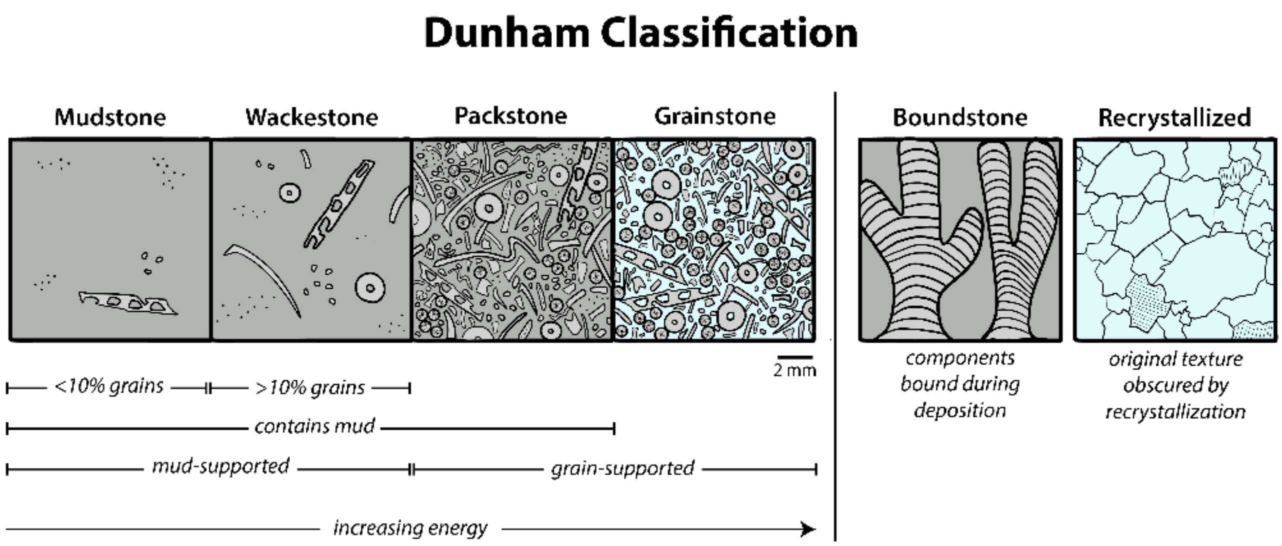

Figure 6.3.9: The Dunham classification scheme assigns names to limestones based on the abundance of mud and the support mechanism of the grains (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

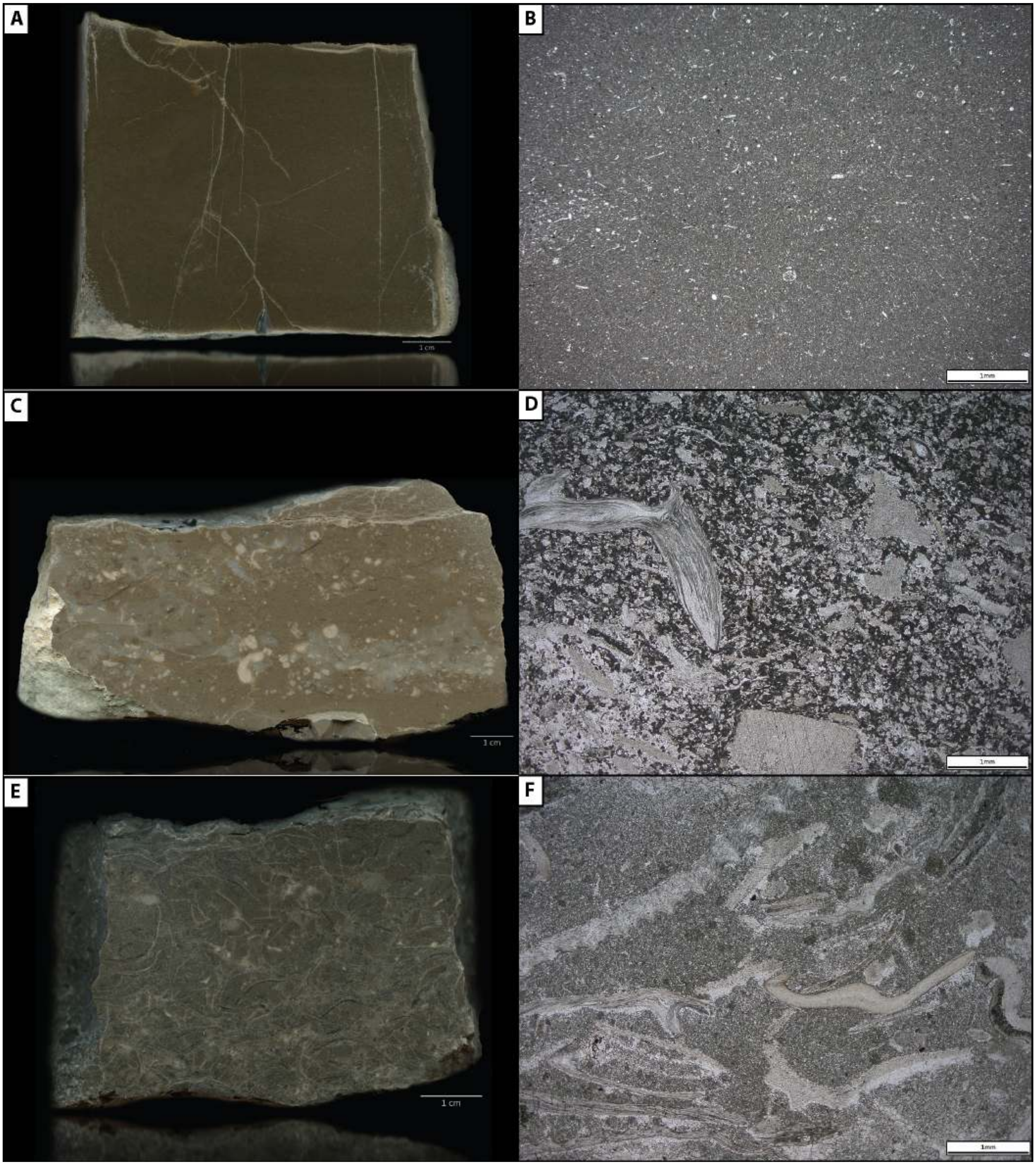

Figure 6.3.10: Paired polished slabs and photomicrographs showing the main types of limestones in the Dunham classification scheme - part 1 (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A & B) Polished slab and photomicrograph (PPL) of a lime mudstone from the Mississippian Lodgepole Formation. C & D) Polished slab and photomicrograph (PPL) of a skeletal wackestone from the Mississippian Lodgepole Formation. Note that the micrite in the thin section has been partially dolomitized and many of the light colored patches between shells are actually dolomite crystals formed within the micrite. E & F) Polished slab and photomicrograph (PPL) of a skeletal packstone. Although the shells in the slab are

somewhat widely spaced, this sample is grain supported because the shells are curved three dimensional forms that are touching each other in the third dimension.

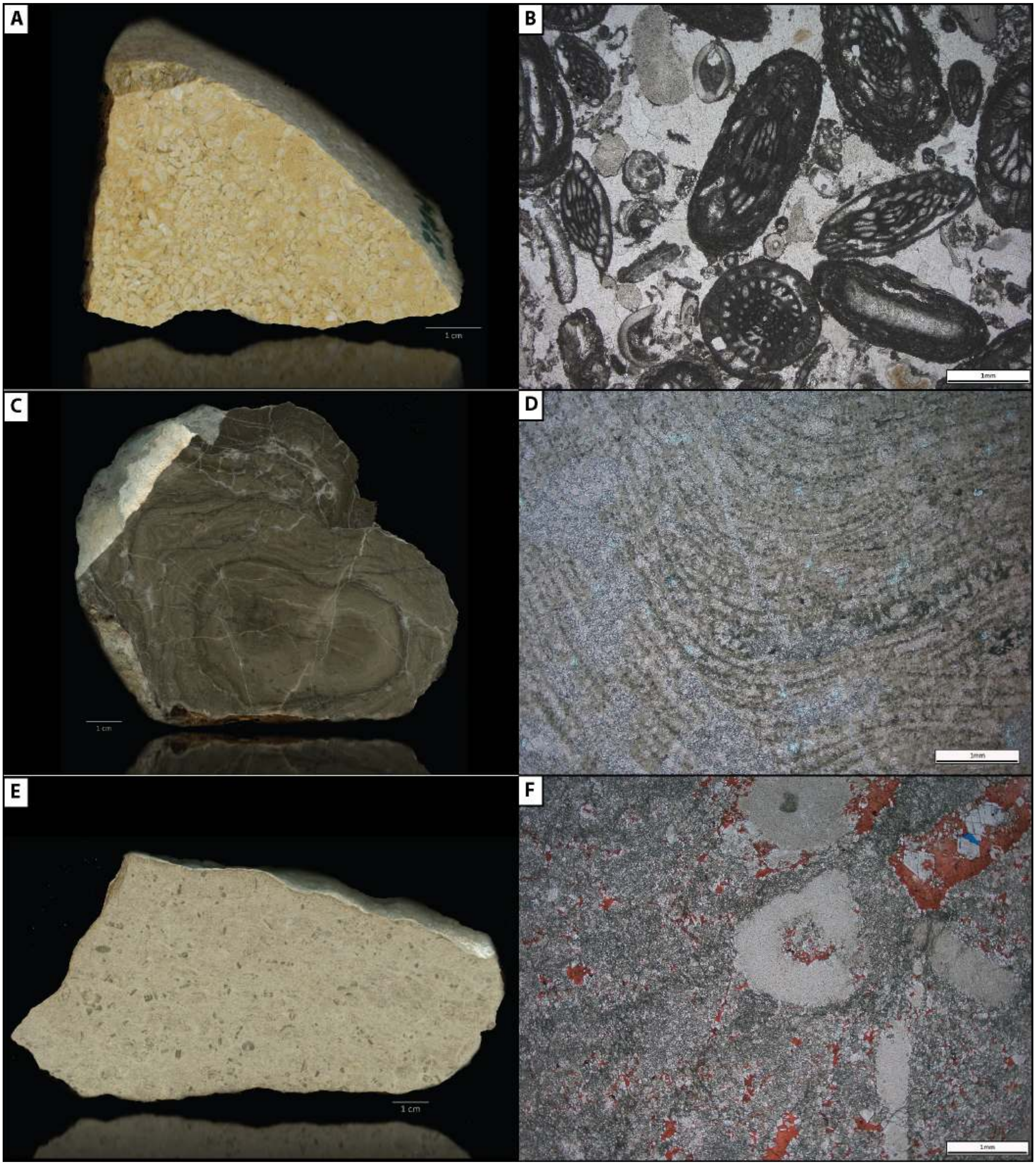

Figure 6.3.11: Paired polished slabs and photomicrographs showing the main types of limestones in the Dunham classification scheme - part 2 (all images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). A & B) Polished slab and photomicrograph (PPL) of a foram-rich skeletal grainstone. As with the packstone, this specimen is clast supported and although some foraminifera appear isolated, they are actually touching each other in the third dimension. C & D) Polished slab and

photomicrograph (PPL) of a stromatoporoid boundstone. E & F) Polished slab and photomicrograph (PPL) of a recrystallized and dolomitized fossiliferous limestone. Although some fossils can still be identified in the polished slab, primary textures have been almost completely destroyed and the rock has a sandy appearance on weathered surfaces from the dolomite rhombs.

CHAPTER OVERVIEW

7: Chemical, Biochemical, and Other Sedimentary Rocks

This chapter focuses on sedimentary rocks other than clastics and carbonates (evaporites, siliceous, and organic-rich rocks). Most of them form either from direct precipitation from fluids or because of the activities of living organisms. Although not as volumetrically significant as carbonates or clastics, the rocks described in this chapter are of great economic importance.

Learning Objectives

- List common evaporite minerals
- Describe how evaporite minerals form and how thick evaporite deposits occur.
- Summarize the special physical properties of evaporites.
- List the basic types of siliceous rocks and explain how they formed.
- Explain how oil, natural gas, and coal form.

Topic hierarchy

Chapter thumbnail shows a worker holding up a piece of coal in front of a coal firing power plant in the Netherlands (Adrem68 via Wikimedia Commons; CC BY-SA 4.0)

7.1: Evaporites

Evaporite minerals and the sedimentary rocks they comprise form from precipitation of mineral from a saline solution. Generally the fluid becomes supersaturated through evaporation and this is easiest to achieve in arid environments.

7.1.1: Evaporite Minerals

Although previously discussed as a stand-alone group, carbonate minerals - particularly calcite ($CaCO_3$) and dolomite ($CaMg(CO_3)_2$) - are also considered evaporite minerals and are some of the first ones to form as concentrations in the brine increase. Trona ($Na_2CO_3 \cdot NaHCO_3 \cdot 2H_2O$) forms in freshwater evaporite settings and is mined as a source of soda ash which is used in the manufacture of glass, paper, detergents, and other chemicals.

Gypsum ($CaSO_4 \cdot 2H_2O$) and anhydrite ($CaSO_4$) are two of the most common evaporite minerals and are members of the sulfate mineral class. The two are closely associated and with increased burial depth gypsum commonly loses the water and transforms into anhydrite. The process can reverse itself if anhydrite returns to suitable conditions in the presence of water. They look very similar, but anhydrite is slightly harder. The two often occur together (sometimes with small amounts of clastic mud) in nodular masses and have a distinctive "chicken wire" texture. Barite ($BaSO_4$) and celestite ($SrSO_4$) are much less common sulfates that occasionally occur as secondary minerals formed from the interaction of gypsum/anhydrite with barium- or strontium-enriched fluids.

Halite ($NaCl$) is the most common halide evaporite mineral and is familiar to anyone who has taken an intro geology lab and was brave enough to lick the samples. Potassium and magnesium salts are late formed minerals that are less common but economically important. Examples include sylvite (KCl), carnallite ($KClMgCl_2 \cdot 6(H_2O)$), kieserite ($MgSO_4 \cdot H_2O$), and polyhalite ($K_2Ca_2Mg(SO_4)_4 \cdot 2H_2O$).

Although rare, some evaporite minerals are economically important when found in significant quantities. They include borax ($Na_2B_4O_5(OH)_4 \cdot 8H_2O$) which is used in laundry and cleaning products and nitre (KNO_3) which is used for making gunpowder and cleaning products.

Evaporite Precipitation

Evaporites form in situations where brines form because the amount of water lost through evaporation is greater than water supplied by rainfall, recharge via rivers, and/or communication with a larger body of water. They can form in a variety of marine, marginal marine, and nonmarine environments that meet these conditions. The minerals listed above don't randomly or simultaneously precipitate, but rather they are formed in a predictable sequence based on the composition of the brine.

Mineral	Percent of the original brine remaining when the mineral starts to precipitate (water depth if you started with 1,000 m)	Thickness of this mineral that would form if you started with 1,000 m of sea water and evaporated it completely
Calcite	50% (500 m)	0.10 m
Gypsum	15% (150 m)	0.61 m
Halite	10% (100 m)	13.30 m
K and Mg salts	5% (50 m)	2.99 m
	Total evaporite thickness	17 m

Table 7.1.1: The evaporite sequence that one could expect if you started with 1,000 m of seawater and evaporated it completely (data from https://www.alexstrekeisen.it/english/sedi/evaporites.php).

7.1.1.1: Evaporite environments

Evaporites can form in a variety of marine and terrestrial environments as long as the rate of evaporation exceeds the rate of recharge. In general, we can lump the spectrum of environments into three main families: platform evaporites, basin-wide evaporites, and non-marine evaporites:

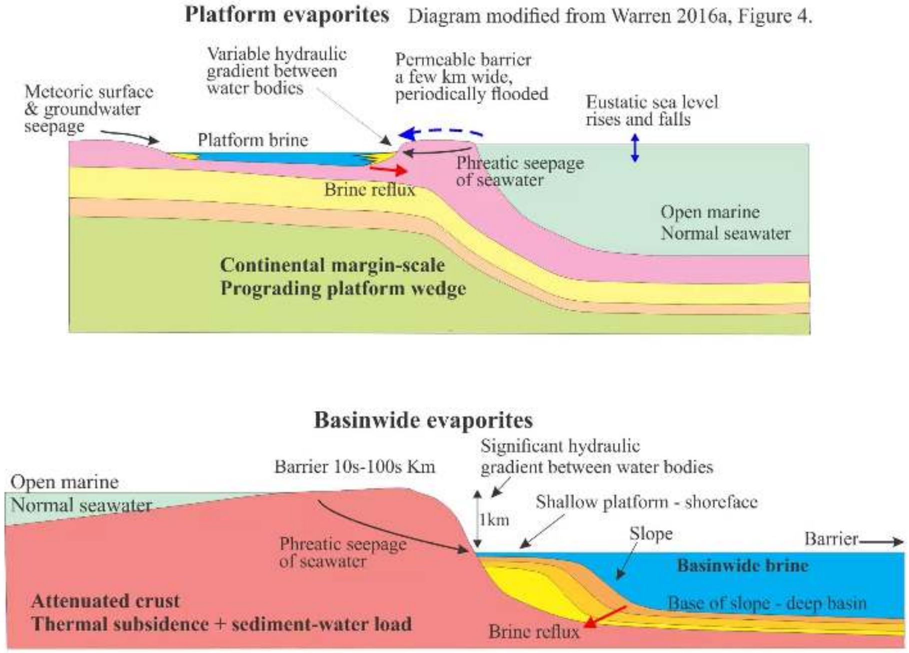

Figure 7.1.2: Two of the three main families or marine evaporite environments are the platform and basin-wide evaporite models (from Brian Ricketts via Geological Digressions - Mineralogy of Evaporites, CC BY).

Platform evaporites model. In this model, evaporites of modest thickness and distribution occur in marginal marine settings where seawater is able to breach a topographic barrier and flood low-lying coastal environments (ex: sabkhas). Evaporite facies are commonly interbedded with shallow water clastics and carbonates.

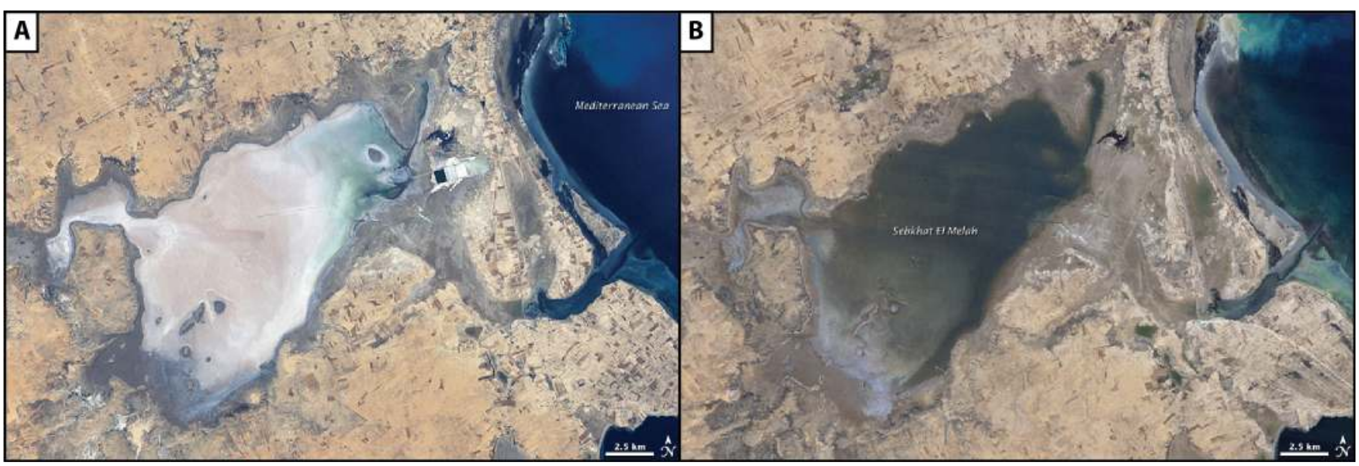

Figure 7.1.2: Landsat images of Sebkhat El Melah, Tunisia a flat sabkah that lies below sea level and is periodically flooded with marine waters from the Mediterranean Sea. A) Image showing the area largely dry in December of 1999 and B) flooded in January, 1987 (NASA image by Jesse Allen via NASA Earth Observatory; public domain).

Basin-wide evaporites. In this model, evaporites accumulate in areas where the crust is low-lying and/or subsiding (ex: early stages of a rift basin or the Mediterranean Sea) and seawater breaches or seeps through a barrier (ex: Straight of Gibraltar) providing a mechanism for a thick succession of evaporites to form along with associated clastic and carbonate deposits. There are no basin-wide deep water evaporites today; the best modern analog might be the Dead Sea and the easiest to visualize ancient examples are the early phases of a rift basin or the Mediterranean Sea which has a thick evaporite succession in the subsurface that formed when the basin was periodically isolated from the Atlantic Ocean between 5 and 6 Ma during the "Messinian Salinity Crisis".

Non-marine evaporites. Non-marine environments include a spectrum of environments including playa lakes, deserts, and various closed basins. As with platform evaporites they are likely to be of limited thickness and distribution. Given that many of them are sourced from non-marine waters they can also have mineralogy that is similar to, but a bit different than marine evaporites.

Figure 7.1.3: Non-marine evaporites forming in the Badwater Basin in Death Valley National Park. Although this closed basin exists in an arid climate today, during the last glacial the climate was much more humid and this area was the site of Lake Manly - a large pluvial lake (Justin Mier via Flickr; CC BY-NC-SA 2.0).

7.1.2: Marine Evaporite Models

If you need 1,000 meters of seawater to make a 17 m thickness of evaporites, how can we possibly explain the presence of evaporite successions that are tens, hundreds, or perhaps even a few thousand meters thick? If it was a single event, it would require an impossible depth of water and would be recorded as a single coherent evaporite succession. It turns out that thick evaporites successions are composed of cyclic packages of evaporites that record a complex history, from which three possible explanations for these thick successions emerge:

Deep-water/deep-basin model. In this case you have a deep body of water filling a deep topographic basin. Because evaporation and recharge are both ongoing, the water remains deep but you are are still able to create a brine of sufficient concentration to cause crystals to settle out from the water column and/or be transported downslope to deeper parts of the basin.

Shallow-water, shallow-basin model. In this model, relatively shallow basins are flooded with modest amounts of water and evaporites form as that water body dries up. Any given event results in only a modest thickness of evaporites, but subsidence allows the process to keep going through time so that thick successions can accumulate.

Deep-basin, shallow-basin model. A deep basin is separated from the open ocean by a barrier. That barrier periodically overflows allowing water to partially flood the basin. Evaporation of water from that event causes a modest thickness of evaporites to form; repeated periodic overflow allows for a thick succession to eventually fill the basin.

7.1.3: Salt Diapirs

The halite, gypsum, and anhydrite that make up the overwhelming majority of evaporite deposits are less dense than most other rocks and they can flow and deform plastically if given enough time (that's especially true when they are buried). This combination of properties explains why evaporite deposits are so prone to deformation and commonly form wall or plume-shaped salt diapirs that appear to rise through and deform adjacent sediments. Older literature attributed the flow of evaporites primarily to density differences, but more recent literature shows that differential loading (often by the progradation of sediments from basin margins) initiates diapirism. Once loading begins and evaporite withdrawal commences in the subsurface, it allows for rapid sediment accumulation on the diapir flanks and deformation of adjacent sediments (think of the sediment like a rolling pin that is rolling over a tube of toothpaste (the evaporites). Because evaporites are easily dissolved, these features are only rarely exposed at the surface - some of the best known examples are in arid deserts (ex: Kavir Desert of Iran) or rapidly eroding coastal exposures.

Figure 7.1.4: Coastal exposure of the Finlay Point Diapir in western Cape Breton, Nova Scotia. Mississippian salt of the Windsor Group deformed as Pennsylvanian clastics prograded into the basin. This feature is only exposed because of rapid coastal erosion; the white rocks on the right side of the image and the gypsum-anhydrite cap atop the diapir (Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

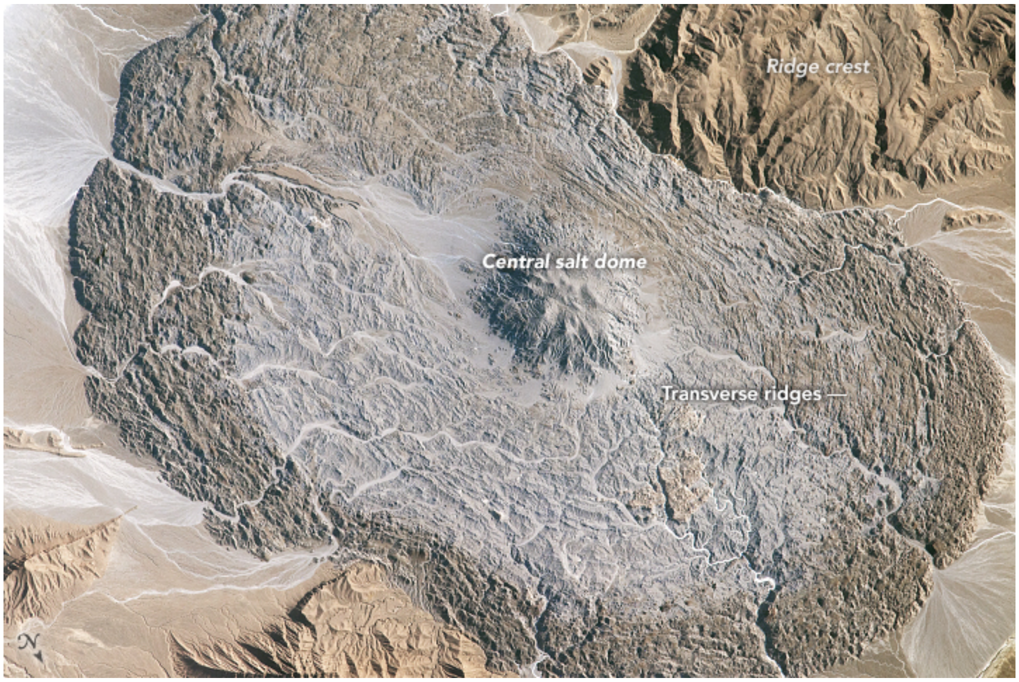

Figure 7.1.5: Outcrop of a salt dome/glacier in the Zagros Mountains of Iran. This feature is approximately 8 miles in diameter. During the rainy season, the salt at the surface can flow as much as a few tens of centimeters in a day (Astronaut photograph ISS052-E-8401 via NASA Earth Observatory; public domain).

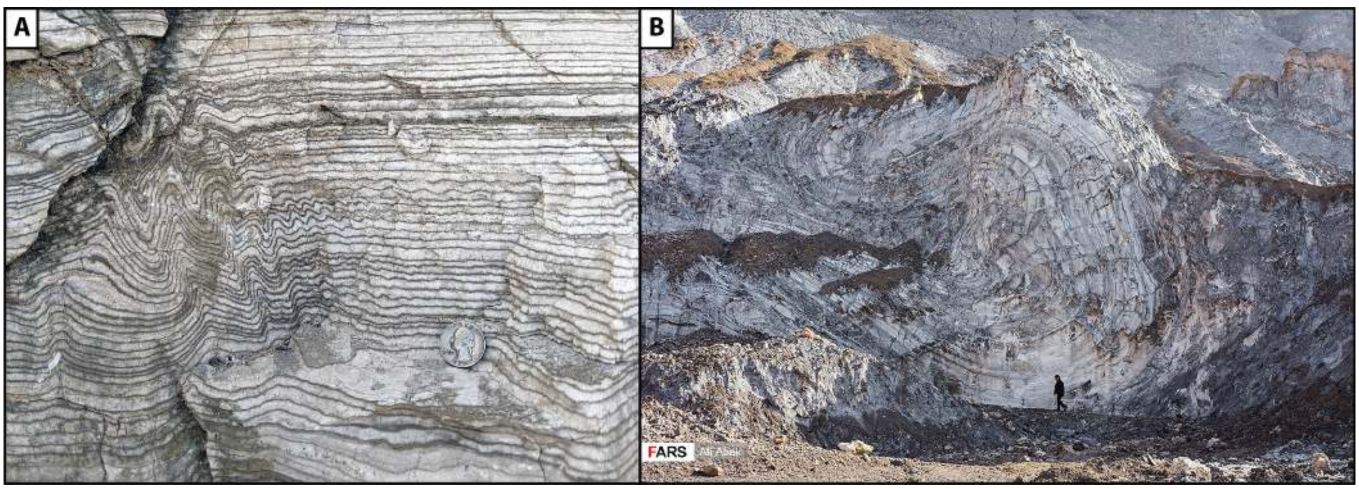

Figure 7.1.6: Deformation in evaporites. A) Small-scale disharmonic deformation in alternating layers of calcite (dark) and gypsum (light) in the Permian Castille Formation in the Delaware Basin of West Texas and New Mexico (Michael Rygel via Wikimedia Commons; CC BY-SA 4.0). B) Recumbent folds in the Shah Alamdar salt dome (Fars Media Corporation via Wikimedia Commons; CC BY 4.0).

7.1.4: Readings and Resources

- Geological Digressions, Mineralogy of carbonates; Sabkhas - https://www.geological-digressions.com/2020/01/16/mineralogy-of-carbonates-sabkhas/
- Geological Digressions, Mineralogy of evaporites: Marine basins - https://www.geological-digressions.com/2020/03/19/mineralogy-of-evaporites-marine-basins/

7.2: Siliceous Sedimentary Rocks

As the name implies, siliceous sedimentary rocks are enriched in silica. We will avoid the numerous mineral and gem names applied to siliceous material (jasper, flint, agate, chalcedony, opal, etc.) and instead use the terms microcrystalline quartz or chert to refer to any/all of them.

7.2.1: Formation of siliceous sediments and sedimentary rocks

Ultimately, all dissolved silica can trace its origin back to the weathering and/or dissolution of existing silicate rocks or input from hydrothermal sources. Once in solution, there are several possibilities for the formation of siliceous sediments:

1. Direct precipitation from water has been reported in modern nonmarine settings and, although not active today, the abundance of chert in Precambrian banded iron formations suggests that this process was active in the ocean in the geologic past.

2. Biogenic silica forms when organisms such as sponges, radiolarians, and/or diatoms extract silica from the water to build their skeletons.

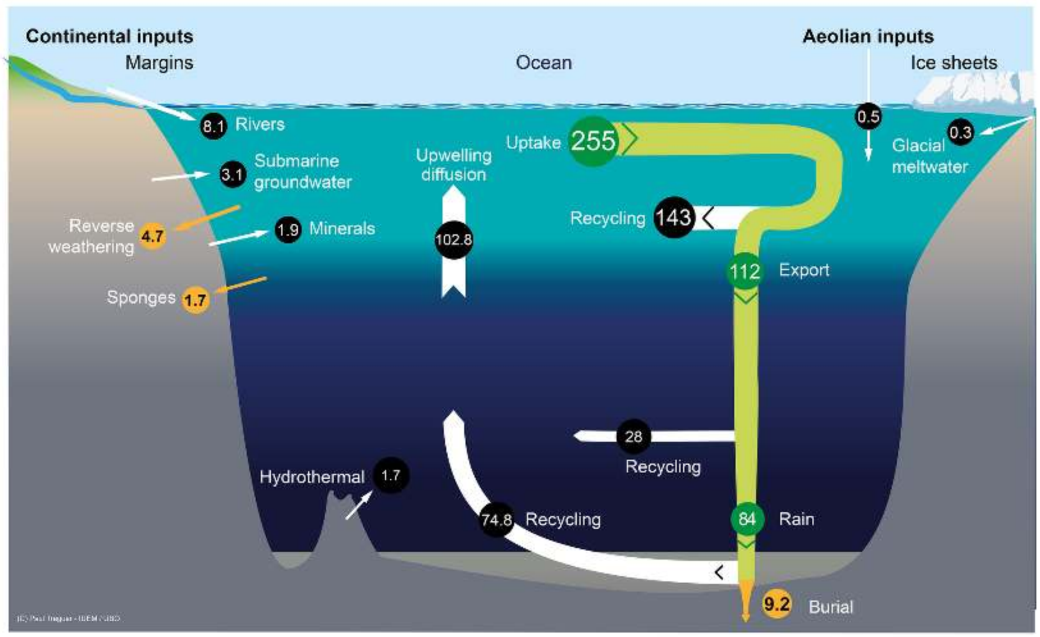

Figure 7.2.1: *Schematic view of the silicon cycle in the modern world ocean (input, output, and biological silicon fluxes). The white arrows represent fluxes of net sources of dissolved silicic acid (dSi) and/or of dissolvable amorphous silica (aSi) and of dSi recycled fluxes. Orange arrows correspond to sink fluxes of silicon (either as biogenic silica or as authigenic silica). Green arrows correspond to biological (pelagic) fluxes. All fluxes are in teramoles of silicon per year (Tmol Si yr^{-1}). Figure from Tréguer et al. via Wikimedia Commons; CC BY SA 4.0.*

7.2.2: Common siliceous sedimentary rocks

Bedded cherts form in the deep ocean as siliceous skeletons rain down from shallow waters. Although siliceous organisms occur across the entirety of the ocean, their contribution is generally overwhelmed by calcareous skeletal material. However, in the deep ocean (below the carbonate compensation depth) or in areas where upwelling occurs, the water chemistry can result in the accumulation of siliceous oozes and environmental conditions that are not favorable for carbonate deposition.

Nodular cherts form in carbonates and other sedimentary rocks when primary components of the rock are replaced by microcrystalline quartz during diagenesis.

Banded Iron Formations (BIFs) are largely Precambrian deposits that are composed of alternating layers of chert and iron oxides (typically hematite or magnetite). Although the exact mechanism for BIF formation is not well understood, most workers agree that they reflect profoundly different ocean chemistry during the Precambrian (less oxygen) that allowed for upwelling of silica and iron-rich waters and regular changes in oxygen levels that resulted in the precipitation of iron oxides. Beyond that, there is ongoing discussion of the nature of diagenetic/metamorphic overprinting of iron-bearing phases and the direct/indirect involvement of microorganisms beyond the production of oxygen.

Silcretes are resistant duricrusts formed at or near the sediment surface by the accumulation of silica through pedogenic processes, groundwater precipitation, or evaporation in ephemeral lakes.

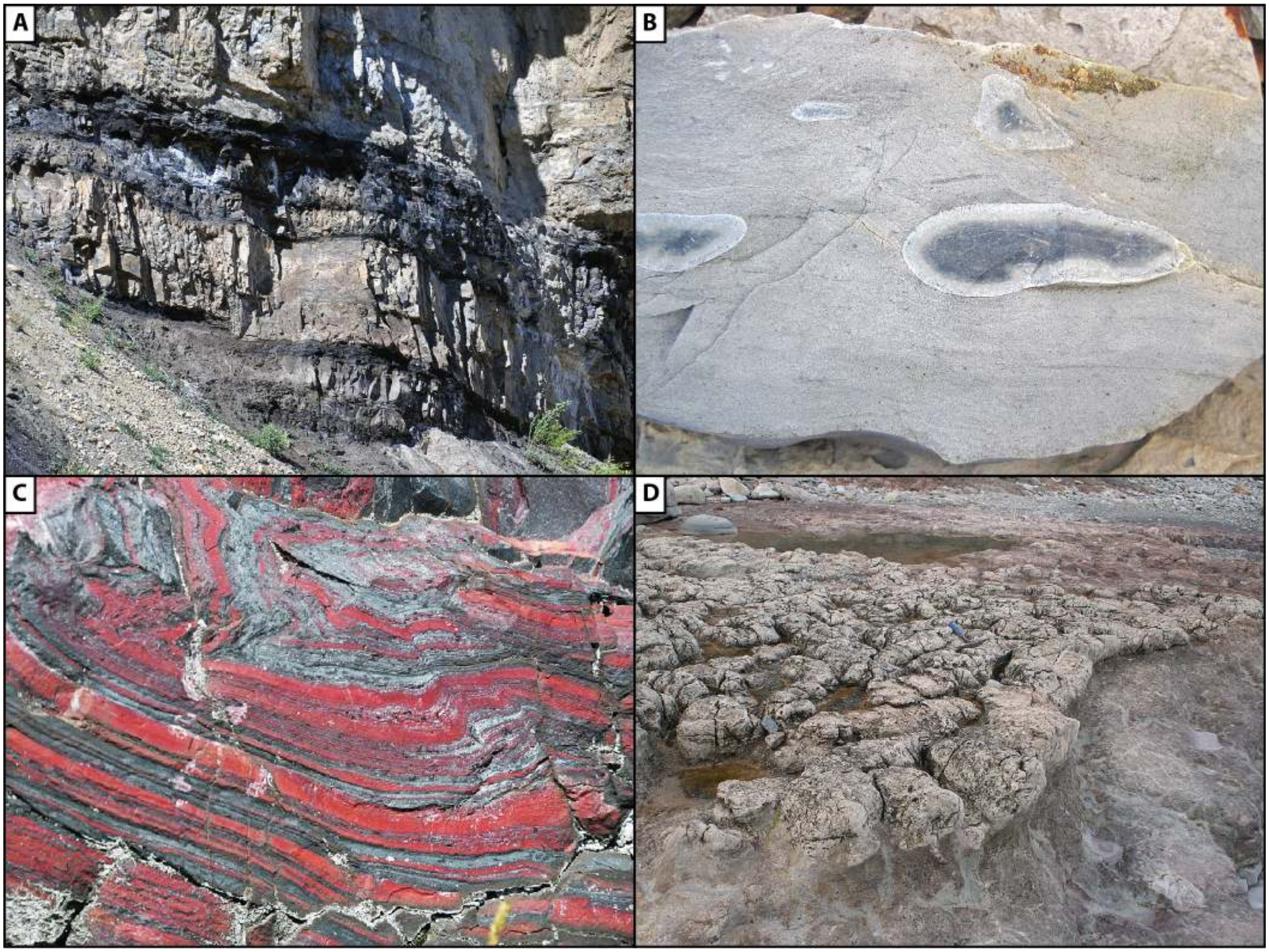

Figure 7.2.1: Outcrop photographs of siliceous sedimentary rocks. A) Gray, tan, and black bedded chert in the Permian Phosphoria Formation (James St. John via Wikimedia Commons, CC BY 2.0). B) Black chert nodules in lthe Devonian Delaware limestone (James St. John via Wikimedia Commons, CC BY 2.0). C) Hematite, red chert ("jasper"), and specular hematite in the Precambrian Negaunee Iron-Formation of Michigan (James St. John via Wikimedia Commons, CC BY 2.0) D) A nodular silcrete paleosol in the Pennsylvanian Waddens Cove Formation of Nova Scotia (Michael C. Rygel via Wikimedia Commons, CC BY-SA 3.0).

7.3: Organic-Rich Sedimentary Rocks

Organic-rich rocks (we can arbitrarily define them as having >5% organics) have 3 main components:

1. Clastic particles - sand, silt, clay and other types of material that were transported by wind or water.

2. Humus – dark fibrous plant material found in terrestrial settings.

3. Sapropel – jelly-like ooze of plant remains (usually algae) putrefying in an anaerobic environment

7.3.1: Petroleum-bearing rocks

For oil and natural gas to accumulate in significant amounts, all components, conditions, and processes of a "petroleum system" must align.

Petroleum is derived from a sapropel-rich source rock that typically formed in aquatic environments where the remains of algae and plankton were able to escape decay through burial in oxygen-poor settings. As sediment continues to accumulate, the organic material is eventually transformed into kerogen, which is a waxy, insoluable organic residue. Given the microscopic nature of kerogen, geochemical analysis is necessary to know what the source was and to predict how gas or oil prone the kerogen is. The three types of kerogen include:

- Type 1 - made mostly of algal remains; likely to produce oil
- Type 2 - a mixture of terrestrial and marine organics; may produce oil
- Type 3 - mixture of woody and algal material; likely to produce natural gas

As pressures and temperatures increase with burial of the source rock, the kerogen is eventually converted to oil and/or natural gas. Peak oil generation (the "oil window") typically occurs between 75 and 150 ^{0}C and peak gas production occurs at temperatures between 125 and 200 ^{0}C.

Accumulation of significant volumes of oil and/or natural gas require numerous other conditions to be met. In the case of a conventional petroleum system, the hydrocarbons must migrate into a reservoir rock that is capped by a seal rock that forms part of a structural or stratigraphic trap. Unconventional systems occur where the hydrocarbons cannot be recovered through traditional vertical drilling techniques; they include tar sands, biodegraded heavy oil, tight reservoirs that have to be horizontally drilled and hydraulically fractured, and others.

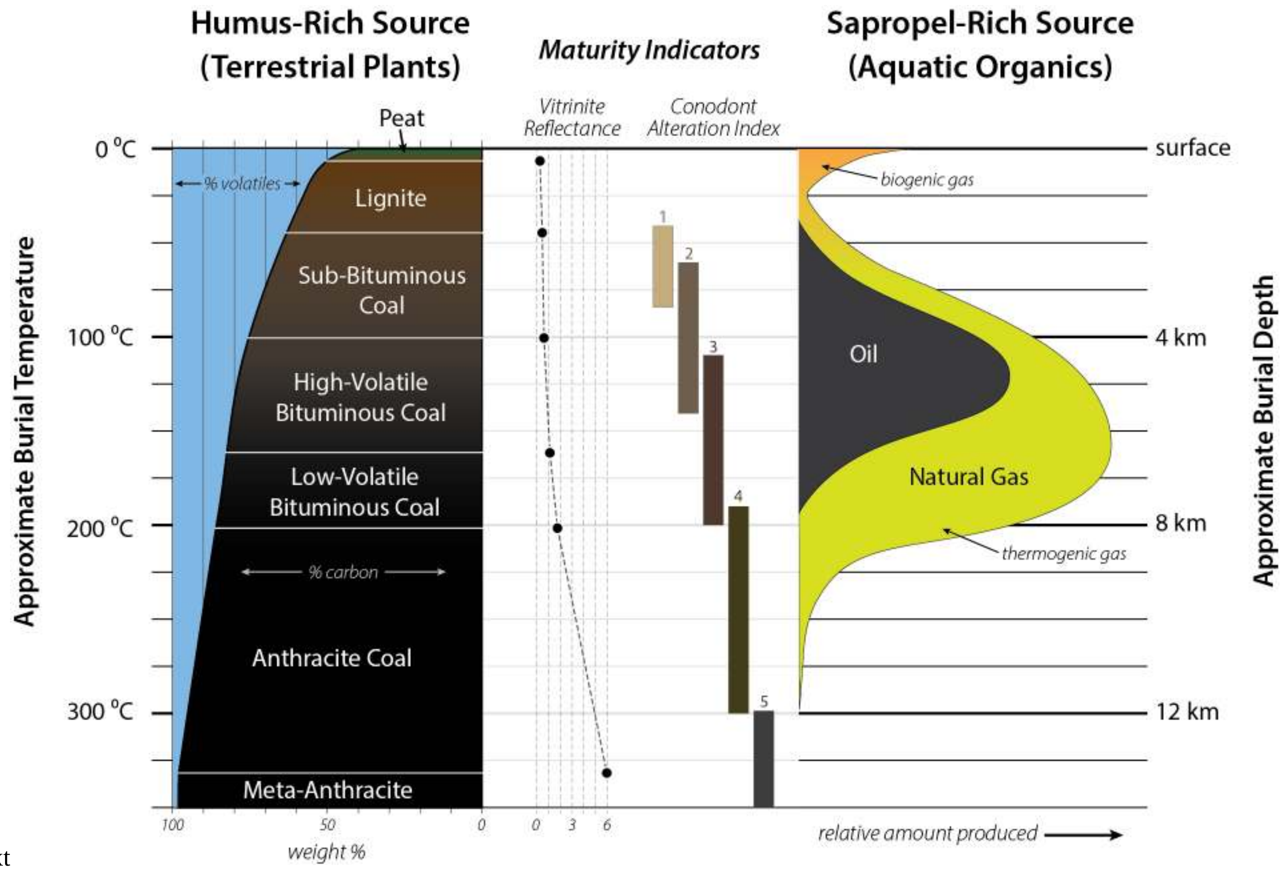

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Figure 7.3.1*:* Diagram showing the approximate relationship between hydrocarbon generations, source rock type, burial depth and temperature for an area with a geothermal gradient of 25 ^{0}C/km. These systems are complex and time dependent; all values and relationships are rough approximations (Michael C. Rygel via Wikimedia Commons; *CC BY SA 4.0)*.

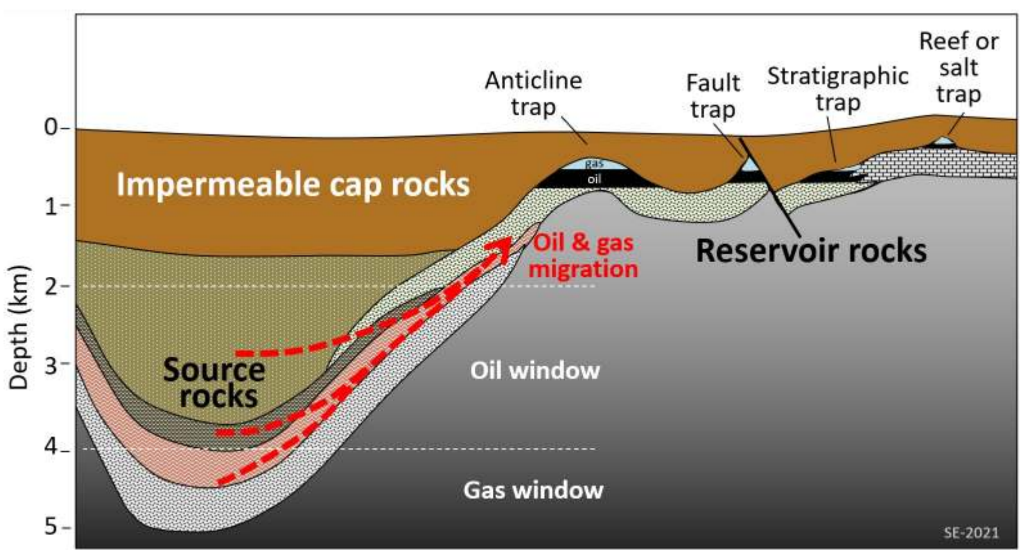

Figure 7.3.2: Elements of a conventional petroleum system include a source, seal (cap rock), reservoir and trap. Note that this area has a higher geothermal gradient than the previous figure causing the oil and gas windows to occur at shallower depths (Steven Earle in 20.3 Fossil Fuels; CC BY 4.0).

Figure 7.3.3: Examples and outcrops of petroleum systems and petroleum-bearing rocks. A) Although many reservoirs have adequate pressure to drive liquid petreoluem to the surface through the well bore, pumpjacks can be used to mechanically lift oil in systems that lack adequate pressure (Antandrus at English Wikipedia via Wikimedia Commons; CC BY-SA 3.0). B) When liquid petroleum reaches or approaches the surface via seeps or erosion of exposed reservoirs, the light component of the crude oil evaporates away and biodegradation occurs leaving behind viscous pitch/tar/liquid asphalt. Perhaps the best known example in North America is the La Brea Tarpits, the famous fossil locality near Los Angeles (Buchanan-Hermit via Wikimeda Commons; CC BY). D) Biodegraded "heavy oil" can also occur in tar sands where biodegradaded oil occurs in the pore spaces of poorly lithified sediment (Michael C. Rygel via Wikimedia Commons; CC BY SA 3.0).

7.3.2: Coal-bearing rocks

The organic material that forms most coal comes from humus (terrestrial plant material) which, under swampy reducing conditions can be preserved as peat. By definition, coal has to be at least 50% weight percent organic material. Economic coal deposits occur in discrete seams (beds) but small fragments of plants can be coalified and preserved within other sedimentary rocks. Although we think of coal as a massive and homogeneous material, many coal seams contain layers of impurities (which turn into ash when burned) and a variety of different types of plant material (macerals) that can be identified with microscopic analysis and sometimes in hand sample.

In certain swampy, clastic-poor environments significant accumulations of plant material can escape decay and accumulate to form relatively thick layers of peat. With increased burial and compaction, volatiles (water and gasses) are driven out and the percentage of carbon increases and the peat can progressively transitions to lignite, bituminous coal, and eventually anthracite coal.

Given its composition, coal occurs only in Devonian or younger rocks; the majority of it accumulated in the Late Paleozoic when vast areas of continental crust were subsiding in response to mountain building and within the humid tropics.

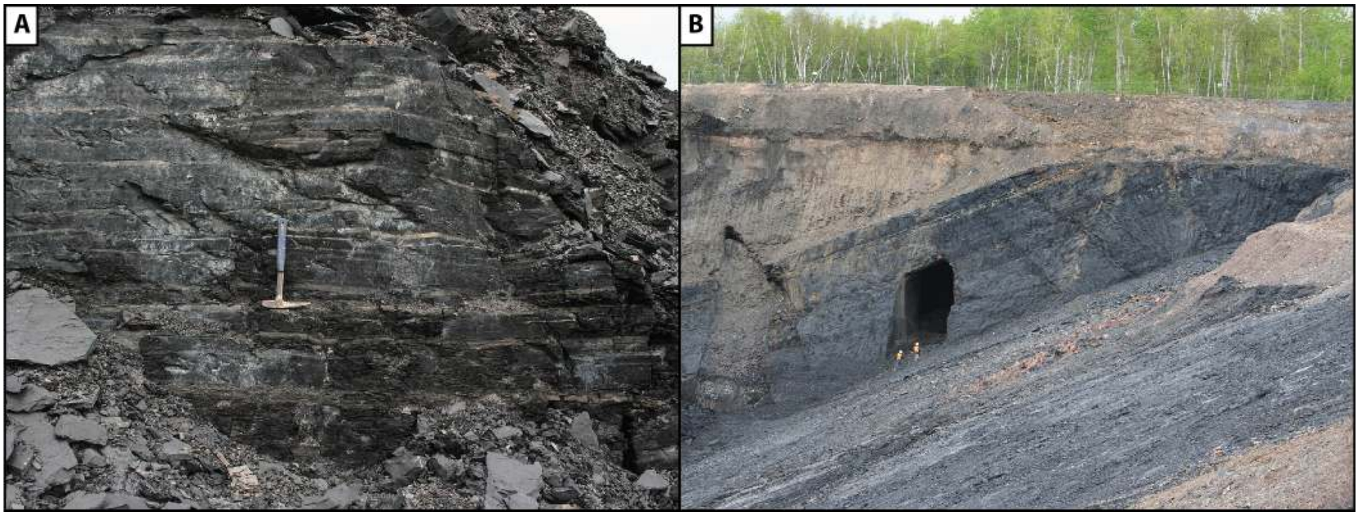

Figure 7.3.4: Outcrops of coal. A) We commonly think of coal as a homogenous black rock, but it often contains bright bands, dull bands, fossils and other identifiable original components, and clastic splits (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). B) Outcrop of the Foord Seam in a strip mine in Nova Scotia that intersected older underground workings (Michael C. Rygel via Wikimedia Commons; CC BY SA 3.0).

7.3.3: Readings and Resources

- Petroleum System on the AAPG Wiki: https://wiki.aapg.org/Petroleum_system
- Kentucky Geologic Survey's webpage on coal: https://www.uky.edu/KGS/coal/

CHAPTER OVERVIEW

8: Diagenesis

Diagenesis is the transformation from sediment into sedimentary rock. It includes everything that is post-depositional, but pre-metamorphic. It's an important time in the history of the rock because of the potential for compaction, textural changes, movement of fluids through pore spaces, as well as the creation, migration, and ultimate fate of fossil fuels.

Learning Objectives

- Explain the most important diagenetic processes
- Identify diagenetic textures and structures and explain their origin.

Chapter thumbnail shows a stylolite cross cutting an ooid in the Cambrian Meagher Formation, SW Montana (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

8.1: Diagenetic Processes

8.1.1: Compaction

Compaction is a reduction in the volume of sediment due to the weight of the overburden and resulting burial pressure. In terms of a starting point, loose sand has a theoretical maximum of 47.6% pore space and wet mud could be 60-90% pore space. Once buried, there can be a marked reduction in volume/pore space; some example values include:

- ~1.1 m sand = 1 m sandstone
- ~2.6 m mud = 1 m mudrock
- ~5 to 10 m peat = 1 m coal

Carbonates are extremely variable, with lime mudstones behaving like shales and reef-type boundstones experiencing very little compaction.

Although significant volume loss can occur simply by squeezing out water and gasses, deformation can also occur at the scale of an individual grain by a) ductile deformation where grains that behave plastically get deformed between more robust grains, b) flexible bending of grains like micas, and c) pressure induced changes in grain contacts. Specifically, grains start out with floating or tangential (point) contacts and that increased pressure with burial causes a progressive shift to long (appears like a straight-ish line), concavo-convex (appears like a curved line), and eventually complexly sutured contacts caused by dissolution and "welding" of grains.

Compaction generally increases with depth and is accompanied, at least initially, by a decrease in porosity caused by grain scale processes and cementation. Secondary porosity can be generated by dissolution of unstable components.

8.1.2: Cementation

Cementation is the second major diagenetic process and it occurs when minerals precipitate from fluids into open pore spaces. Well sorted sediments with open and interconnected pores are particularly prone. The most common cements are calcite (which forms from solutions with a pH >7) and silica cement which forms from solutions with a pH<7). Pores can also fill with water, gas, or oil.

8.1.3: Other processes

Although compaction and cementation are the most important processes, other diagenetic changes can also occur and may be locally significant. Examples include:

- Replacement happens when one mineral takes the place of another. Common examples include the alteration of feldspars to clay minerals and pyrite replacing calcite.
- Recrystallization is the transformation of a mineral to its polymorph or another form of itself (change in crystal size or morphology). An example would be the transformation of aragonite to calcite.
- Bioturbation is the disruption of sediment caused by the activity of organisms. It results in everything from discrete tracks, trails, and burrows to the complete destruction of bedding.
- Pedogenesis is a term used for all of the soil forming processes, including wetting and drying, the action of plants and animals, and weathering which can destroy original structure of a rock. During pedogenesis (and after) the movement of water and changes in oxidation may also cause mottling of rock.
- Dissolution happens when select components are dissolved from the rock.

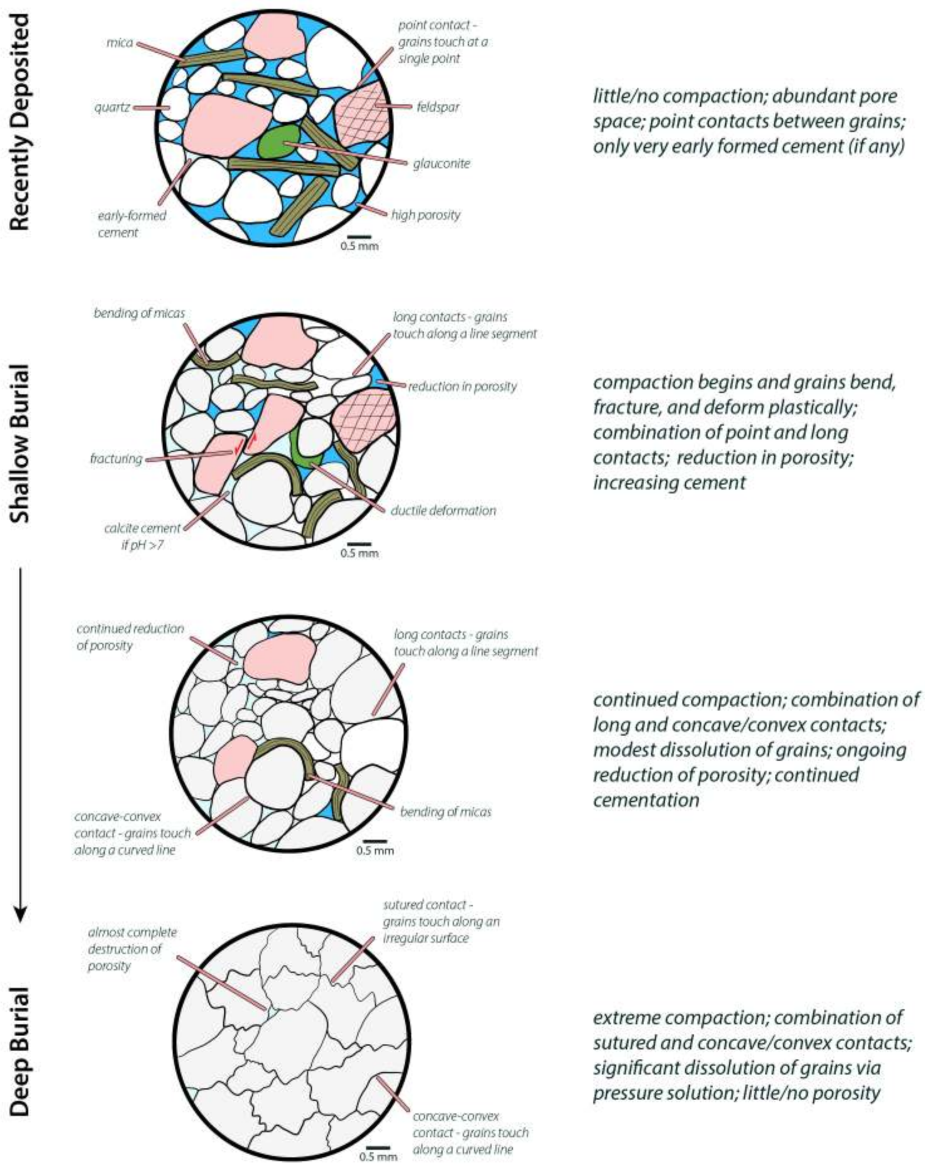

Figure 8.1.1: Grain-scale diagenetic changes to a sandstone caused largely by compaction and cementation (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

8.1.4: Readings and Resources

- Taylor, J.M., 1950, Pore-Space Reduction in Sandstones. *AAPG Bulletin*; 34 (4): 701–716.
- https://www.geological-digressions.com/tag/stylolites-stylolitization/

8.2: Diagenetic Structures

Admittedly there is no clear temporal cut-off between the secondary structures described in Chapter 4 (soft-sediment deformation structures, mudcracks, etc.) and the diagenetic structures list below - except to say that the diagenetitc structures generally form somewhat later and might be more disconnected from the processes that were active when the sediment was being deposited.

8.2.1: Diagenetic features formed by the growth of minerals/cement

Concretions are concentric accumulations of mineral cement. They come in a wide variety of shapes and sizes. Rhizoconcretions (aka rhizoliths) are a specific type of concretion developed around roots and commonly made of calcite or siderite.

Nodules are similar to concretions, except that they are internally massive. Distinctive types include:

- Caliche nodules – composed of calcite formed in association with soil forming processes
- Bauxites and laterites – nodular soils composed of iron- and alumina-rich minerals that form in response to intense tropical weathering.
- Septarian nodules – nodules with radial or concentric cracks, commonly filled with calcite or siderite crystals
- Desert rose – although it does not exactly fit the definition of a nodule, it is a rose- or flower shaped accumulation of gypsum

Liesegang rings – colored bands that formed as minerals precipitated from groundwater; commonly cut across bedding and may form elaborate curved patterns.

Manganese dendrites – chemical precipitate of MnO_2 that forms on rock faces or bedding planes as a result of subaerial weathering. Commonly takes on a fern- or plant-like appearance.

Mottling – zones of distinctive color within rocks formed as a result of geochemical variations within the rock. May develop around organic-rich zones or areas with differences in porosity/permeability. Some may form circular or spherical reduction halos.

8.2.2: Features associated with dissolution

Geodes – layered accumulations of minerals that infill voids. They can form in voids of any origin and commonly form in cavities formed by dissolution of carbonates

Geopetal structures - small cavities formed beneath shells, etc. that contain crystals that are in the formerly open spaces. Useful for determining way up.

Secondary porosity - pores develop from the removal of material; commonly by selectively targeting certain grains or fossils.

Stylolites – irregular, jagged lines caused by pressure solution and the accumulation of insoluble residues.

8.2.3: Structures formed at the surface

Pedogenic slickensides – polished surfaces in mudrock formed in response shrinking and swelling associated with soil-forming processes. A hallmark of Vertisols.

Bioturbation – tracks, trails, burrows, or any feature formed from the activity of organisms.

8.2.4: Recrystallization structures

Cone-in-cone structures - zones within limestones that have fibrous calcite crystals that form a zig-zag pattern. They form in response to compaction and recrystallization.

Pseudospar – Coarsely crystalline calcite formed from the recrystallization of fine-grained lime mud (micrite).

8.2.5: Replacement features

Dolomitization – calcite/aragonite is replaced by dolomite. Dolomite shows up as rhombohedral crystals in thin section.

Replaced fossils (or other material) - forms when silica, pyrite, or some other mineral replacement occurs. Carbonate skeletons are prone to replacement.

Chickenwire structures – distinctive nodular zones in evaporates, commonly formed in response to the transition from gypsum to anhydrite and the incorporation of clastic sediment.

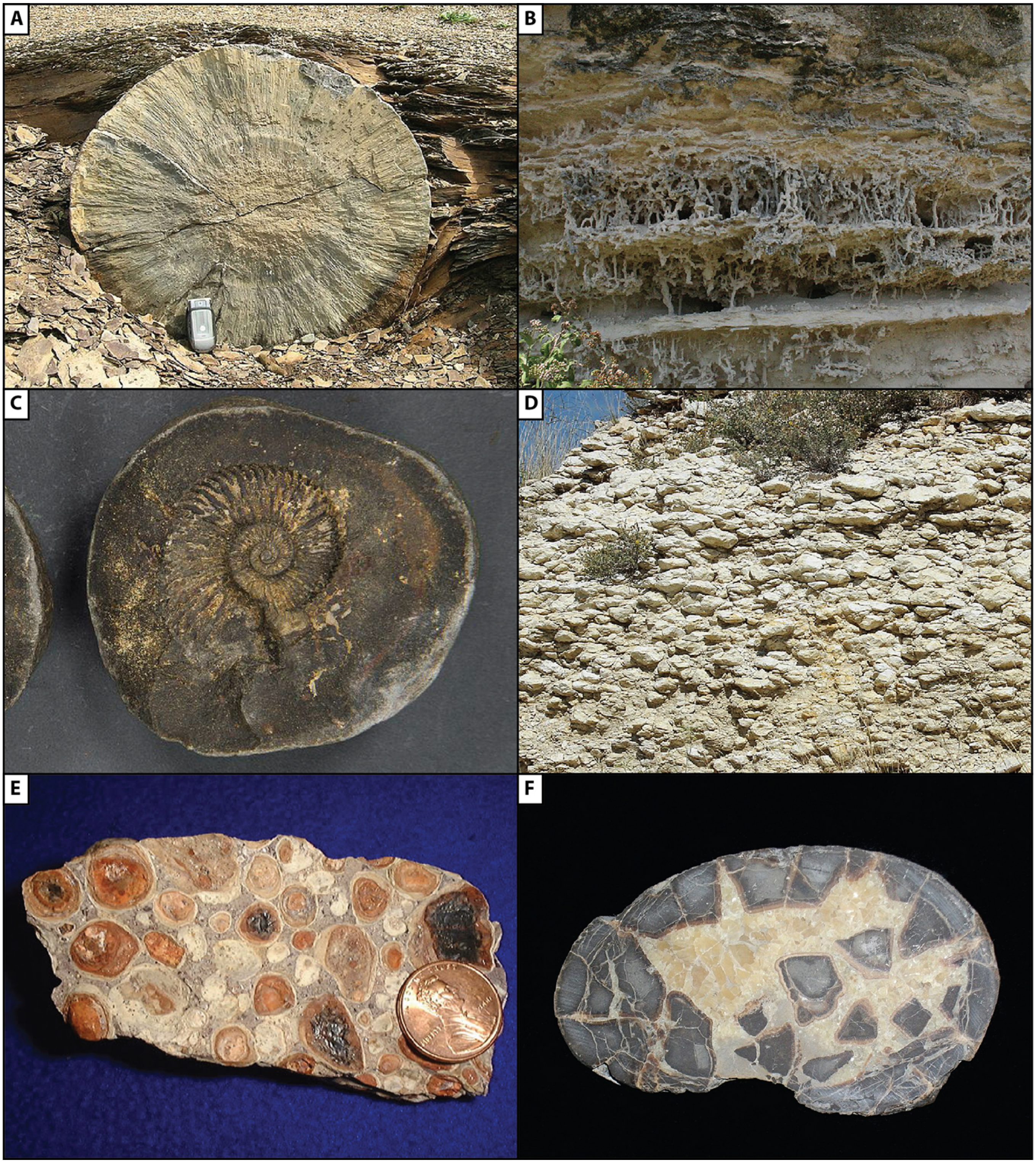

Figure 8.2.1: Diagnetic features formed in association with the growth of minerals or cement, part 1. A) Cross-section through a concretion. Although this example has a striking radial fabric, you can still see faint concentric layering (Mark Buchanan via Wikimedia Commons; CC BY-SA 3.0). B) Vertical rhizoconcretions developed around roots (Jfoote via Wikimedia Commons; CC BY-SA 3.0). C) Internally massive nodule developed around a coiled cephalopod (Hannes Grobe/AWI via Wikimedia Commons;

CC BY-SA 4.0). D) Dense accumulation of modern caliche nodules in Texas (David R. Tribble via Wikimedia Commons; CC BY-SA 3.0). E) Nodules in a sample of bauxite (USGS via Wikimedia Commons; public domain). F) Septarian nodule with distinctive cracked texture (Neptunerover via Wikimedia Commons; public domain).

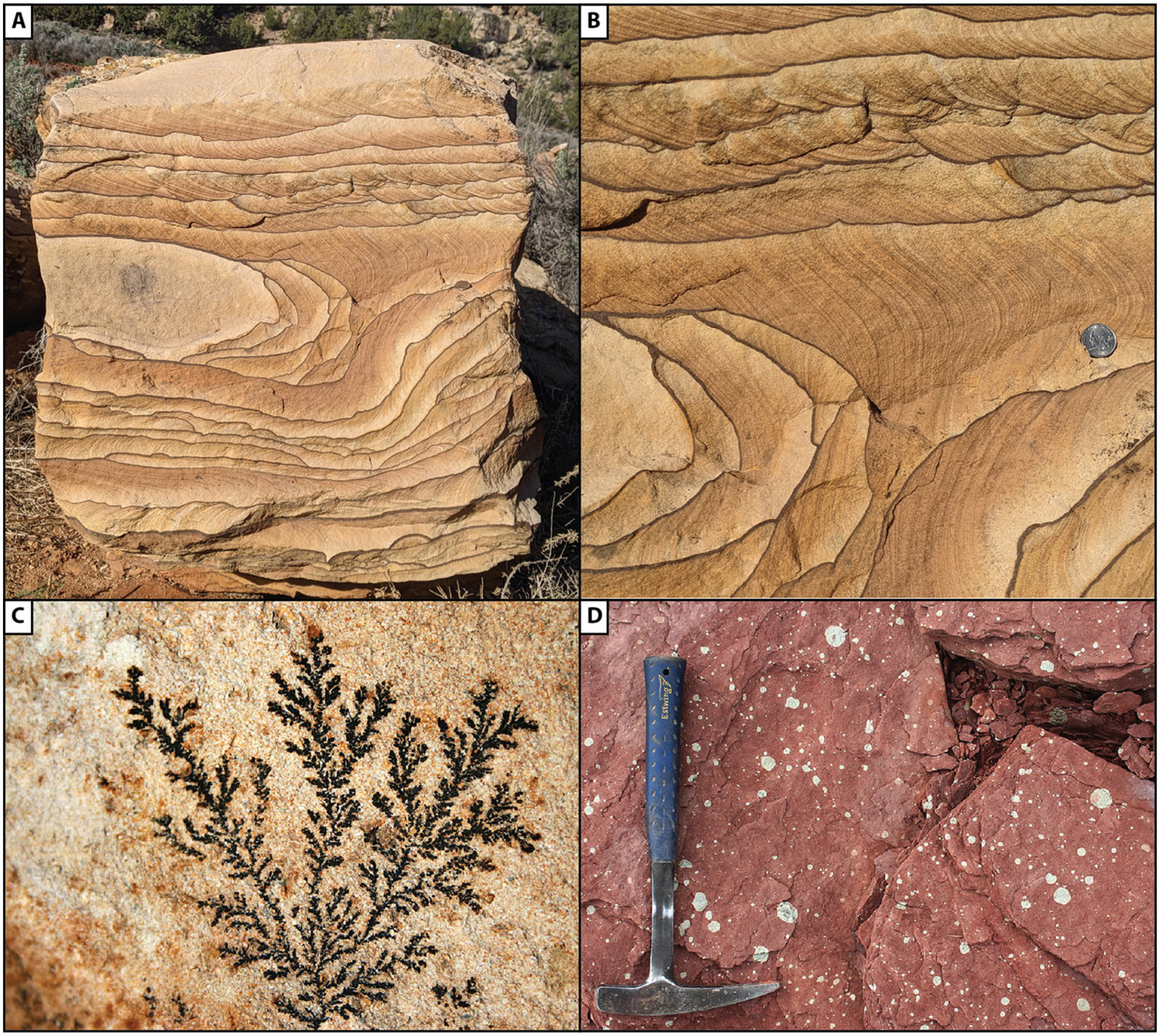

Figure 8.2.2: Diagnetic features formed in association with the growth of minerals or cement, part 2. A & B) Liesegang rings in sandstone. B shows a detail of the right center of the boulder in A. C) Manganese dendrite in finely crystalline volcanic rock. Unknown unit and age. D) Green mottles (reduction halos) developed around organic material in a sandstone from the Mississippian Mabou Group, Nova Scotia. All images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0.

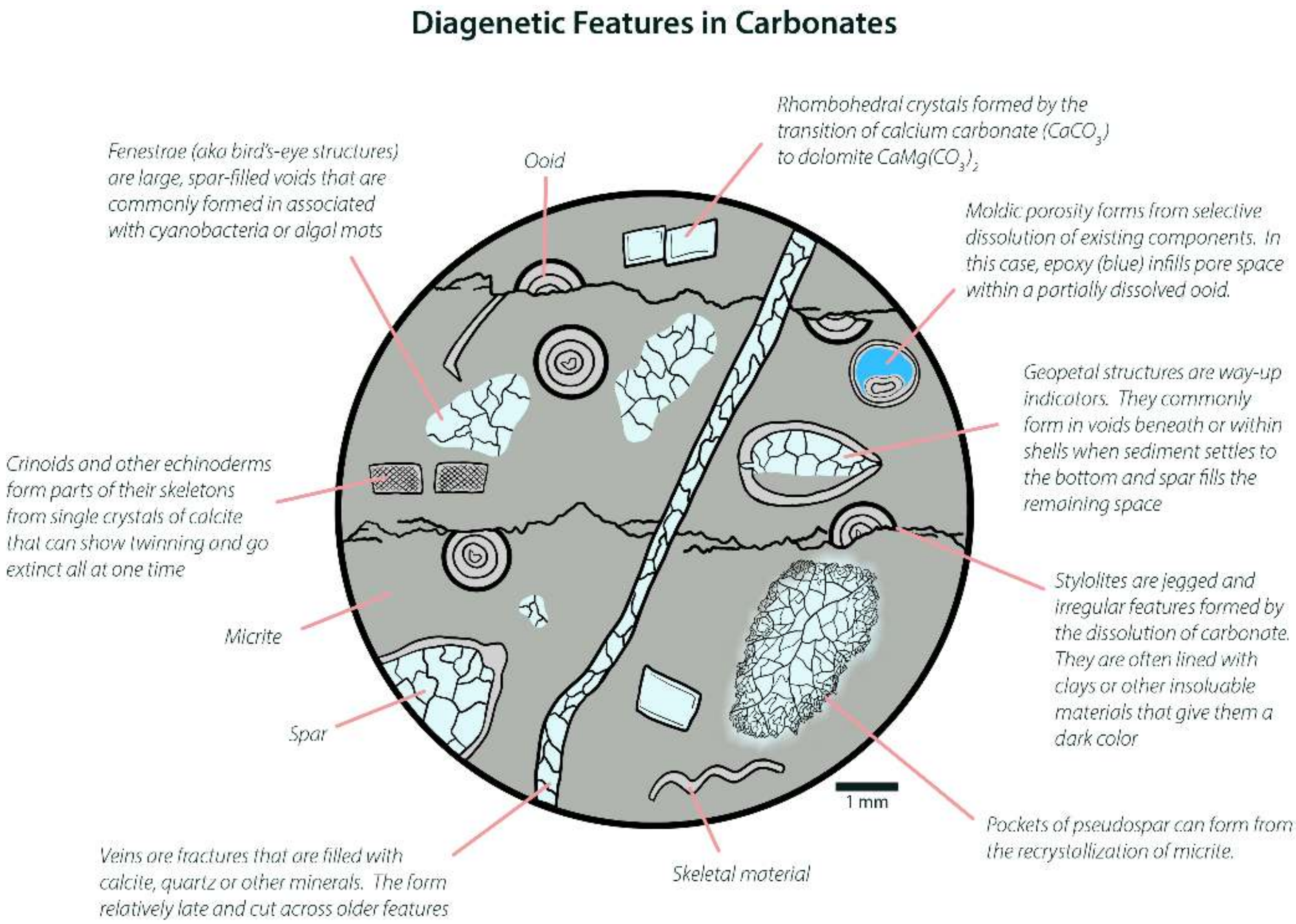

Figure 8.2.3: Carbonate diagenetic features in thin section. (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

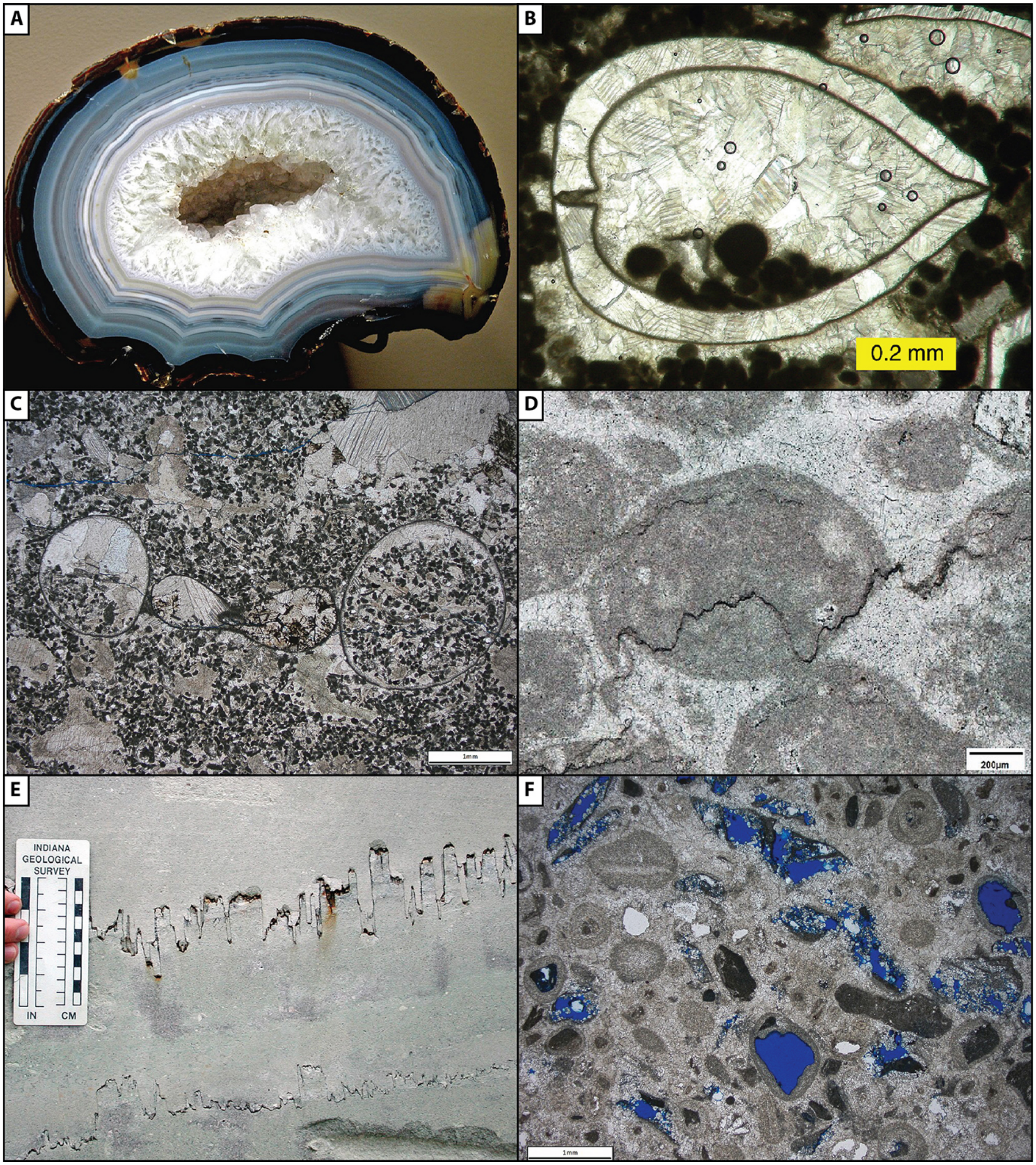
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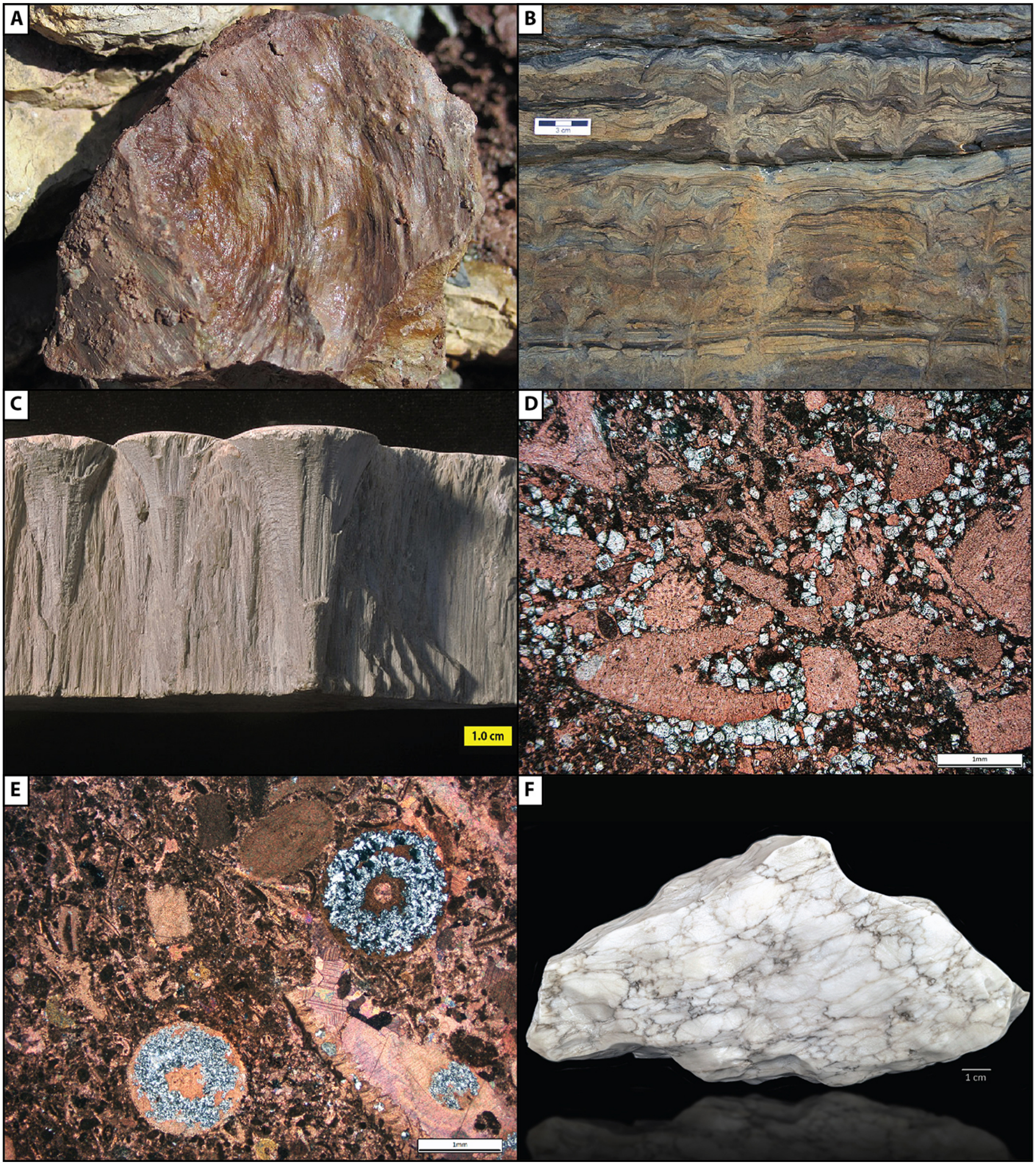
Figure 8.2.4: Diagenetic features formed in association with dissolution. A) Geode infilled with quartz crystals (James St. John via Wikimedia Commons; CC BY 2.0). B) Thin-section image of peloids and calcite cement inside of a recrystallized bivalve shell. The peloids settled to the bottom of the shell before the calcite was precipitated and thus are geopetal structures showing that up was toward the top of the page (Mark Wilson via Wikimedia Commons; public domain). C) Similar geopetal structure formed by sediment inside of a coiled gastropod or cephalopod shell (Michael C. Rygel via Wikimedia Commons; CC BY -SA 4.0). D & E show stylolites in thin section and in outcrop, respectively. F) Moldic porosity in a limestone formed by the selective dissolution of ooids. D-F from Michael C. Rygel via Wikimedia Commons; CC BY -SA 4.0.

Next page »
Figure 8.2.5: Diagenetic features formed at the surface, in association with recrystallization (A-B) or replacement (C-F). A) Pedogenic slickensides in a Mississippian paleosol (James St. John via Wikimedia Commons; CC BY 2.0). B) Trace fossils in Permian-aged shallow marine/deltaic sediments. C) Cone-in-cone limestone (Mark Wilson via Wikimedia Commons; public domain). D) Partially dolomitized fossiliferous limestone (PPL). Slide was stained for calcite which makes dolomite crystals (white) stand out from the fossils (light red) and micrite (dark reddish brown). E) Photomicrograph (XPL) of partially silicified fossils in limestone; slide was stained for calcite. F) Gypsum and anhydrite showing "chicken wire" texture. B, D, E, and F from Michael C. Rygel via Wikimedia Commons; CC BY -SA 4.0.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CHAPTER OVERVIEW

9: Fossils

Fossils are defined as any evidence of past life; for our purposes we will consider something to be a fossil if it is unusual in that it is either ancient or has been altered from its original form in a way that makes it more likely to be preserved. Fossils are very important because they tell us about how organisms changed and evolved through time, how they interacted with each other and their environment, and how Earth’s climate and environments have changed through time.

Learning Objectives

- List the major types of fossils and modes of preservation
- Describe the morphology of fossils using the appropriate terminology
- Identify the major types of invertebrate body fossils in hand sample and thin section
- Identify the most common trace fossils and apply the ichnofacies concept to interpret depositional environments

Chapter thumbnail shows a fossil gastropod (Masha Milshina via Wikimedia Commons; CC BY-SA 4.0).

9.1: Types of Fossils

Despite the incredible diversity and complexity of the fossil record, we can lump fossils into three main types:

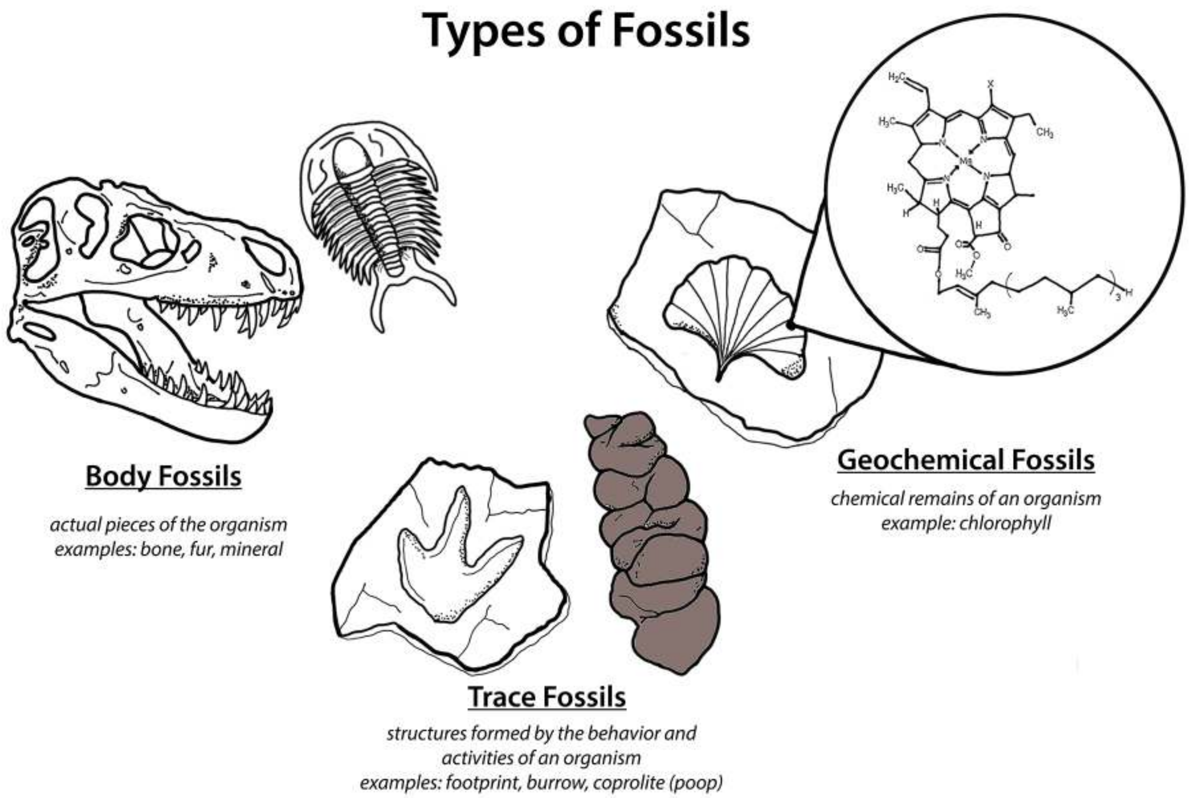


Figure 9.1.1: Types of fossils (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

Body fossils are preserved portions of the actual body parts of plants, animals, and other organisms. They include a wide variety of types including bones, teeth, scales, feathers, hair, shells, soft tissue (skin, muscle, etc.), as well as wood, roots, leaves, seeds, and many others.

Trace fossils are structures that are formed by the activity of an organism and commonly include features that disrupt layers of sediment (tracks, trails, burrows), erosion (boring, scraping, orbiting), the creation of layering or organization (various types), and even fossilized excrement (coprolites). Trace fossils record a wide variety of behaviors including feeding, resting, movement, dwelling and many others. It's worth noting that growth is not a behavior, so things like roots and stromatolites are best considered as body fossils.

Geochemical fossils are subtle chemical clues that are created by organisms and preserved in the geologic record. Perhaps the most common example is a mineral or rock that is enriched in carbon-12 because of the activity of photosynthetic organisms.

With practice, you will learn to identify pseudofossils - naturally occurring crystals, sedimentary structures, and weathering features that resemble body or trace fossils.

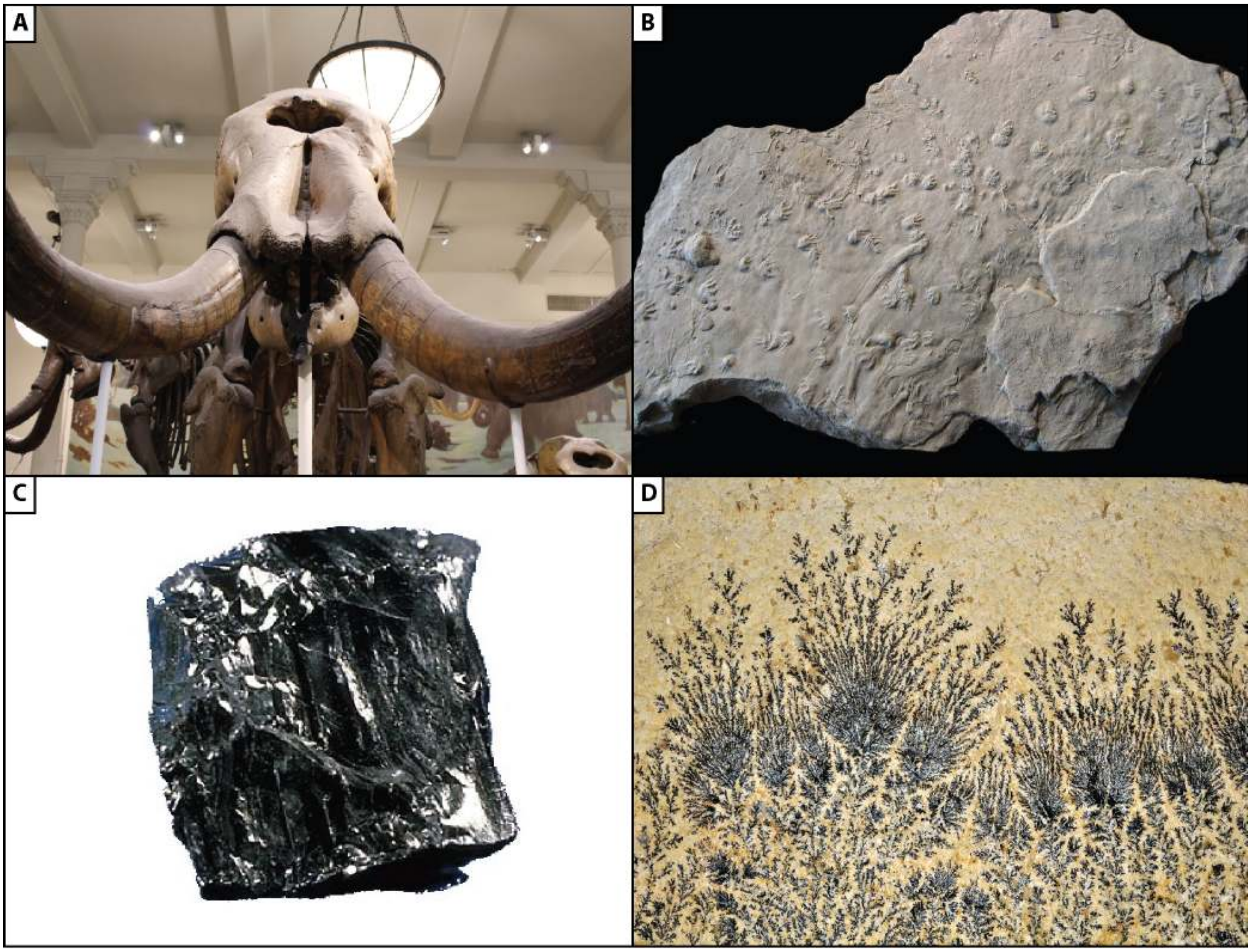


Figure 9.1.2: The three main types of fossils and something that isn't. A) Body fossils include things like this mammoth skull (Rodrigo Alomía Díaz via Wikimedia Commons; CC BY-SA 4.0). B) Trace fossils include features like these amphibian trackways preserved on the base of a bed of sandstone (Michael Rygel via Wikimedia Commons; CC BY-SA 3.0). C) Even if no recognizable plant material is preserved, coal could be considered a geochemical fossil because it is enriched in ^{12}C (U.S. Geological Survey via Wikimedia Commons; public domain). D) Although they look like plants, these manganese dendrites are pseudofossils composed of completely inorganic branching crystals. (Géry PARENT via Wikimedia Commons; CC BY-SA 3.0).

9.2: Types of Preservation

Preserving a recently deceased organism in the geologic record is challenging. This process typically requires one of the following:

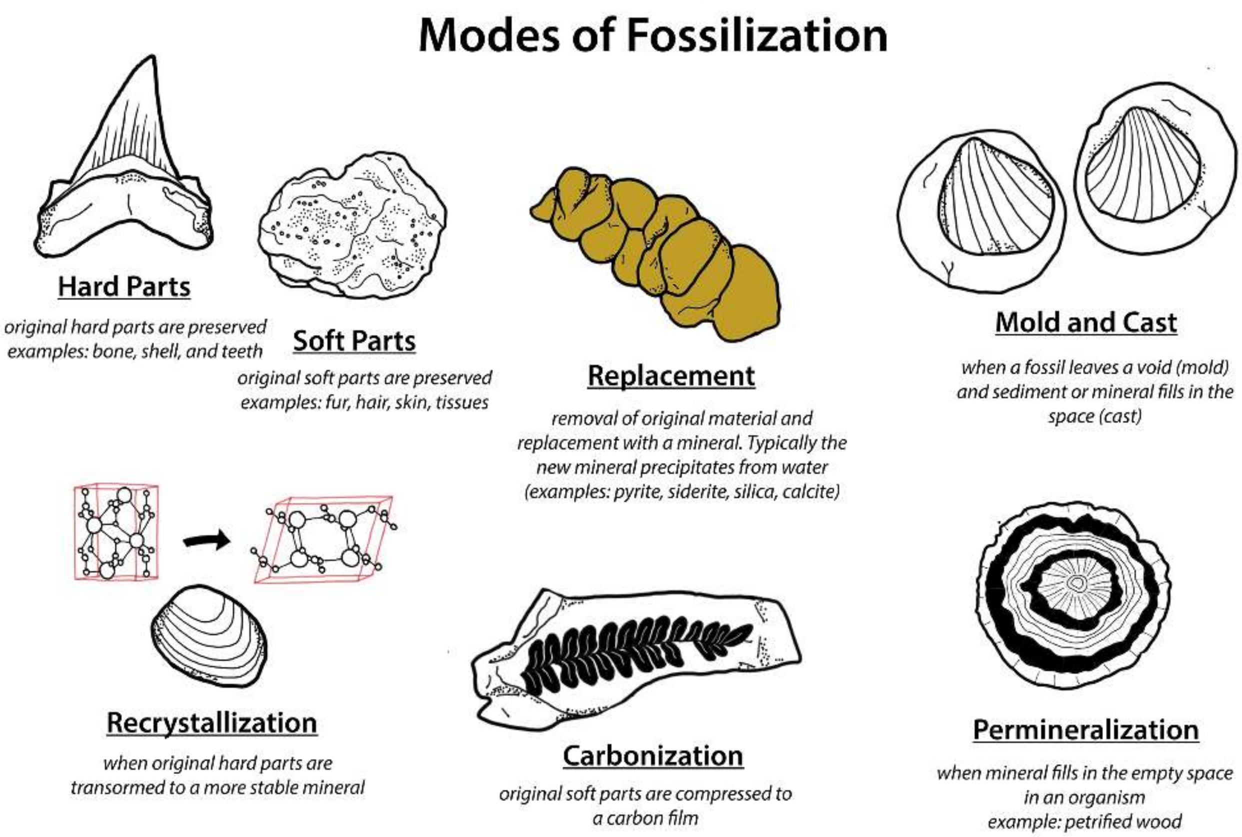


Figure 9.2.1: In order to be preserved as a fossil, organisms must undergo one of the modes of preservation shown above (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

9.2.1: Hard and Soft Parts

Preservation without alteration can happen when particularly robust hard parts of organisms like bones, shells, teeth, etc. are buried and protected from decay and damage. Under very special conditions such as freezing, entrapment in amber, or mummification the soft parts of organisms like hair, skin, or muscle tissue can be preserved.

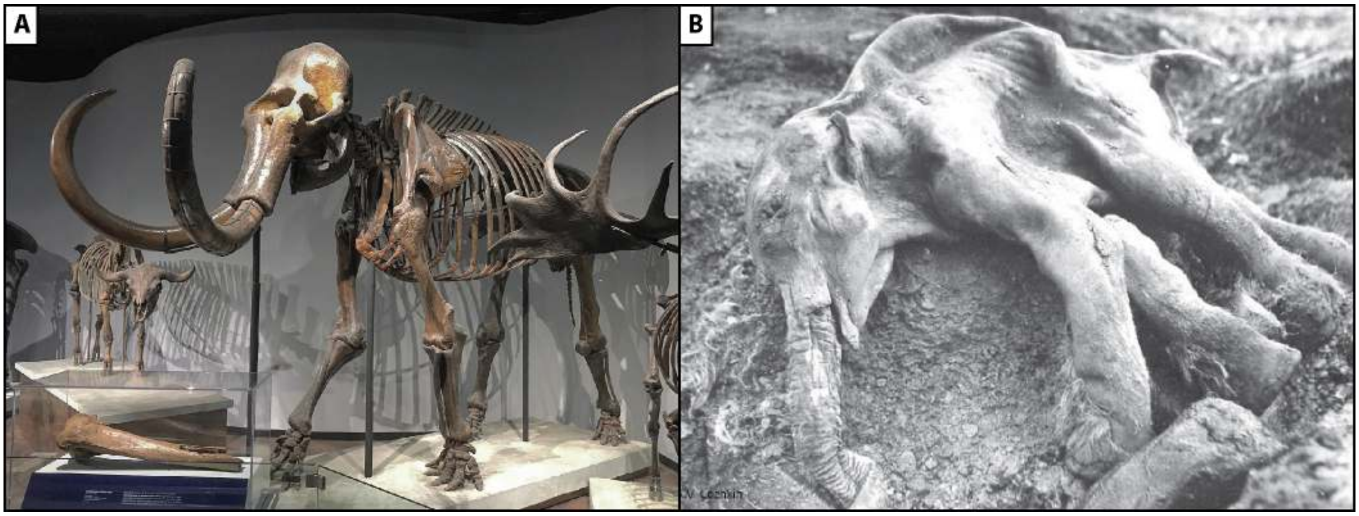


Figure 9.2.2: Hard parts and soft parts. A) The hard parts of this mammoth that are easily preserved include its bones and tusks (Zissoudistrucker via Wikimedia Commons; CC BY-SA 4.0). B) Under certain circumstances (like being preserved in permafrost), soft parts like the hide, hair, and flesh of this baby mammoth can be preserved (NOAA v*ia Wikimedia Commons; public domain).*

9.2.2: Preservation with Alteration

Preservation with alteration is the most common way of preserving body fossils and it happens when the original tissue is changed in some way from its original form. Recrystallization happens when the original material is transformed into a more stable crystal form; the most common example of this is the transition from aragonite (iridescent nacre or mother of pearl) into chalky calcite. Carbonization happens when the organism is compressed, and water/gas are driven off so that only a thin carbon film remains. When minerals precipitate out of groundwater to fill in open spaces in wood, bone or other porous tissue the process of permineralization takes place (ex: "petrified" wood formed when microcrystalline quartz fills in woody tissue). And lastly, replacement takes place when one mineral is precipitated in place of another (ex: pyrite replacing calcite).

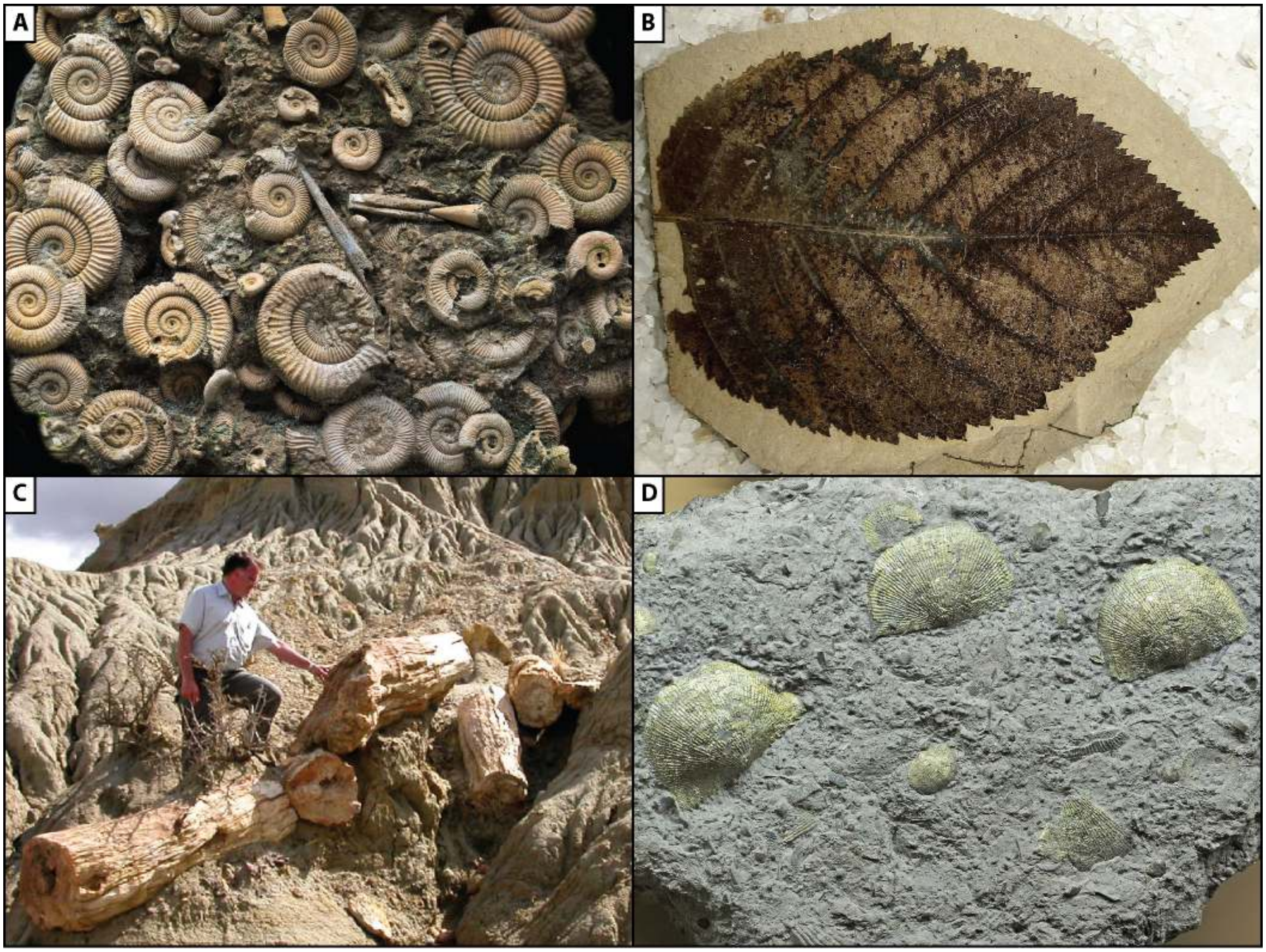


Figure 9.2.3: Preservation with alteration. A) Calcite replaced the aragonite that originally made up these fossil ammonites (Reinhold Möller via Wikimedia Commons; CC BY-SA 3.0). B) A carbonized leaf (Kevmin via Wikimedia Commons; CC BY-SA 3.0). C) Permineralized wood (Dhzanette via Wikimedia Commons; public domain). D) These shells were made of carbonate that was later replaced by pyrite (Luis Fernández García via Wikimedia Commons; CC BY-SA 4.0).

9.2.3: Casts, Molds, and Impressions

Sometimes even when the tissue itself isn't preserved, a fossil may be still be there in the form of a cast, mold, or impression. A cast occurs when sediment or minerals fill in a void space left by a decayed or dissolved organism. A mold is the name we use for the actual void. An impression happens when organic material is pressed into sediment and the texture of that material is preserved.

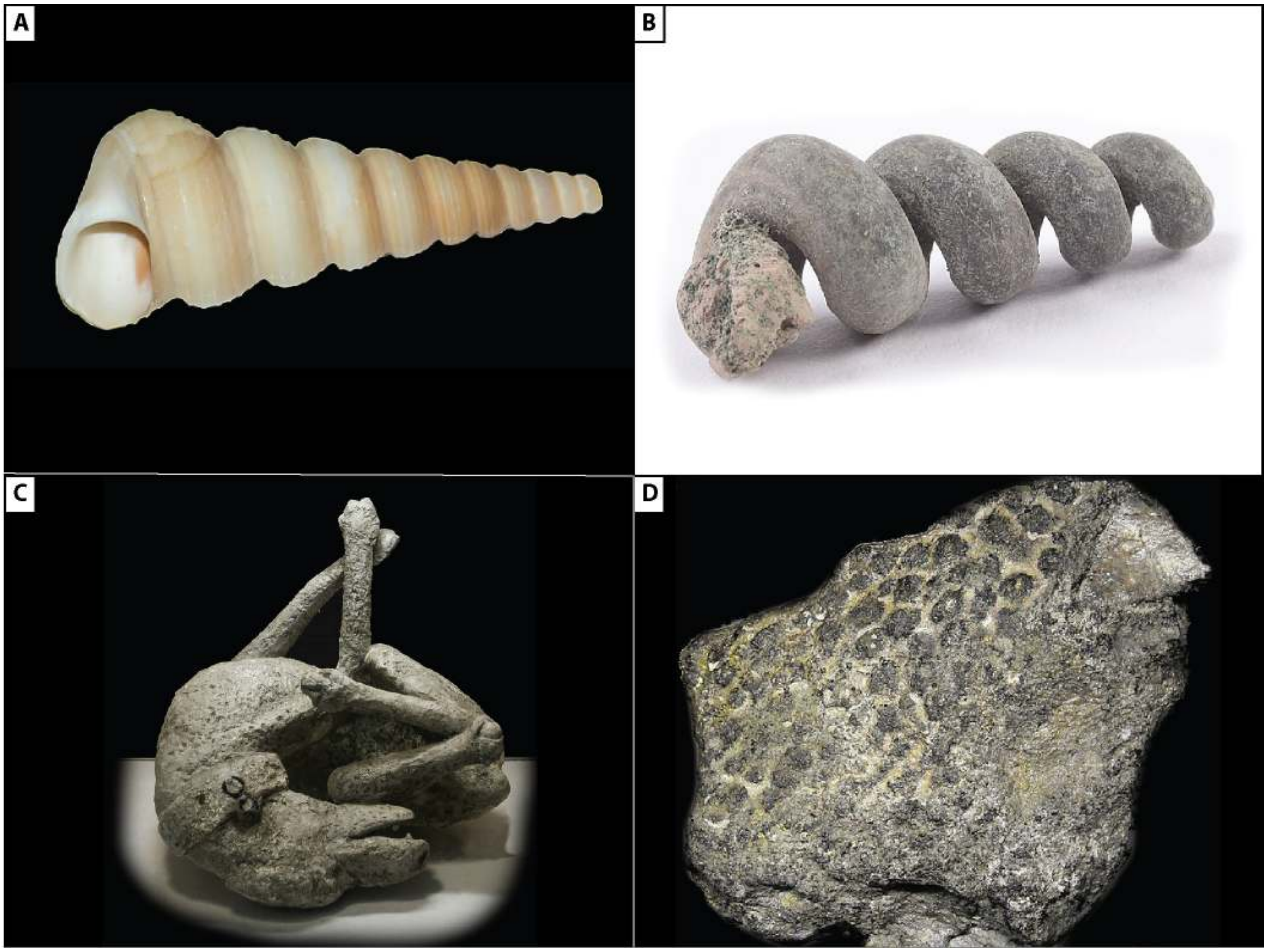


Figure 9.2.4: Casts, molds, and impressions. A) The void inside of this shell could form a mold if infilled with sediment or minerals as in the next figure (Dr. Zachi Evenor via Wikimedia Commons; CC BY 2.0). B) This cast formed when sediment infilled the void inside of a gastropod shell like the one in the previous figure (Rowan-earth via Wikimedia Commons; CC BY-SA 4.0). C) Cast of a sleeping dog killed and entombed by the eruption of Mount Vesuvius in 79 AD. D) Dinosaur skin impression (replica) from the Jurassic Morrison Formation (James St. John via Wikimedia Commons; CC BY 2.0).

9.3: Describing Fossils

Before we can start to identify and interpret fossils, we need to know what features to look for and what words to use to describe these features. Some of the most important characteristics are discussed below.

9.3.1: Holes

With certain types of invertebrate fossils, holes in the organism represent cavities where parts of an organism, or sometimes the whole organism lived. The number (single versus multiple), size (large versus small), and internal structure (simple versus complex) of these holes can all provide important clues about the type of organism that the fossil represents.

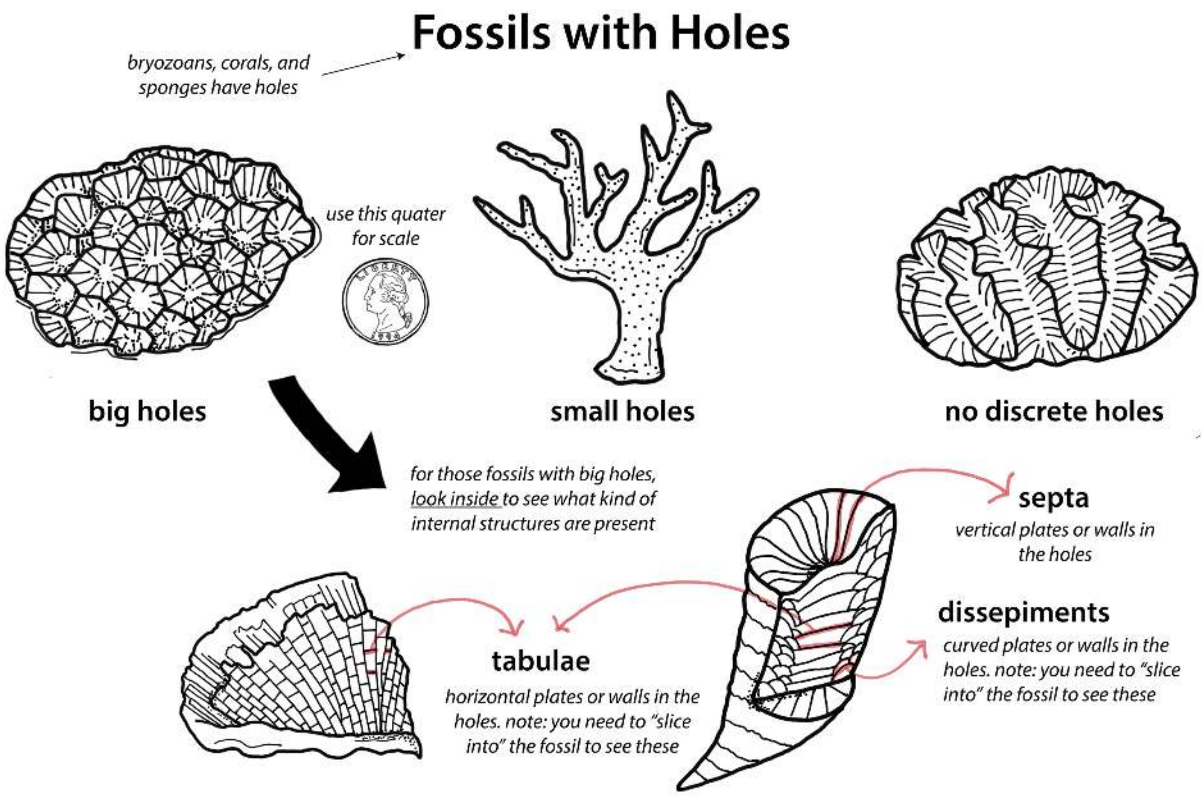


Figure 9.3.1: When describing fossils with holes we focus on the size of the holes and the internal structures within the holes. (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

9.3.2: Symmetry

Symmetry in a fossil occurs when you can draw a line through the organism in a way that splits it into two halves that are mirror images of each other. Radial symmetry happens with circular objects that have an almost infinite number of symmetry planes, penta-radial symmetry is when there are five symmetry planes, bilateral symmetry is when there is one symmetry plane that splits the organism into two halves, and some organisms have no symmetry at all.

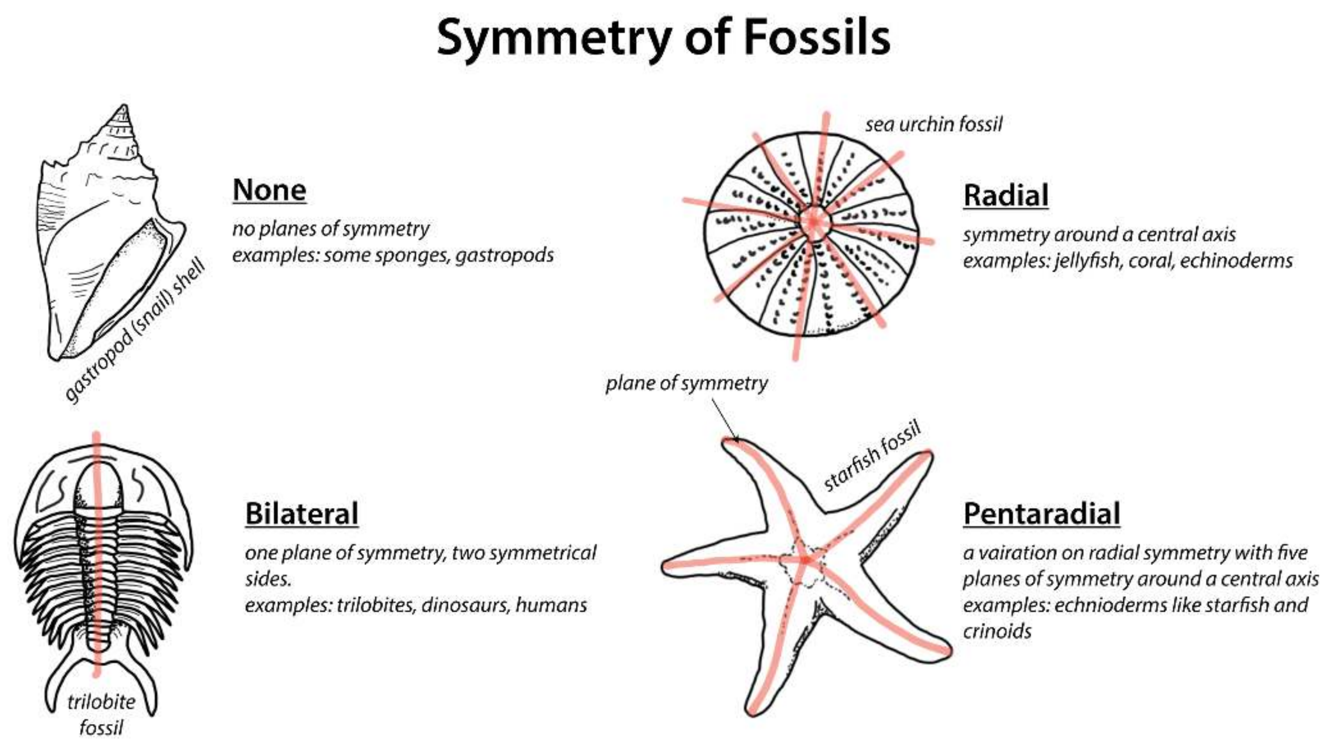


Figure 9.3.2: Describing the symmetry of fossils. (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

9.3.3: Symmetry in organisms with valves

With organisms that live inside of a shell with two halves (valves), some may have shell halves that are mirror images of each other (bilateral symmetry with similar valves) and others might have symmetry that cuts through the shell halves (bilateral symmetry with dissimilar valves).

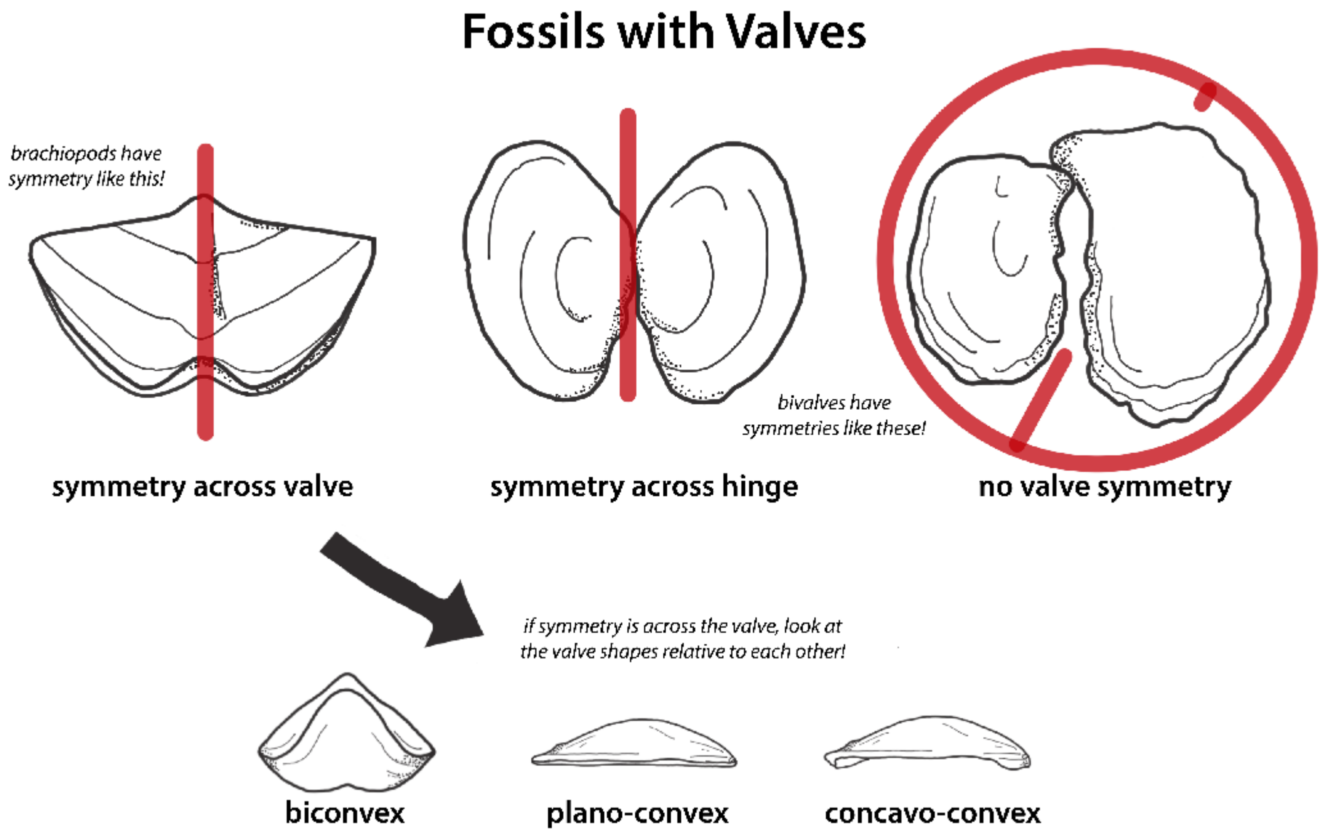


Figure 9.3.3: When describing the symmetry of fossils with valves we focus on the symmetry across the hinge and the symmetry of the valves relative to each *other. (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)*

9.3.4: Shells

For organisms that have shells we describe the shape and internal structure. Shells can spiral upwards (conisprial), spiral along a single plane (planispiral), be conical to flat (patellate), or be straight coned (orthocone). Some shells will have chambers separated by walls (septa). The place where those septa meet the outside of the shell are called sutures.

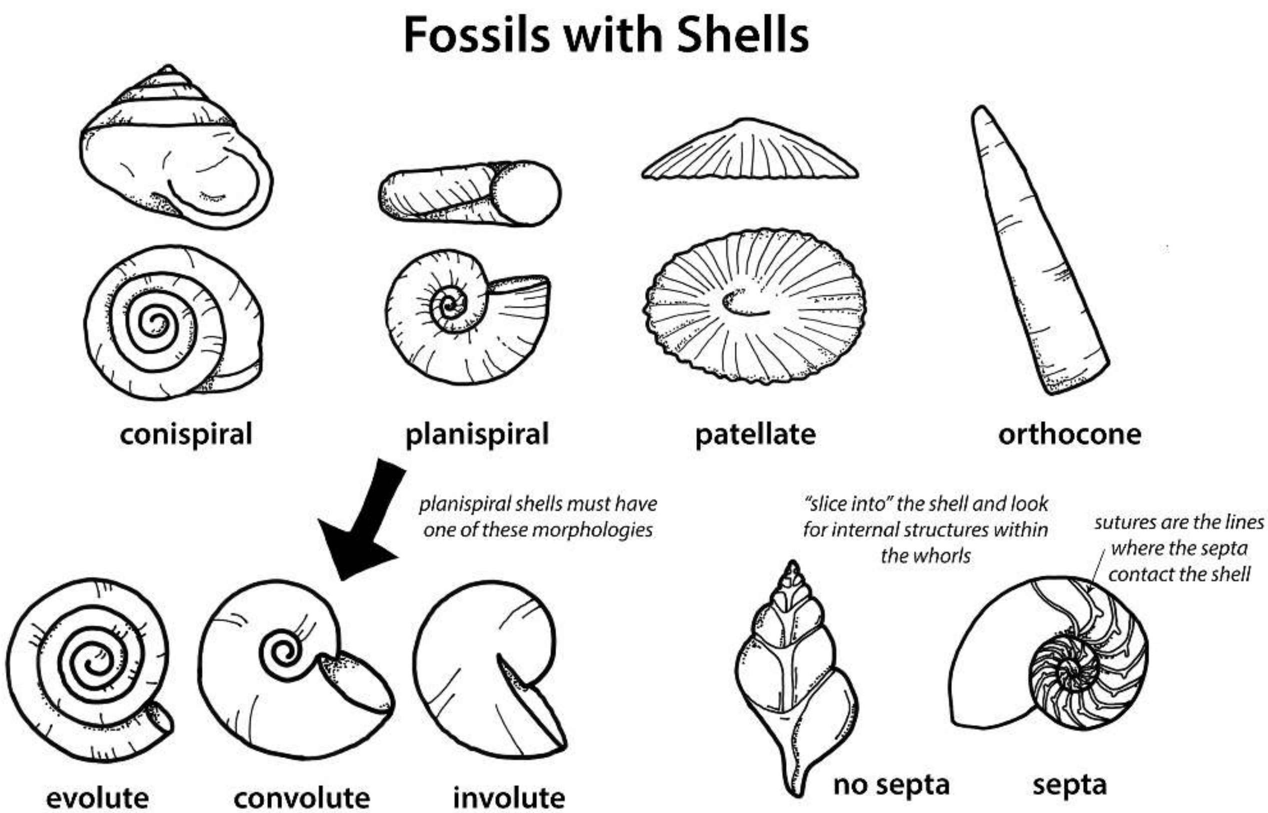


Figure 9.3.4: Describing shell morphology and internal structure. (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

9.4: Major Fossil-Forming Groups (Invertebrates)

Now that we know how to describe fossils, we can start to identify them. In this section we provide a brief overview of the major invertebrate fossil forming groups organized taxonomically. In most cases the information presented here can get you to the taxonomic level of Order. If you want to learn more, consider taking an invertebrate paleontology course, attempting Lab 8: Fossils in "The Story of Earth - An Observational Guide", or browsing the detailed explanations, photographs, and three dimensional models available at the Paleontological Research Institute's Digital Atlas of Ancient Life webpage or fossilid.info

9.4.1: Porifera (Sponges)

Figure 9.4.1: Photograph showing sponge biodiversity and morphotypes in the Caribbean Sea, Cayman Islands. May 23, 2007. Included are the yellow tube sponge (Aplysina fistularis), the purple vase sponge (Niphates digitalis), the red encrusting sponge (Spiratrella coccinea), and the gray rope sponge (Callyspongia sp). Caribbean Sea, Cayman Islands. May 23, 2007 (NOAA Photo Library via Flickr; CC BY 2.0).

Sponges evolved in the Precambrian but their fossil record doesn't really start until the Cambrian Period. There are three important characteristics used for classification of sponges: 1) the body plan of the soft tissue, 2) the composition of the skeleton, and 3) spicule morphology (if present). Sponge skeletons can be composed of spongin, a fibrous protein (think of a bath sponge that can soak up water), carbonate, and/or silica. In some cases these mineral components exist as small needle like crystals called spicules. Sponge fossils can be divided up into the following Classes: Archaeocyatha, Demospongiae, Hexactinellida, Stromatoporoidea, Homocleromorpha, and Calcarea. The preceding hyperlinks provide links to the relevant webpages in the Digital Atlas of Ancient Life; there is also an excellent collection of sponge photographs at fossilid.info.

Sponge fossils are mostly found in marine settings but some groups live in freshwater environments. Their fossils range from the Cambrian to the present.

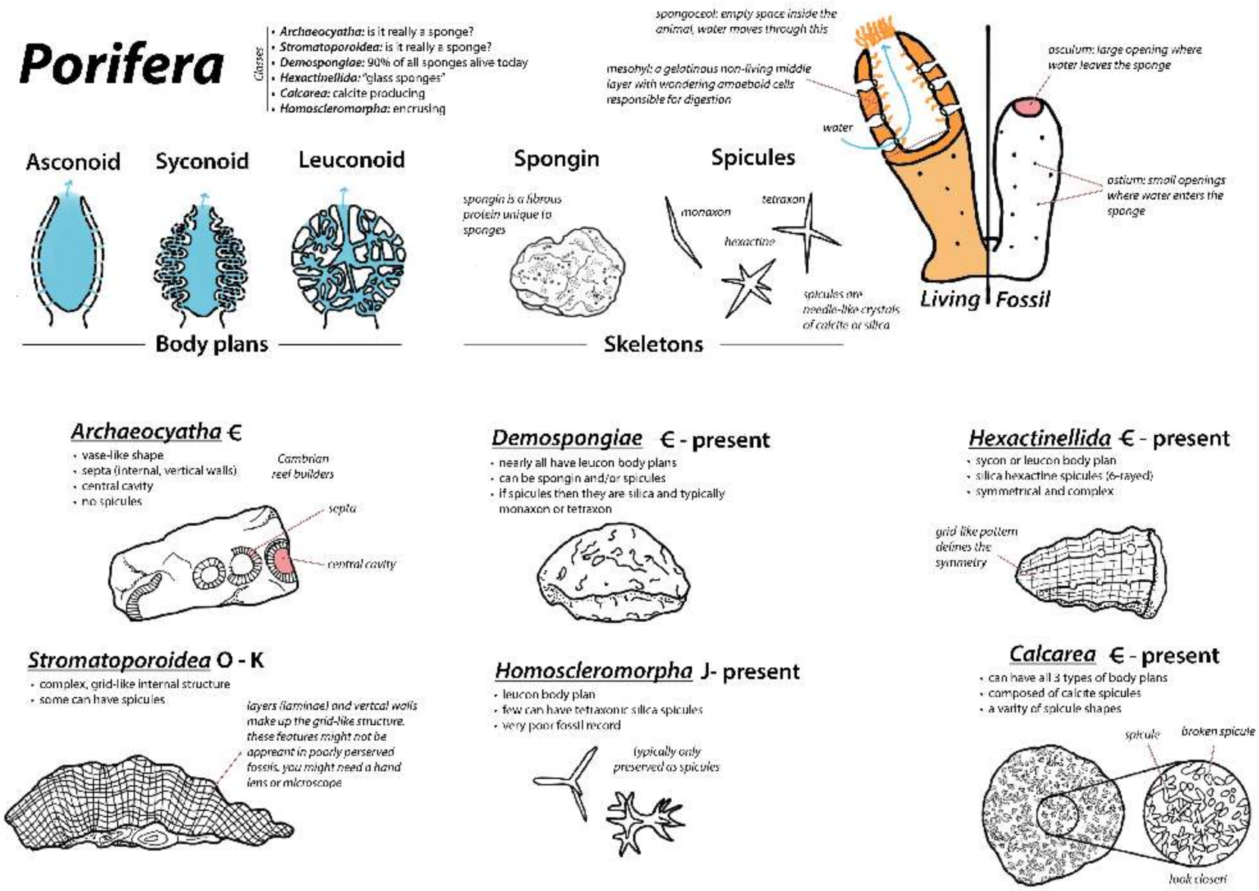


Figure 9.4.2: Classes of Porifera and how to identify them (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

9.4.2: Cnidaria (Corals)

Figure 9.4.3: Photograph the sun coral (*Tubastraea sp.* - a scleractinian coral). This non-native coral that has successfully colonized artificial habitats in the Gulf of Mexico along the southeast coast of Florida, and the Florida Keys (FWC Fish and Wildlife Research Institute via Flickr; CC BY-NC-ND).

Corals belong to the Phylum Cnidaria which also includes jellyfish, sea fans, and sea anemones (among others). These organisms all share the following characteristics: radial symmetry, two tissue types (they are diploblastic), and the presence of venomous stinging cells called nematocysts. The group evolved in the Precambrian but the major coral Orders did not appear until the Ordovician. Corals make up the bulk of the fossil record for Cnidaria because they are the only group that produces a carbonate skeleton. Fossil identification of corals is based on those mineralized skeletons by looking at life mode (did the coral live as a colony or as a solitary individual) and internal structure within the holes that coral animal lived in. The major fossil forming groups of corals include tabulate corals, rugose corals, and scleractinian corals. For more detailed information, pictures of coral fossils, and 3D models, please visit the Digital Atlas of Ancient Life webpage which contains details about rugose, tabulate, and scleractinian corals or fossilid.info's pages on rugose and tabulate corals.

Coral fossils are only found in marine settings and range from the Cambrian to the present.

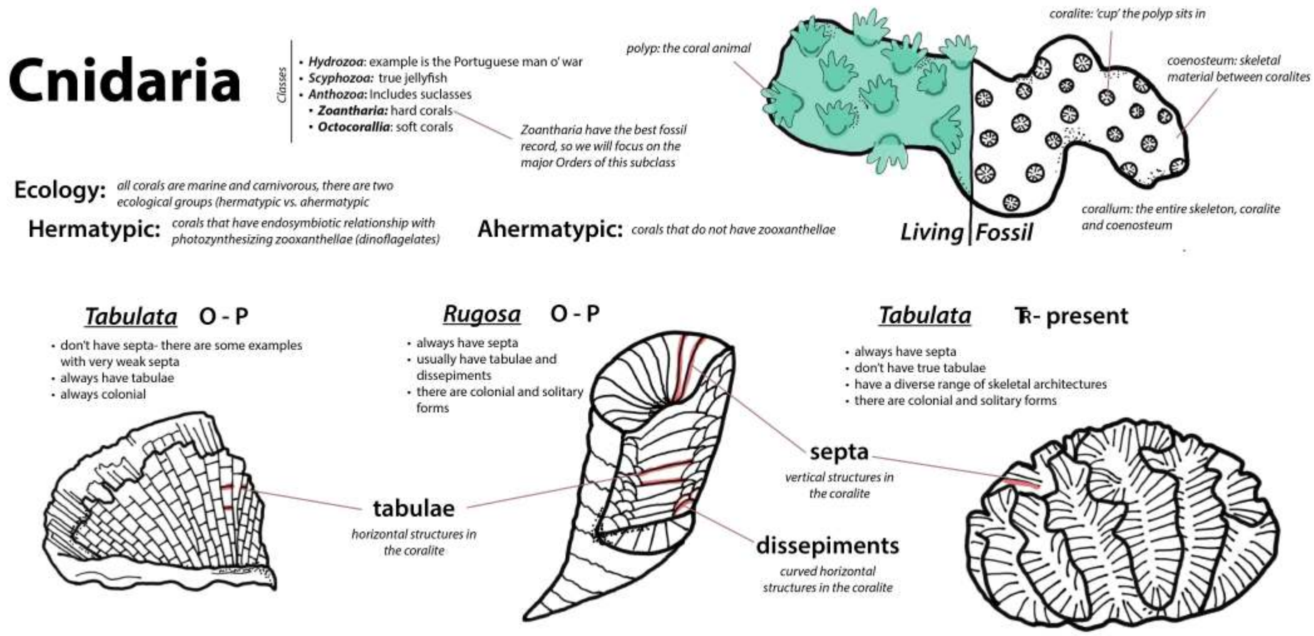


Figure 9.4.4: Orders of coral and how to identify them (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

9.4.3: Lophoporata (Brachiopods and Bryozoans)

The Phylum Lophophorata includes brachiopods and bryozoans. These groups are united in that they have a special feeding organ called a lophophore and exhibit bilateral symmetry. Otherwise it can be hard to believe the two groups are related because they look so different! Brachiopods are shelled solitary organisms while bryozoans are colonial organism that can look similar to corals! Because both groups produce mineralized skeletons, they have excellent fossil records. The Digital Atlas of Ancient Life webpage has an excellent discussion of brachiopods vs bivalves.

9.4.3.1: Brachiopoda

Figure 9.4.5: Photograph the extant inarticulate brachiopod *Lingula anatina* (Wilson44691 via Wikimedia Commons; public domain).

Figure 9.4.6: Photograph of North Pacific Lampshell (*Terebratalia transversa*), an extant articulate brachiopod (Marilynne Box *via Wikimedia Commons;* CC BY 4.0*)*.

Brachiopods are valved (two-shelled) organisms that live in aquatic settings. While they superficially look similar to bivalves, brachiopod valves are not mirror images of each other (that is, the line of symmetry is not across the hinge). Instead, the line of symmetry is across the valve. Brachiopod classification is based on multiple morphological characteristics of the shells. Most important among these characteristics is the length of the hinge line (the line the two valves make where the shell rotates open/close) and the symmetry of the two valves relative to each other. The brachiopod webpages in the Digital Atlas of Ancient Life and fossilid.info provide pictures, models, and a fuller discussion of these organisms.

Brachiopod fossils are only found in marine settings. While the group does range from the Cambrian to the present, only a few make it into the Cenozoic.

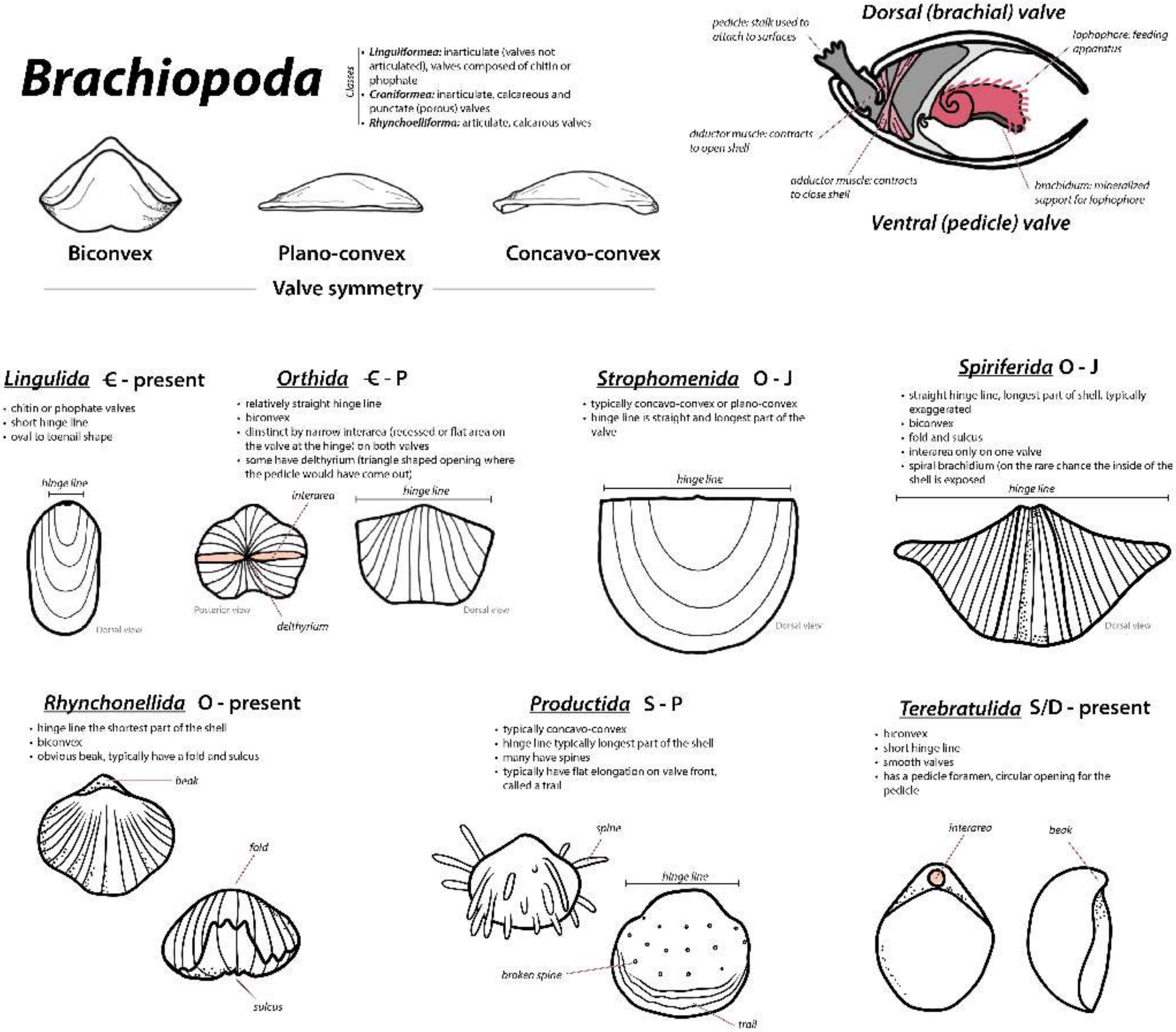


Figure 9.4.7: Brachiopod Orders and how to identify them (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

9.4.3.2: Bryozoa

Figure 9.4.8: Photograph of a colony of the modern marine bryozoan *Flustra foliacea* (Hans Hillewaert *via Wikimedia Commons*; CC BY-SA 4.0*)*.

Bryozoans are colonial organisms where each animal in the colony is a clone. Each of these clones is called a zooid and each zooid lives inside a small chamber called a zoecium. All the zoecium are arranged together to form a colony. Classification of bryozoans is based on the shape of the zooid and the colony arrangement. For proper identification of bryozoans, it is best to look at colony structure and arrangement under the microscope. For field identification, using the growth forms can be very useful for identifying most of the major bryozoan Orders. Bryozoans can exhibit massive, branching, lacy, or encrusting growth forms. The bryozoan pages at the Digital Atlas of Ancient Life and fossilid.info provide pictures, models, and a fuller discussion of these organisms.

Bryozoans are mostly found in marine environments but some groups exist in freshwater settings. The group ranges from the Ordovician to the present but many of the groups go extinct in the Triassic.

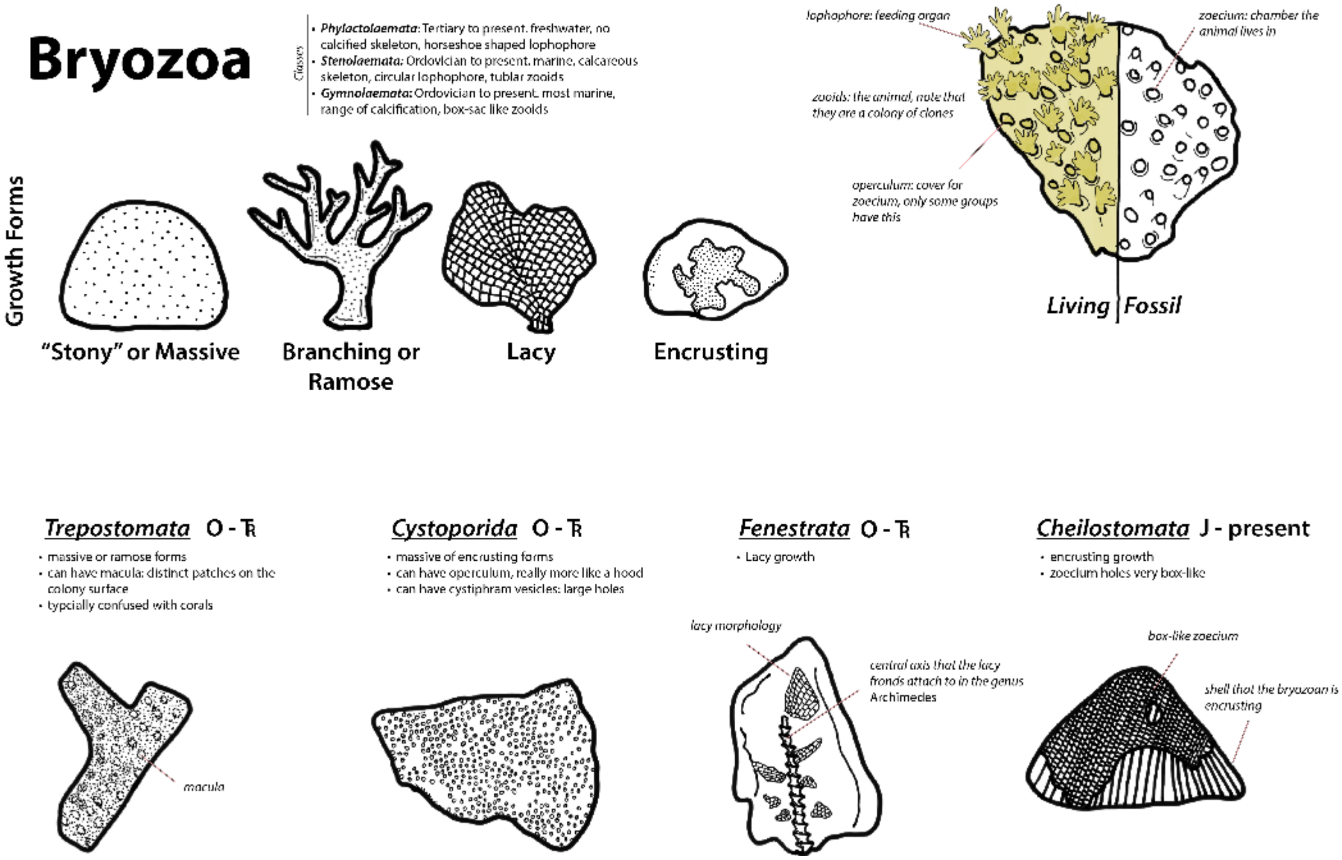


Figure 9.4.9: Fossil bryozoan Orders and how to identify them (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

9.4.4: Mollusca

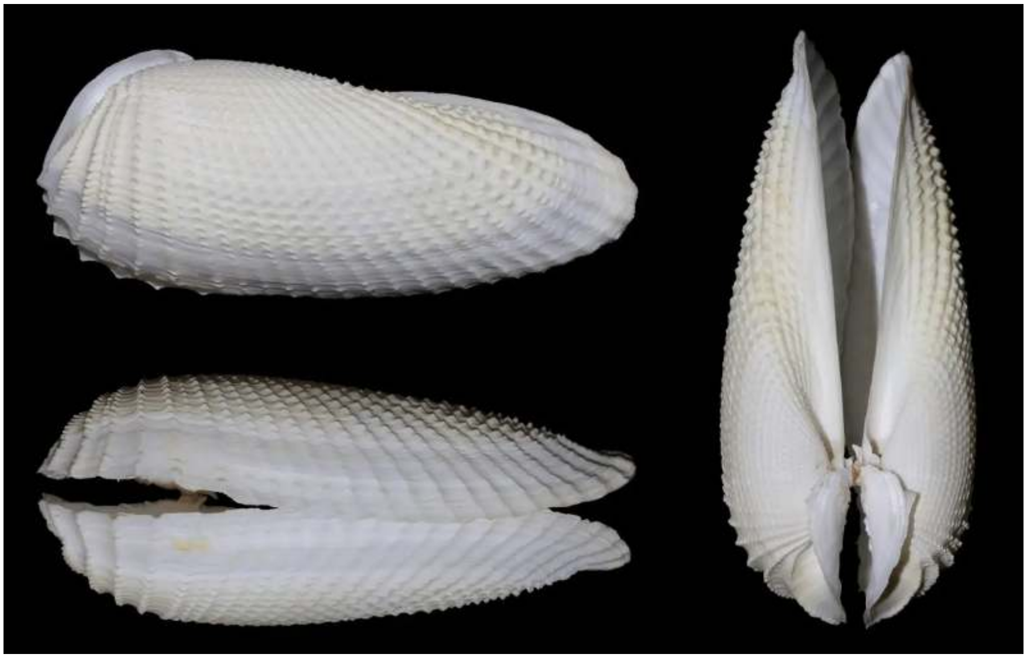


Figure 9.4.10: Photograph of an angel wing bivalve (Cyrtopleura costata), an elongate bivalve that burrows into hard substrates by slowly rotating its shell like a drill bit (MerlinCharon via https://animalia.bio/cyrtopleura-costata; CC0).

Figure 9.4.11: Photograph of a grapevine snail (*Helix pomatia*), a species of air-breathing land snail (Jürgen Schoner *via Wikimedia Commons*; CC BY-SA 3.0).

Figure 9.4.12: Photograph of a nautilus, an extant coiled nautiloid cephalopod (Manuae *via Wikimedia Commons;* CC BY-SA 3.0*).*

Many mollusks have mineralized hard parts, their fossil record is extensive but identification can be difficult as detailed classification is partially based on soft tissue. Here we will focus on some of the most common and easily identifiable fossil groups: Bivalvia, Gastropoda, and Cephalopoda. Bivalves consist of two valves and are classified based on soft tissue features like their gill system and siphon (an organ used for respiration and reproduction). Because these features do not fossilize well, fossil identification is based on scars the soft tissue leaves on the shells as well as how the valves are articulated (dentition). Gastropods, which includes snails and slugs, suffer from the same difficultly with some groups losing the shell all together. Cephalopods, are the most straightforward group to identify. We can use the shape of their shell, the complexity of the internal walls (septa) in those shells, and the positioning of the siphuncle (a canal that connects the chambers of the shell) to identify four of the five Subclasses that belong to the group.

For pictures and more information please visit the following websites:

- Bivalves - Digital Atlas of Ancient Life and fossilid.info
- Gastropods - Digital Atlas of Ancient Life and fossilid.info
- Cephalopods - Digital Atlas of Ancient Life and fossilid.info

Mollusca are found in all types of environments but cephalopods are exclusively marine. The group ranges from the Cambrian to the present.

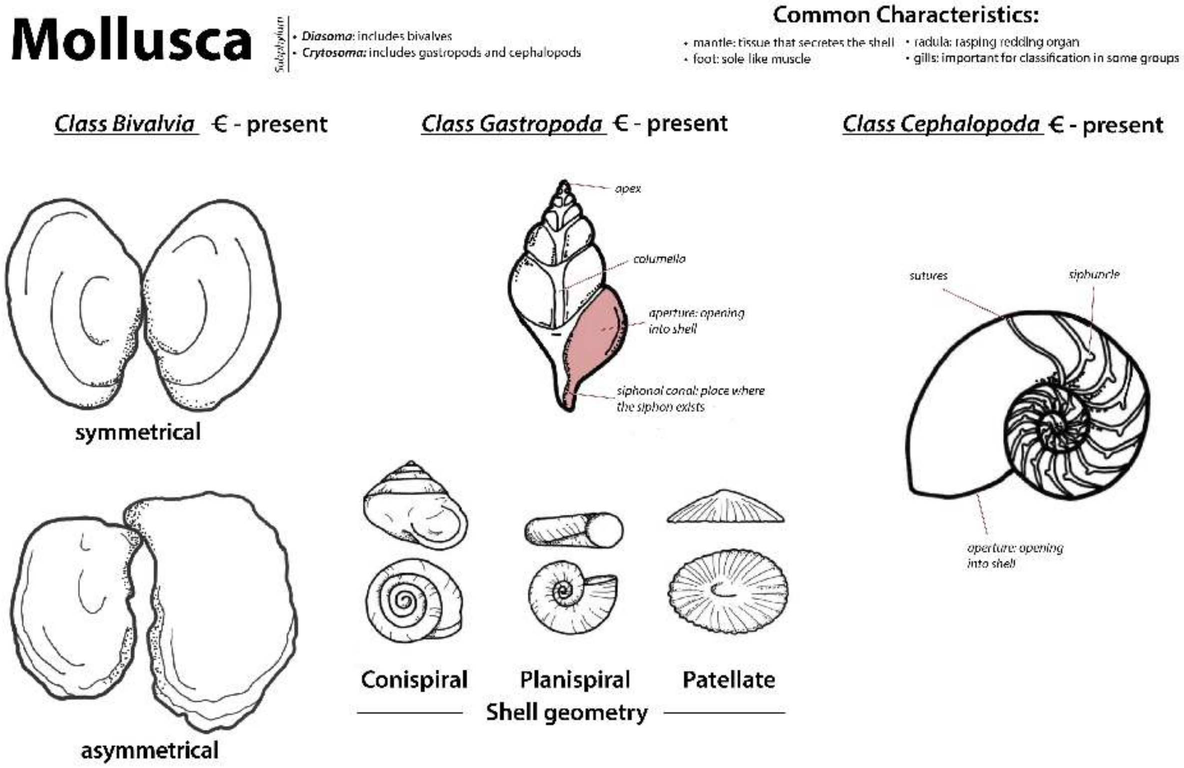


Figure 9.4.13a: Common fossil forming mollusca Classes (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

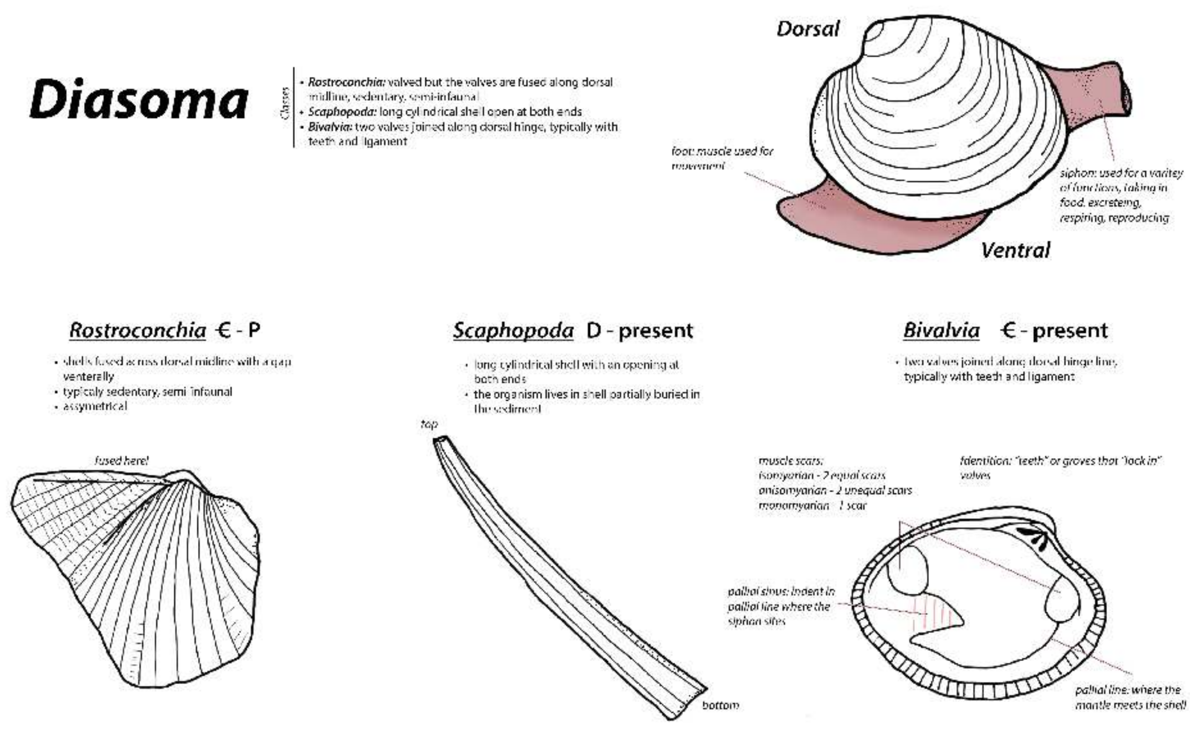


Figure 9.4.13b: Some common fossil-forming Classes that belong to the Phylum Mollusca, Subphylum Diasoma (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

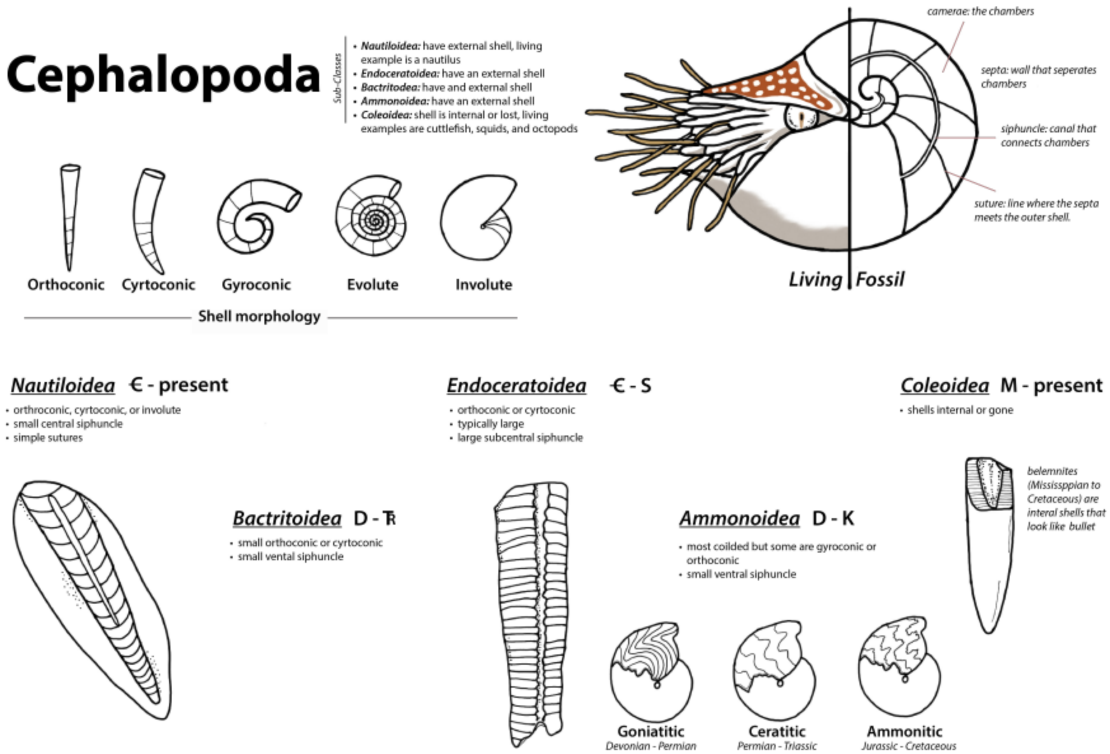


Figure 9.4.13*c*: Cephalopod Subclasses and how to identify them (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

Figure 9.4.14: Living ostracod (a type of arthropod) showing real time swimming motion (Ikjbagl via Wikimedia Commons; CC BY-SA 4.0).

Figure 9.4.15: Living barnacles (a type of arthropod) surround themselves with a ring of calcareous plates (paweesit via Flickr; CC BY-ND 2.0).

Figure 9.4.16: Photograph of a trilobite fossil (*Selenopeltis sp.*); trilobites are an extinct type of arthropod (Kevin Walsh via Flickr; CC BY-ND 2.0).

Arthropods are the most diverse of the animal kingdom and includes groups like insects (Hexapoda), crustaceans, and trilobites. These groups are united in that they have jointed appendages, segmented bodies, and an exoskeleton. Because many of those exoskeletons are not exceptionally robust, the fossil record of this group does not quite match the Phylum's abundance and diversity in the modern. The most notable fossil group is the Subphylum Trilobitomorpha. Classification in this group is based on the number of thorax segments, the shape and size of the third body segment (pygidium), the shape and size of the head (cephalon), the shape and position of the stomach (glabella), among other characteristics. Because many of the trilobite Orders have relatively short stratigraphic ranges, this group is commonly used for biostratigraphy. The Digital Atlas of Ancient Life's arthropod, trilobite, and crustacean webpages and fossilid.info's trilobite, chelicerate, and insect pages provide images and more discussion of these organisms.

Arthropods are found in all types of environments and range from the Cambrian to present. Trilobites are only found in marine settings and go extinct at the end of the Permian.

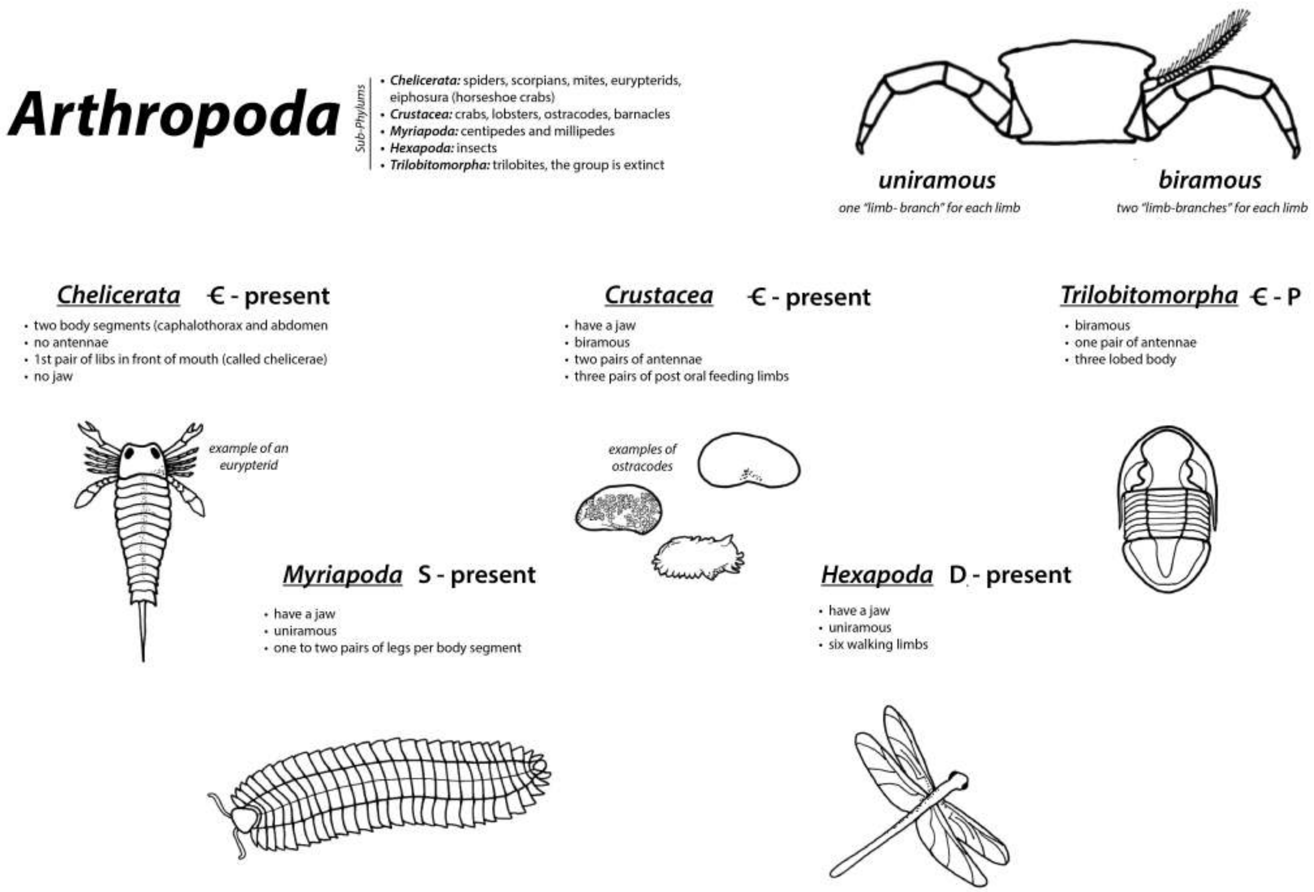


Figure 9.4.17a: The Subphyla of arthropods (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

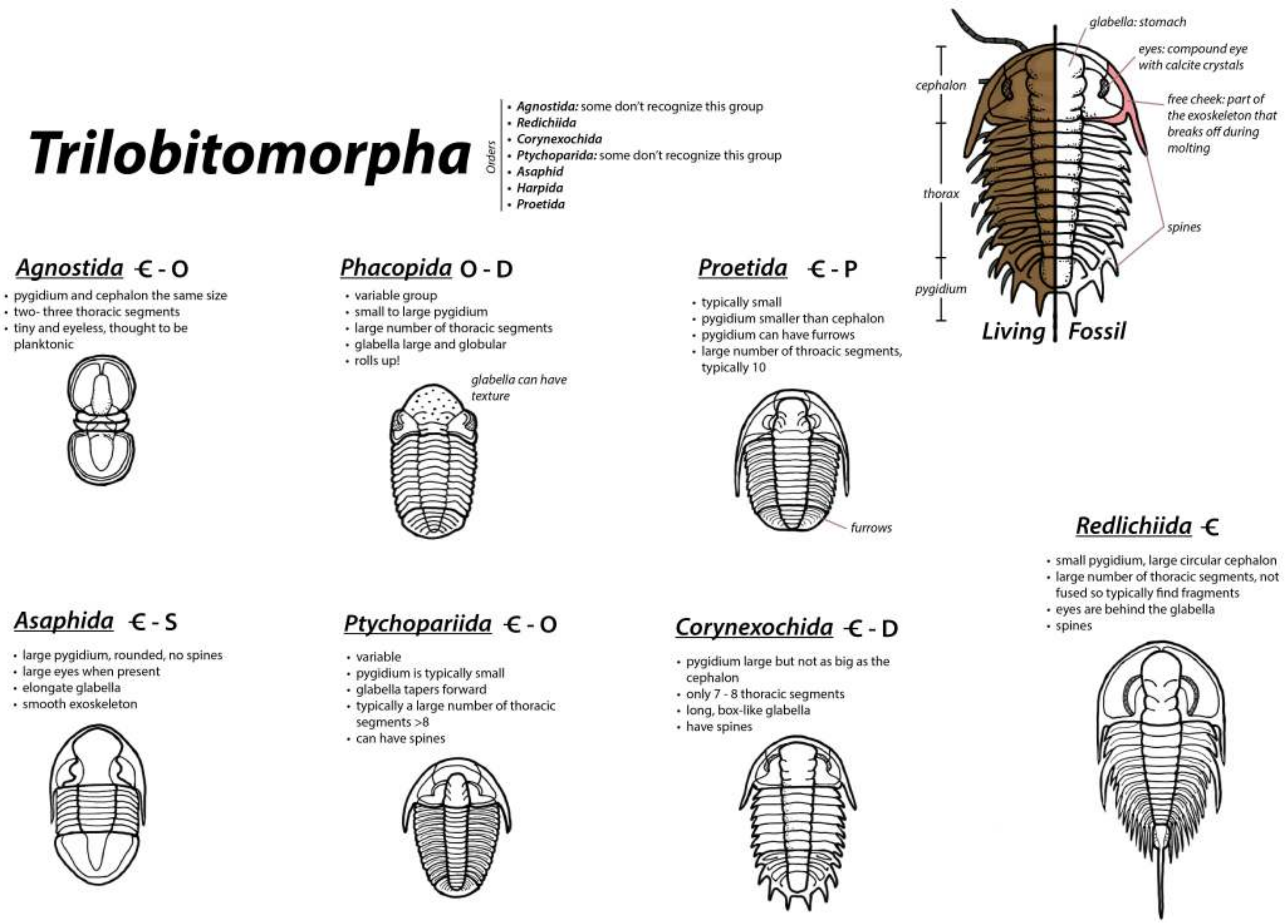


Figure 9.4.17*b*: The Orders of trilobites and how to identify them (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

Figure 9.4.18: Photograph of a living reticulated starfish (*Oreaster reticulatus*). Many starfish are distinctively pentaradial, with five arms. Numerous tube feet occur in the ambulacral grooves running down the axis of each arm's underside. Many starfish are predatory and consume prey by everting their stomachs through their mouths (center structure on underside) and digest food externally. *Oreaster reticulatus* is predatory on a variety of invertebrates and is also a deposit feeder (James St. John via Flickr; CC BY 2.0).

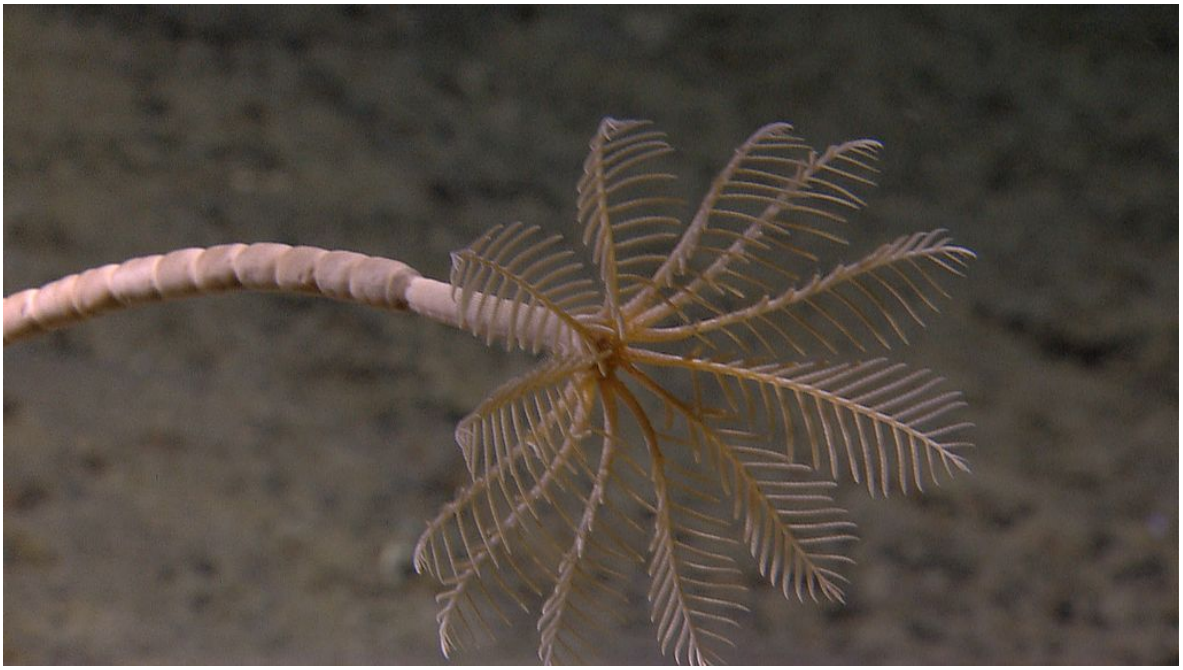


Figure 9.4.19: Photograph of a living stalked crinoid (NOAA Photo Library via Flickr; CC BY 2.0).

Figure 9.4.20: Photograph of living sand dollars (arbyreed via Flickr; CC BY-NC-SA).

Echinoderms include starfish, sea urchins, and crinoids (among others). The shared characteristics of this Phylum include five-fold (pentaradial) symmetry, a water vascular system (water is pumped through the body for respiration and locomotion), and a mesoderm skeleton composed of porous plates of calcite crystals. Echinoderms are divided into two Subphyla: Eleutherozoa (non-stemmed) and Pelmatozoa (stemmed). Many of the calcite plates that make up the mesoderm skeleton are not fussed together so the fossil record of this group includes lots of disaggregated echinoderm plates in additional to complete body fossils. The echinoderm webpages at the The Digital Atlas of Ancient Life and fossilid.info provide more pictures and discussion of these organisms.

Echinoderms are only found in marine environments and range from the Cambrian to the present.

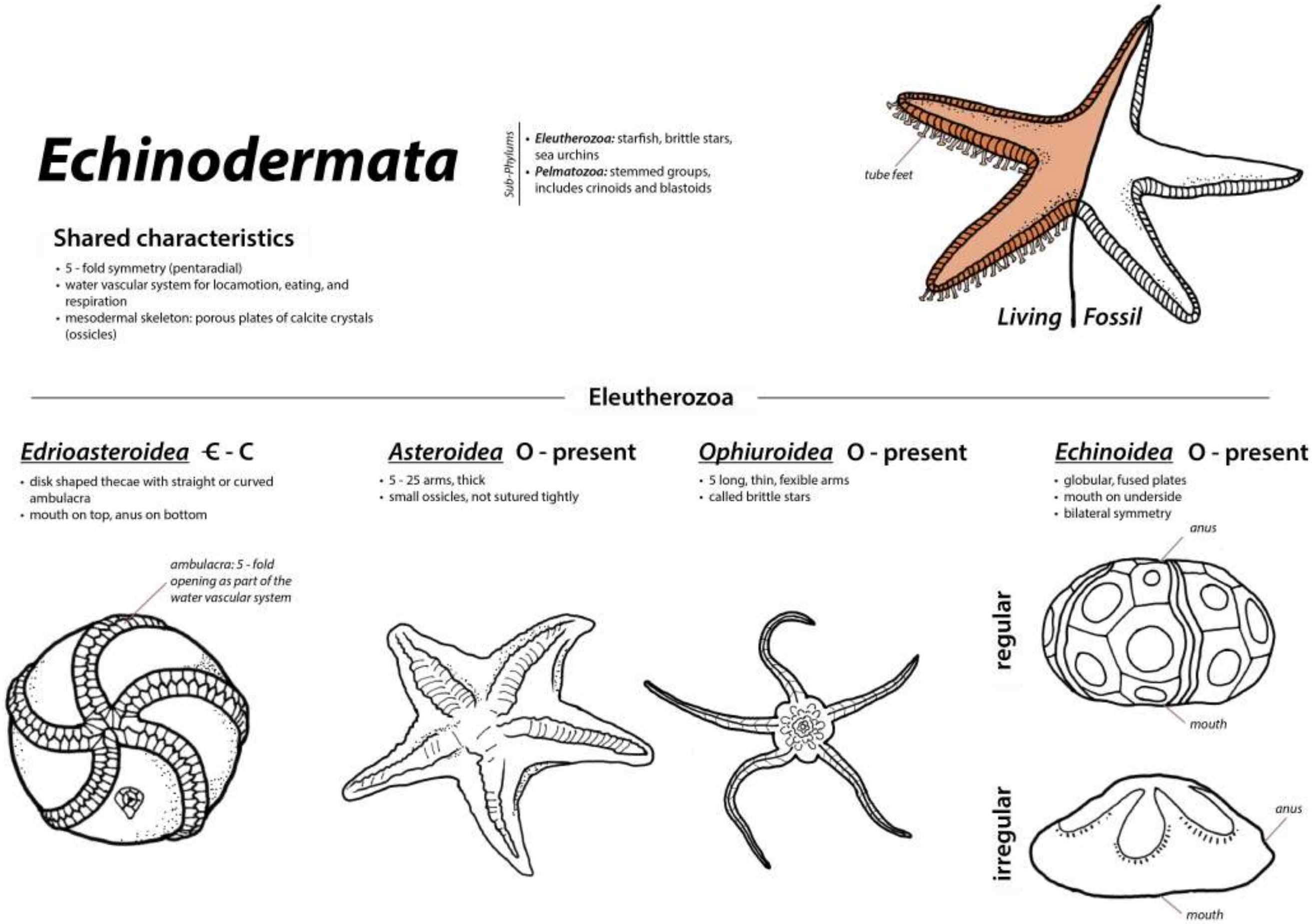


Figure 9.4.21a: Classes that belong to the un-stemmed (Eleutherozoa) echinoderms and how to identify them (Page Quinton via Wikimedia Commons; CC BY-SA)

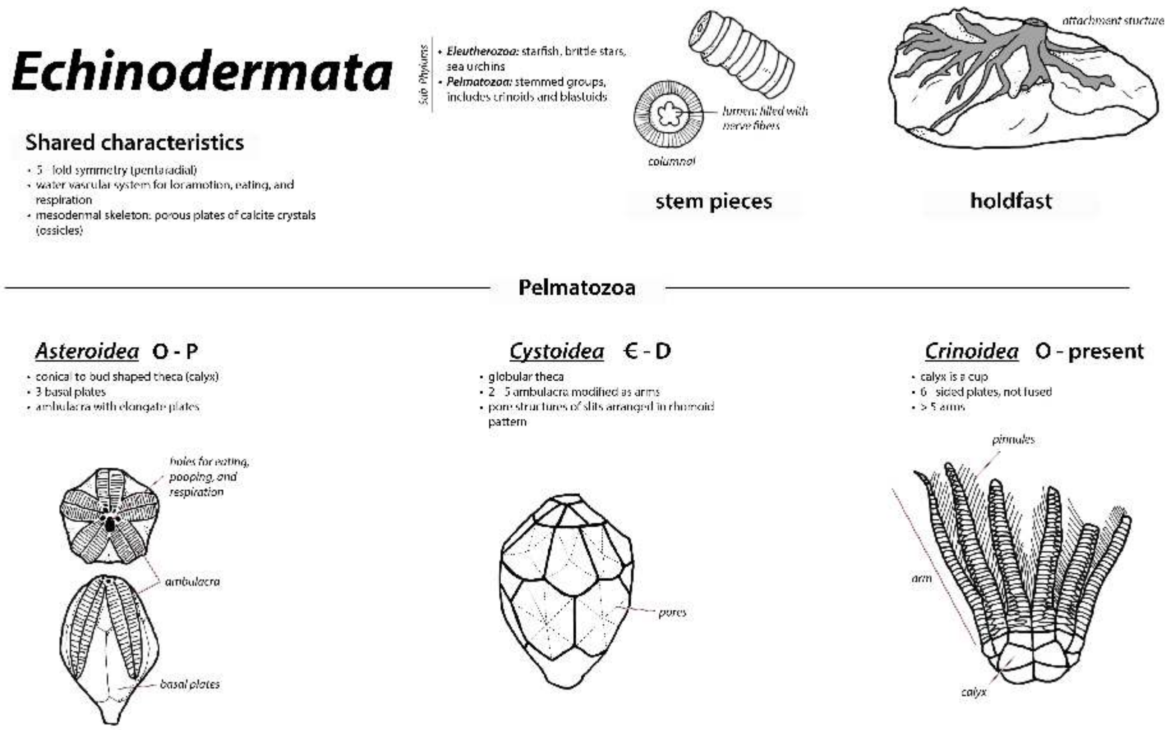


Figure 9.4.21b: Classes that belong to the stemmed (Pelmatoza) echinoderms and how to identify them (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

9.5: Microfossils

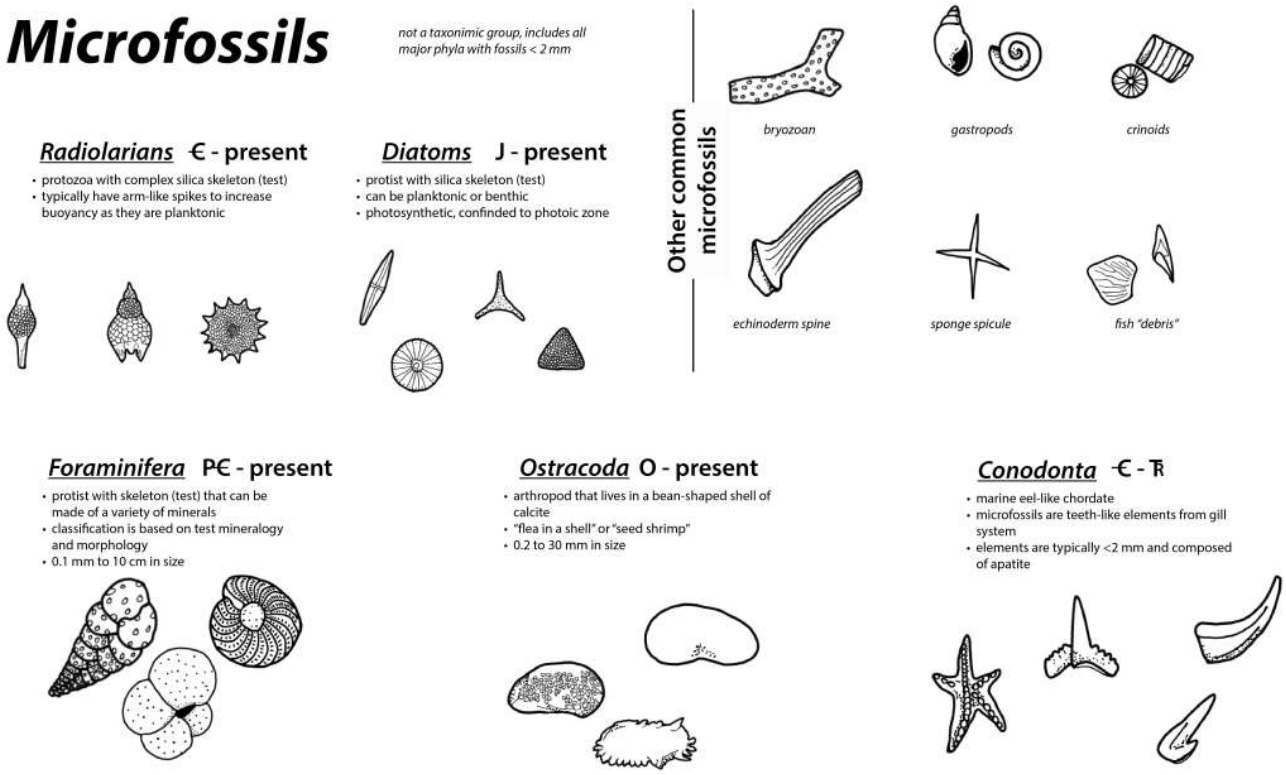


Figure 9.5.1: Major microfossil groups. (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

9.5.1: Major Microfossil Groups

Microfossils are fossils that are less then 2 mm in size and require the aid of a microscope to study. Many taxonomic groups are represented in the microfossil record but here we will focus on some of the common mineralized microfossil groups. These include foraminifera, radiolarians, diatoms, ostracods, and conodonts. These microfossils are important tools for biostratigraphy and paleoclimatology. Other notable nonmineralized microfossils include spores and pollen.

9.5.2: Foraminifera

Foraminifera are an Order of single celled protista that have an extensive fossil record. These organism are aquatic and live inside a shell referred to as a test. Classification of the foraminifera genera is based on shell mineralogy, shell morphology, and internal shell structure. Foraminifera with shells composed of calcium carbonate are more readily preserved than organic (chitin) based groups and make up the bulk of the foraminifera fossil record. Some common fossil forming Genera are Textulariids, Fusulinids, Miliolids, Rotaliids, and Globigerinids. For more information please visit the Digital Atlas of Ancient Life's foraminifera webpage.

Most foraminifera are marine but there are a few freshwater varieties. The group ranges from the Cambrian to the present.

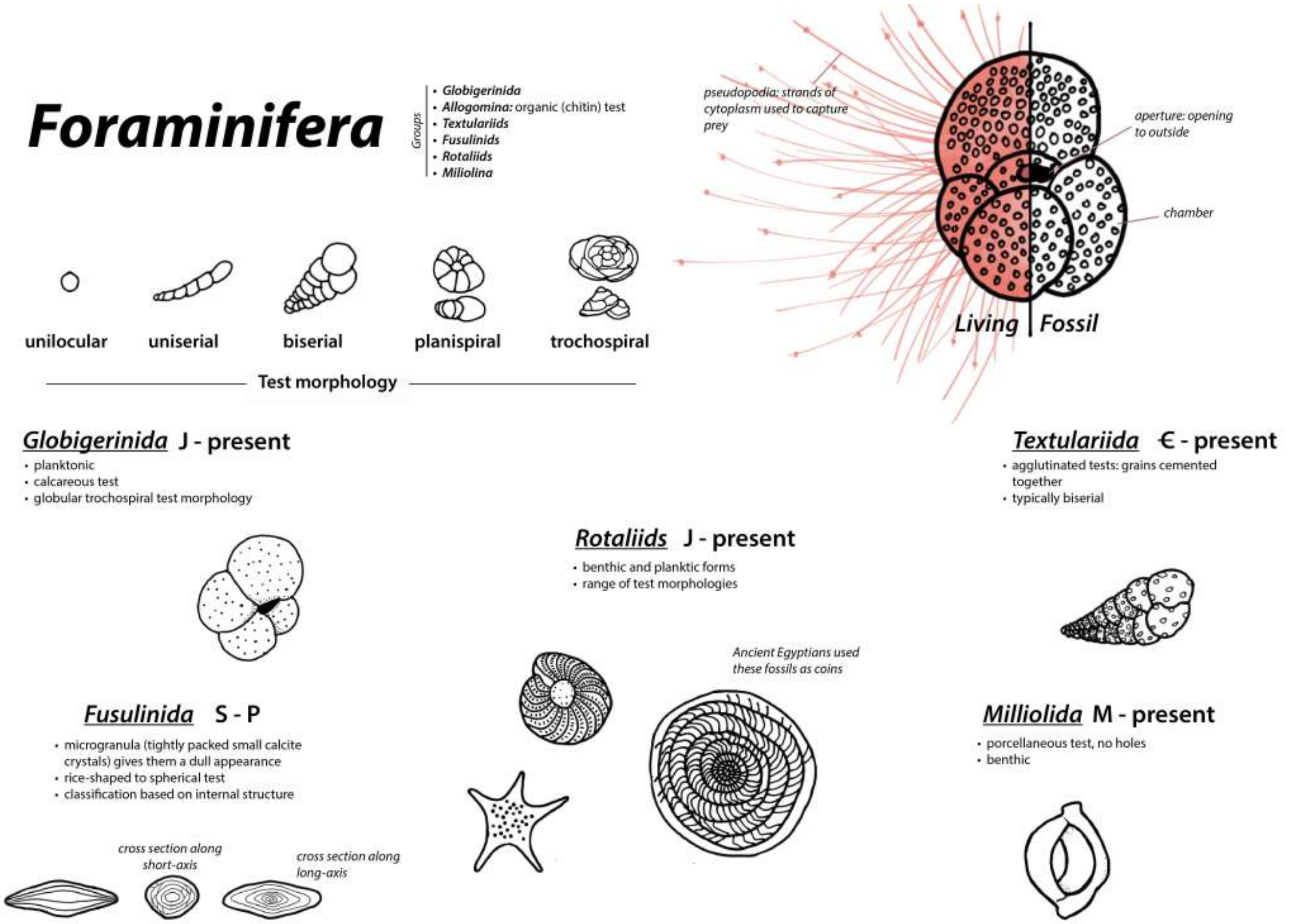


Figure 9.5.2: Major fossil forming groups of foraminifera. (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

9.6: Trace Fossils

Trace fossils are sedimentary structures formed as the result of the activity of an organism, other than growth (Jackson, 1997, Glossary of Geology). They represent fossilized behavior and thus provide important insight into the conditions that existed during/immediately after the deposition of sediment. The study of trace fossils is called ichnology.

9.6.1: Types of Trace Fossils

Trace fossils can be grouped into several major types of structures based on what the organism was doing:

- Bioturbation structures record disruption of bedding by burrows, tracks, and trails.
- Bioerosion structures record erosion of substrate by boring, scraping, or biting.
- Biostratification structures occur when organisms impart an organization to sediment by making layers or structures.
- Fecal material also counts as a trace fossil! It can tell us about the organism's diet and possibly its biological affinity.

Figure 9.6.1: Bird and human footprints in the mud disrupt the sediment and thus represent bioturbation structures (Julian Dowse via Geograph; CC BY-SA 2.0 DEED).

Figure 9.6.2: Marine bivalves of the genus *Teredinidae* (aka "shipworms") use their valves as a rasp and are able to slowly burrow into woody substrate and thus create bioerosion structures (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0).

Figure 9.6.3: Sand bubbler crabs eat by processing sand through their mouthparts and discard the sand as small round balls which imparts an organization to the sediment and thus represents biostratification.

Figure 9.6.4: Fecal pile from an acorn worm exposed in the intertidal zone. Fecal material ranges from mundane to spectacular and tells us bout the diet and presence of an organism. (Doug Greenberg via Flickr; CC BY-NC-ND 2.0 DEED).

Although they might appear quite similar to trace fossils, there are several structures that don’t technically count. These would include any sorts of tool marks formed by passive contact between an object (even an animal) and the bed as well as growth structures like stromatolites and roots.

Figure 9.6.5: Scratch circles can form when the wind blows plants causing them to rotate and inscribe scratches into the adjacent substrate. Because they record passive contact and not a behavior, they do not count as trace fossils (David Marvin via Flickr; CC BY-NC-ND 2.0 DEED).

9.6.2: Behavioral Classification (Ethology)

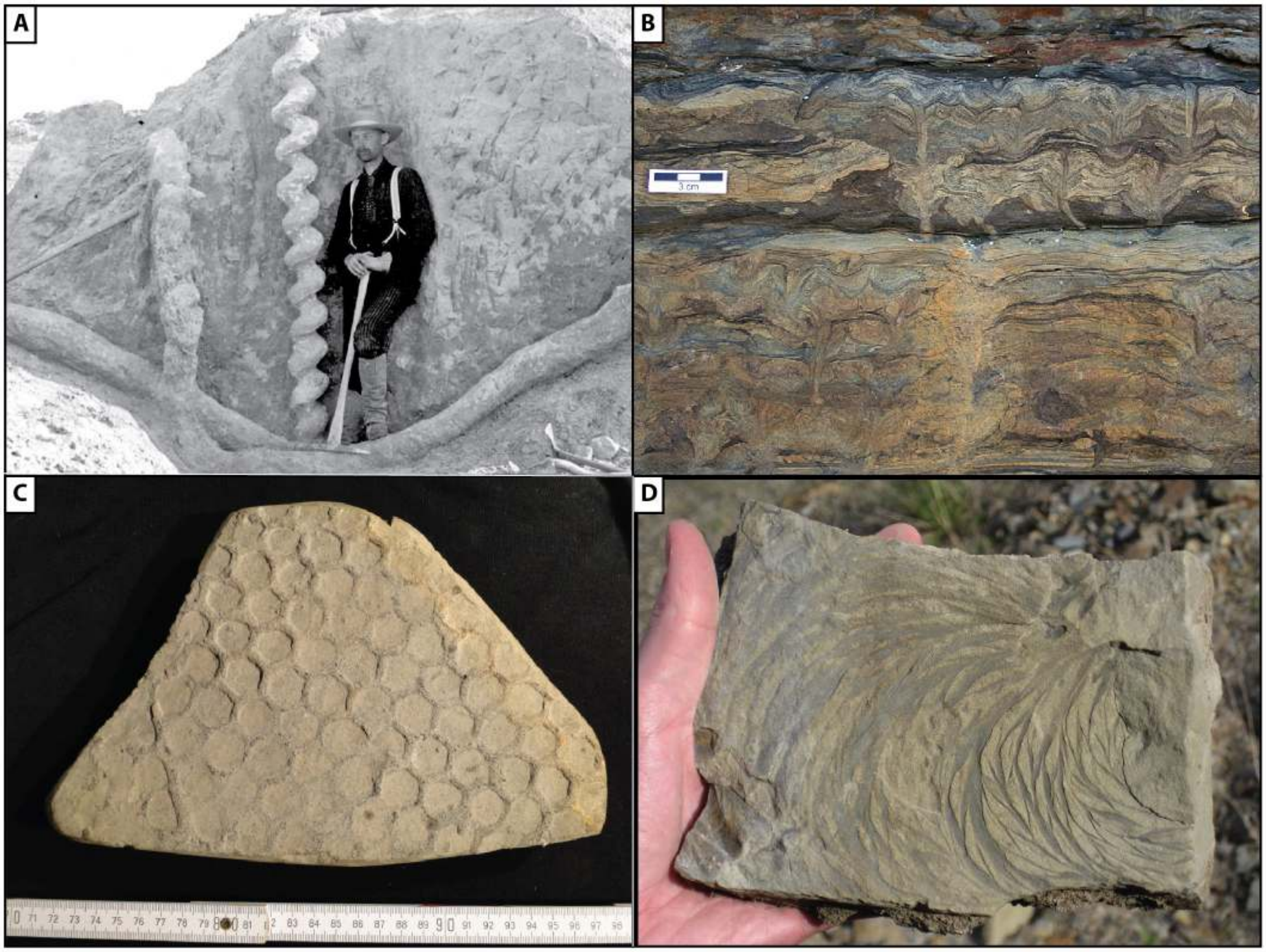


Figure 9.6.6: A) *Daemonelix* is the dwelling trace of the extinct beaver *Paleocaster* (James St. John via Flickr; CC BY 2.0 DEED). B) *Lingulichnus* is formed as inarcticulate brachiopods moved progressively upward through rapidly accumulating sediment to maintain their equilibrium with the sediment surface (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). C) *Paleodictyon* is considered by some to be a deepwater farming trace composed of a series of interconnected hexagonal burrows that allow water to flow through and bacteria to grow (Dr. Markus Bertling, via https://westfalen.museum-digital.de ; CC BY-NC-SA). D) *Zoophycos* is a feeding trace formed by repeated, offset probing of the sediment (James St. John via Flickr; CC BY 2.0 DEED).

Regardless of what an organism was and when it lived, certain behaviors are very common and produce trace fossils with comparable features. Some of the most common behaviors include:

- Dwelling traces (Domichnia) – are formed when the organism constructs a home in the subsurface. They include a wide range of morphologies including bifurcating or u-shaped burrows that are commonly perpendicular to bedding. Some may have thick linings and a horizontal component.
- Escape (Fugichnia) and equilibrium (Equillibrichnia) structures – are formed when an organism moves vertically through the substrate to escape burial or maintain it's equilibrium with the sediment surface. They are commonly similar to dwelling burrows except that they appear nested or stacked.
- Farming traces (Agrichnia) - are burrows that are formed to allow an organism to grow algae or bacteria for later consumption. They typically consist of complicated geometric patterns.
- Feeding traces (Fodichnia) – are formed as an organism moves through sediment in order to process it for food. They often have a significant horizontal component and may be composed of single burrows that branch, bifurcate, or show a slight offset to form a more complicated pattern.

- Grazing traces (Pascichnia) - are formed at the sediment surface as an organism moves along as it feeds. They are typically bedding plane parallel, rarely crosscut each other, and may have complex ornamentation or patterns
- Locomotion traces (Repichnia) – are linear to sinuous horizontal (bedding plane parallel) footprints, tracks, or trails produced by movement of an organism.
- Predation traces (Praedichnia) - form when one organism eats, or attempts to eat, another. They can include bite marks, borings, repair of crushed shells, and a variety of other forms.
- Resting traces (Cubichnia) are generally preserved as shallow depressions made when organisms settle or dig into substrate. They commonly preserve the shape of the organism or parts of the organism that are in contact with the substrate.

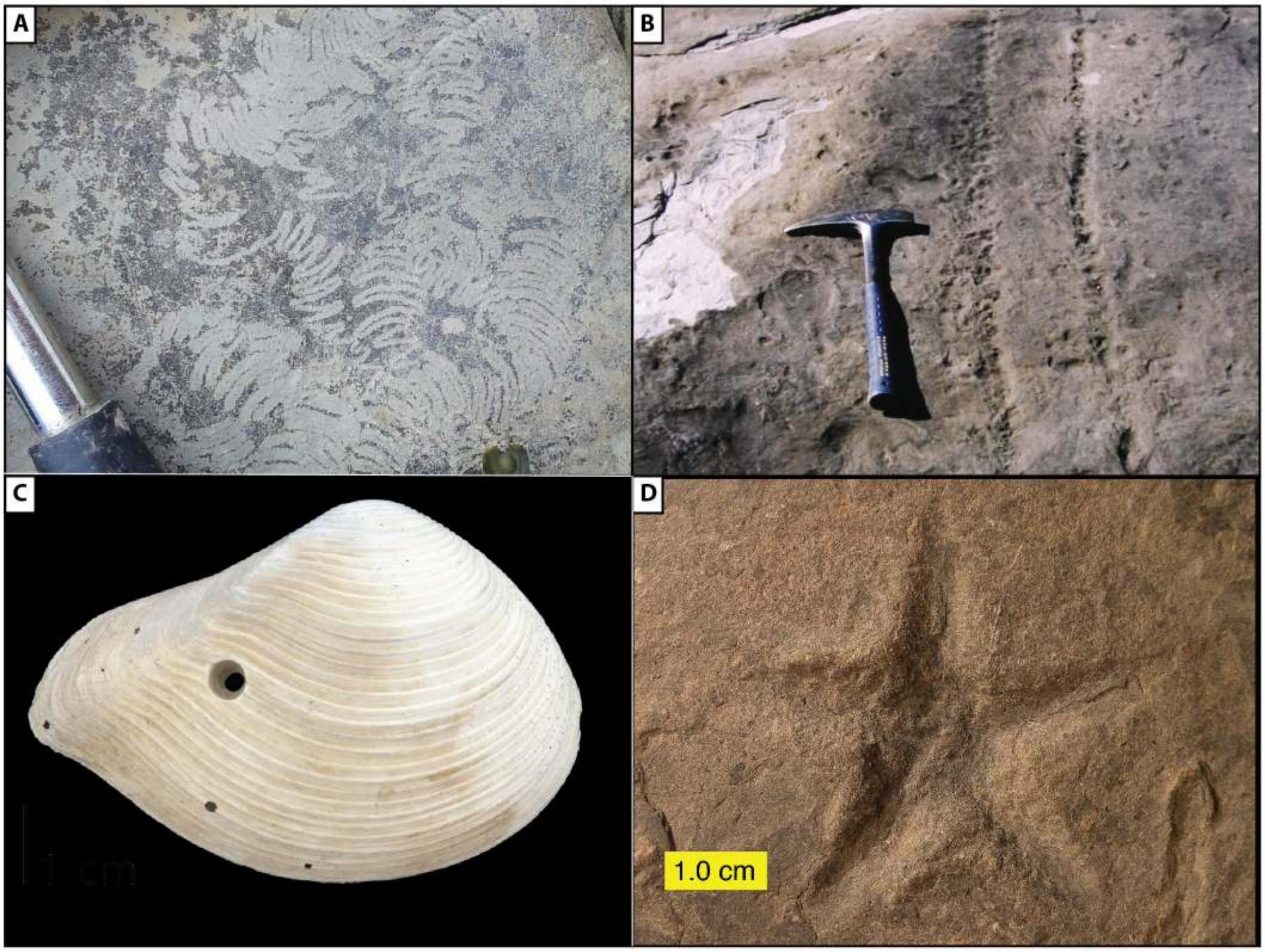


Figure 9.6.7: A) Nereites is a deepwater grazing trace (Richdebtomdom via Wikimedia Commons; CC BY-SA 3.0). B) Diplichnites is a locomotion trace; this example was made by a giant Pennsylvanian millipede (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). C) Predatory gastropod drill hole in the shell of a bivalve (Jonathan R. Hendricks via Digital Atlas of Ancient Life ; CC BY-SA 4.0). D) Asteriacites is the resting trace of a starfish (Mark Wilson via Wikimedia Commons; public domain).

9.6.3: Taxonomic Classification

Trace fossils are classified into ichnogenera and ichnospecies based on the physical traits of the burrow. We do it this way because different trace-makers may produce similar burrows when behaving similarly and because substrate can profoundly influence the appearance of a track, even if from the same organism. Some of the most common and/or noteworthy trace fossils are pictured and described below.

Image	Name (Ichnogenus) and Description
	***Arenicolites*** Vertical u-shaped burrow that lacks concentric spriete and may have funnel-shaped tops. Broadly similar to *Diplocraterion* but without stacked spriete. Interpreted as a feeding/dwelling burrow of worms, crustaceans, beetle larvae, and a variety of other organisms. Image from Gennadi Baranov via Fossilid.info; CC BY-NC.
	***Asteriacites*** Star-shaped resting traces that may be preserved as depressions on the top of beds or in relief on the base of beds. Thought to be the resting trace of a starfish. Image from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0.
	***Chondrites*** A vertical to horizontal downward branching burrow with a pattern similar to the roots of a plant. Fill can be slightly more coarse-grained than the surrounding sediment. <1 mm in diameter and up to several inches tall. Thought to be a feeding structure of an annelid worm or a broadly similar organism. Image from Gennadi Baranov via Fossilid.info; CC BY-NC.

Cochlichnus

Bedding parallel trace that roughly approximtes a sine curve. Interpreted as an invertebrate locomotion trace. Image from Falconaumanni via Wikimedia Commons; CC BY-SA 3.0

Conichnus

Cone-shaped depression with a round bottom. Coarser clasts can be concentrated around the burrow margin. Interpreted as a dwelling or resting trace. Although a variety of tracemkers are possible, many were formed by anemones. Image from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0.

Cosmorhaphe

Bedding plane parallel meandering trace that does not cross-cut itself. Interpreted as a grazing trace. Image from Massimilianogalardi via Wikimedia Commons; CC BY-SA 3.0

Cruziana

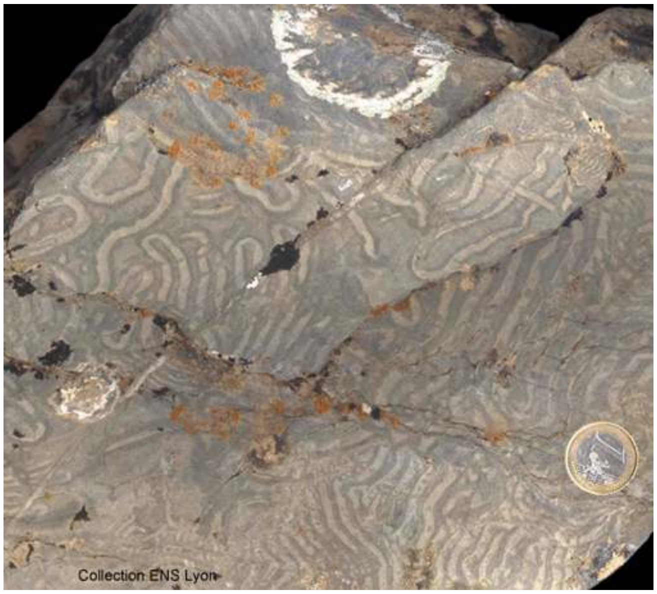
A horizontal, bilobed burrow with a pronounced medial ridge, commonly with herringbone-like scratches that converge towards the medial ridge. Interpreted as a locomotion trace of an arthropod. Image from PePeEfe via Wikimedia Commons; CC BY-SA 3.0.

Diplichnites

Elongate trackway with two parallel rows of tracks; the tracks themselves consist of numerous small, closely spaced blunt to elongate impressions caused by footfalls. Interpreted as the locomotion trace of an arthropod. Image from Kenneth C Gass via Wikimedia Commons; CC BY-SA 3.0.

Diplocraterion

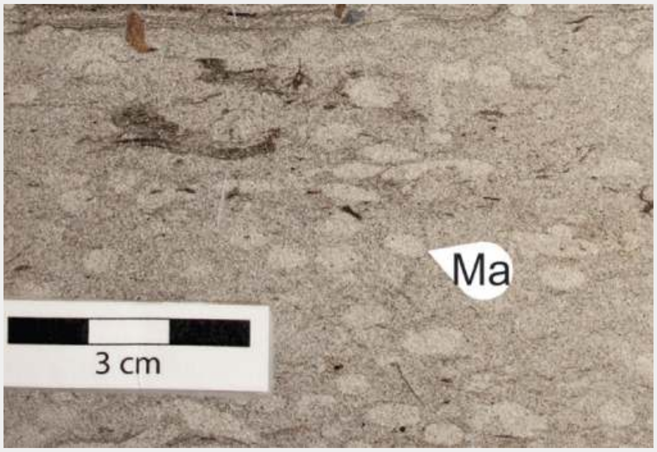
Vertically oriented, u-shaped burrow that has a stacked appearance caused by curved concentric "spreite" (backfills) connecting the arms of the burrow. In bedding plane view they are commonly preserved as paired circular burrows with a "dumbbell" morphology caused by the connecting spreite. Superficially similar to *Skolithos* but it is u-shaped with paired limbs. Roughly the diameter of a straw. Interpreted as the dwelling burrow of a crustacean. Image from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0.



Helminthoida
Trace consists of tight, closely spaced bedding plane parallel coils that do not crosscut one another. *Helminthopsis* is very similar but has a more irregular and open pattern of loops/coils. Interpreted as the feeding trace of a worm. Image reproduced with written permission for the educational purposes (NC ND) of this LibreText contribution by ENS de Lyon. Original image by Pierre Thomas - Lithothèque ENS de Lyon; https://planet-terre.ens-lyon.fr/ressource/Img38-2003-04-14.xml

Helminthopsis
Irregular, bedding plane parallel traces that do not crosscut one another; it appears more like loose coils than meanders or loops. The fill is commonly different than the surrounding material. *Helminthoida* is very similar but has a more regular curved pattern with tighter loops. Interpreted as the feeding trace of a worm. Image from Mark Wilson via Wikimedia Commons; public domain.

Lockeia
An oval or teardrop-shaped trace that is commonly preserved in positive relief on the bottom of a bed. It is bilaterally symmetrical and looks similar to the bottom of a bed. Broadly similar to *Rusophycus* but lacks the medial ridge. Roughly the size and shape of a kidney bean; commonly very dense occurrences. Interpreted as a bivalve resting trace. Image from James St. John via Flickr; CC BY 2.0 DEED.



Macaronichnus
Variably oriented burrows that are longitudinally sinuous and roughly circular in cross section. Burrows are "cleaner" and have less orgnic material than surrounding sediment. Most are <1.5 mm diameter, unlined, and unbranching (although they can cross-cut). Interpreted as feeding traces of polychaete worms. Image from Dafoe, L. T. & Williams, G. L. (2020). Lithological, sedimentological, ichnological, and palynological analysis of 37 conventional core intervals from 15 wells, offshore Labrador (Newfoundland and Labrador) and southeast Baffin Island (Nunavut). *Geological Survey of Canada, Bulletin*, 613. https://doi.org/10.4095/315362 via NRCan Photo Database; Open Government Licence - Canada

See Figure 7c in https://www.researchgate.net/publication/262559003_Chapter_4_The_Ichnofacies_Paradigm/figures?lo=1

Mermia
Burrows are bedding plane parallel depressions or ridges organized into chaotic loops or circular forms. Interpreted as locomotion traces of worms, midges, or larvae

Nereites
A bedding plane parellel meandering trace consisting of a central furrow flanked by small lobes on both sides of the furrow. Interpreted as a grazing trace. Image from *Richdebtomdom via Wikimedia Commons; CC BY-SA 3.0)*

Olivellites
Sinuous, unbranched, bedding plane parallel trace that has two lobes separated by a medial ridge that runs parallel to the long axis of the trace. The lobes are ornamented with fine curved meniscate-like ornamentation that runs roughly parallel to the long axis of the trace. Interpreted as a gastropod locomotion or grazing trace. Image from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0.

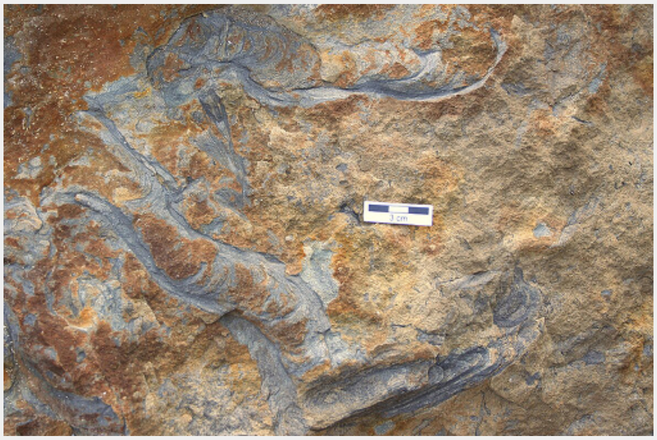
Ophiomorpha
Vertical to horizontal cylindrical burrow with a smooth interior wall and and an outer wall that has a knobby or "corn-cob" like appearance (lined with fecal pellets). Tunnels can change diameter and branch. Interpreted as a crustacean dwelling trace. Image from James St. John via Flickr; CC BY 2.0 DEED.

Paleodictyon
A bedding plane parallel network of hexagonal burrows organized in a honeycomb-like arrangement; vertical shafts common in the modern but rarely preserved. Interpreted as a gardening trace for cultivating microbes; trace maker uncertain. Image from Falconaumanni via Wikimedia Commons; CC BY-SA 3.0
.

Phycosiphon
Looping burrows that extend away from a central tunnel; individual loops build outward via small, subtle spriete. Interpreted as a deposit feeding trace. Image by West Virginia Geologic Survey via https://www.wvgs.wvnet.edu/www/museum/seenfossil/seenfossil.htm

Planolites

A roughly bedding plane parallel unlined burrow that is typically less than 0.5 cm in diameter. Some specimens branch. Burrows are typically linear, cross-cutting, and preserved in positive relief on the base of sandstone beds. *Planolites* and *Thallasinoides* are similar, the main difference is size - the former is more the size of a noodle and the latter is more the size of sausage. Interpreted as the feeding trace of a worm. Image from Gennadi Baranov via Fossilid.info; CC BY-NC.

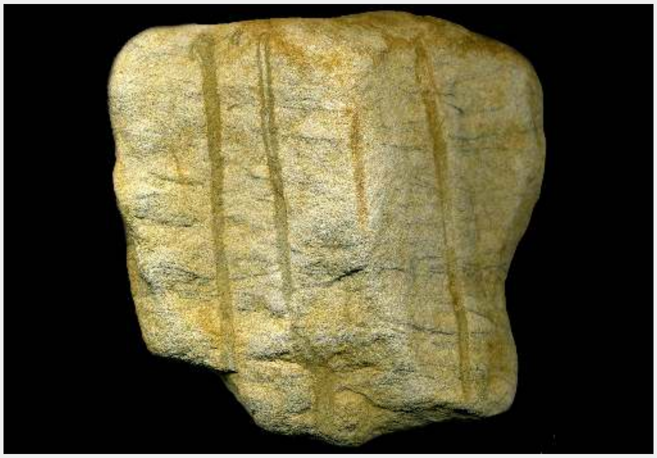


Rhizocorrallium

A horizontal, oblique, or U- or J-shaped burrow that shows muddy spreite between the limbs of the burrow. It is distinguished from *Diplocraterion* because it is typically bent, inclined, or horizontal. It is interpreted as a dwelling or feeding burrow. Image from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0.

Rosselia

A vertically-oriented, downward-tapering funnel-shaped burrow that may be lined with mud. Upper part was originally bulb-shaped but later eroded to form more of a funnel. Interpreted as a dwelling burrow of a worm-like detritus feeder. Image from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0.

	***Rusophycus*** A bilobate depression that may have a medial ridge and scratches in the depressions. Looks like a coffee been. Most commonly preserved in positive relief on the base of beds. Most are interpreted as resting traces of arthropods. Image from James St. John via Flickr; CC BY 2.0 DEED.
You can view an image at https://ichnology.ku.edu/invertebrate_traces/tfimages/scoyenia.html	***Scoyenia*** Variably oriented linear burrow with a rope-like or wrinkled texture. Unbranched but may cross-cut. Interpreted as the feeding trace of terrestrial larvae.
	***Skolithos*** Vertical, typically unlined burrow that is generally filled with material identical to the surrounding substrate. Roughly the diameter of a straw. Similar to *Diplocraterion* but not demonstrably paired. Interpreted as the dwelling trace of a variety of invertebrates. Image from James St. John via Flickr; CC BY 2.0 DEED.
	***Spirorhaphe*** A roughly bedding-plane parallel burrow that spirals inward and then reverses in the center to form a matched pair of coils. Interpreted as the farming trace of a polychaete worm. Image from Falconaumanni via Wikimedia Commons; CC BY-SA 3.0

	***Taenidium*** Unlined sinuous burrow with pronounced meniscate backfill. Interpreted as the feeding trace of a worm-like organism. Image by Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0.
	**Tetrapod trackway** Any number of named trackways that have footprints from front and back feet. Marks from toes commonly preserved. Morphology highly dependent upon substrate. Tetrapod locomotion trace. Image from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0.
	***Thallasinoides*** A relatively large horizontal to vertical unlined burrow; may branch to form a Y- or a T-shape. *Planolites* and *Thallasinoides* are similar, the main difference is size - the former is more the size of a noodle and the latter is more the size of a hot dog or sausage. Note: if burrow has a knobby texture via fecal pellet lining, it is called *Ophiomorpha*. Interpreted as a feeding or dwelling burrow of a crustacean. Image from Mark Wilson via Wikimedia Commons; public domain.

Undichnia
Assemblage of sinuous grooves formed by dragging fish fins. Tail trace is commonly sinusoidal; pectoral fin traces are parallel to each other and more gently curved. Interpreted as a fish locomotion trace. Image from Figure 7 in Ronchi, A.; Marchetti, L.; Klein, H.; Groenewald, G.H. A Middle Permian Oasis for Vertebrate and Invertebrate Life in a High-Energy Fluvial Palaeoecosystem of Southern Gondwana (Karoo, Republic of South Africa). *Geosciences* 2023, *13*, 325. https://doi.org/10.3390/geosciences13110325; CC BY 4.0 DEED

Zoophycos
A central vertical burrow that has corkscrew-shaped tiers of sheet-like subhorizontal burrows that make a "rooster tail" or "pinwheel" pattern on bedding planes. Three dimensional form rarely preserved; flattened preservation roughly along bedding planes is much more common. Most are intereted as the feeding burrow of a worm-like organism. Image from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0.

9.6.4: Seilacherian (Behavioral) Ichnofacies

Trace fossils record behaviors and those behaviors are dictated by environmental conditions; consequently certain types and assemblages of trace fossils tend to occur together in time and space because the behaviors they record work in a given environment. These assemblages are called ichnofacies and are named for a distinctive trace fossil that is typical of that environment. It is worth noting that a given trace fossil can be present in a variety of environments and the focus is on the assemblage of traces, not a single diagnostic trace fossil. Although nearly two dozen ichnofacies have been described, the diagrams provided below highlight the four most important/common marine ichnofacies (*Skolithos*, *Cruziana*, *Zoophycos*, and *Nereites*), two continental ichnofacies (*Scoyenia* and *Mermia*), and three substrate-dependent ichnofacies (*Glossofungites*, *Teredolites*, and *Trypanites*).

9.6.4.1: Marine Ichnofacies

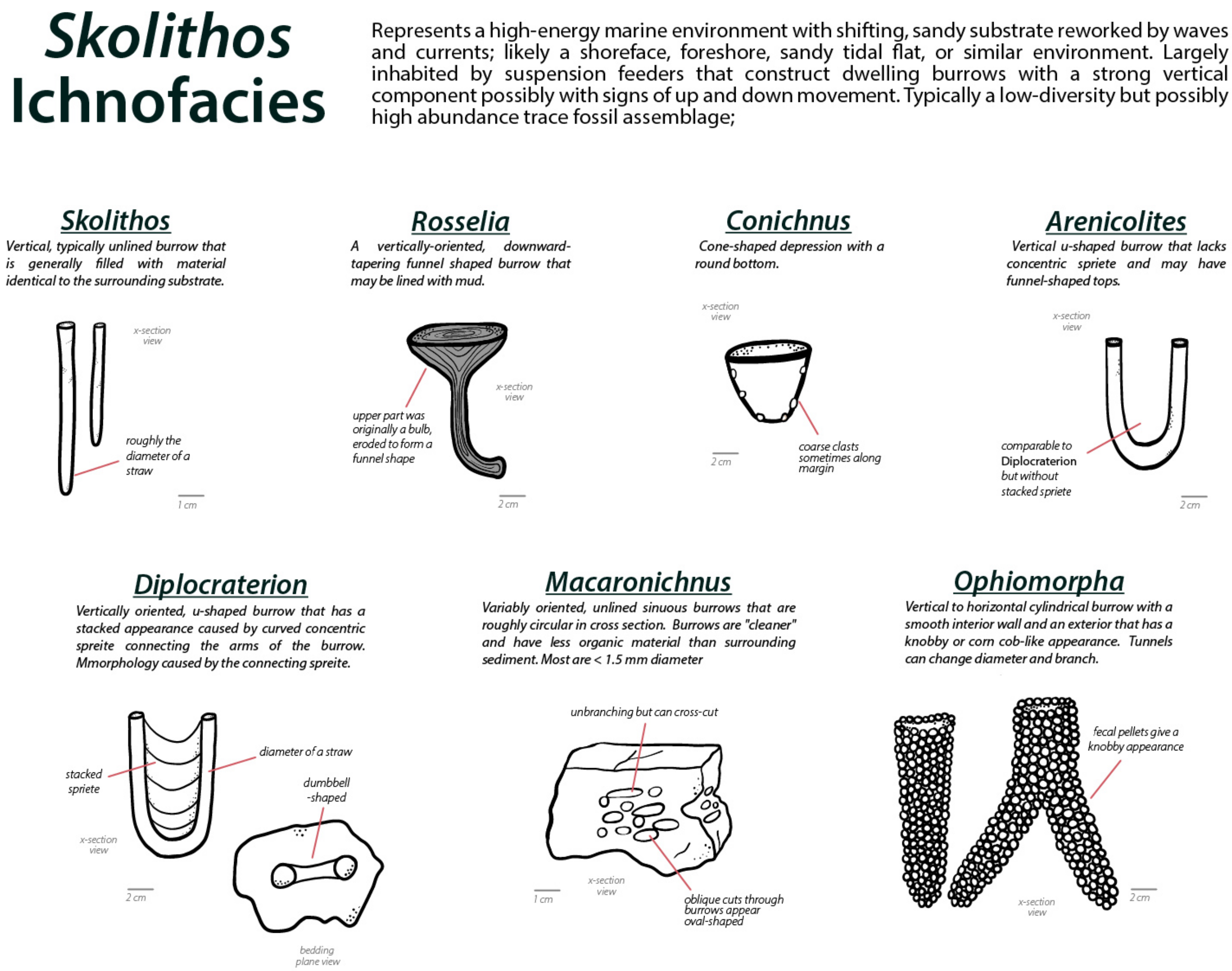


Figure 9.6.8: Explanation of the *Skolithos* Ichnofacies and illustrations of some of the most common fossils in it (Page C. Quinton via Wikimedia Commons; CC BY-SA 4.0).

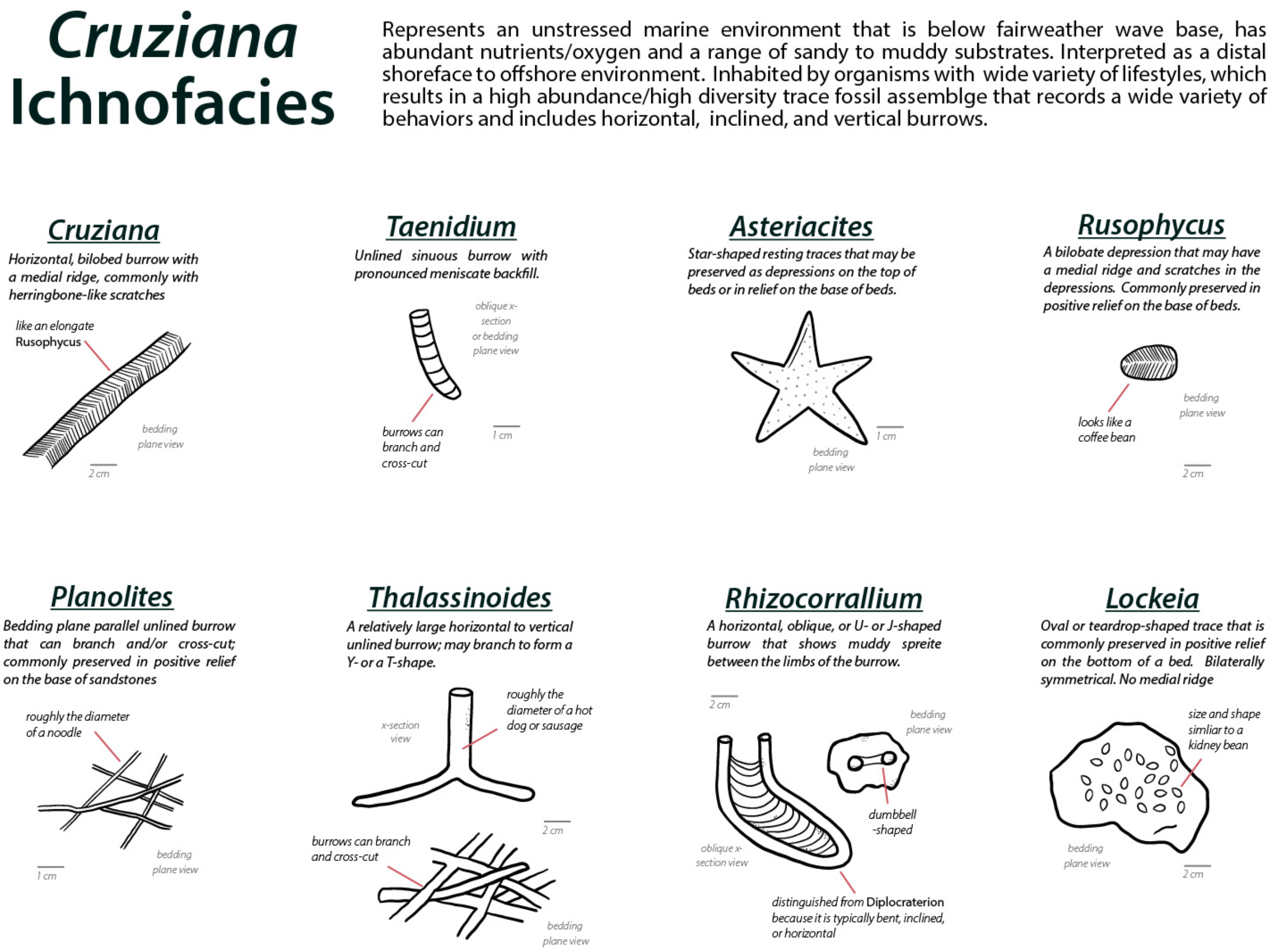


Figure 9.6.9: Explanation of the *Cruziana* Ichnofacies and illustrations of some of the most common fossils in it (Page C. Quinton via Wikimedia Commons; CC BY-SA 4.0).

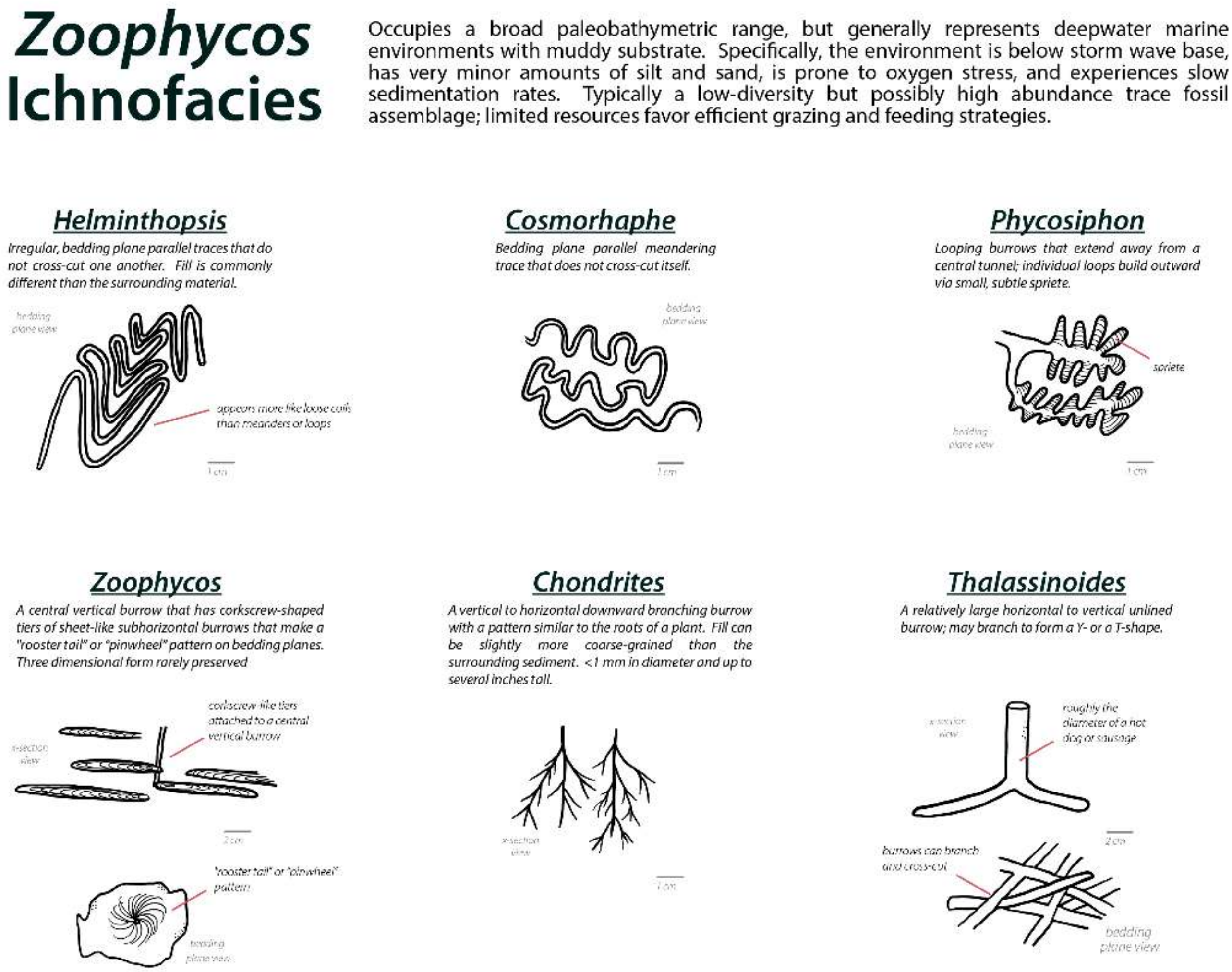


Figure 9.6.10: Explanation of the *Zoophycos* Ichnofacies and illustrations of some of the most common fossils in it (Page C. Quinton via Wikimedia Commons; CC BY-SA 4.0).

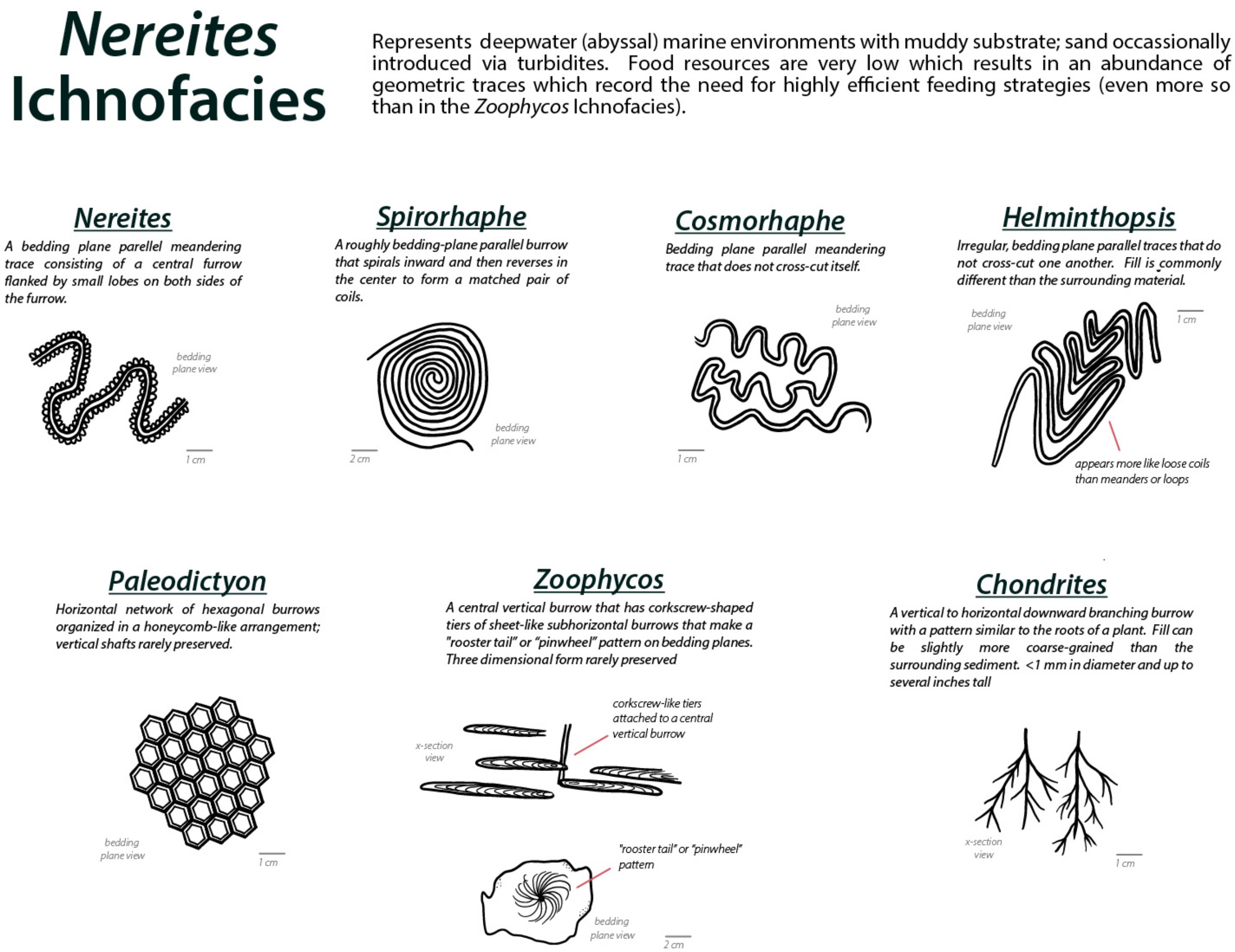


Figure 9.6.11: Explanation of the *Nereites* Ichnofacies and illustrations of some of the most common fossils in it (Page C. Quinton via Wikimedia Commons; CC BY-SA 4.0).

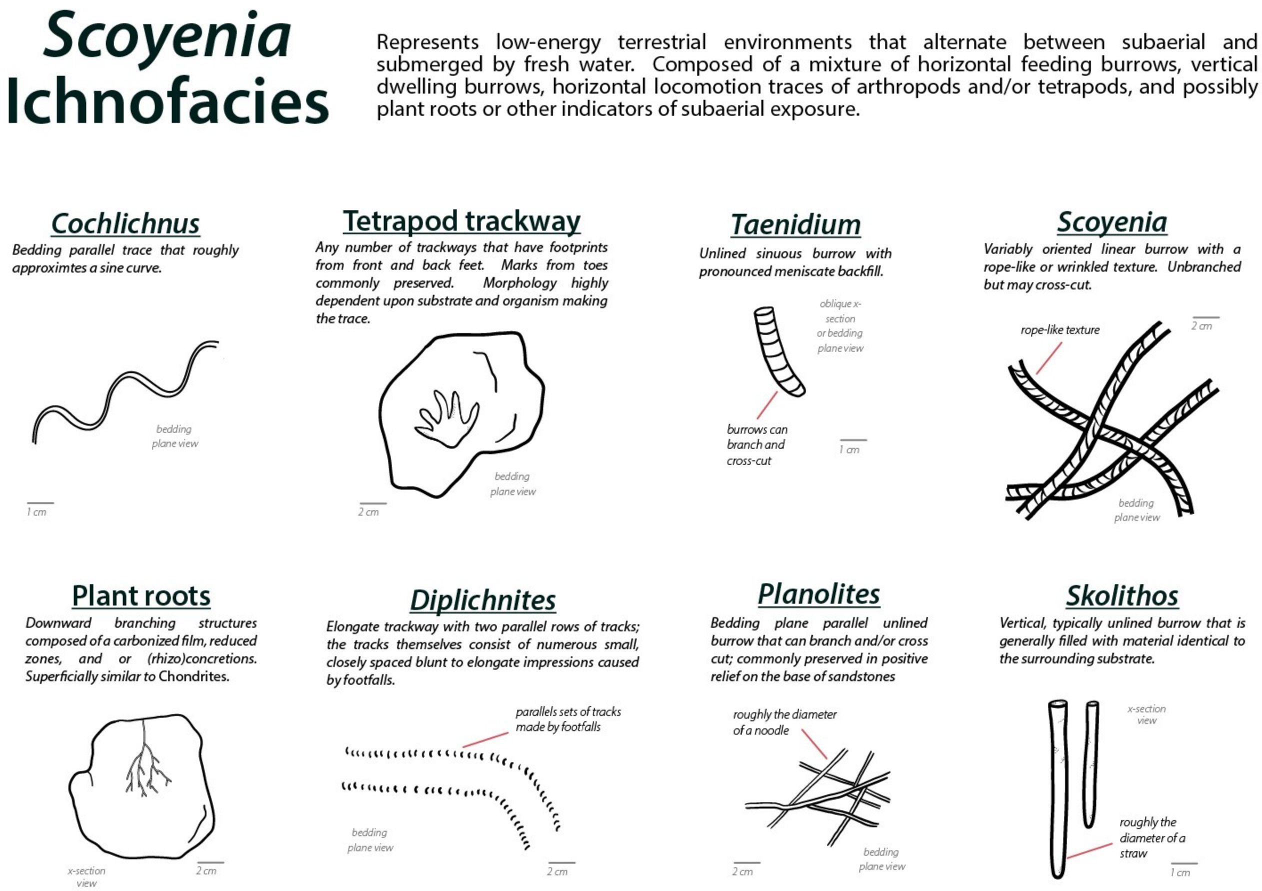


Figure $9.6.12$: Explanation of the *Scoyenia* Ichnofacies and illustrations of some of the most common fossils in it (Page C. Quinton via Wikimedia Commons; CC BY-SA 4.0).

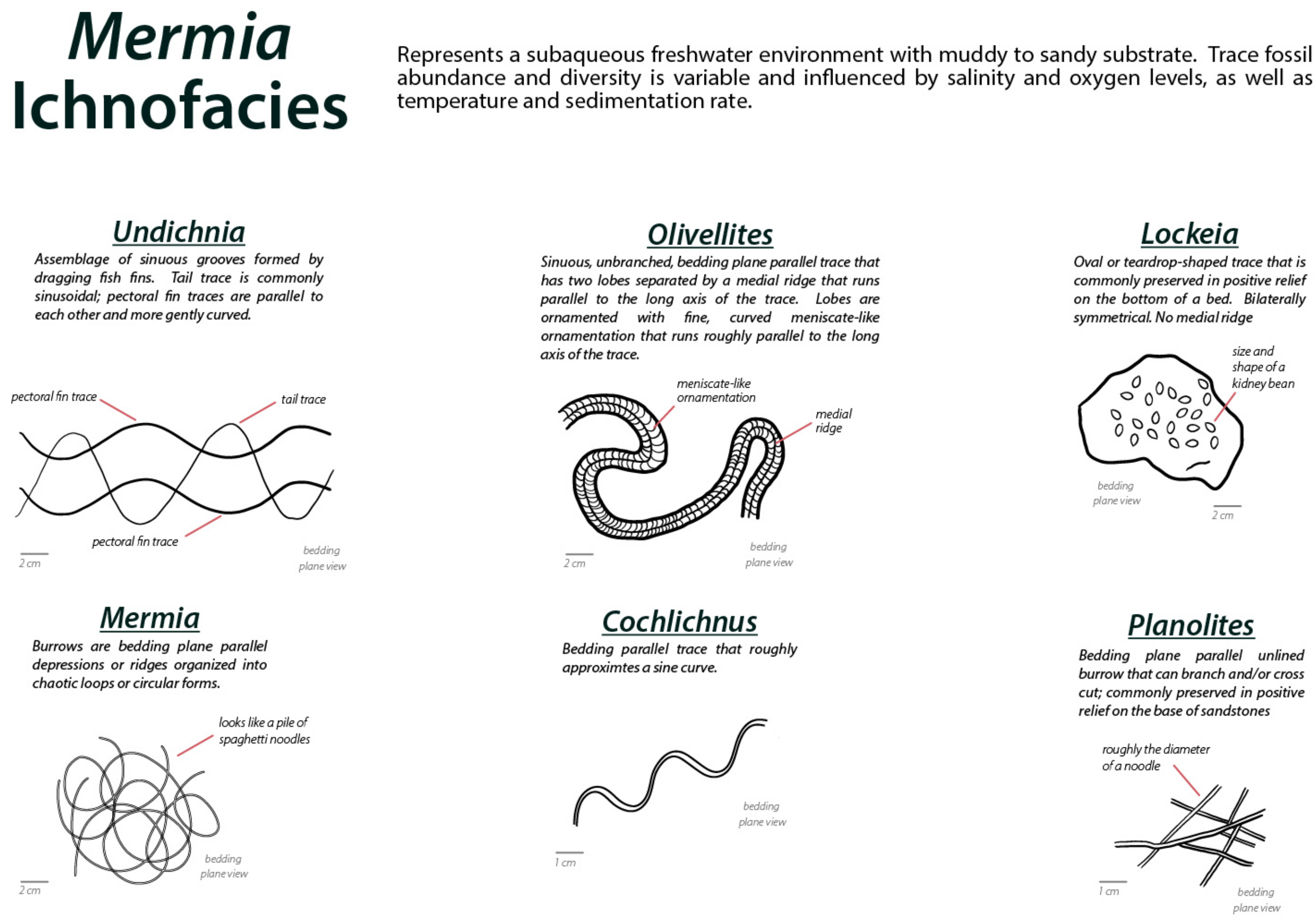


Figure 9.6.13: Explanation of the *Mermia* Ichnofacies and illustrations of some of the most common fossils in it (Page C. Quinton via Wikimedia Commons; CC BY-SA 4.0).

9.6.4.3: Substrate-Dependent Ichnofacies

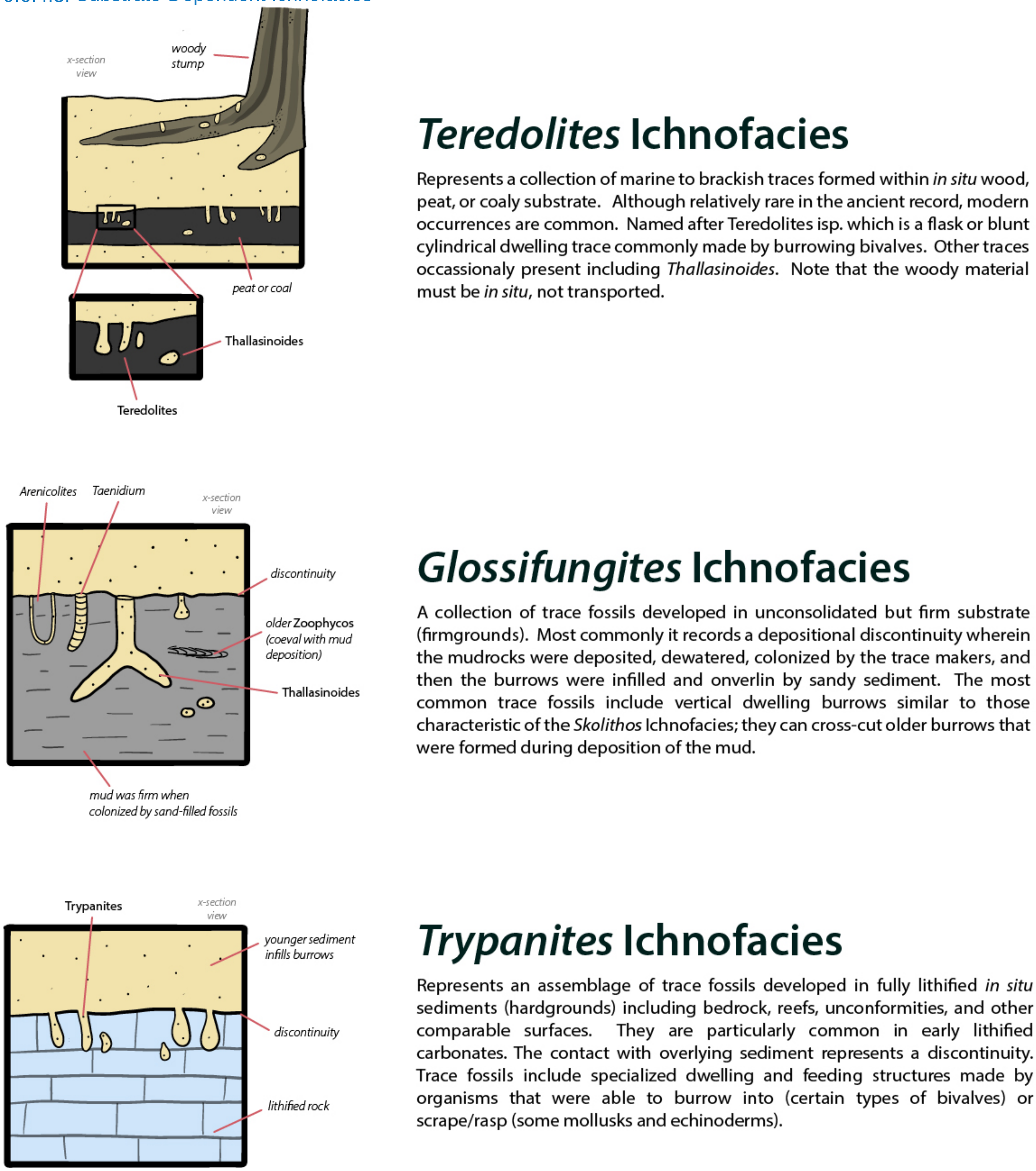


Figure 9.6.14: Explanation of the three substrate dependent ichnofacies and illustrations of some of the most common fossils in it (Page C. Quinton via Wikimedia Commons; CC BY-SA 4.0).

Figure 9.6.15: Marine bivalves of the genus *Pholas* created these circular burrows in peat deposits exposed in the intertidal zone and thus represent the *Teredolites* Ichnofacies. Image copyright Jessica M. Winder and Jessica's Nature Blog at https://natureinfocus.blog/2018/05/13/holes-in-peat-clay-at-broughton. Reproduced with written permission of the author for the educational purposes of this Libretexts contribution (NC ND + no further reproduction).

Figure 9.6.16: Sand-filled trace fossils in black shales of the lower Pebbly Beach Formation (Permian). These palimpsest trace fossils were probably excavated in firm substrate and could be interpreted as part of the *Glossofungites* Ichnofacies (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

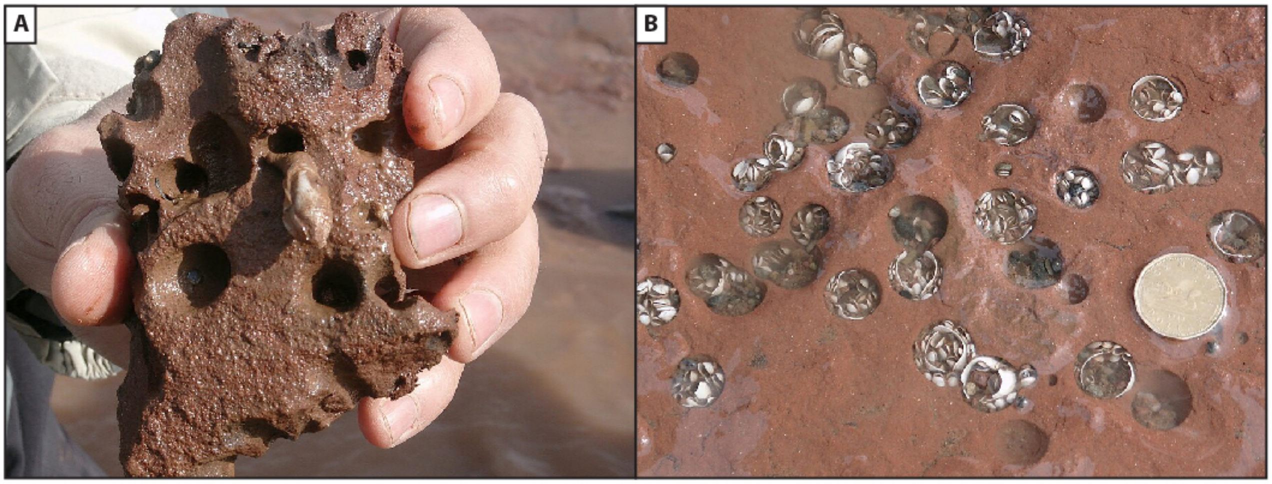


Figure 9.6.17: Angelwing bivalves and associated burrows in poorly-consolidated sandstones exposed in the intertidal zone represent the *Trypanites* Ichnofacies. These bivalves use their shells like a rasp and are able to slowly burrow into firm substrate by opening and closing their shells to slowly rotate and abrade the surrounding material (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0).

9.6.5: Readings and Resources

- PBS Eons, Something Has Been Making This Mark For 500 Million Years (Video on *Paleodictyon*) - https://www.youtube.com/watch?v=Pz1fccY3S84
- KU Ichnology webpage - https://ichnology.ku.edu/
- Ichnofossils at FossilID.info - https://fossiilid.info/112?mode=in_baltoscandia
- Digital Atlas of Ancient Life's trace fossil webpage - https://www.digitalatlasofancientlife.org/vc/trace-fossils/

- Bromley, R.G., 1996, Trace fossils—Biology, taphonomy, and applications [2d ed.]: London, Chapman and Hall, 384 p. - https://www.taylorfrancis.com/books/mono/10.4324/9780203059890/trace-fossils-richard-bromley
- Ekdale, A.A., Bromley, R.G., and Pemberton, S.G., 1984, Ichnology; the use of trace fossils in sedimentology and stratigraphy: Society of Economic Paleontologists and Mineralogists, Short Course No. 15, 317 p.
- Frey, R.W., 1973, Concepts in the study of biogenic sedimentary structures: Journal of Sedimentary Research, v. 43, no. 1, p. 6–19. - https://chooser.crossref.org/?doi=10.1306%2F74D726C1-2B21-11D7-8648000102C1865D
- MacEachern, J.A., Bann, K.L., Gingras, M.K., and Pemberton, S.G., 2007, Applied Ichnology, SEPM Short Course Notes, No. 52, 145 p. - https://sedimentary-geology-store.com/catalog/book/applied-ichnology
- Knaust,D., and Bromley, R.G., 2012, Trace Fossils as Indicators of Sedimentary Environments; Developments in Sedimentology v. 64, 924 p. - https://www.sciencedirect.com/bookseries/developments-in-sedimentology/vol/64/suppl/C
- Seilacher, A., 1953, Studien zur Palichnologie I. Uber die methoden der Palichnologie: Neues Jahrbuch Geologie Palaontologie, Abh. 96, p. 421–452.
- Seilacher, A., 2007, Trace Fossil Analysis, Springer-Verlag, Berlin, 226 p. - https://link.springer.com/book/10.1007/978-3-540-47226-1

9.7: Fossils in Thin Section

9.7.1: Diagrams Showing Common Fossils

The following diagrams show schematic representations of some of the most common fossils encountered in thin sections; they are not organized taxonomically, but rather based on gross morphology. Please note that the scale bar in the following four diagrams is approximate and that the actual sizes of the fossils can vary by an order of magnitude or more.

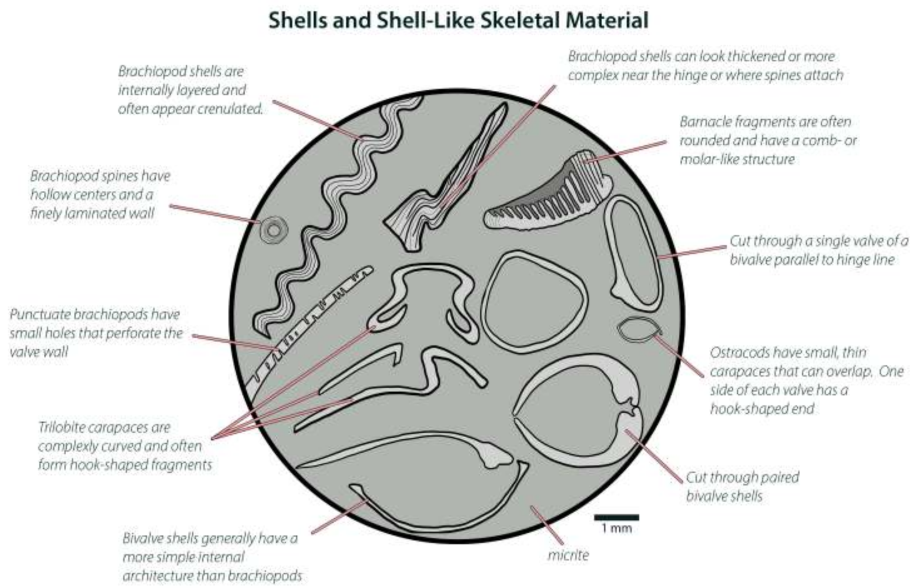


Figure 9.7.1: Fossils that are, or appear to be, made out of a walled shell or exoskeleton include brachiopods, bivalves, barnacles, and trilobites (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

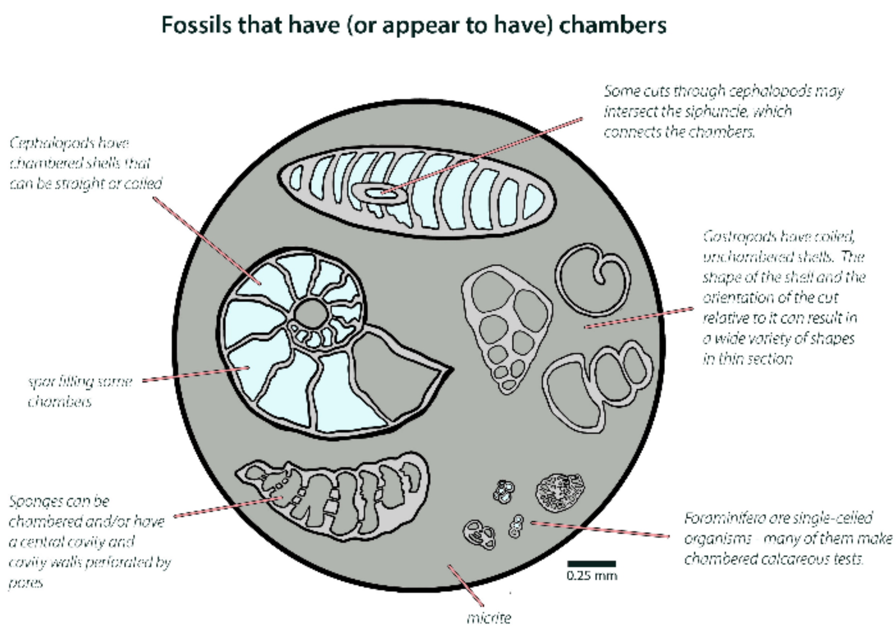


Figure 9.7.2: Fossils that have, or appear to have, internal chambers include cephalopods, sponges, gastropods, and foraminifera (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

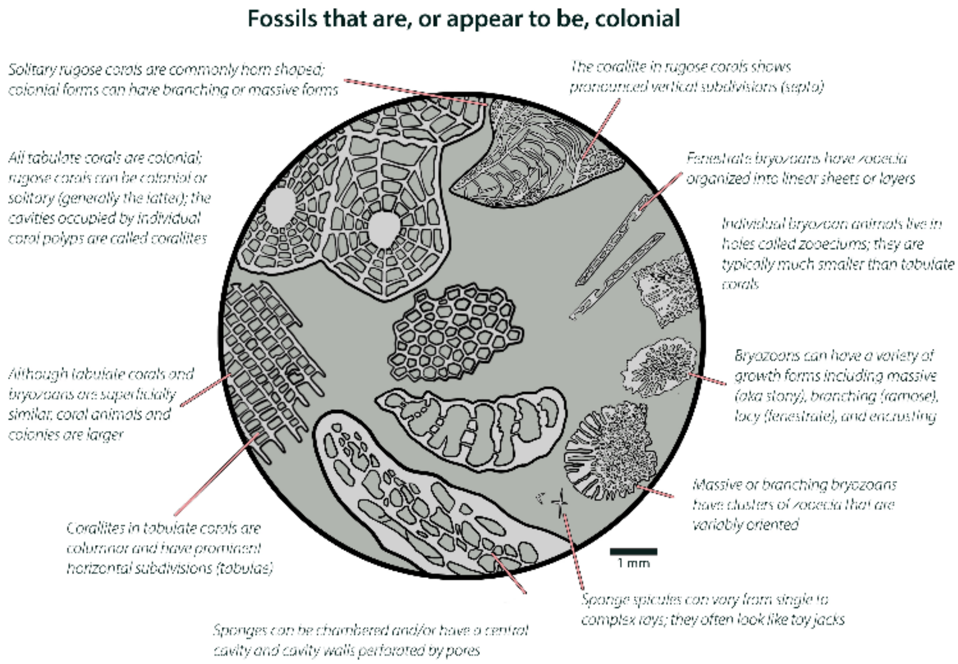


Figure 9.7.3: Fossils that are, or appear to be, colonial include tabulate corals, rugose corals, bryozoans, and sponges (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

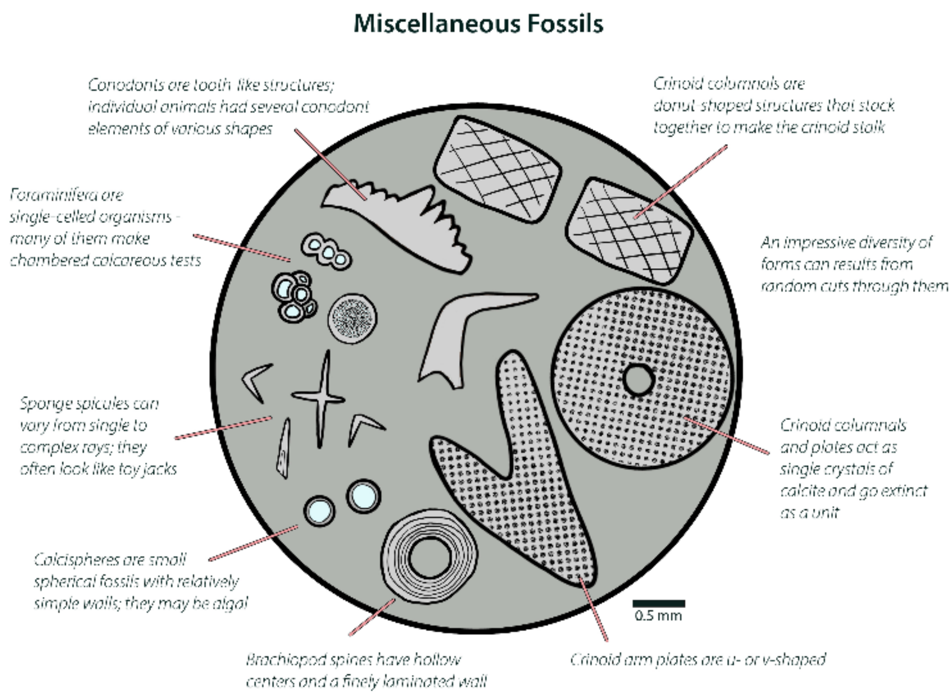


Figure 9.7.4: Other distinctive or noteworthy fossils encountered in thin section include crinoid plates, crinoid columnals, brachiopod spines, calcispheres, sponge spicules, and conodonts (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

9.7.2: Porifera (Sponges)

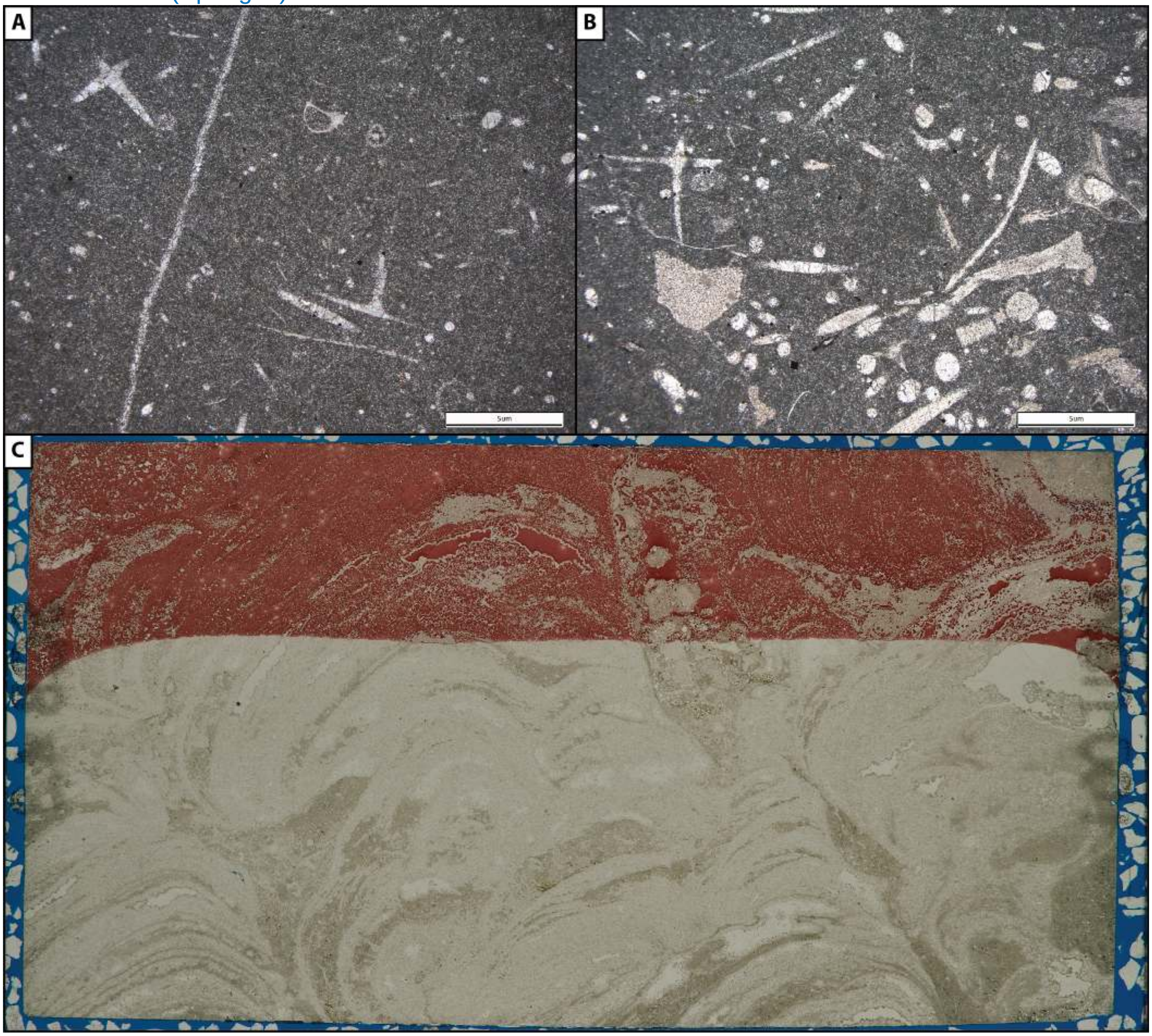


Figure 9.7.5: Examples of sponges in thin section. A and B) Sponge spicules are the most commonly type of sponge fossils in most settings; they vary from simple to complex rays and commonly resemble toy jacks. C) Entire sponges are much less commonly preserved and, as is the case with these encrusting examples, can be difficult to distinguish from stromatolites (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

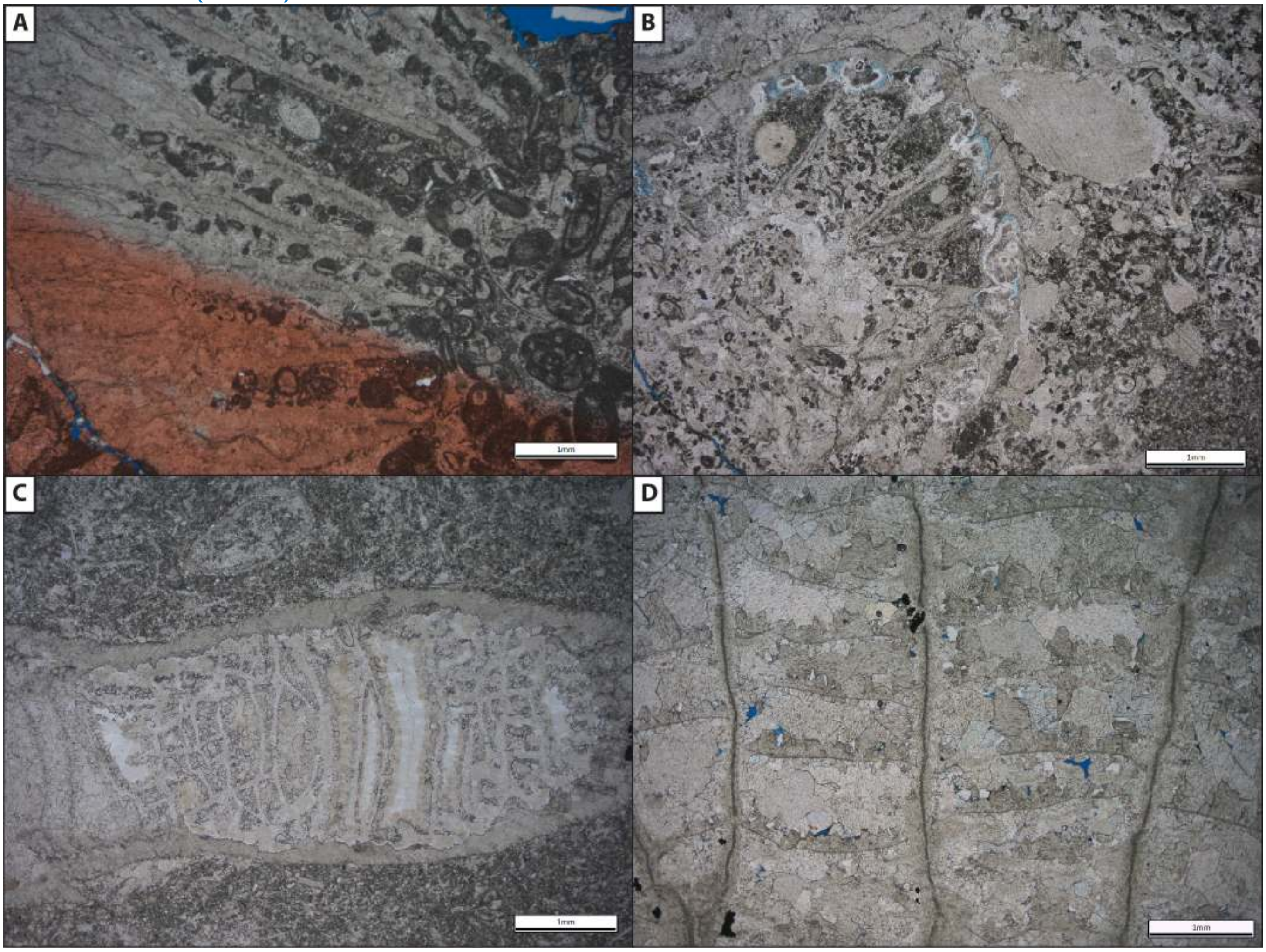


Figure 9.7.6: Examples of rugose (A & B) and tabulate (C & D) corals in thin section. Rugose corals can be solitary or colonial and have corallites with prominent vertical subdivisions (septa). Tabulate corals are colonial and have corallites with pronounced horizontal subdivisions that makes colonies look like storeys in a building (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

9.7.4: Lophoporhata

9.7.4.1: Brachiopoda

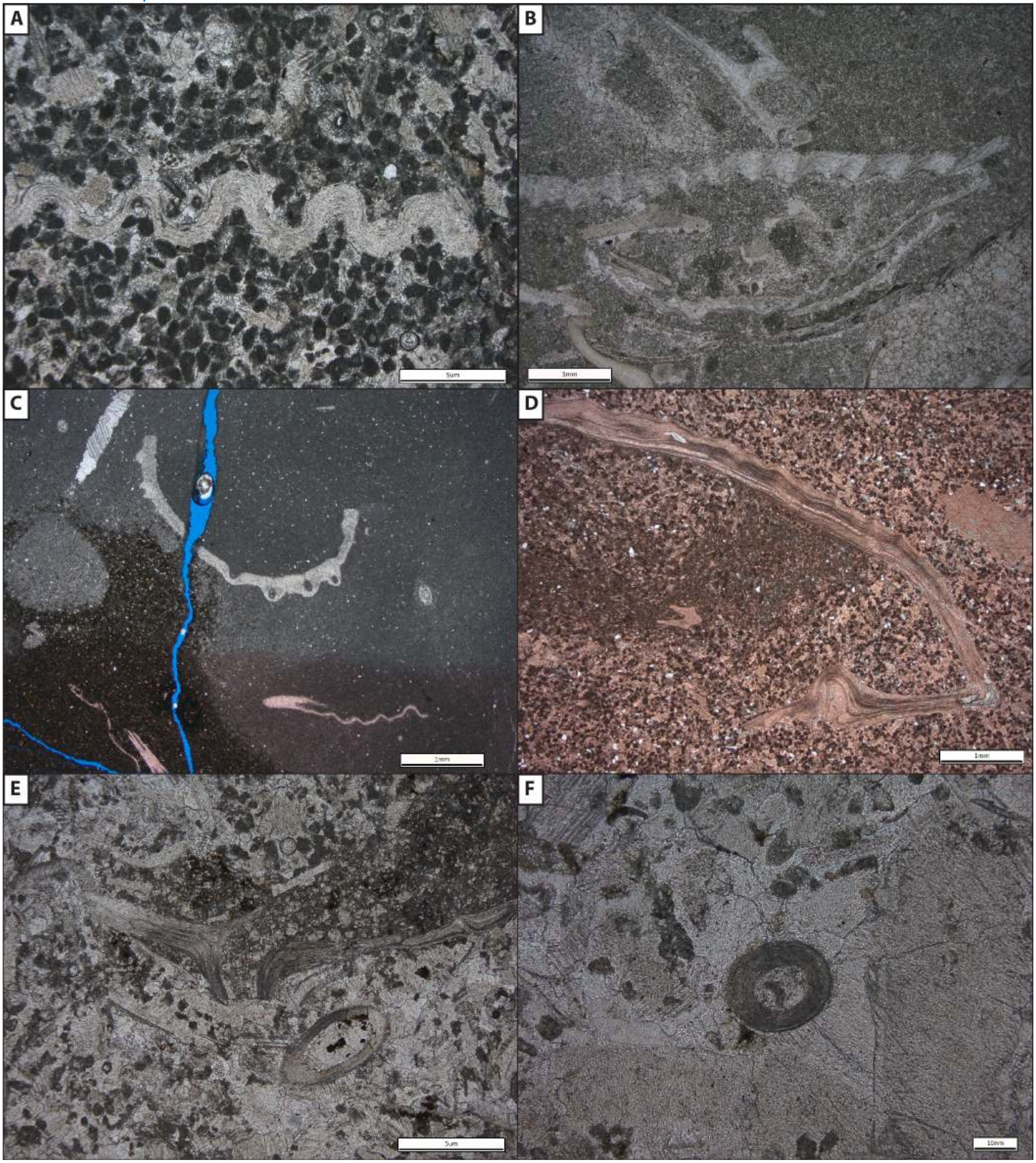


Figure 9.7.7: Brachiopod fossils in thin section. A-D) Brachiopod valves often appear crenulated and have a complex or ornamented appearance. E & F) show brachiopod spines which have hollow interiors and finely laminated walls (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

9.7.4.2: Bryozoa

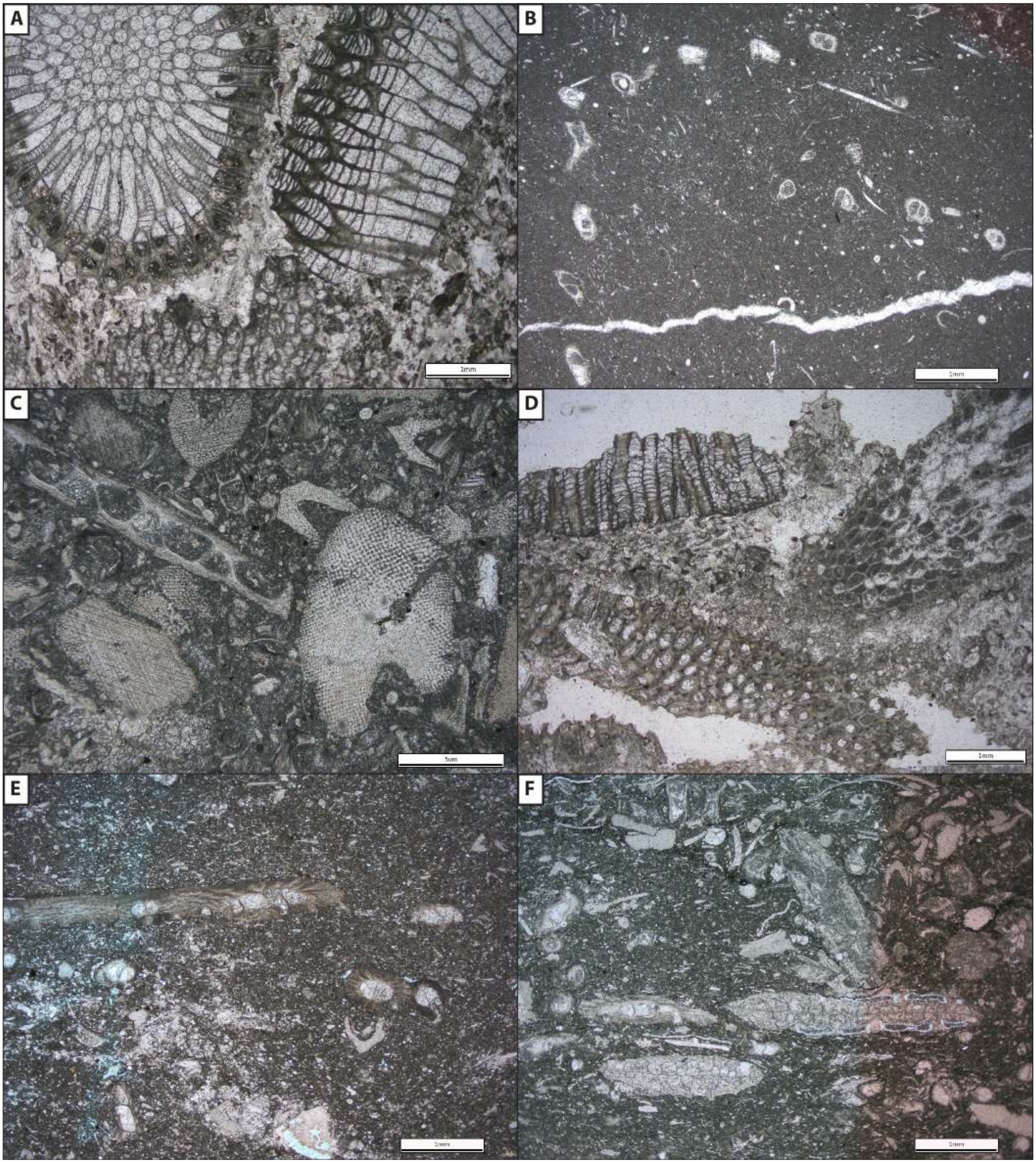


Figure 9.7.8: Bryozoans can show a variety of morphologies (stony, branching, lacy, etc.) and show an equally wide range of morphologies in thin section. Although they superficially resemble corals, they are not closely related and generally have much smaller openings (typically < 1 mm) than corals. A) shows a cross section through branching bryozoan, B and C) show sections through lacy forms (note that the bryozoan is the chambered feature in the NW quadrant of C; much of the rest of the image is filled with crinoid plates). E & F) show variable cuts through probably lacey bryozoans (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

9.7.5: Mollusca

9.7.5.1: Gastropods

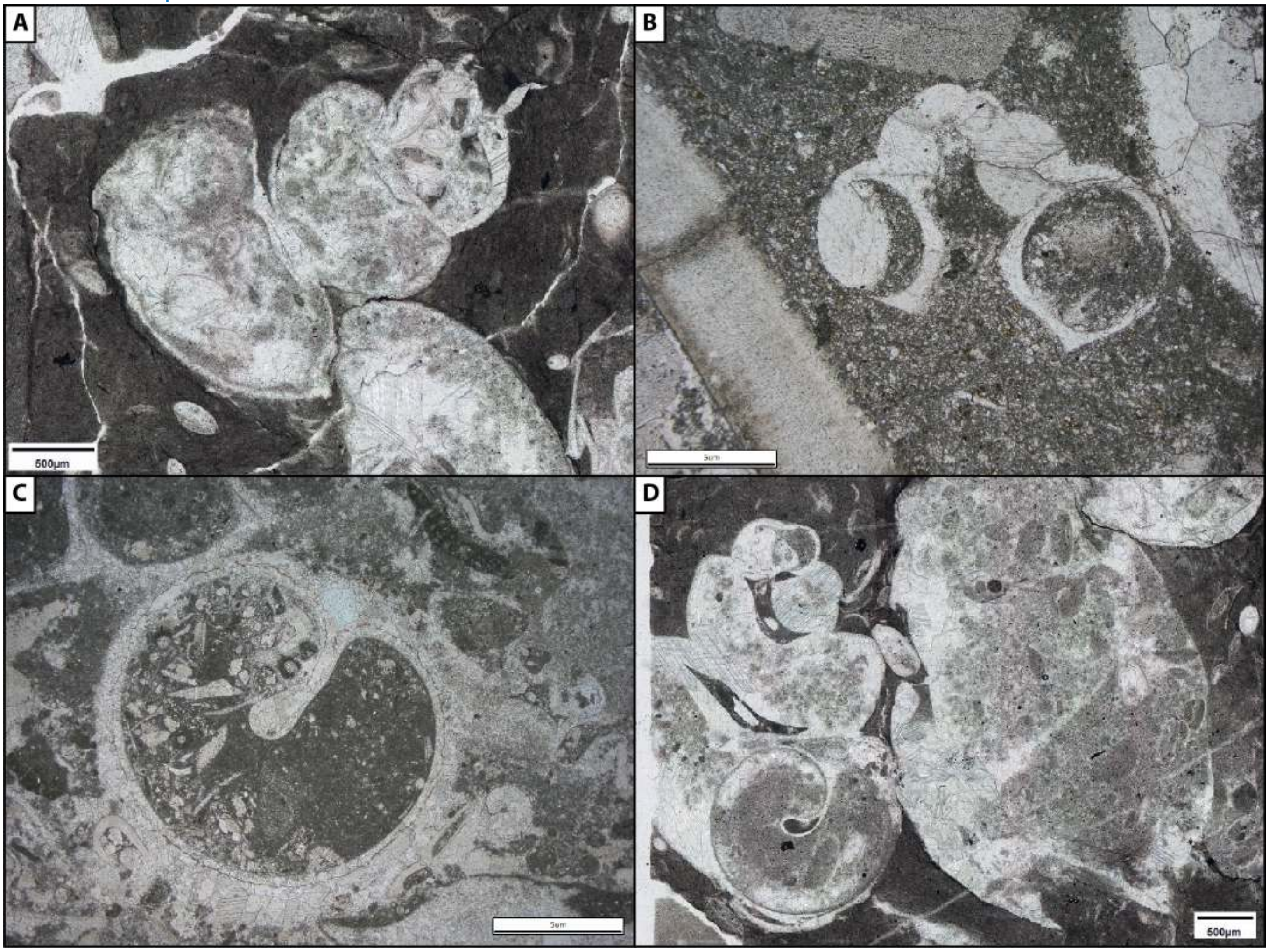


Figure 9.7.9: Examples of gastropods in thin section. A and B) show longitudinal cuts through gastropods with medium height and squat spirals. C) Shows a transverse cut with the spirals extending into and out of the plane of the photo. D) The left half of the image shows a transverse cut through a gastropod contained in the outermost spiral of a larger gastropod that is cut in longitudinal section (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

9.7.5.2: Cephalopods

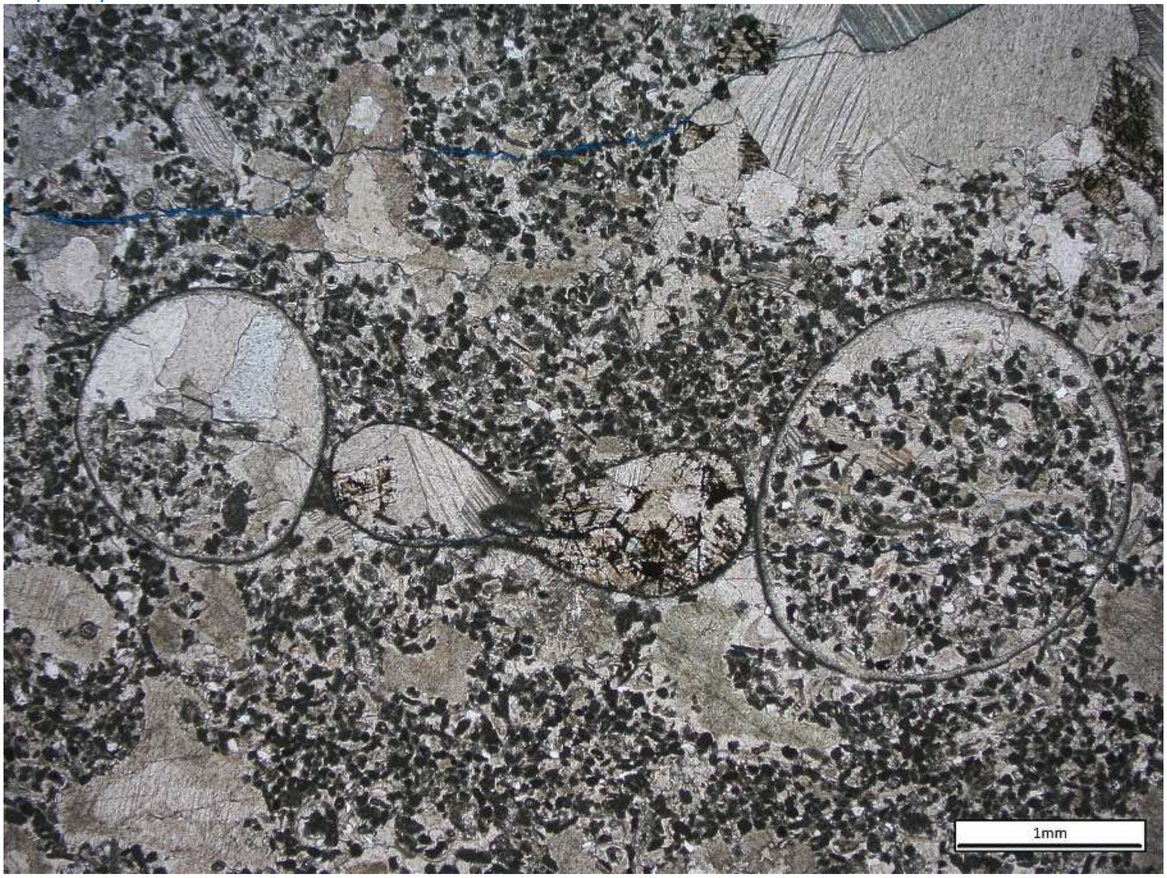


Figure 9.7.10: Axial cross section through a possible coiled cephalopod showing progressive decrease in the diameter of the chamber. Without preservation of a siphuncle or internal septum, it is also possible that this is a gastropod that does not coil out of plane. Note the geopetal fill with sediment at the bottom and spar at the top (Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

9.7.6: Arthropoda

9.7.6.1: Ostracods

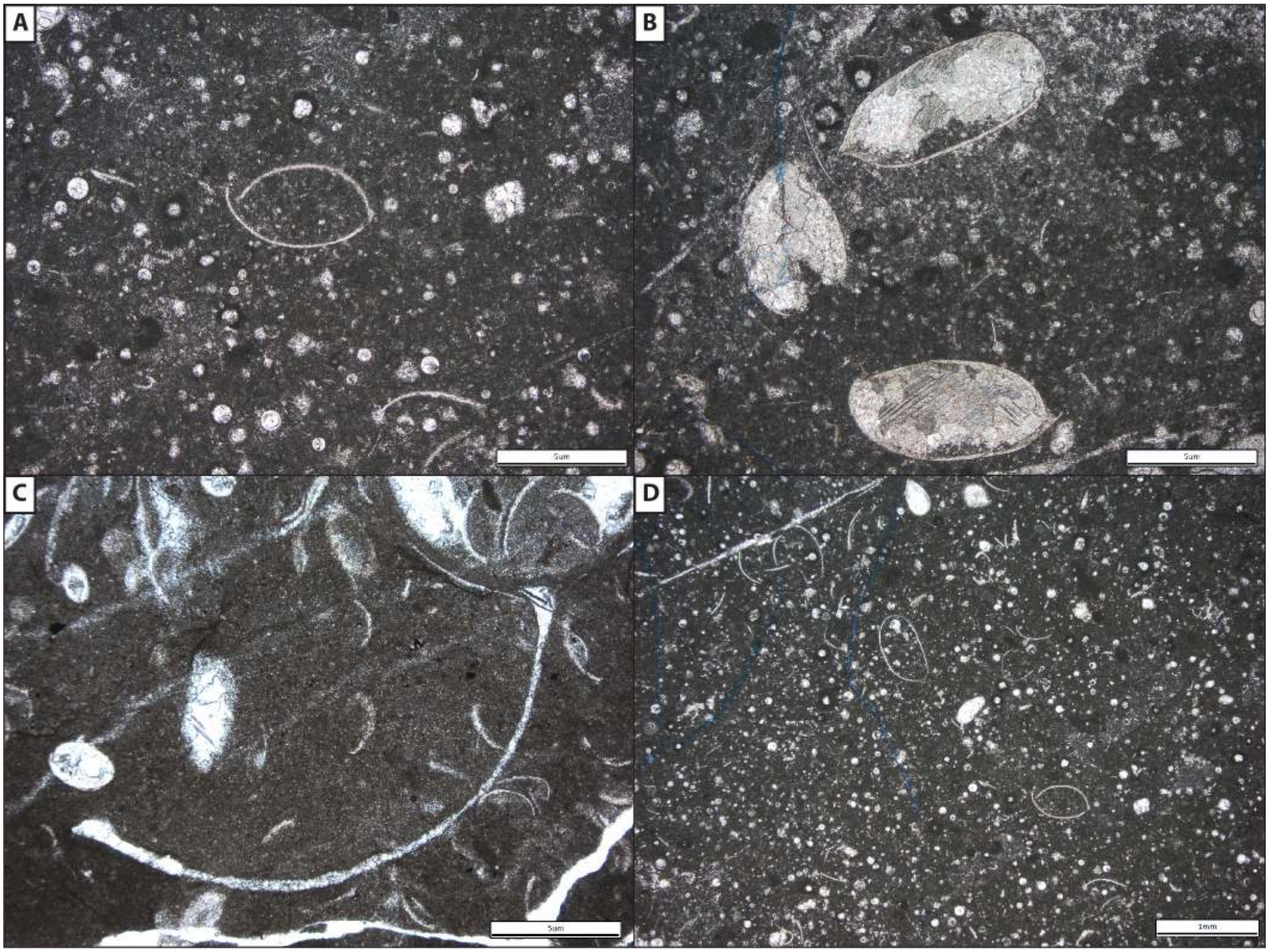


Figure 9.7.11: Ostracods are small arthropods with thin valves. The valves halves are held together with soft tissue, partially overlap, and have a hook-shaped edge opposite the hinge. Note the large bivalve in C and numerous calcispheres in both A and D (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

9.7.6.2: Trilobites

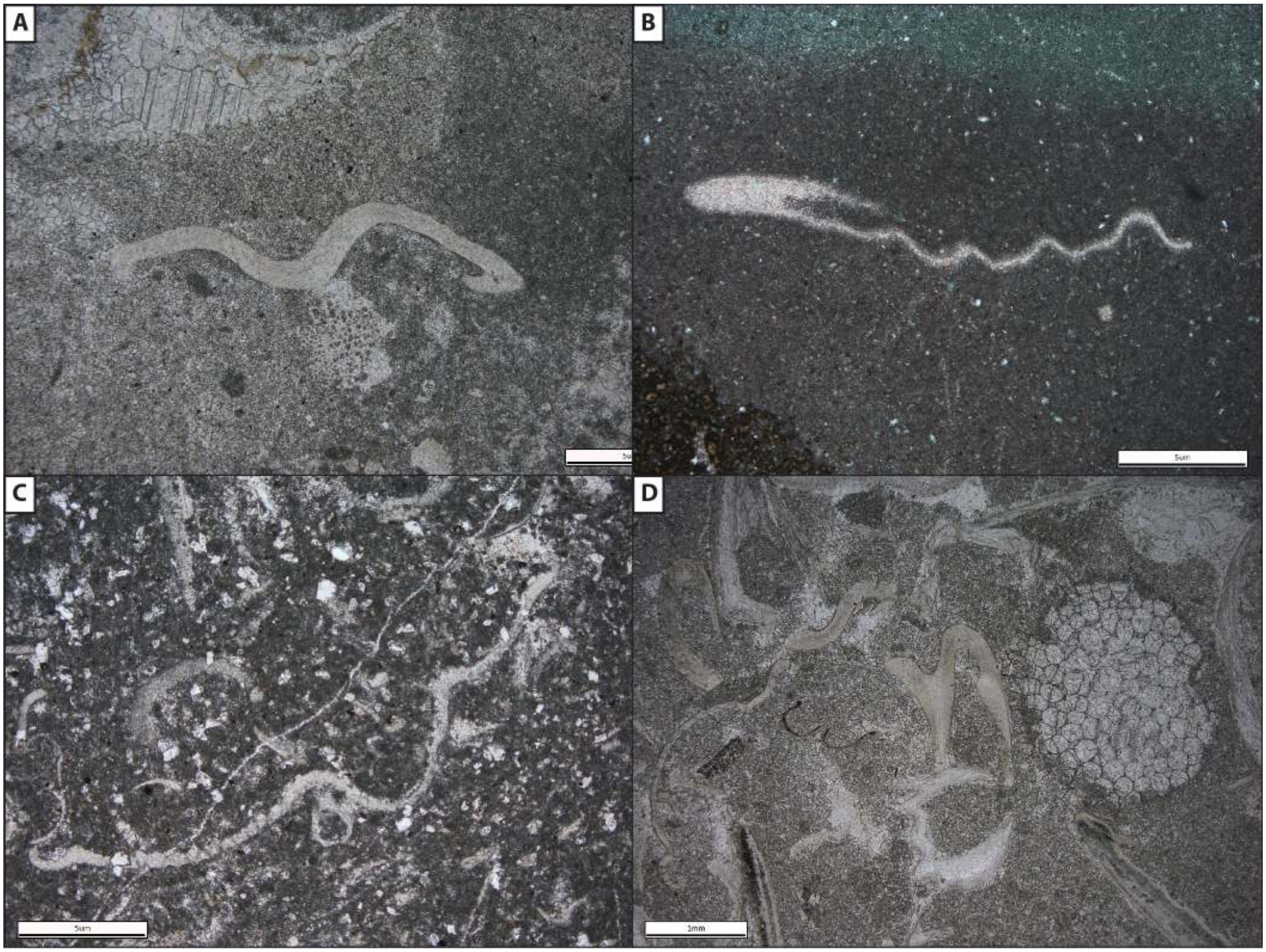


Figure 9.7.12: Trilobites were segmented arthropods that molted as they grew which makes their fossils particularly abundant in some beds. In thin section they are preserved as curved fossils that can terminate with a distinctive shepard's hook that is morphologically similar to, but larger than that of ostracods (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

9.7.6.3: Barnacles

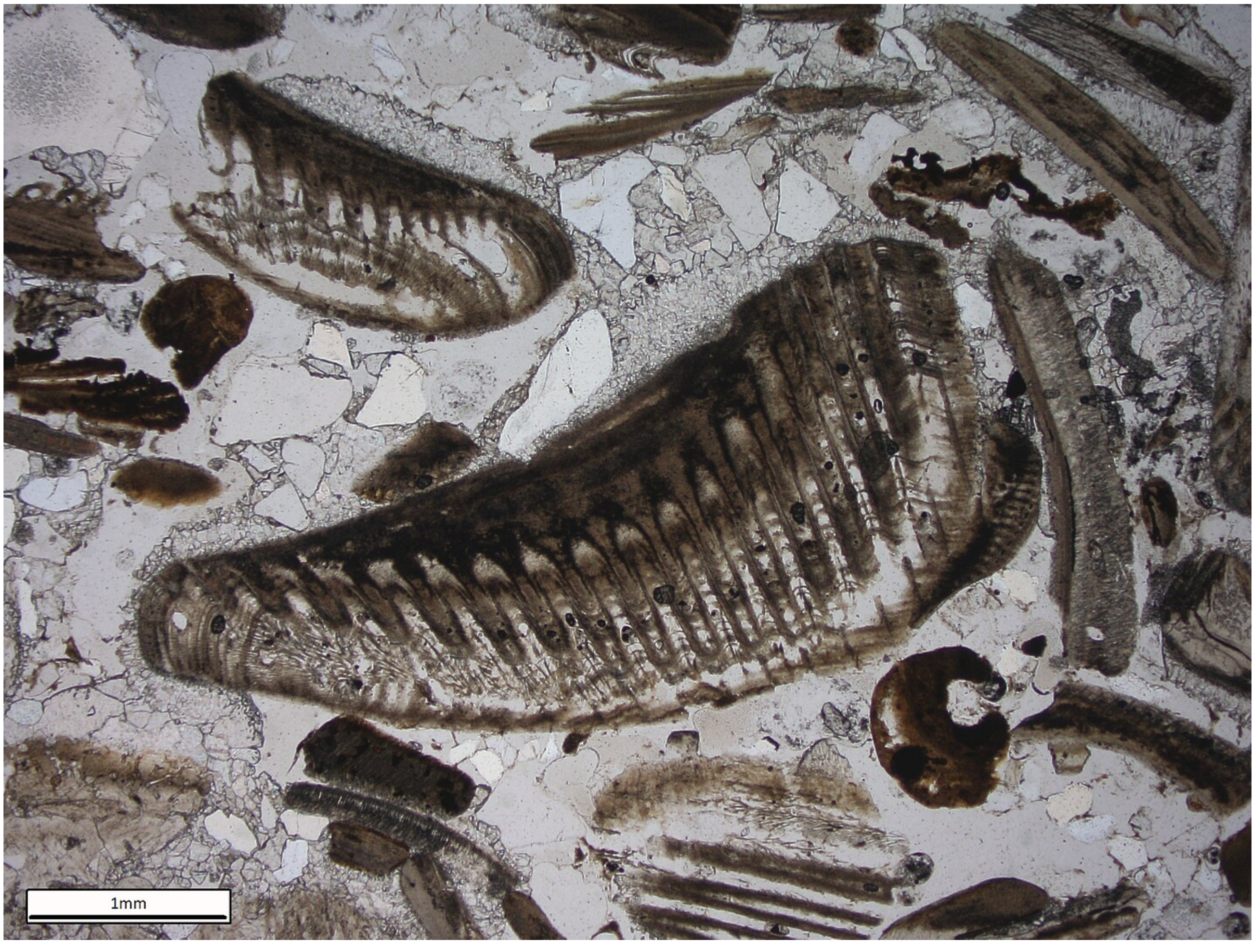


Figure 9.7.13: Barnacles are small to medium-sized crustaceans that, in their adult form, are attached to a hard substrate. They surround their bodies with a series of complexly-shaped plates that have a ribbed architecture that can appear folded and tooth-like in cross section (Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

9.7.7: Echinodermata

9.7.7.1: Crinoids

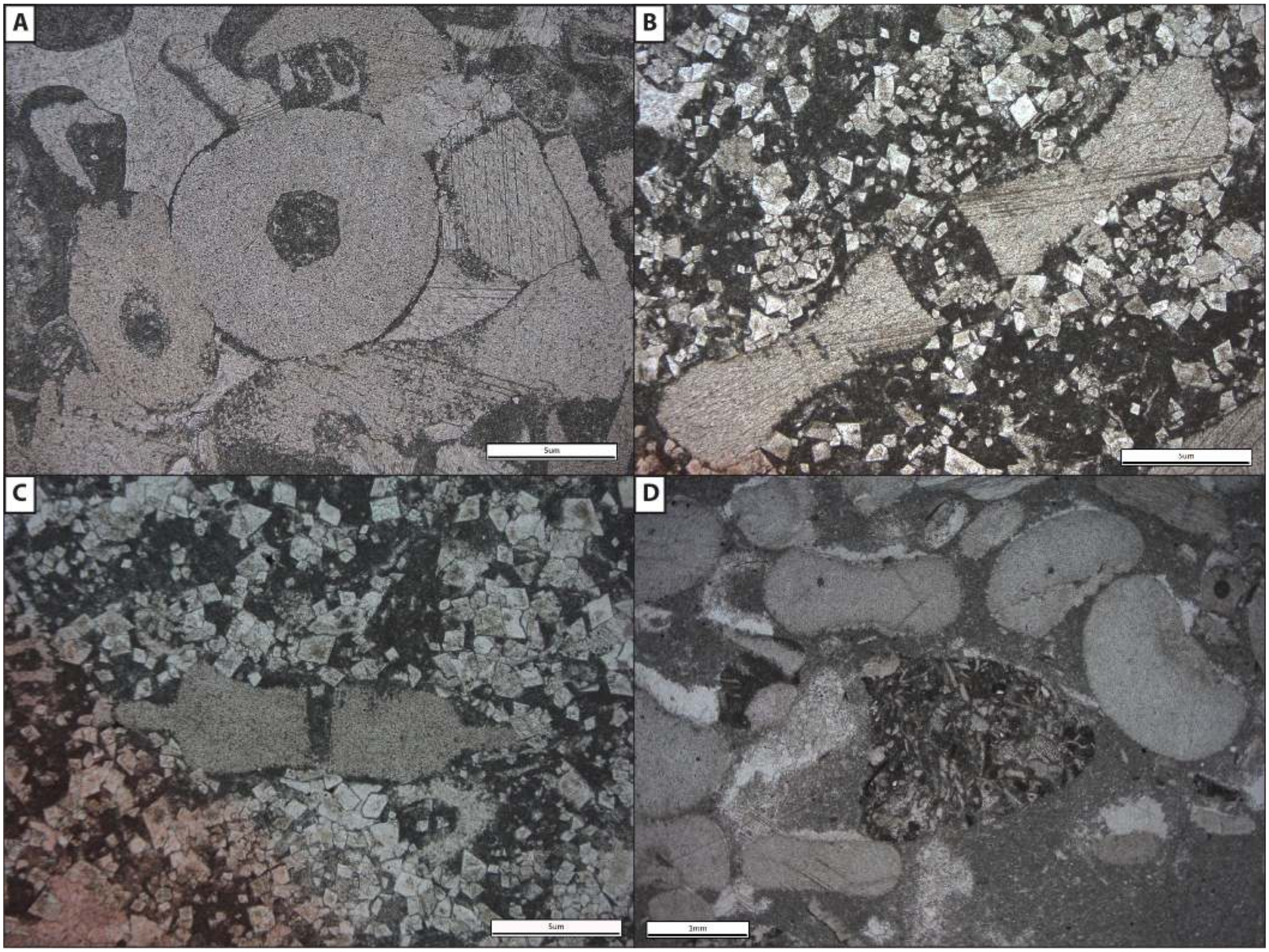


Figure 9.7.14: Crinoid columnals are donut-shaped segments that make up the stalk of crinoids, they have a central canal and small arms (cirri) can attach to them. Two dimensional cuts through them can result in wide variety of morphologies. A) shows a transverse cut through a columnal with results in a circular shape with a clearly visible central canal. B and C) show transverse cuts through columnals that intersect the central canal. D) shows a small intraclast surrounded by several transverse cuts through columnals that don't intersect the canal (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

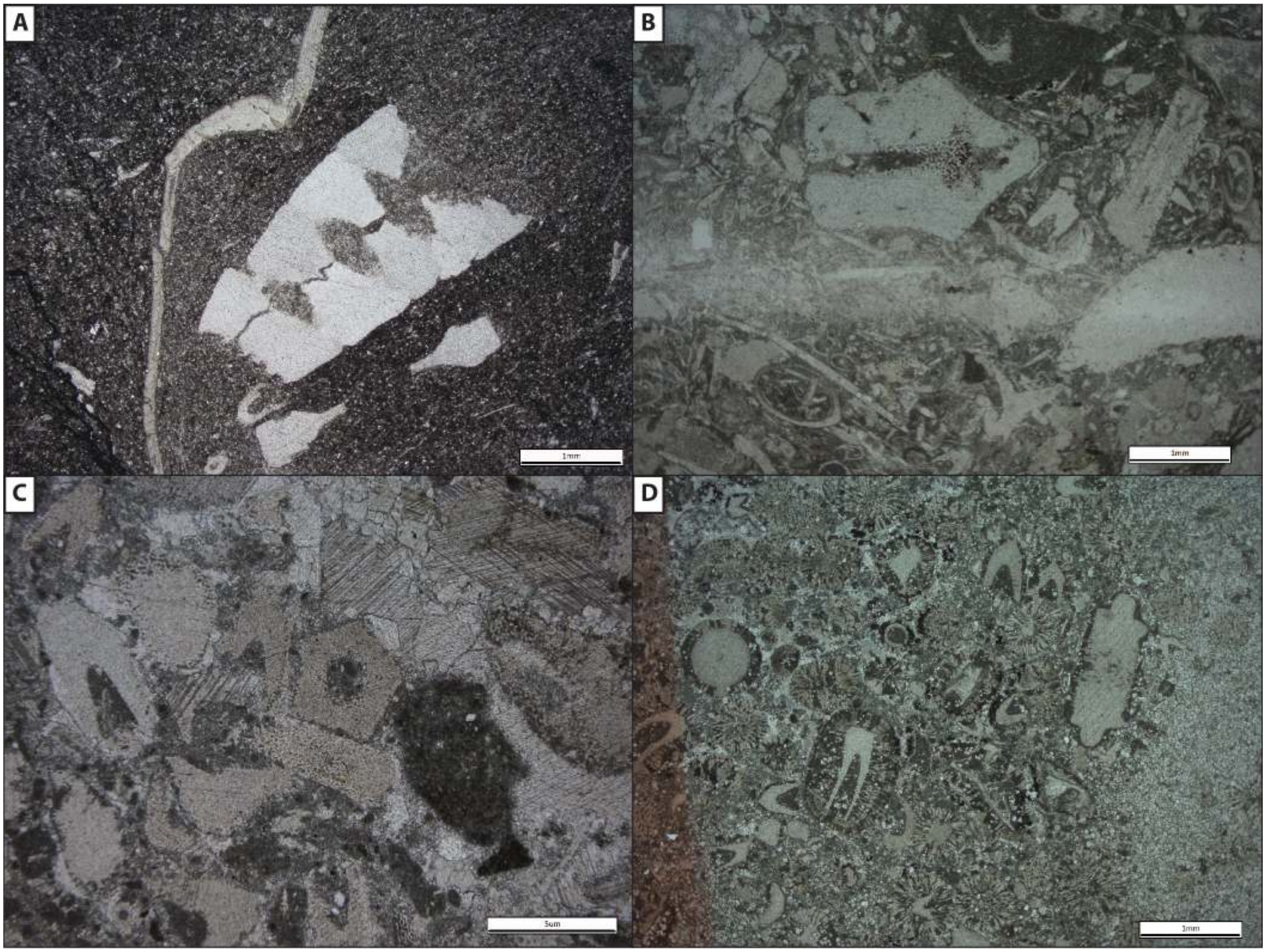


Figure 9.7.15: A) Slightly oblique longitudinal cut through several stacked, articulated crinoid columnals. B) Arm plate of a crinoid. C) Numerous crinoid plates and a columnal with five-fold symmetry in the central canal. D) A variety of crinoid skeletal elements, many of which form the cores of ooids (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

9.7.8: Forminifera

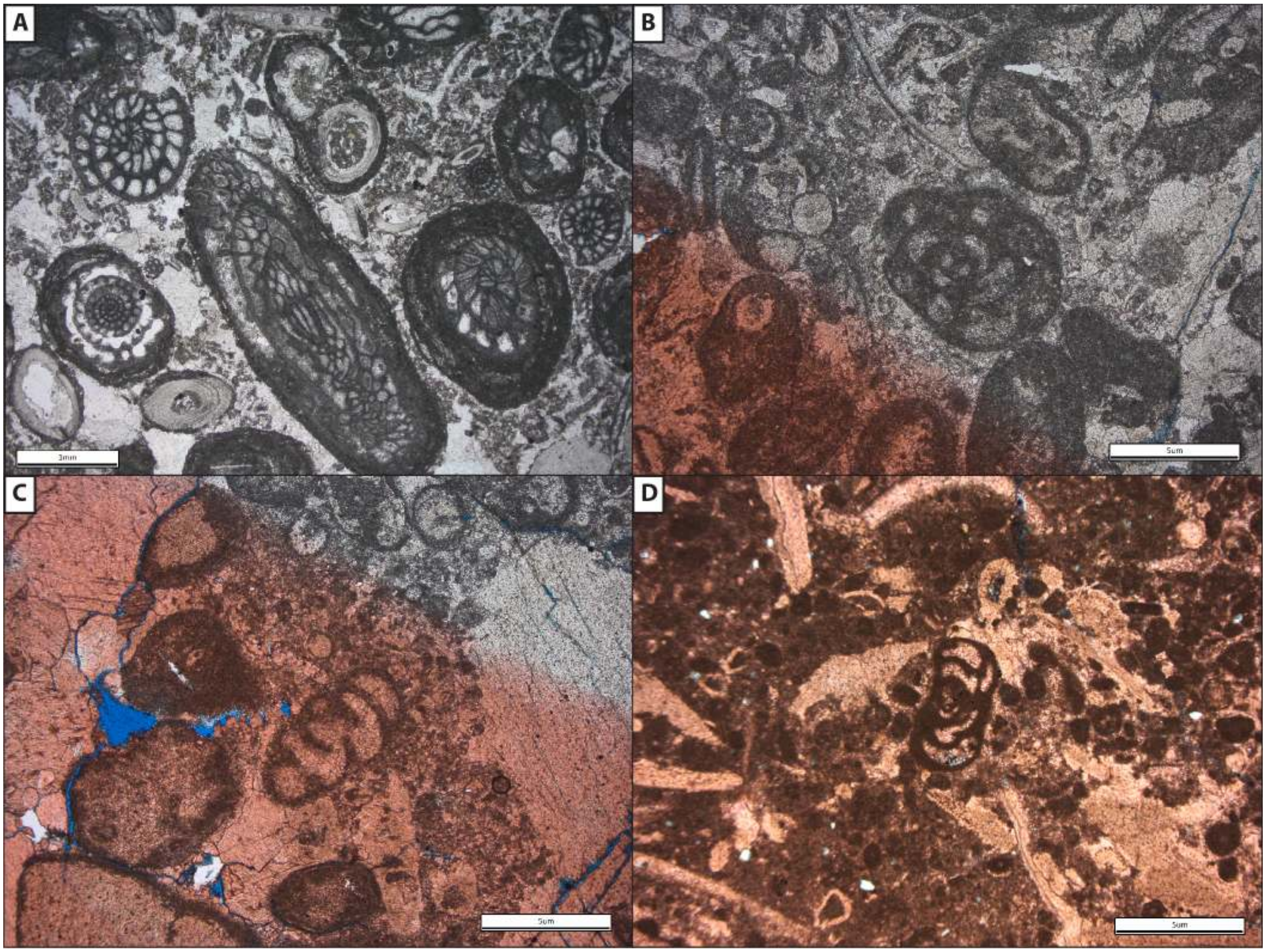


Figure 9.7.15: Foraminifera are single-celled organisms, many of which build a calcareous test with internal chambers (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

9.7.9: Miscellanea and Algae

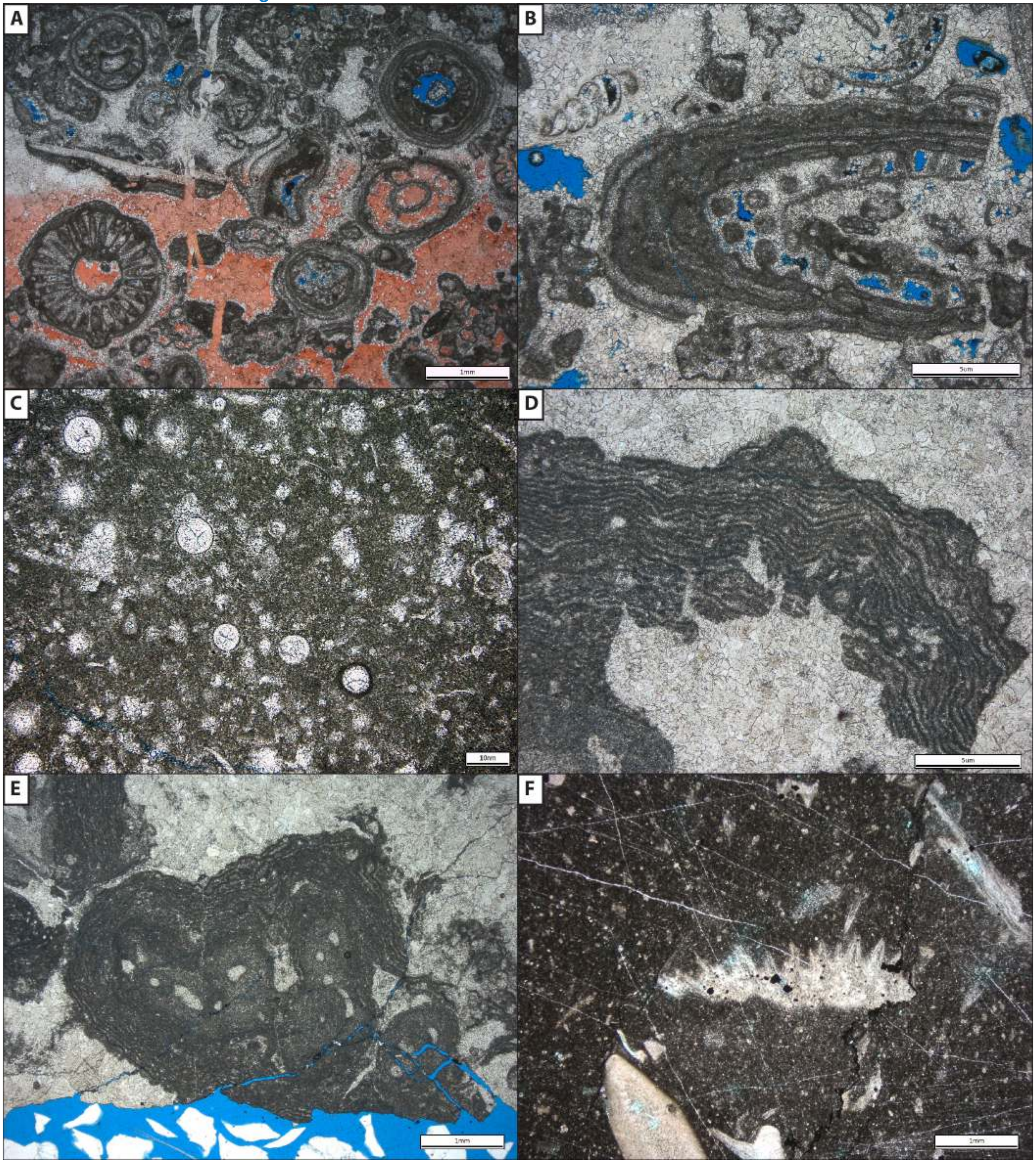


Figure 9.7.15: A selection of other distinctive fossils include A and B) the Permian marine green algae *Mizzia* sp., C) Mississippian calcispheres which are likely of algal origin, D) Permian *Archaeolithoporella* - an encrusting organism of uncertain affinity, E) *Tubiphytes*, an encrusting organism of uncertain affinity, and F) conodonts are tooth-like structures (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

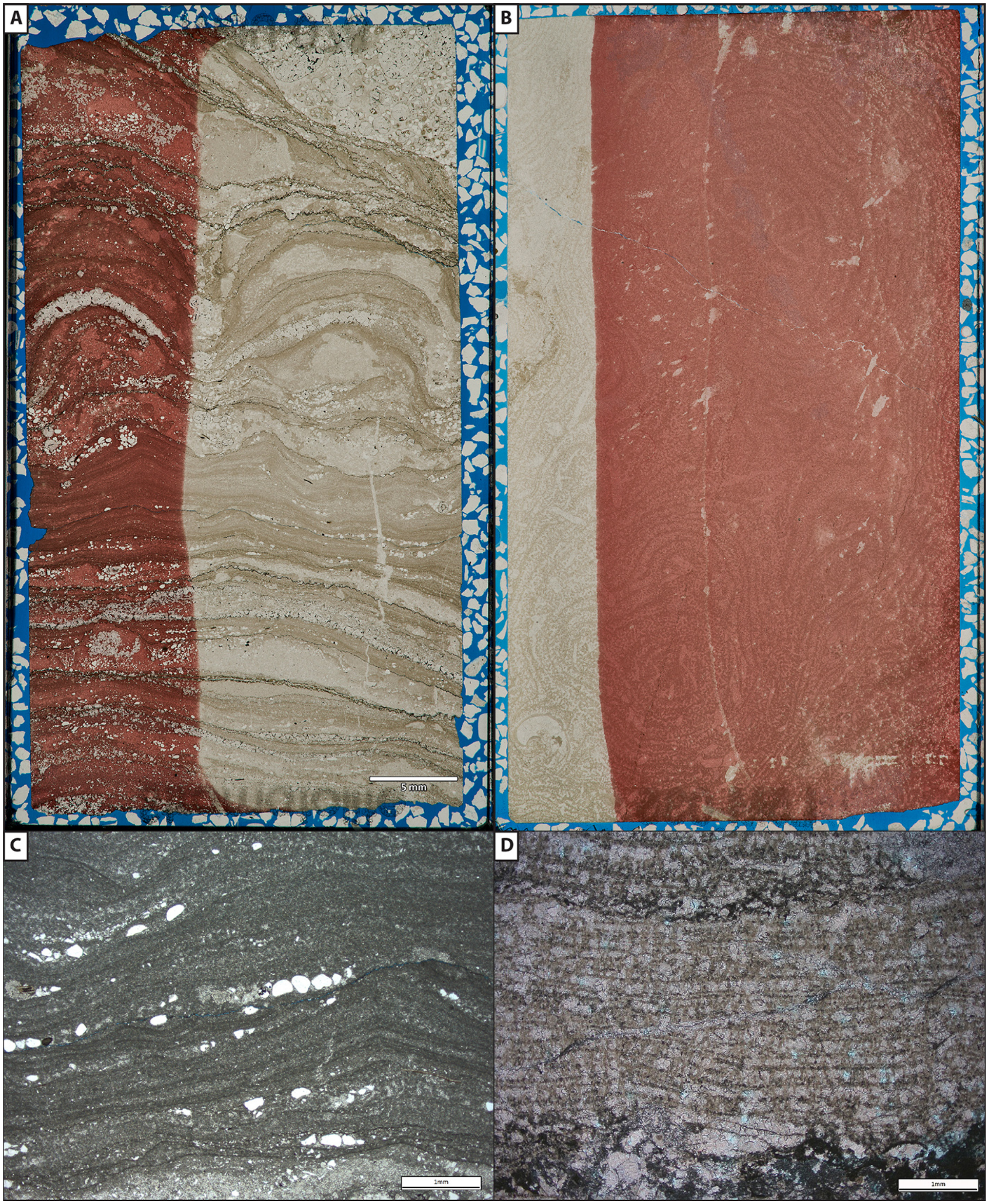


Figure 9.7.16: A and C) Stromatolites from as photosynthetic microbes grow and trap sediment; although local conditions may result in distinctive overall morphologies, they lack an internal structure aside from the laminae. B and D) Stromatoporoids might be related to sponges; although they superficially appear similar to stromatolites, well-preserved specimens show a more complex internal architecture with a lattice-like network of horizontal laminae and vertical pillars (all images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0).

9.7.10: Readings and Resources

- Microfacies of Carbonate Rocks by Erik Flugel
- A Color Guide to the Petrography of Carbonate Rocks by Peter Scholle and Dana Ulmer-Scholle
- Carbonate World website

CHAPTER OVERVIEW

10: Depositional Environments

In this chapter we provide a survey of the most common clastic and carbonate environments and talk about their geometry, occurrence, typical lithologies and structures, and provide numerous pictures of each. Given the scale of these features, the kind of facies analysis needed to determine depositional environments is best learned in the field or by looking at drill core.

Learning Objectives

- Describe the most common lithologies, sedimentary structures, and fossils in major clastic and carbonate depositional environments.
- Explain what processes are important in these environments.

Chapter thumbnail shows a cast of a standing lycopsid in floodplain deposits of the Pennsylvanian Joggins Formation, Nova Scotia (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

Topic hierarchy

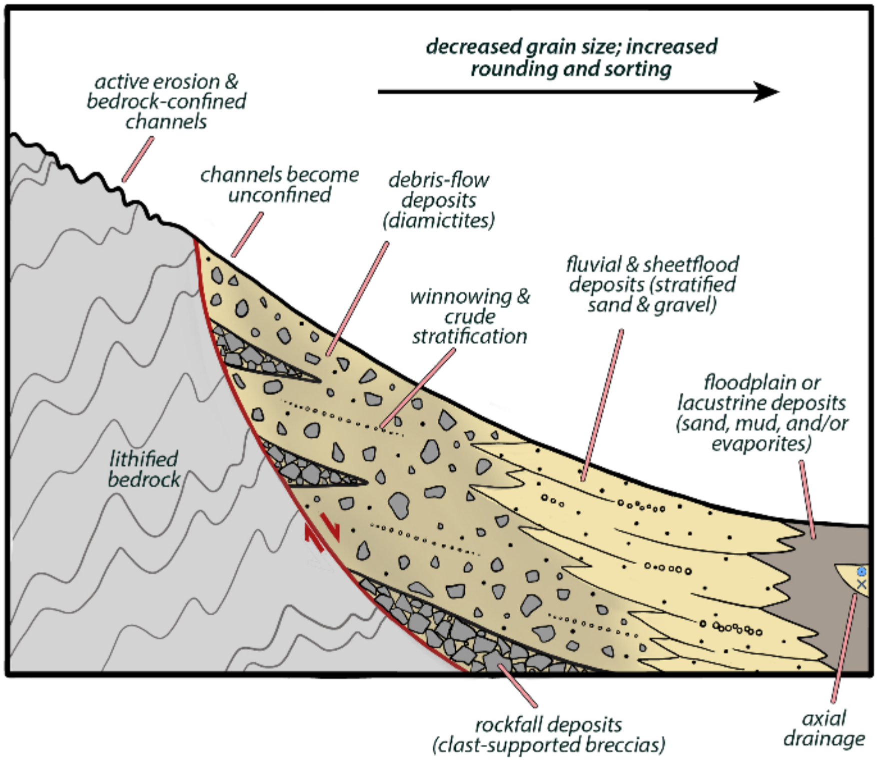
10.1: Alluvial Systems

10.1.1: Alluvial Fans

Alluvial fans are mounds of coarse grained sediments formed when a confined stream disgorges into an unconfined area. They typically occur along the margins of mountain ranges where bedrock incised channels draining uplands spill out on to broad open valley floors. Alluvial fans occur in areas with significant topographic relief caused by rapid subsidence or uplift (rift basins, foreland basins, fold-and-thrust belts, etc.). They are semi-circular in map view, form mounds in transverse cross section, and form basinward-thinning wedges in lateral cross sections.

Figure 10.1.1: Oblique image of an alluvial fan building into the Badwater Basin in Death Valley, National Park shown with 1.5x vertical exaggeration. Notice that the fan originates at the point where the flow emerges from the bedrock-confined channel. The pronounced line around the edge of the alluvial fan is Badwater Road (Courtesy Google Earth, image is ©2024 Maxar Technologies; exported images from Google Earth can be embedded on websites for educational and non-commercial use).

Internally, alluvial fans typically include debris flow deposits composed of poorly-sorted, possibly matrix-supported conglomerates, breccias, and/or diamictites and stream flow deposits composed of sorted and stratified sand and gravel. Grain size generally increases toward the mountain front and decreases in a basinward direction. Sediments commonly occur in coarsening upward cycles that record tectonic and sedimentary pulses. (Paleo)flow is generally in a radial pattern away from the source.



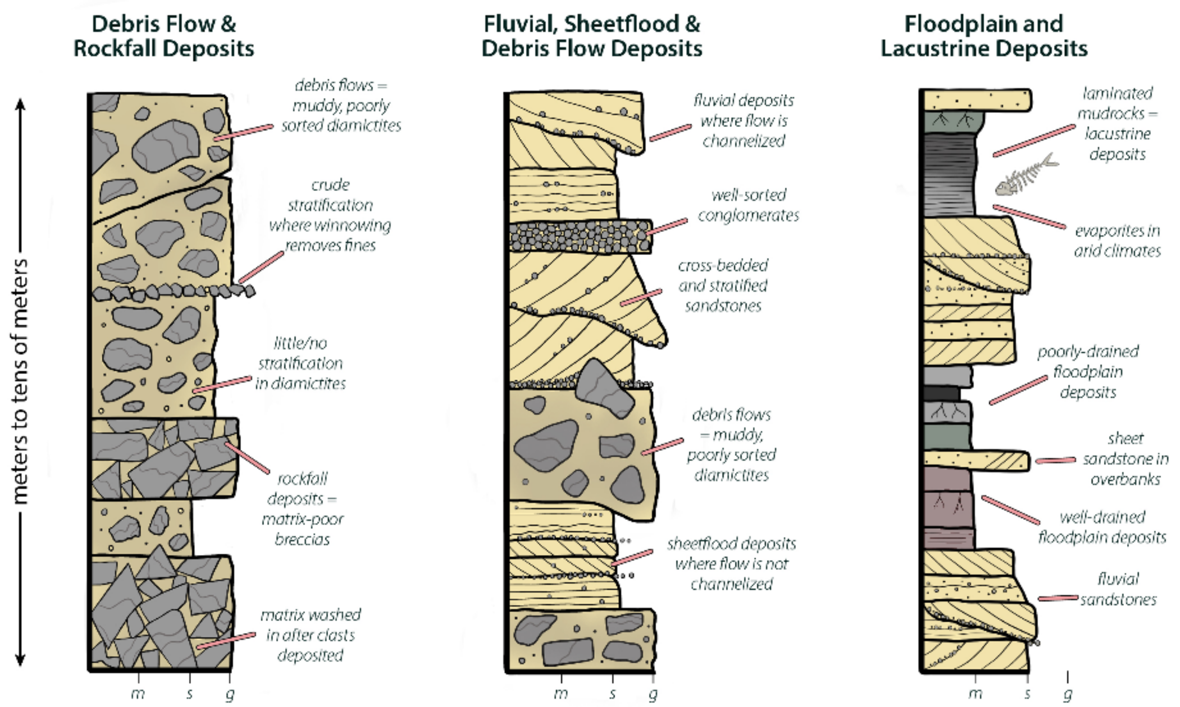


Figure 10.1.2: Cross-sectional sketch of an alluvial fan and schematic measured sections through alluvial fan deposits (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

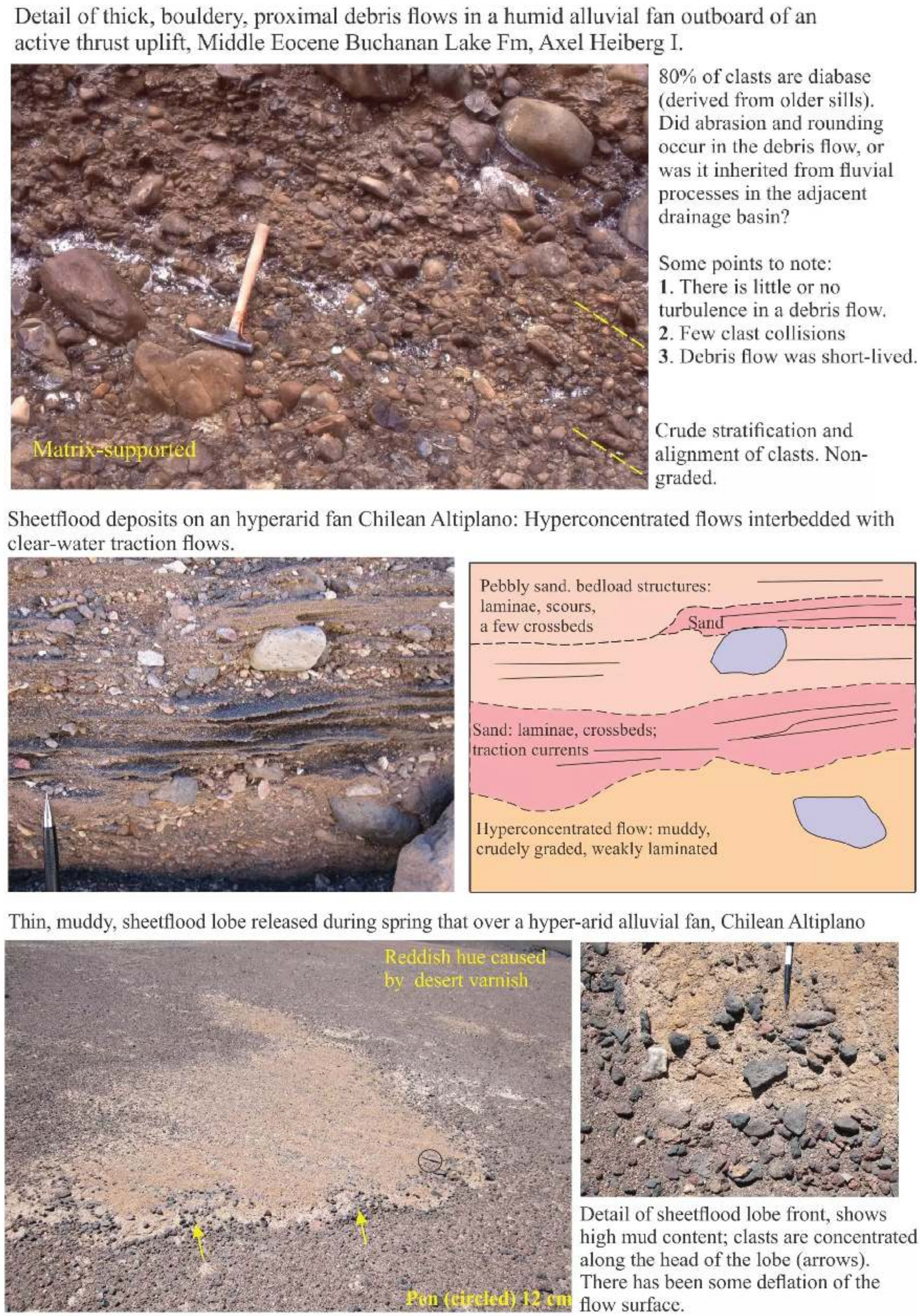


Figure 10.1.3: A variety of debris flow, stream flow, and sheet flood deposits in alluvial fans (Brian Ricketts via https://www.geological-digressions.com/sedimentary-structures-alluvial-fans; CC BY-SA 4.0).

10.1.2: **Fluvial Deposits**

Rivers are areas where overland flow is concentrated in a channel. Floodplains are the areas adjacent to channels that are periodically inundated during floods. Although we commonly think of river channels as u-shaped features, they exhibit a wide range of geomorphic variability and are commonly preserved as complex, amalgamated bodies in the geologic record. The orientation of the outcrop relative to paleoflow direction can also influence one's perception of channel body geometry. A simple channel that appeared u-shaped when viewed perpendicular to paleoflow direction could appear like an elongate sand ribbon if viewed parallel to paleoflow. Modern rivers can be characterized using a variety of criteria including sinuosity, number of active channels, types of bars, and numerous other characteristics. For our purposes, we will take an admittedly oversimplified approach and focus on three points in the continuum of fluvial style: braided, meandering, and anastomosed (aka fixed-multichannel).

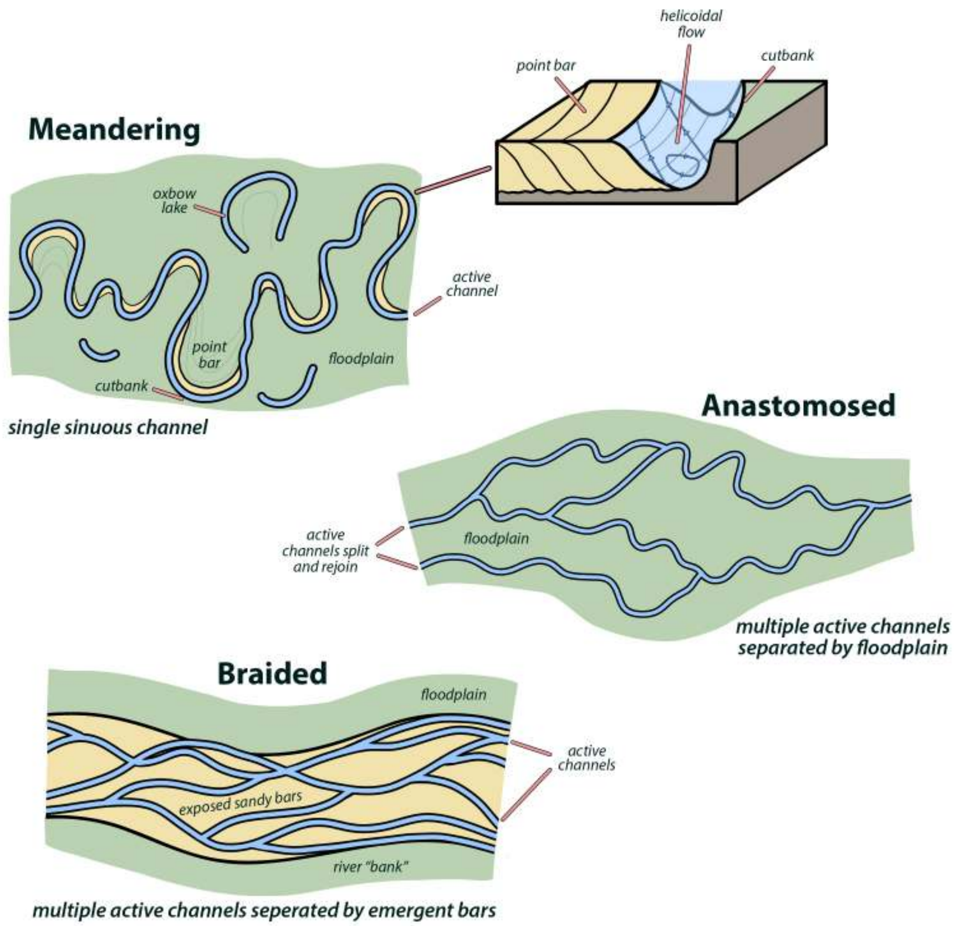


Figure 10.1.4: Points in the continuum of fluvial form. Green areas represent floodplains, yellow areas are emergent sandy bars, and blue areas are active channels. Diagram from Page Quinton via Wikimedia Commons; CC BY-SA 4.0 after Miall (1977) and Leeder (2009).

10.1.2.1: Braided Rivers

Braided river systems consist of a broad "river" with an interwoven network of active channels separated by emergent bars. They are typically bedload-choked sand- and/or gravel-dominated systems with bars that typically migrate in a downstream direction. Given the coarse nature of the sediment and the overabundance of bedload, most sediment transport happens during flood stage when the entire river is bankfull and within-channel bars are submerged.

Overall, abundant bedload makes for unstable banks and sand- or gravel deposits with only modest amounts of mud and numerous erosion surfaces. Active channel deposits are much more abundant than floodplain deposits. The Platte River of Nebraska is one of the best known modern braided rivers; well exposed and documented systems include the Castlegate Sandstone (Cretaceous) and the South Bar Formation (Pennsylvanian).

Figure 10.1.5: Google Earth images showing the same segment of the braided Platte River under low/normal flow conditions in August 2006 (above) and at bankfull stage in June 2001. In these bedload-choked systems, most sediment transport happens during flood and bankfull flows (courtesy Google Earth, exported images from Google Earth can be embedded on websites for educational and non-commercial use).

Figure 10.1.6: The Cretaceous Castlegate Sandstone is a well known and well studied sandy braided fluvial deposit (Loco Steve via Flickr; CC BY 2.0). Notice the abundance of sandstone, the numerous internal erosion surfaces, and the paucity of mudrock.

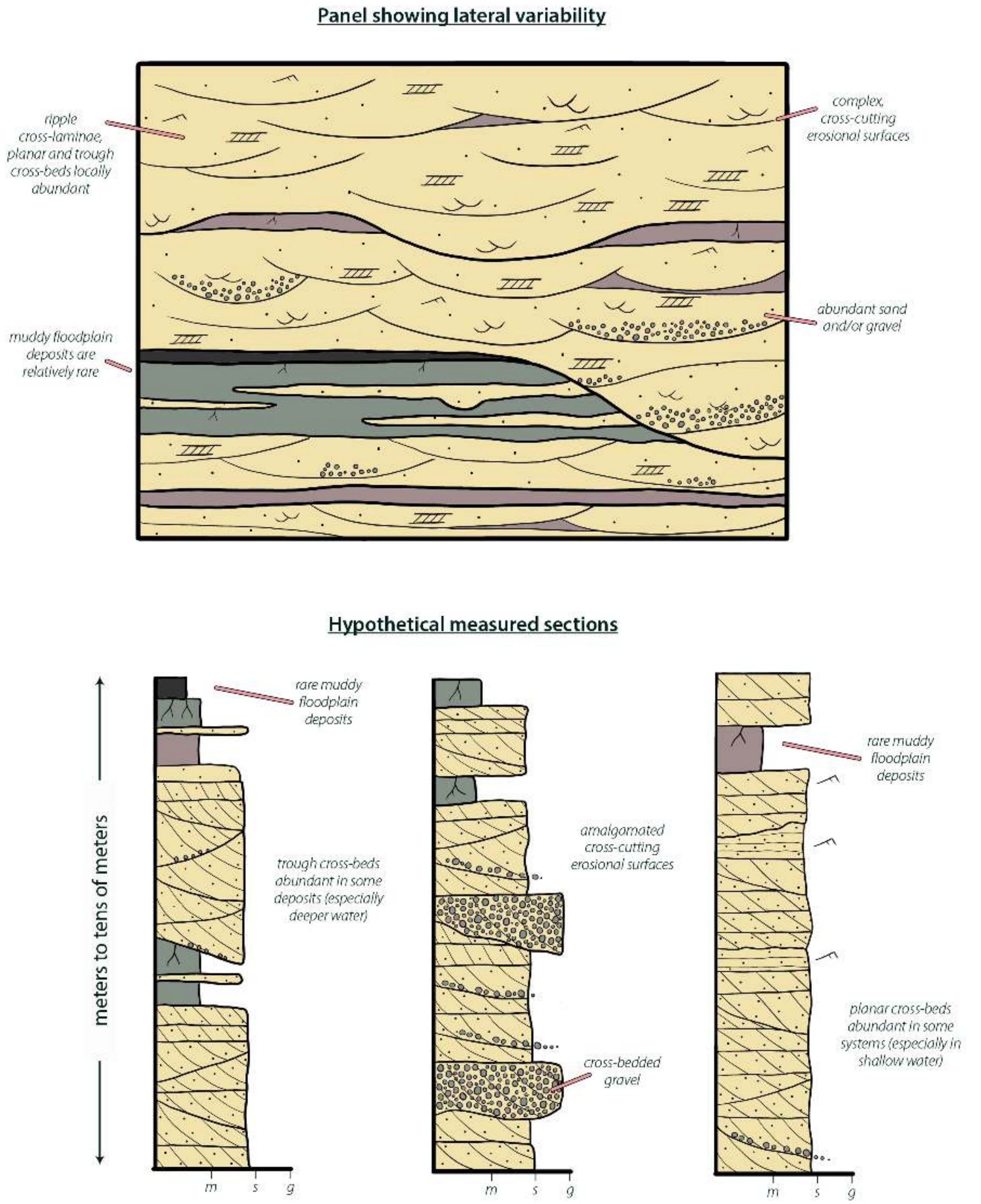


Figure 10.1.7: Overview of the sedimentology of braided river deposits with particular emphasis on the features that are most likely to be preserved in the geologic record (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

10.1.2.2: Meandering Rivers

Meandering rivers have a sinuous main channel with a broad, often muddy, floodplain. They typically occur in low gradient areas, carry a mixture of bedload and suspended load, and have more energy than they need to carry the sediment load. They expend the excess energy winding across the floodplain and distributing sediment across it. Internally, the water in channels actually moves downstream in a corkscrew fashion (helicoidal flow) which facilitates erosion on the cutbank and deposition on the opposing pointbar. Sinuosity increases with continued cutbank erosion and pointbar deposition, which gives meandering systems their distinctive sinuous patter. Eventually curvature increases to the point where a reach of the river gets cut off and isolated forming an oxbow lake. The oxbow lake is filled with standing water and eventually fills in with mud and/or organic material.

Overall, meandering river systems generally consist of sandy or heterolithic channel deposits encased within finer floodplain deposits. In a vertical section, a fining-upward trend may be formed by the progression from the (relatively) coarse basal channel deposits, through sandy and heterolithic point bar deposits (inclined heterolithic strata), and eventually into finer grained floodplain deposits. The Mississippi River is, by far, the best know modern meandering river; noteworthy ancient examples include portions of the Devonian Catskill Formation of PA and NY and the Cretaceous McMurray Formation of Alberta.

Figure 10.1.8: Google Earth image showing a meandering portion of the Mississippi River. The light blue lines represent the boundary between the states of Mississippi (north) and Louisiana (south); the border was originally drawn down the middle of the active river channel - that course has now been abandoned in numerous places (courtesy Google Earth, image is from Landsat/Copernicus; exported images from Google Earth can be embedded on websites for educational and non-commercial use).

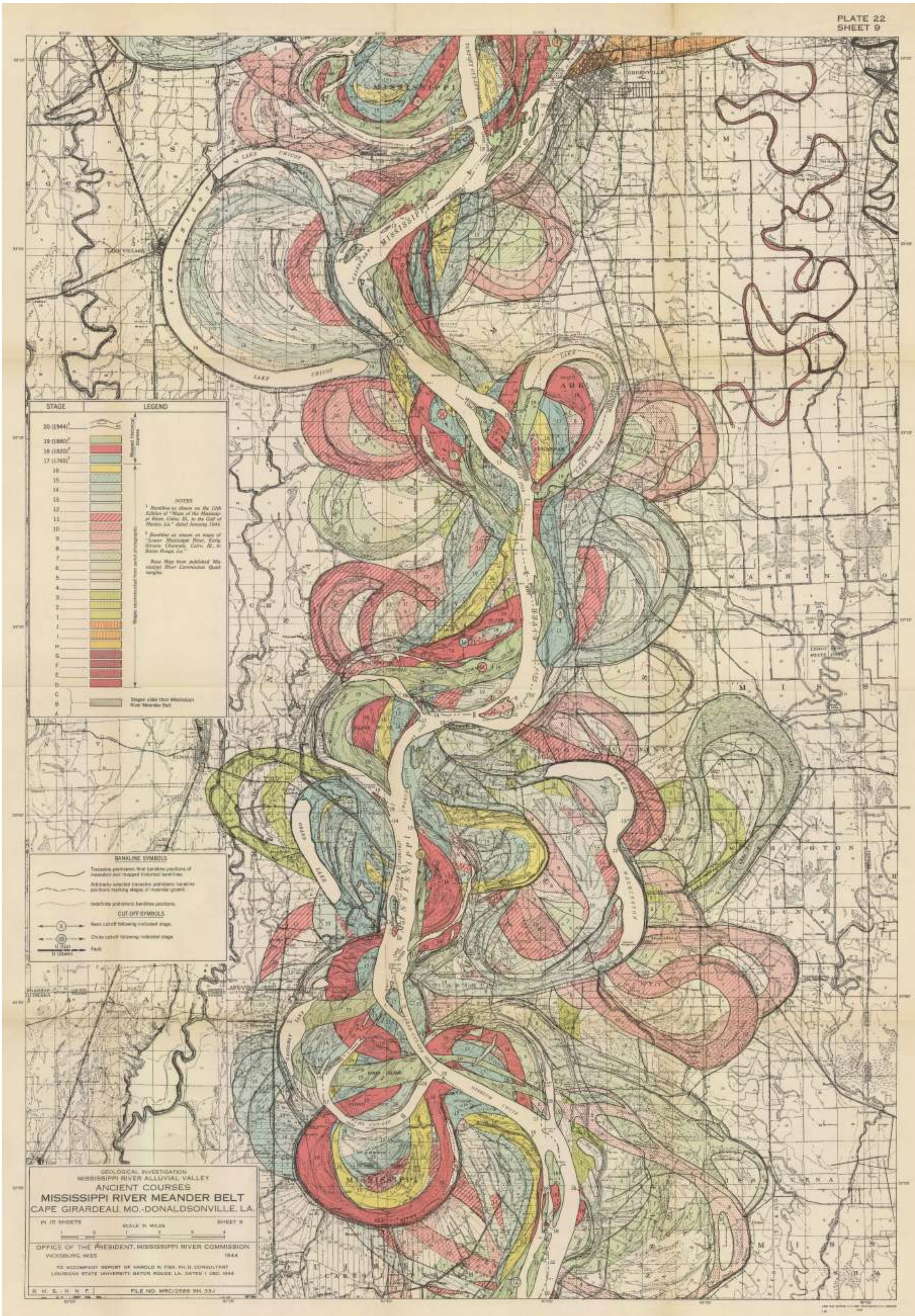


Figure 10.1.9: One of Harold Fisk's (1944) maps showing former courses of the Mississippi River; the different colors represent the path of the river at different points in time since the last glacial maximum (U.S. Army Corps of Engineers via Radical Cartography; public domain).

Figure 10.1.10: Google Earth image showing anchored ships aligned by helicoidal flow. Flow is from the top of the image toward the bottom. Ships are anchored on their upstream side and aligned by helicoidal flow so that the long axis of the ship points away from the cut bank and obliquely up the point bar (courtesy Google Earth; exported images from Google Earth can be embedded on websites for educational and non-commercial use).

Figure 10.1.11: Inclined heterolithic strata making up point bar deposits in a meandering channel body, Devonian Catskill Formation near North Bend, PA (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0).

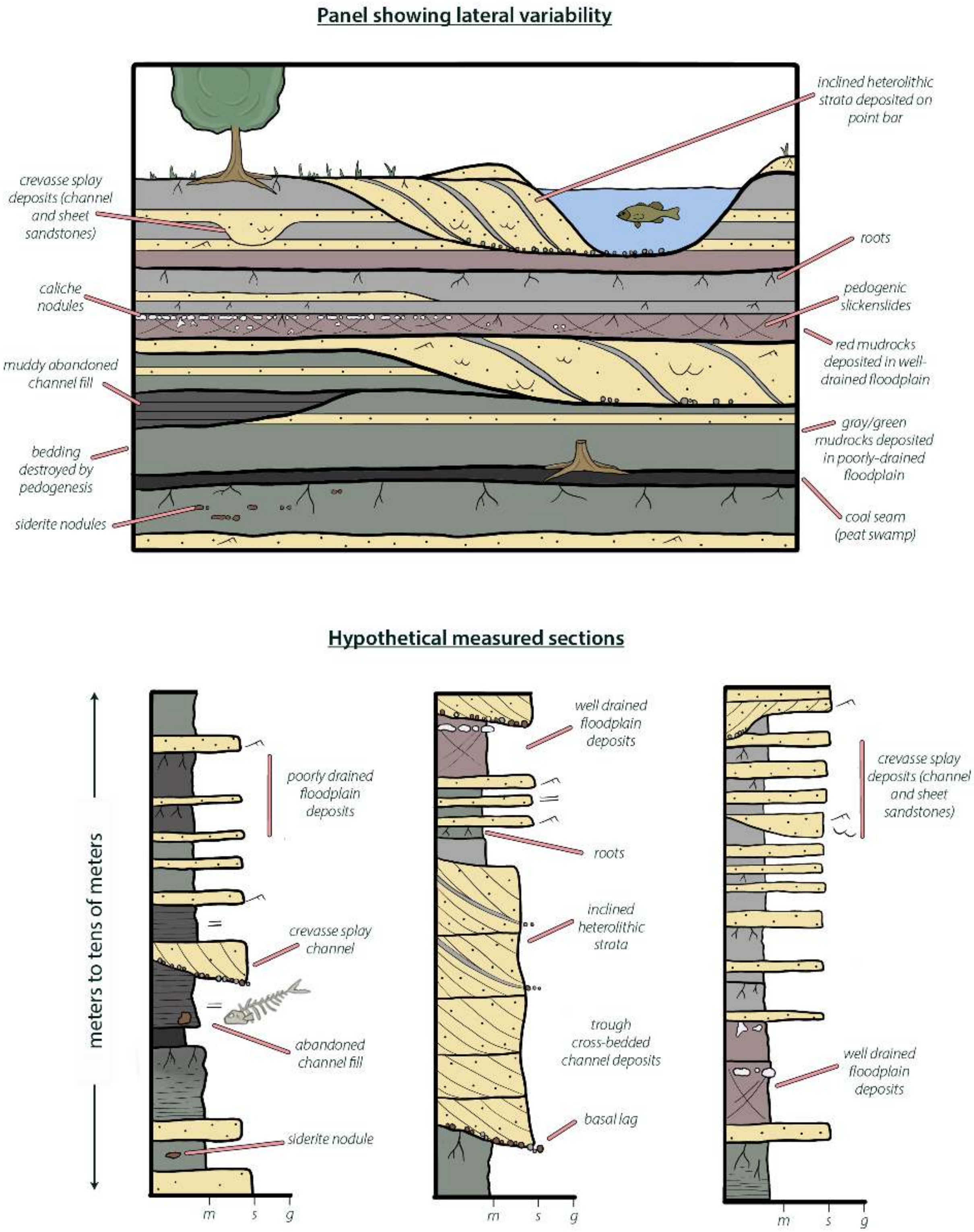


Figure 10.1.12: Overview of the sedimentology of meandering river deposits with particular emphasis on the features that are most likely to be preserved in the geologic record (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

10.1.2.3: Anastomosed Rivers

Anastomosed river systems have channels that are largely fixed in place without much lateral movement. They are commonly part of a larger drainage network with multiple interconnected channels separated by floodplains. In the geologic record it is difficult to prove that multiple channels were active simultaneously (especially given outcrops of limited lateral extent) so the best one can hope for is to find stable channels - ideally at the same level - isolated within floodplain deposits.

Figure 10.1.13: Google Earth image showing an anastomosed portion of the Columbia River near Spillimacheen, British Columbia (courtesy Google Earth, image is ©Province of British Columbia; exported images from Google Earth can be embedded on websites for educational and non-commercial use).

Figure 10.1.14: The u-shaped, sandstone-dominated channel bodies of the Pennsylvanian Springhill Mines Formation are interpreted as the deposits of an ancient anastomosed river system that delivered sediment into a rapidly subsiding basin (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

10.1.3: Floodplain Deposits

Floodplain deposits consist of muddy sediment formed from suspension deposition following flooding events and sandy deposits in areas that were closer to higher velocity flows that were capable of moving bedload. Floodplain deposits may show evidence of subaerial exposure (desiccation cracks, raindrop imprints), evidence of plant growth or other terrestrial fossils, and/or evidence of pedogenesis including the destruction of bedding or the development of soil horizons.

Levees are areas of positive topographic relief on the floodplain that form adjacent to channels. They form when channelized flows overtop the river banks and rapidly deposit sediment as the flows become unconfined. Crevasse splays are lobe-shaped sand bodies that form where a channel breaches the levee and deposits a lobe of relatively coarse grained sediment.

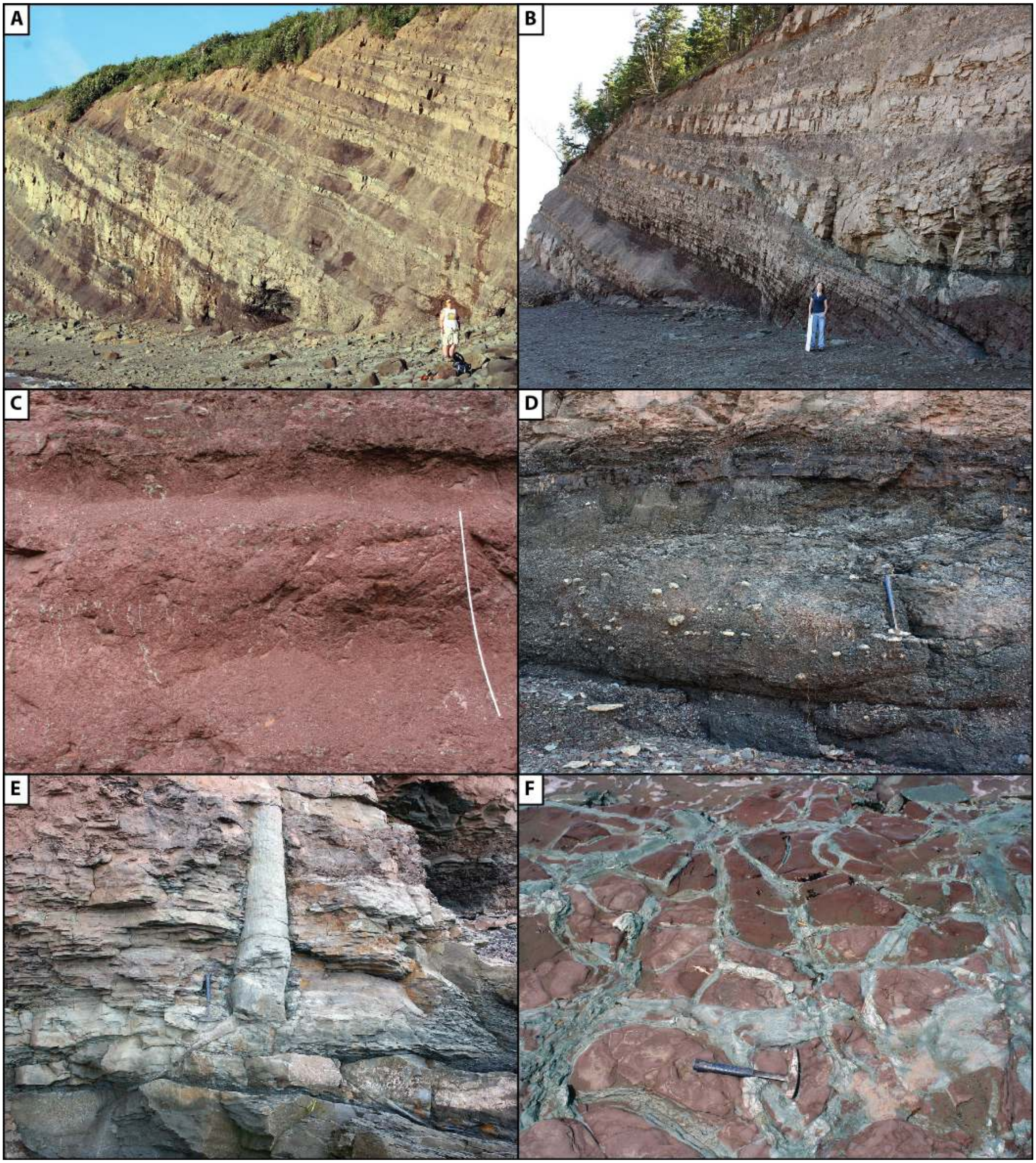


Figure 10.1.15: Floodplain deposits. A) Interbedded sheet sandstones and mudrocks are common floodplain deposits. The sheet sandstones represent energetic flow when riverbanks are overtopped and the mudrocks represent suspension deposition from standing water. B) Some sheet sandstones thicken toward and can be traced back to the channels that sourced the sediment; stacked packages of sheet sandstones can represent progradation of crevasse splay deposits. C) Although floodplain mudrocks start off as laminated or platy mudrocks, pedogenic processes can destroy bedding and red colors represent oxidizing conditions. D) Floodplain deposits can also record waterlogged conditions as evidenced by gray and green mudrocks and organic-rich horizons, and even thin coals. E & F) Fossils and sedimentary structures can provide evidence of terrestrial conditions and subaerial exposure (all images from Michael C. Rygel via Wikimedia Commons; various CC BY-SA licenses).

10.1.4: Avulsion and Aggradation

Avulsion is the process by which an active channel rapidly shifts to a different location on a topographically lower part of the floodplain. It happens when primary flow breaches a levee and diverts through a crevasse channel, usually during floods. If the new route provides a gradient advantage the active channel may shift to the new course.

Aggradation is the increase in land elevation due to sediment deposition. River channels and levees commonly aggrade together and become elevated above the floodplain. This process eventually raises the downstream elevation and decreases the gradient. If/when the levee is breached during a flood, the active channel will preferentially shift to a the new (steeper gradient) course.

Together, the avulsion and aggradation can profoundly influence the preservation of fluvial deposits. Low avulsion rates or high subsidence rates tend to produce isolated channel bodies; high avulsion rates or low subsidence rates tend to produce highly amalgamated channel bodies. These relationships get much more complex when the processes interact.

Figure 10.1.16: Google Earth image showing an active channel of the Columbia River, swampy floodplain areas, tree-covered levees, and a lobe-shaped sandy crevasse-splay deposit formed during a flood event. If there is a gradient advantage, the main channel could divert through the breach in the levee and avulse to a new course (courtesy Google Earth, image is ©Province of British Columbia; exported images from Google Earth can be embedded on websites for educational and non-commercial use).

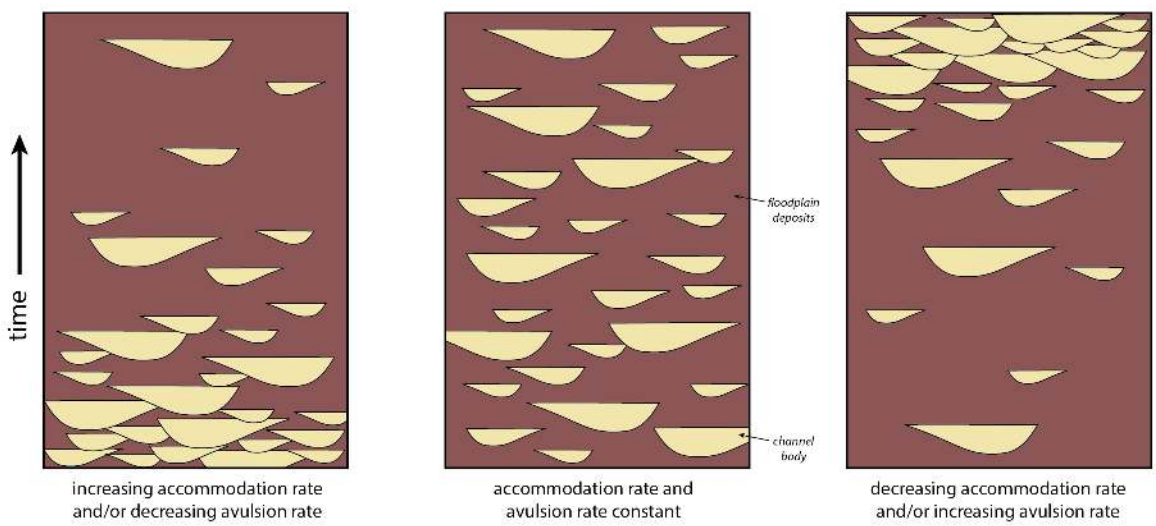


Figure 10.1.17: Diagrams showing the relationship between channel body amalgamation, avulsion rate, and sedimentation rate. Diagram from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0 which is after Bryant et al. (1995).

10.1.5: Overview

Fluvial deposits are generally preserved as sand or gravel channel bodies that have erosional bases and may be amalgamated. Channel bodies may be encased in finer-grained floodplain deposits. Both river and floodplain deposits can contain evidence of fresh-water and/or terrestrial fossils. Floodplain deposits may show evidence of soil formation and subaerial exposure.

10.1.6: Readings and Resources

- Bryant, M., Falk, P., & Paola, C., 1995, Experimental study of avulsion frequency and rate of deposition. Geology, 23(4), 365-368.
- Harold Fisk's 1944 maps of the Mississippi River: http://www.radicalcartography.net/index.html?fisk
- Leeder, M. R., 2009, Sedimentology and sedimentary basins: from turbulence to tectonics. John Wiley & Sons.
- Miall, A. D., 1977, A review of the braided-river depositional environment. Earth-Science Reviews, 13(1), 1-62.
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- Miall, A. D. (2010). Alluvial deposits. In N. P. James & R. W. Dalrymple (Eds.), *Facies Models 4* (pp. 105–137). Geological Association of Canada.

10.2: Deserts

10.2.1: Definition and Distribution

By definition, deserts are dry areas that receive less than 10 inches of rainfall per year. Although these conditions can occur at just about any geographic location, areas that are ~30° from the equator tend to be at least semiarid because this is where we have convection cells bringing dry air down from altitude. Deserts are also particularly likely to form in "rainshadow" areas in the downwind side of mountain ranges.

Figure 10.2.1: Photograph of Death Valley National Park showing many of the striking landforms present in desert areas (Brocken Inaglory via Wikimedia Commons; CC BY-SA 4.0). The geomorphology of these areas is beyond the scope of this chapter; instead we will focus on the features of desert landscapes that are most commonly preserved in the geologic record.

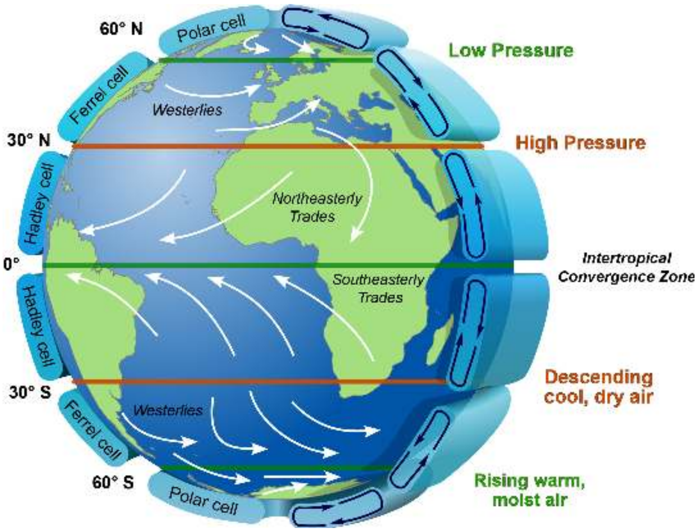

Figure 10.2.2: Atmospheric circulation patterns showing how circulation cells can influence precipitation patterns at different latitudes (modified from Kaidor via Wikimedia Commons; CC BY-SA 3.0).

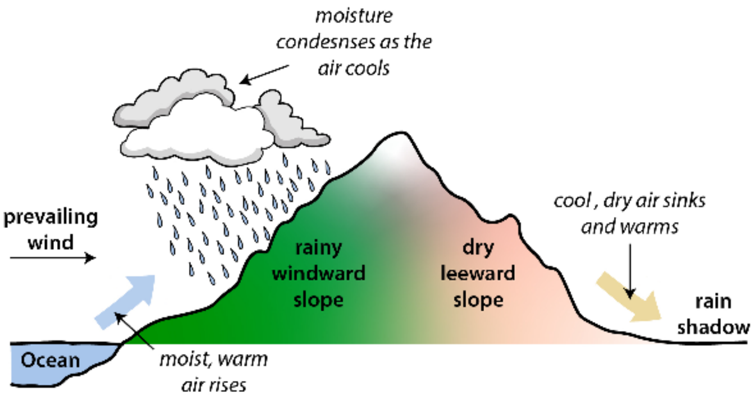

Figure 10.2.3: Illustration of the rainshadow effect on the downwind side of mountain ranges.

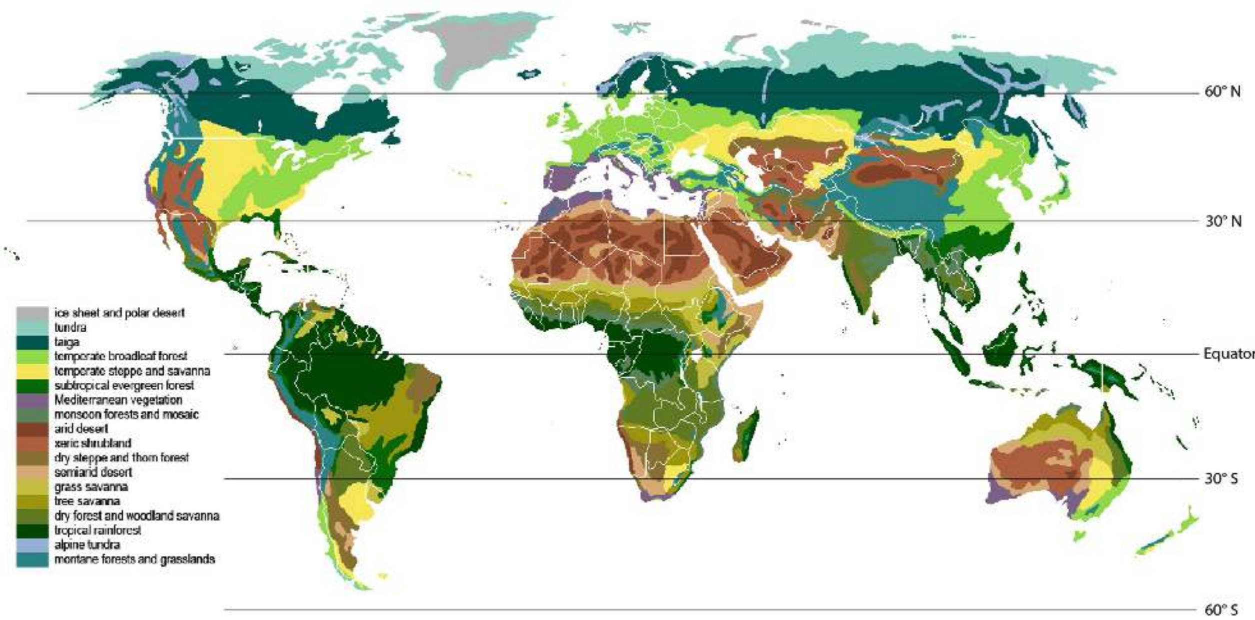

Figure 10.2.4: Map showing the distribution of vegetation types; deserts are represented as browns and tans (modified from Ville Koistinen (user Vzb83) via Wikimedia Commons; CC BY-SA 3.0). Notice that deserts are particularly abundant and on the downwind side of mountain ranges.

10.2.2: Processes and Deposits

In the geologic record, desert deposits can cover thousands of square kilometers and be many hundreds of meters thick. The Jurassic Navajo Sandstone and correlative units reach thicknesses of up to 700 m and widespread occurrences across Wyoming, Utah, Colorado, and Arizona represents under half of the original >340,000 mi^2 extent of the Navajo Sand Sea. And even this vast ancient desert is dwarfed by the modern Sahara Desert which covers >3.6 million mi^2 in northern Africa.

Figure 10.2.5: Exposures of hundreds of meters of the Navajo Sandstone (Jurassic) in the walls of Pine Creek and Zion Canyons, Zion National Park (Steve Boland via Flickr; CC BY-NC-ND 2.0).

In terms of the origin of sediment within deserts, it may be derived from weathering of older (underlying) rock and sediment, fluvial transport or mass wasting from adjacent uplands, or moved by wind from other areas within the desert. The lack of water in these arid areas limits plant growth, which means that under normal conditions sediment transport is dominated by wind. Silt- and clay-sized particles can be suspended within the air and larger sand-sized particles can be temporarily suspended as well as moved by traction and saltation. Windblown sand can be organized into large sand dunes with variable morphologies; they can be up to tens of meters tall and are often preserved as very large, wedge-shaped sets of cross beds. Low-amplitude wind-ripples migrate across dune faces. When sand accumulations become oversteepened or disturbed, grain flow can happen down dune surfaces which causes inverse grading and cross-laminae that thicken downdip.

Figure 10.2.6: Photograph of a sandstorm about to overtake a military base in Iraq (Tobin via Flickr; CC BY-NC-SA 2.0). These impressive storms form when winds grow strong enough to suspend and/or saltate silt- and clay-sized particles exposed at the surface.

Figure 10.2.7: Collection of videos showing common sediment transport processes in deserts. Clockwise from the top left they include suspension of silt and sand in a storm, saltation of sand grains, grainflow down the slipface of a dune, and movement of sediment into and across a desert by flash floods.

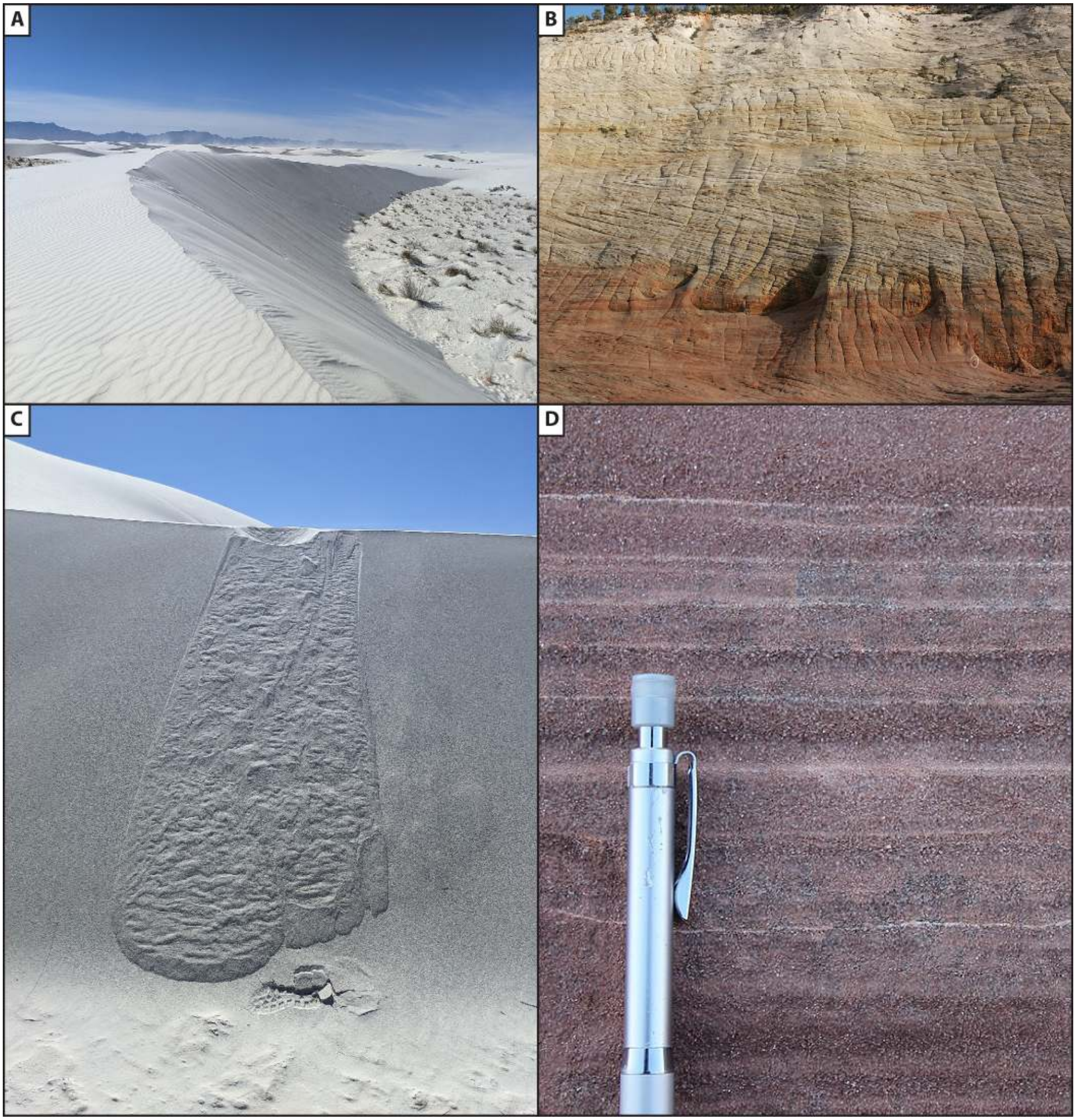

Figure 10.2.7: Modern sedimentary structures in deserts (left) and corresponding sedimentary structure preserved in the rock record (right). A) Large sand dunes in White Sands National Park ornamented with low-amplitude wind ripples oblique to the crest of the dune (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0). Note that the dune is migrating into an interdune area with relatively abundant vegetation. B) Large eolian cross-beds in the Jurassic Najavo Sandstone in Zion National Park (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0). Circled area shows a person for scale. Because the ultimate limiting factor on the height of dunes is the depth of the fluid, windblown dunes in deserts can grow to impressive heights. C) Grainflow in cohesionless sand on the slipface of a dune; footprint for scale (Michael C. Rygel via Wikimedia Commons; CC BY-SA). These features would be preserved as wedge shaped layers of sediment that thickened in a downslope direction; they might contain inverse grading. D) Inverse grading in eolian sandstone of the Jurassic Carmel Formation (Michael C. Rygel via Wikimedia Commons; CC BY-SA).

Interdune areas can record a variety of processes and are more sedimentologically complex. Common features in interdune areas include:

- Deflation lags (or desert pavements) - accumulations of gravel that form when finer sediment is winnowed away leaving a gravel-armored surface behind.
- Ventifacts - faceted clasts that are created by wind erosion
- Playa lakes - temporary bodies of water that can leave behind evaporites, mud, and evidence of desiccation.
- Interdune vegetation - plants adapted to arid conditions can take root in the more stable substrate of interdune areas
- Fluvial systems - can be locally or seasonally present, especially where rivers are sourced from more distant or upland areas. Fluvial transport of sediment into deserts provides input for sediment that will eventually be reworded and transported by the wind.

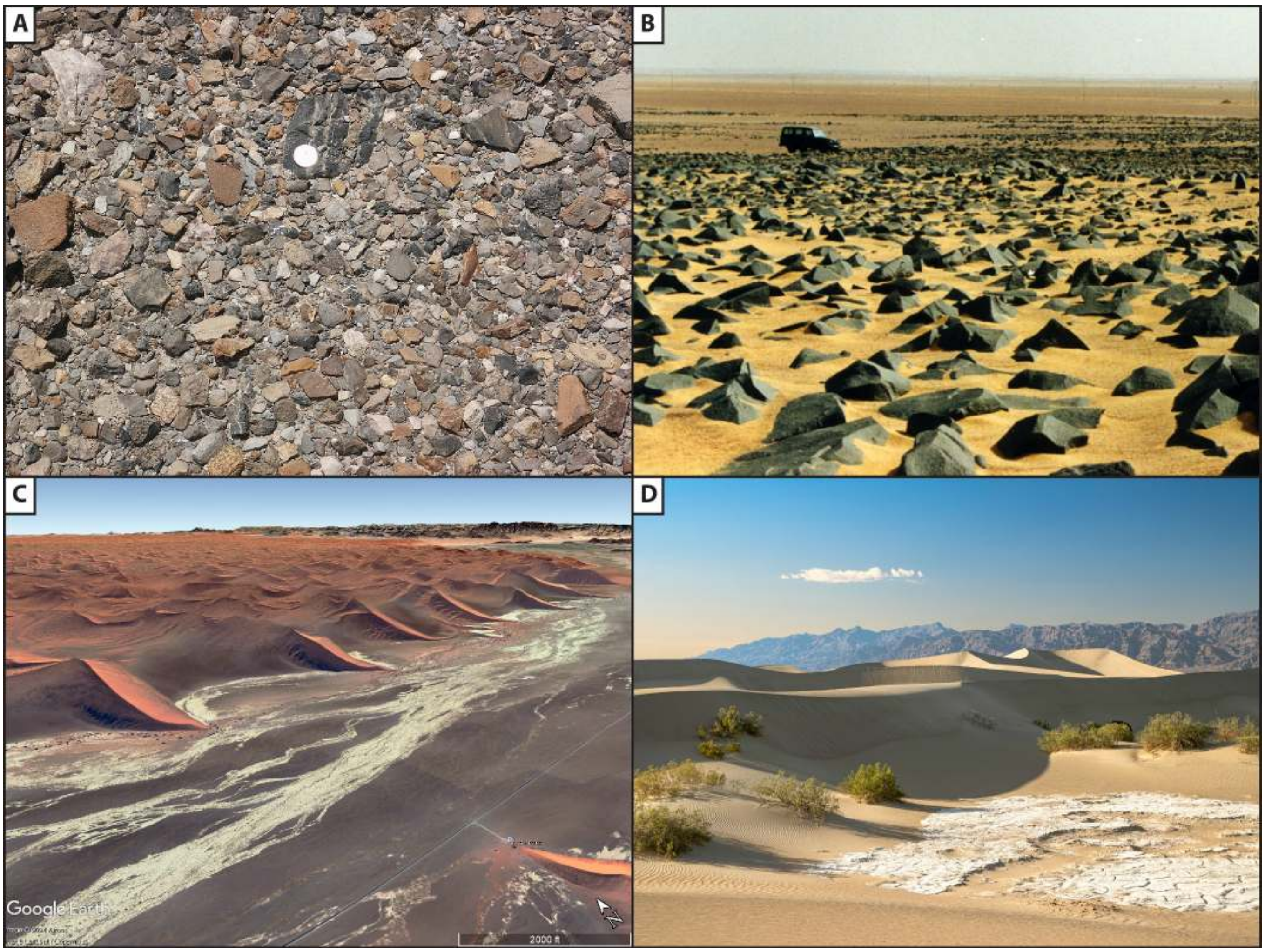

Figure 10.2.8: Modern sedimentological feature of interdune areas. A) Deflation lag (desert pavement) in Death Valley National Park (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0). B) Ventifacts with flat, faceted faces formed by wind erosion in a desert in Libya (calind via Wikimedia Commons; CC BY-SA 3.0). C) Oblique Google Earth image showing an interdune area in the Namib Sand Sea with a seasonal fluvial system and interdune vegetation (courtesy Google Earth, exported images from Google Earth can be embedded on websites for educational and non-commercial use). D) Photograph of interdune area with vegetation and desiccation cracks in fine grained sediment deposited in an emphemeral lake (Anthony Brown via Pexels; Pexels license).

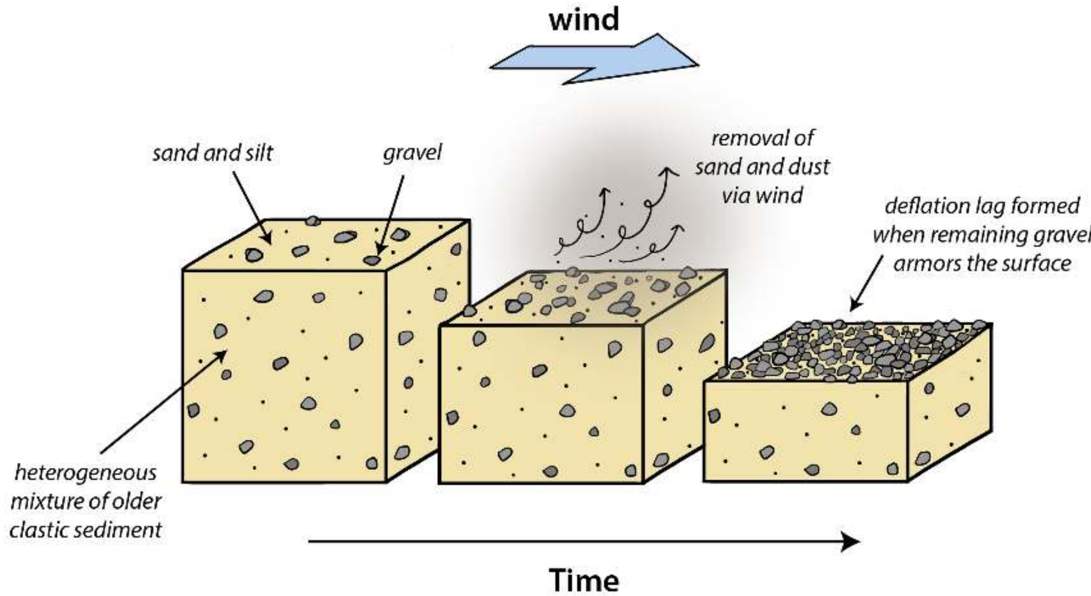

Figure 10.2.9: Diagram showing the formation of a deflation lag (desert pavement) by the removal of fines via wind and concentration of remaining coarse fraction (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

10.2.3: Overview

Overall, desert deposits commonly consist of large, eolian cross-beds that can be much thicker than subaqueous cross-beds, low amplitude wind ripple cross laminae, deflation lags, evidence of ephemeral lakes and rivers, evaporite minerals in playas or paleosols.

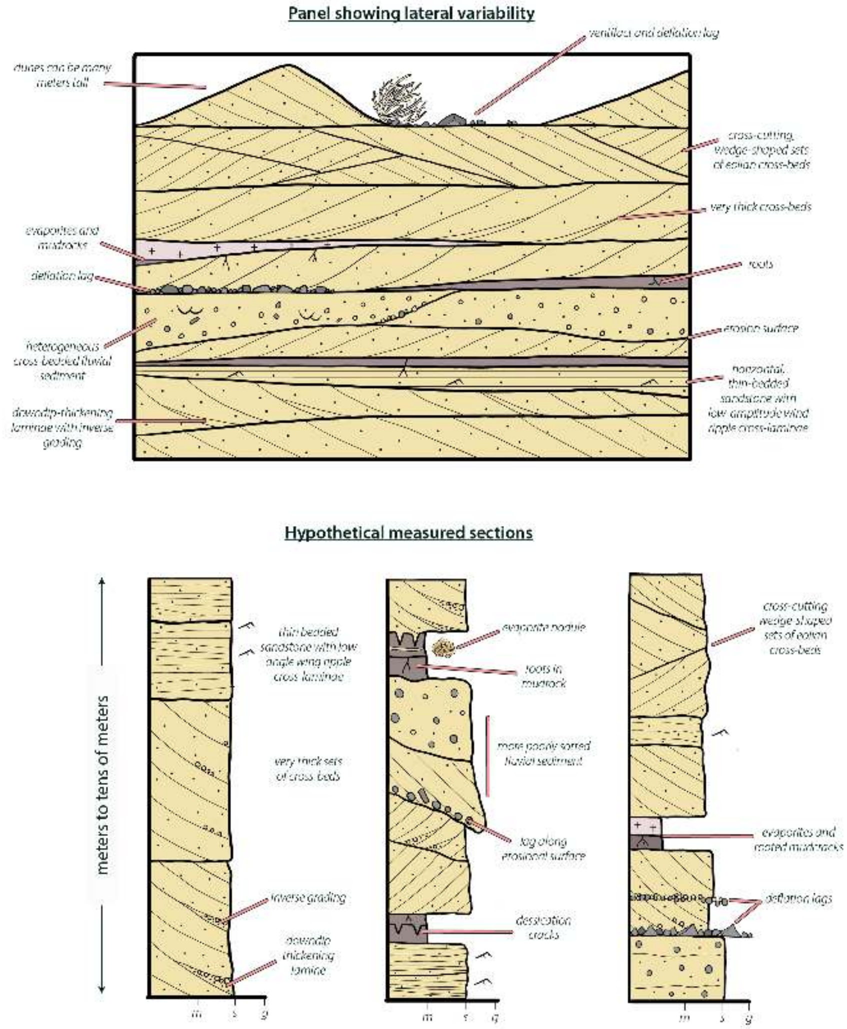

Figure 10.2.10: Overview of the sedimentology of desert deposits with particular emphasis on features that are likely to be preserved in the geologic record (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

10.2.4: Readings and Resources

- Laity, J.J., 2008, Deserts and Desert Environments, Wiley-Blackwell, 368 p., ISBN: 978-1-577-18033-3.
- Brookfield, M. E., & Silvestro, S. (2010). Eolian systems. In N. P. James & R. W. Dalrymple (Eds.), *Facies Models* (4th ed., pp. 7–7). Geological Association of Canada.

10.3: Clastic Marginal Marine Environments

Marginal marine environments form at the intersection of land and sea, where terrestrial and marine processes converge. In regions with a moderate supply of clastic sediment, these environments are shaped by the balance between waves, rivers, and tides which results in the development of beaches, deltas, and tidal flats, respectively. Relative sea level also plays a critical role in influencing these processes and the resulting depositional architecture. For instance, coastal areas dominated by fluvial systems are more likely to form deltas in regressive settings and estuaries in transgressive settings.

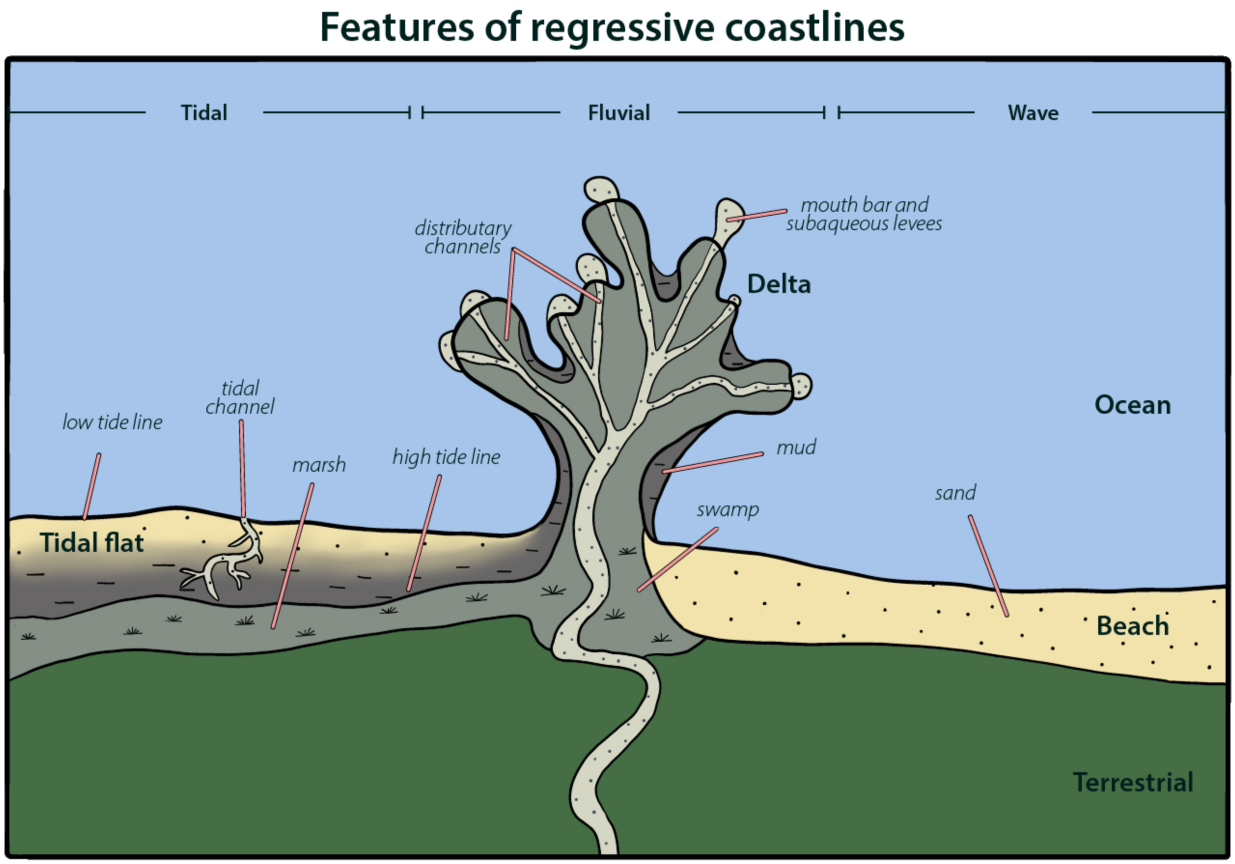

Figure 10.3.1: Features of regressive coastlines emphasizing geomorphic features associated with tide-, river-, and wave-dominated areas (from Page Quinton via Wikimedia Commons; CC BY-SA 4.0 which is after Boyd et al. (1992).

10.3.1: Beaches

A beach is a sandy, wave-dominated environment that is attached to the mainland. Waves are created when wind blows across the surface of the ocean; the resulting friction generates waves. Remember that waves cause circular motion in the water column and that motion translates down to a depth that is equal to one half of the wavelength. Fairweather wave base is the depth at which waves “feel” bottom under normal conditions; it marks the transition from marginal marine “beach” type deposits to the fully marine environments of the shelf.

10.3.1.1: Environments and Processes

Beaches are characterized by a distinct succession of environments, processes, and sedimentary structures that are summarized and illustrated below.

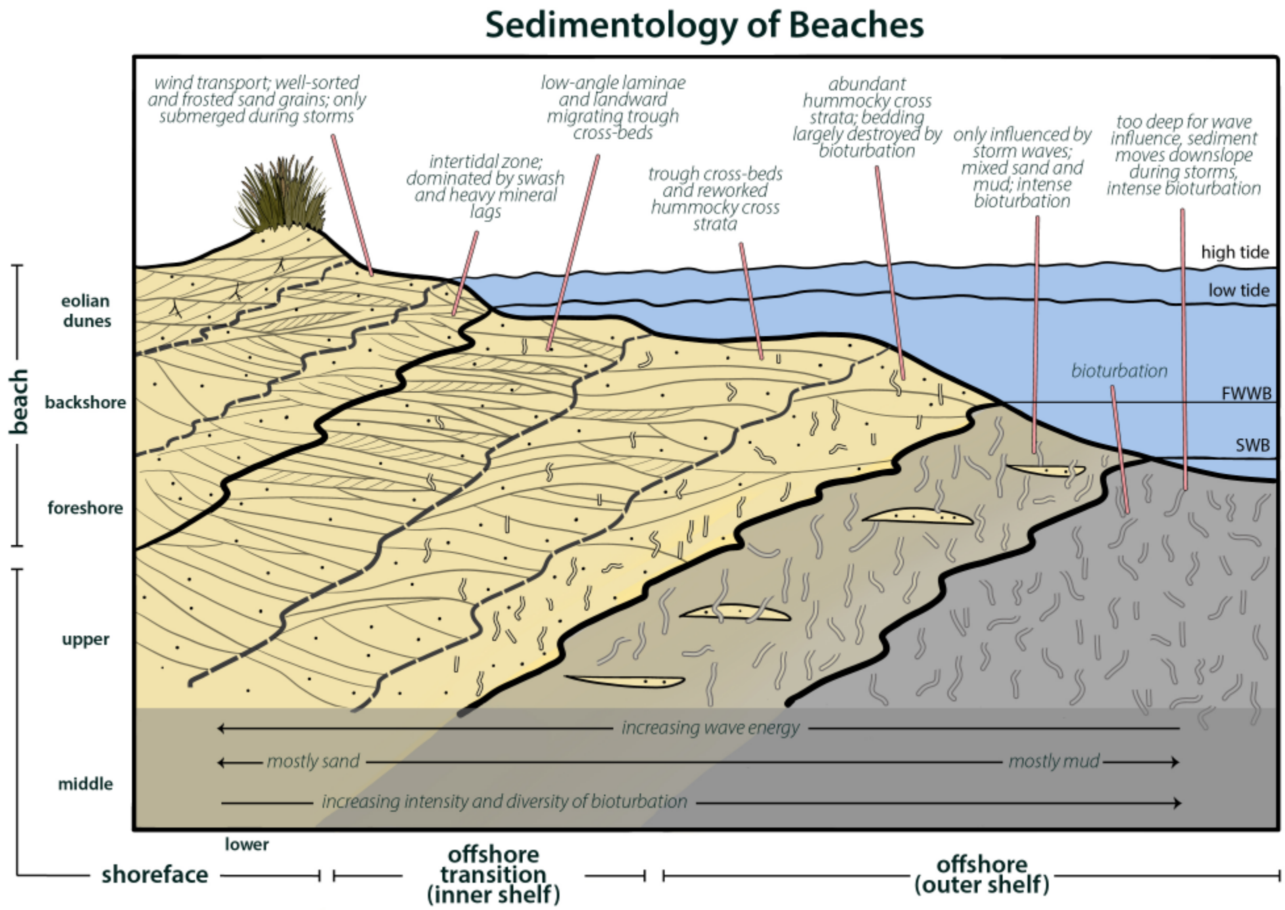

Figure 10.3.2: Cross-sectional view illustrating the major sedimentological features of beaches (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

Backshore areas only submerged during storms or extremely high tides and are dominated by eolian processes as wind blows inland across the beach. Eolian dunes and vegetation are common and the resulting deposits are characterized by landward- and seaward dipping low angle laminae, cross-beds, well-sorted, wind-reworked sand, and possibly roots and minor bioturbation.

The foreshore lies within the intertidal zone, between high and low tide levels. Here, sedimentary processes are dominated by the action of swash – where breaking waves send shallow/high velocity pulses of water and sand up the beach profile. Some water and sediment might flow back downslope. This results in low-angle laminae and the formation of heavy mineral lags. Bioturbation is very minor as few organisms can live in this dynamic and high energy environment.

Figure 10.3.3: Photographs of modern beaches. A) Backshore area with eolian dunes and vegetation on Ocracoke Island, North Carolina (Carol Highsmith, Library of Congress; no known restrictions on publication). B) Foreshore zone showing laminated sand cause by swash (Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

The shoreface is always submerged and is regularly reworked by wave activity. The shoreface can be divided into three distinct subzones based on sedimentary structures and biological activity:

1. Upper shoreface: This zone is characterized by trough cross-bedding from landward-migrating dunes, low-angle laminae, and abundant erosion surfaces.
2. Middle shoreface: Sedimentary structures in this zone include trough cross-bedding and reworked hummocky cross-stratification (HCS). Bioturbation becomes more abundant and diverse.
3. Lower shoreface: This deeper zone is dominated by abundant HCS formed by oscillatory wave conditions during storms. Bioturbation becomes both intense and diverse - often to the point of destroying primary bedding structures. Fairweather wave base marks the boundary between the lower shoreface and deeper water deposits of the offshore transition (inner shelf).

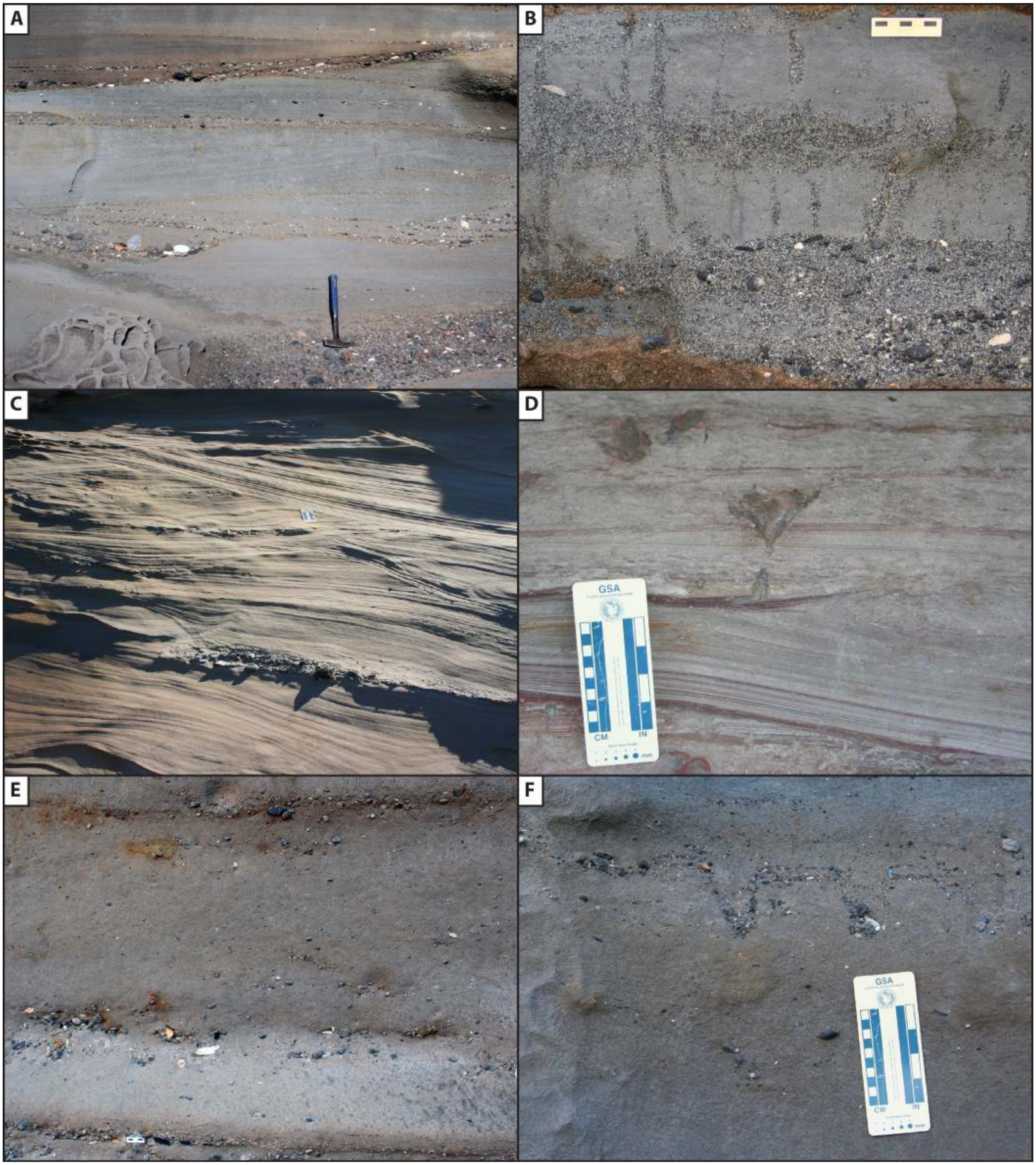

Figure 10.3.4: Photographs of shoreface deposits from the Wasp Head Formation (Permian), New South Wales, Australia (all photos from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0). Images on the left are outcrop scale and focus on stratification and sedimentary structures; images on the right are more detailed and show representitive bioturbation. A & B) Low-angle laminated pebbly sandstones of the upper shoreface, C & D) Hummocky cross-stratified middle shoreface deposits with Rosselia, E & F) Intensely bioturbated pebbly sandstone of the lower shoreface with *Conichnus* (anemone resting trace).

10.3.1.2: Barrier Islands

<u>Barrier islands</u> are detached beaches separated from the mainland by low-lying brackish wetlands called <u>lagoons</u>. Barrier island deposits are largely similar to the beach deposits described above and are best diagnosed through their context and association with lagoons. Lagoonal deposits include muds deposited from suspension deposition, beach sands introduced via washover events, and tidal inlet channels. Lagoons are typically organic-rich and experience both oxygen and salinity stress.

Most barrier islands form via one of two models:

1. Transgression by shoreface retreat: As sea level rises, storm washover processes move sand landward, resulting in the gradual migration of the barrier island toward the mainland. New sand is supplied by longshore drift of sediment supplied to the coast via rivers.
2. In-place drowning: If sea level rises rapidly, waves may deposit sand in situ, leading to the development of a barrier island without significant lateral migration.

Figure 10.3.5: Barrier island (foreground) and lagoon (middle of the image) near Carboneras, Mexico, which is visible near the top left of the image (courtesy Google Earth, exported images from Google Earth can be embedded on websites for educational and non-commercial use). Note the lobe-shaped washover deposits on the lagoon side of the barrier island.

10.3.2: Deltas

Deltas are marginal marine systems formed where fluvial systems build into standing bodies of water; they have both subaerial and subaqueous deposits. Deltas grow through the progradation of lobes, with new lobes forming as rivers shift to steeper gradients during avulsions. This process distributes sediment evenly across the delta system. Although they can be shaped by waves and tides, rivers are the dominant force shaping deltas. The largest examples typically occur on passive margins or in slowly subsiding areas where sediment supply exceeds the rate of subsidence.

10.3.2.1: **Environments and Processes**

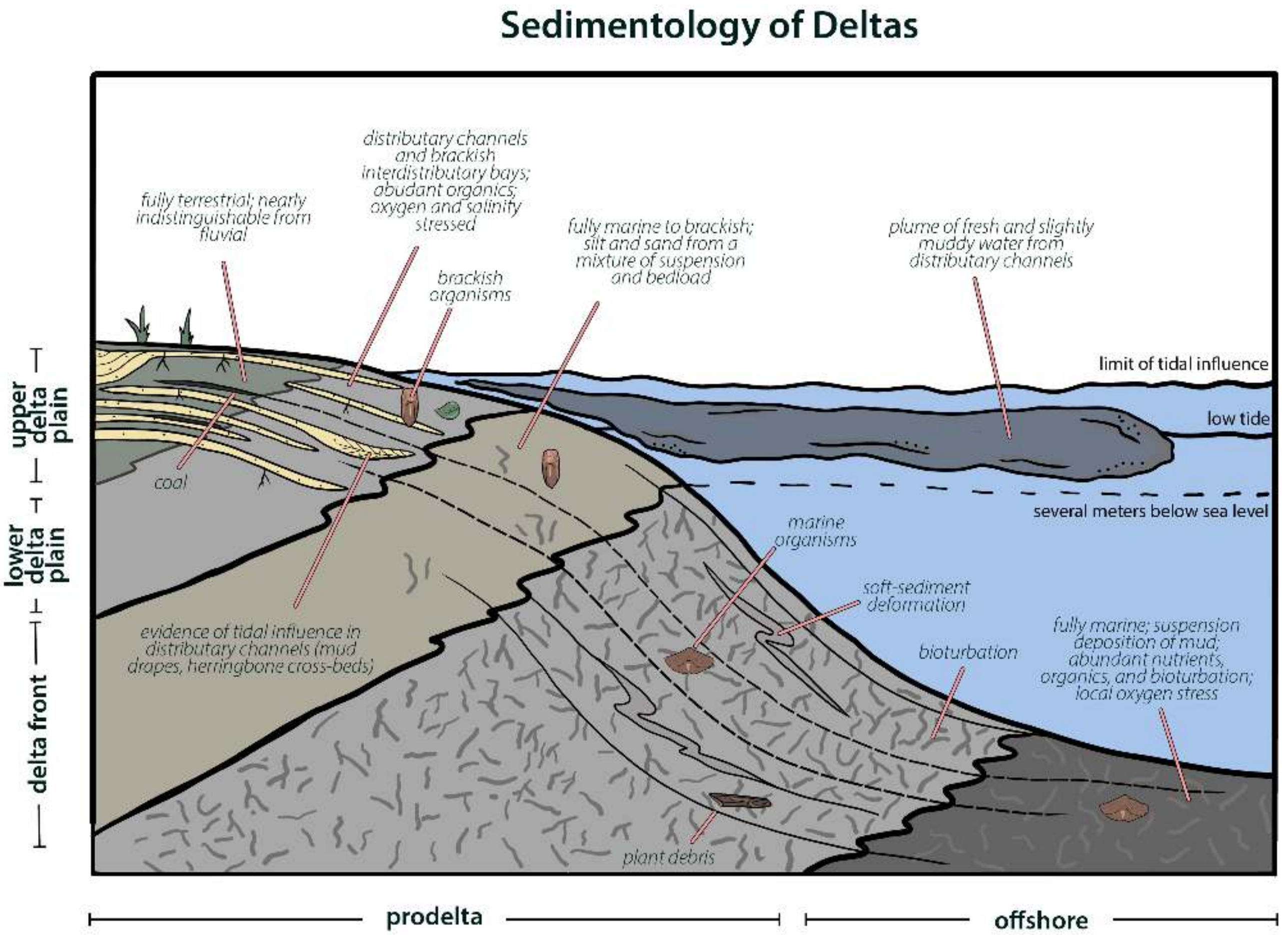

Figure 10.3.6: Cross-sectional sketch showing the major sedimentological subdivisions of a delta (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

The delta plain is the subaerial portion of the delta, which can be divided into two zones. The upper delta plain lies above the limit of tidal influence and is shaped primarily by fluvial processes. It is composed of river and floodplain deposits and they commonly contain peat swamps (coal). The lower delta plain, between the low tide mark and the furthest extent of tidal influence, experiences a mix of fluvial, tidal, and marine processes. This area is a stressed environment due to salinity, oxygen, and water level fluctuations. It includes distributary channels and interdistributary bays and wetlands.

Distributary channels transport water and sediment to the sea, often featuring river-like deposits with tidal mud drapes. These channels may cut into underlying deposits. Interdistributary intervals are broad, shallow bays with a mix of fresh and marine waters. Sediments in these intervals include muddy, organic-rich deposits and sandy crevasse splays. Faunal diversity and the intensity of bioturbation is limited by environmental stresses.

The delta front is a shallow marine zone (typically < 10 m deep) where clastic sediments are deposited by both bedload and suspension processes. This area may experience additional modification by wave or tidal processes, resulting in conditions that range from fully marine to brackish. Mouth bars and subaqueous levees form where sediment-laden channelized flow slows down where it meets standing water.

The prodelta lies seaward of the delta front, where fine silts and clays settle. These fines are supplied by plumes of fresh, muddy water that are less dense than salt water and may extend considerable distances from where distributary channels meet the ocean. This zone is less influenced by tides and waves, creating stable, open marine conditions with abundant nutrients and organic matter. Bioturbation is common, although localized stresses from decaying organics can reduce oxygen levels.

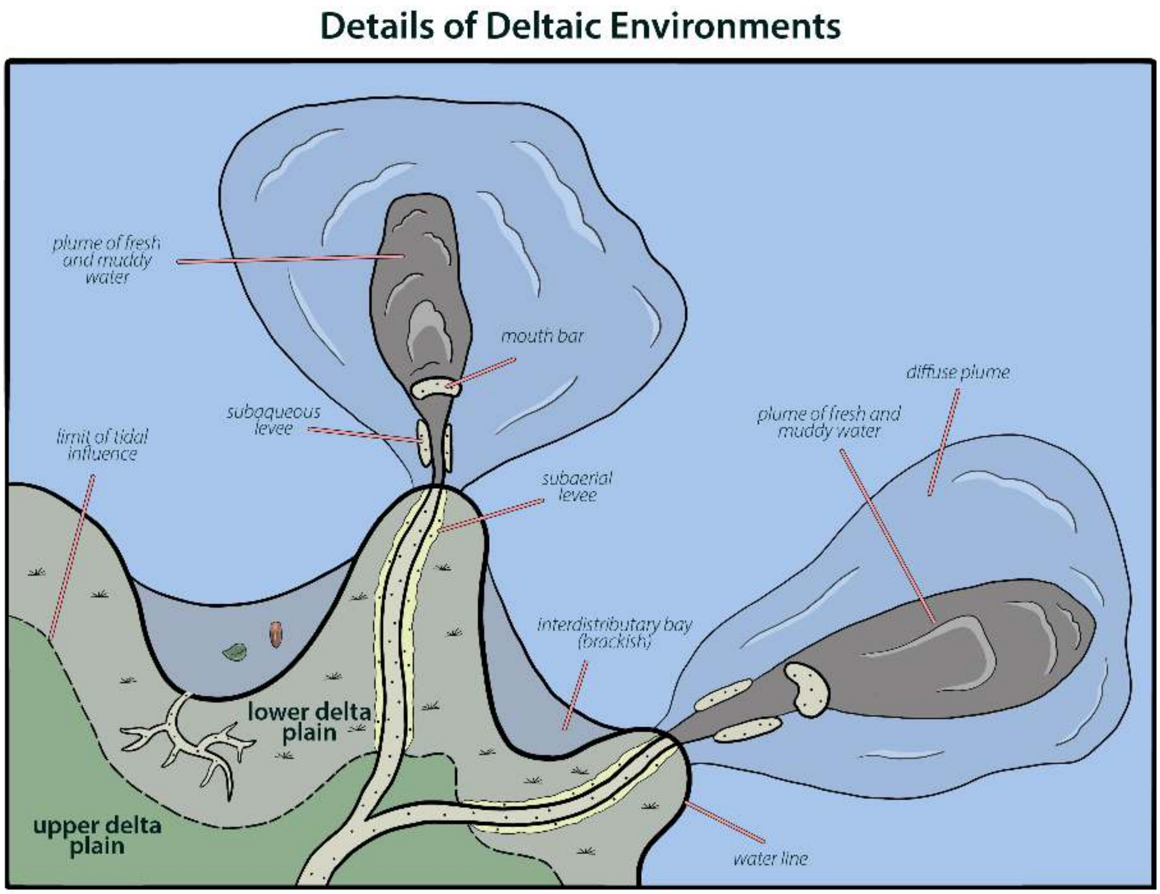

Figure 10.3.7: Map view sketch showing the major sedimentological subdivisions of the terminal portion of a delta (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

10.3.2.2: Morphology and Geometry

In map view, deltas commonly display a triangular to lobe-shaped morphology and in cross-section view they exhibit a wedge-shaped profile that thins in both updip and downdip directions. Although dominated by fluvial processes, the shape of and processes within deltas can be modified by tides and waves. Overwhelmingly fluvial-dominated deltas, such as the Mississippi River Delta, form lobate shapes with sandy distributary fingers. These types of deltaic successions commonly show stacked shallowing upward successions that record the progradation, and eventual abandonment, of individual lobes. Wave-dominated deltas, like the Nile Delta, are shaped by wave action that results in a blunt overall shape with flanking beaches. Tide-dominated deltas, such as the Mahakam Delta, feature funnel-shaped, sand-filled distributary channels and prominent tidal indicators.

Deltaic deposits on progradational coastlines commonly exhibit coarsening-upward successions formed from shallowing and the advance of the delta into the standing body of water. Given the overlap with purely fluvial systems, deltaic interpretations are best made from multiple outcrops where one can more confidently identify sandy or silty fluvial deposits prograding into lacustrine or marine mudrocks or seismic lines where one can see seaward-dipping clinoforms (when viewed perpendicular to the coast) or mounded clinoforms (when viewed parallel to the coast).

Figure 10.3.8: Airphotos showing the spectrum of delta morphologies. A) The Mississippi River delta complex includes multiple inactive lobes as well as the active Balize lobe (the "Bird foot" delta) near the right edge of the image. The photo was taken during high flow and highlights the influx of muddy water into the Gulf of Mexico. The Atchafalaya River is an active distributary channel and is flowing into the Gulf of Mexico near the left edge of the image. (NASA via Visible Earth; public domain). B) Detail of the Bird Foot delta (NASA via Visible Earth; public domain), C) The tide-dominated Mahakam Delta with the funnel-shaped distributary channels characteristic of these types of deltas (NASA via Earth Observatory; public domain). D) The wave-dominated Pariba Del Sul Delta shows a blunted morphology and a series of beach ridges (courtesy Google Earth, exported images from Google Earth can be embedded on websites for educational and non-commercial use).

10.3.3: Tidal Flats

Tides are periodic vertical changes in water level, caused by the combined effects of centrifugal and gravitational forces in the Earth-Moon and Earth-Sun systems. Most coastlines experience a tidal range (the difference in height between high and low tide) of just a few meters, but local conditions can cause it to be much higher in some locations. The world's highest tides are present in the Bay of Fundy in Atlantic Canada where the average tidal range in some parts of the bay are ~15 m (50 ft).

Figure 10.3.9: Photographs of a wharf on the Bay of Fundy at high tide (left) and low tide (right) taken approximately six hours apart (Samuel Wantman via Wikimedia Commons, CC BY-SA 3.0).

10.3.3.1: Causes of the tides

As mentioned above, tides are caused by gravitational and centrifugal forces caused by the interaction of the Earth-Moon and Earth-Sun systems. Although the moon is much smaller than the Sun, it is much closer and exerts approximately twice the tidal influence of the Sun. For the sake of simplicity, we will first consider the interaction of the Earth and Moon in our initial discussion; the interaction of the Earth and Sun is much the same and is best considered in the discussion of neap and spring tides. A fuller discussion of the causes of the tides is available in Chapter 11.1 of Webb's Introduction to Oceanography (also an OER resource).

The Moon's gravity pulls water toward it which creates a tidal bulge that is greatest directly beneath it. A second, smaller tidal bulge is present on the opposite side of the Earth; it is caused by centrifugal force from the rotation of the Earth and Moon around their shared center of mass (the barycenter). The tidal bulges remain beneath and opposite the position of the Moon, but because the Earth spins on it's rotational axis and the moon rotates in the same direction, it takes 24.84 hours for a tidal bulge to return to the same spot on the surface of the Earth. The unevenness of the two high tides (and intervening low tides) is further enhanced by the inclination of the Moon's orbital plane relative to the equator. Thus, many places experience two unequal high tides and two unequal low tides every 24.84 hours (about 50 minutes later each day).

Tidal range also changes because of the interaction of the lunar and solar tidal bulges. Spring tides are times with relatively high tidal ranges that occur when the Earth, Moon, and Sun are in alignment and the resulting tidal bulges experience constructive interference. Conversely, neap tides are periods of lower tidal range caused by the Moon being at 90° to the Earth-Sun line which results in destructive interference. Thus, there is a ~28 day cyclicity with spring tides occurring during new and full moons (about 14 days apart) and neap tides occur midway between them when the moon is at first quarter and last quarter, respectively.

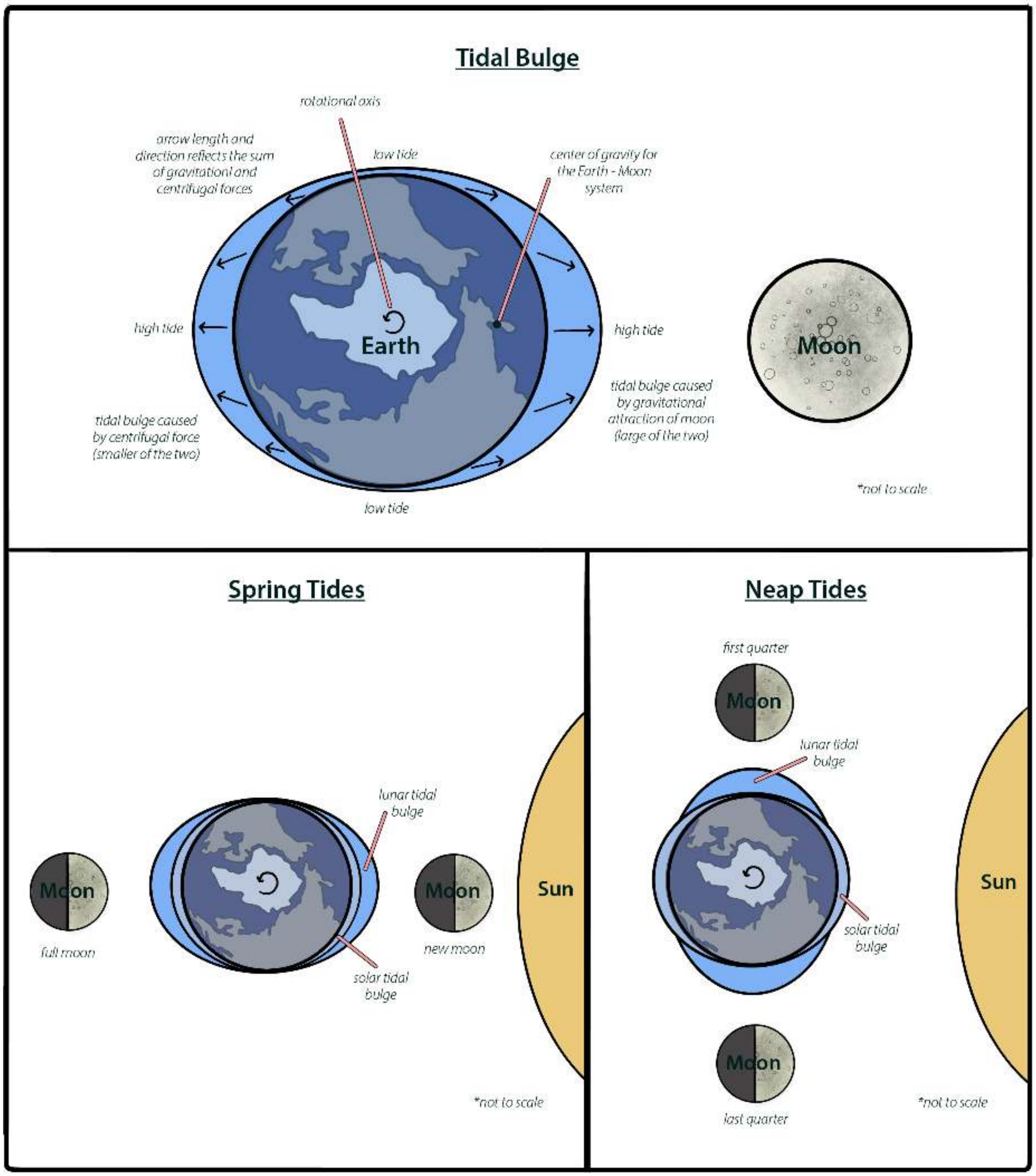

Figure 10.3.10: Diagrams showing gravitational and centrifugal tidal bulges (above) and the interaction fo the Earth-Moon-Sun system to cause neap and spring tides (below). From Page Quinton via Wikimedia Commons; CC BY-SA 4.0.

10.3.3.2: Tidal Deposits

The influence of tides can result in regular changes in current velocity and direction which can, in turn, cause the formation of distinctive sedimentary structures that include:

- Tidal rhythmites: These deposits consist of sand-mud couplets, where mud is deposited during periods of low water movement (high and low tide) and sand is deposited when currents are stronger during the rising and falling tides. In some tidal systems, the ebb and flood currents are weaker, resulting in mud-sand couplets that are unequal in size.
- Heterolithic bedding: Tidal deposits often show more complexity than simple rhythmites. Alternations between sand and mud reflect fluctuations in flow strength due to changes in current velocity, water depth, neap-spring cyclity, and other factors.

- Mud drapes: In tidal environments, periods of slack water may allow the deposition of thin mud drapes over coarser sediments, even in areas with strong currents capable of forming dunes.
- Herringbone cross-beds: In situations where currents are consistently strong, sedimentary structures such as herringbone cross-beds may form. These features reflect changes in current direction, typical of tidal environments where the flow alternates between flood and ebb tides.

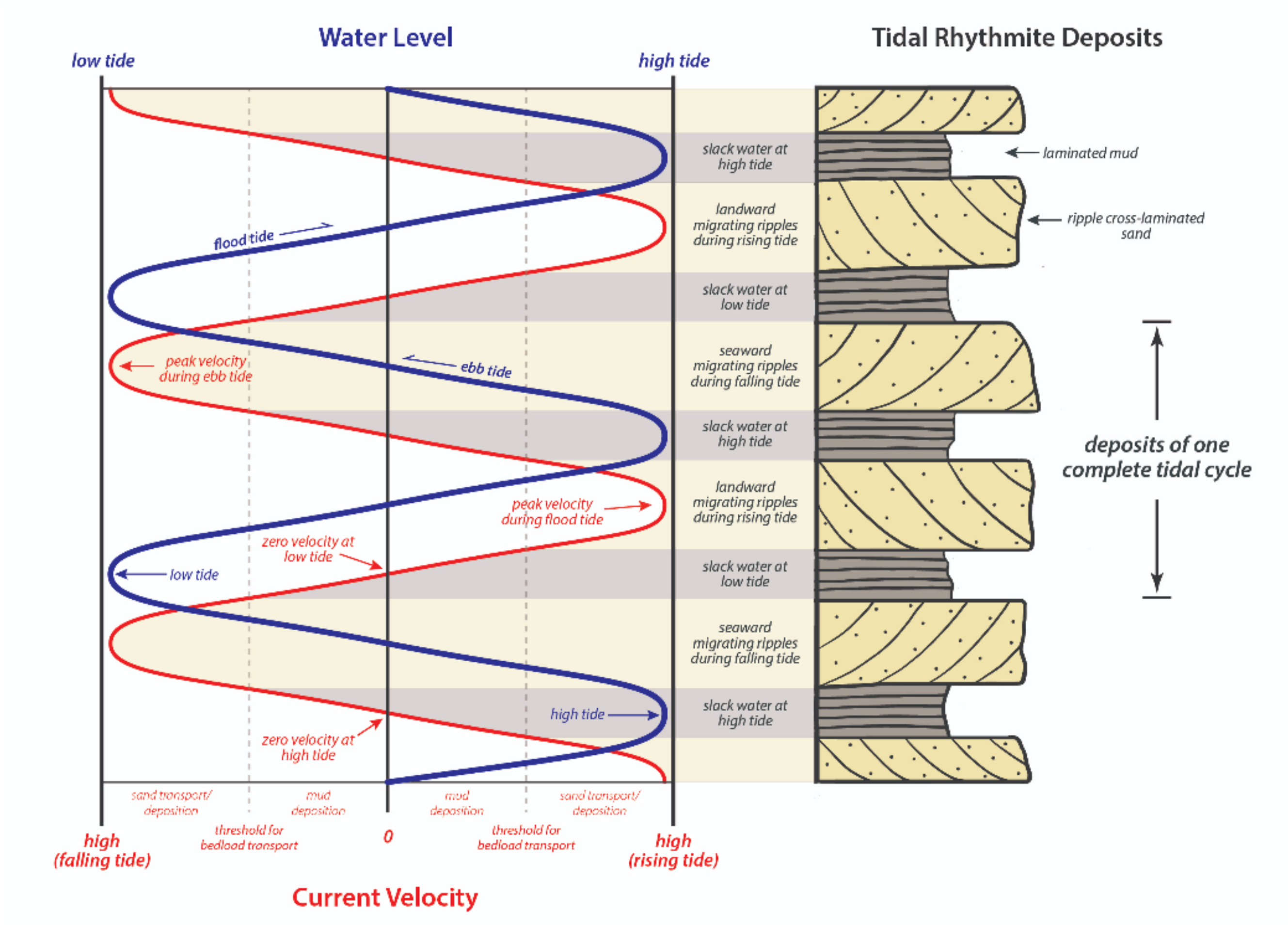

Figure 10.3.11: Illustration showing the relationship between tides, tidal currents, and tidal rhythmite deposits. From Page Quinton via Wikimedia Commons; CC BY-SA 4.0 which is after Dalrymple, R.W. (1992) and Kvale et al. (1998).

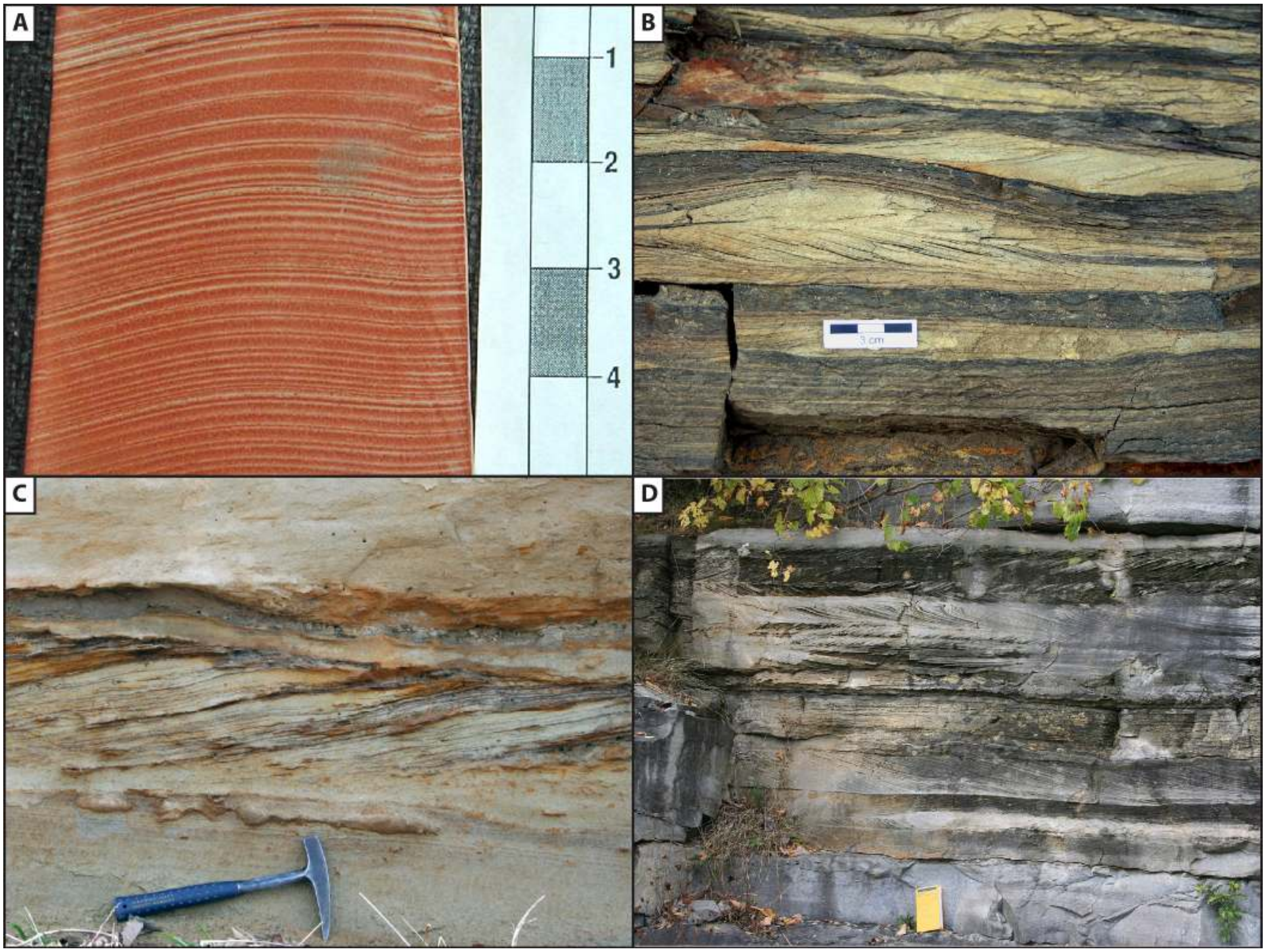

Figure 10.3.12: Tidal deposits include A) Tidal rhythmites (scale bar is in inches), B) heterolithic bedding, C) mud drapes, and D) herringbone cross-bedding. A) is from James St. John (Flickr; CC BY 2.0) and B-D) are from Michael C. Rygel via Wikimedia Commons; CC BY-SA 4.0).

10.3.3.3: Tide-dominated coasts

In coastal areas dominated by tides we can consider three main zones:

- Supratidal areas are located above the high tide mark and are only inundated during exceptionally high tides or storms. These areas consist of muddy salt marshes that are colonized by salt-tolerant grasses, succulents, and herbs.
- The intertidal zone is effectively synonymous with the term "tidal flat" and is the low-gradient area that is variously exposed and submerged during a tidal cycle. Updip areas are typically muddy because they are only inundated around high tide when water velocity is relatively low. Tidal flats become progressively sandier downdip because these areas experience bedload transport during relatively high velocity flows during falling and rising tides. The intertidal zone is the area that is exposed and submerged with each tidal cycle. They often have a dendritic network of tidal channels that help drain water during low tide. These areas contain both sand and mud and heterolithic bedding is common; lenticular bedding dominates in updip areas and flaser bedding dominates in downdip areas.
- Subtidal areas are always submerged and are profoundly influence by currents from rising and falling tides and/or currents that parallel the shoreline. They are composed of sandy sediment organized into a variety of large dune and sandbar morphologies.

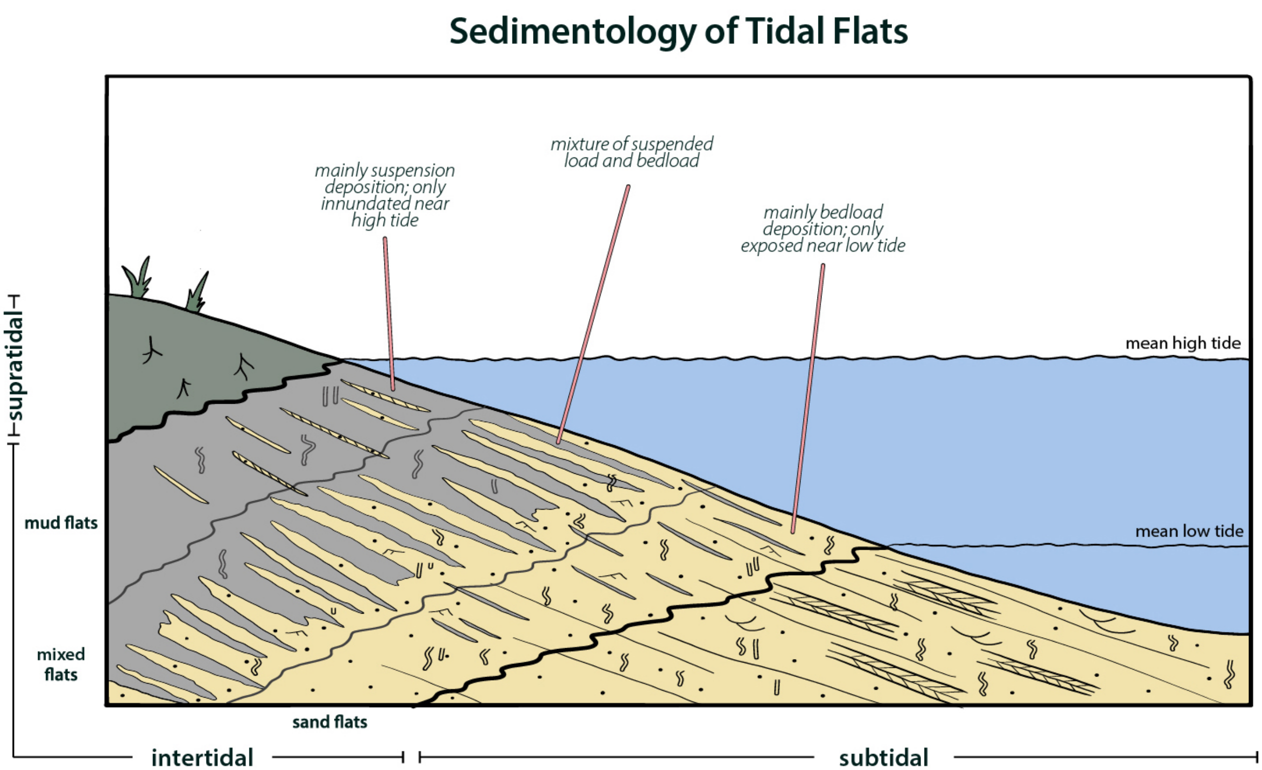

Figure 10.3.13: Cross-sectional sketch showing the major sedimentological subdivisions of a tide-dominated coast (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

10.3.3.3.1: Tidally-influenced rivers

Rivers that drain into areas with high tides can be profoundly impacted by tidal cycles; in extreme cases the channel might be bankfull during high tide and flow inland and nearly empty and flow seaward during low tide. In certain regions, rising tides can create a tidal bore - wave that travels upstream, marking the point where flow reversal occurs. Tidal influence can extend a surprising distance upstream in larger systems and might be expressed through the influx of brackish waters or decreased flow velocity because of ponding during high tide. As mentioned above, tidally-influenced deltas often take on a funnel-like morphology with relatively straight, radiating distributary channels

10.3.4: Readings and Resources

- Boyd, R., Dalrymple, R., and Zaitlin, B., 1992, Classification of clastic coastal depositional environments. Sedimentary Geology. v. 80. p. 139-150.
- Dalrymple, R.W., 1992, Tidal Depositional Systems in Walker, R.G. and James, N.P. (eds.), Facies Models (2nd ed), Geological Association of Canada, St. John's, p. 195-218 a
- Kvale E.P., Sowder K.H., Hill B.T., 1998, Modern and ancient tides - Poster and explanatory notes, SEPM, Tulsa, OK, and Indiana Geological Survey, Bloomington, IN.
- Webb, P., 11.1 - Tidal forces in "Introduction to Oceanography". https://rwu.pressbooks.pub/webboceanography/chapter/11-1-tidal-forces/

10.4: Clastic Marine Environments

10.4.1: Nature of Ocean Basins

The depth of ocean basins is largely controlled by tectonic activity and the geological setting in which they form. Pericontinental seas are found at the edges of continents, adjacent to very deep oceanic waters. These seas contrast with epicontinental seas, which occur when continental areas become submerged. Epicontinental seas are relatively rare in the modern world, with Hudson Bay being one of the few examples. However, they were far more prevalent during the Cretaceous and other greenhouse climate phases when ocean basins held less water (rapid seafloor spreading made them warmer and more buoyant) and because there was more liquid water present because there was little/no glacial ice. This lead to widespread, but relatively shallow flooding of continental landmasses.

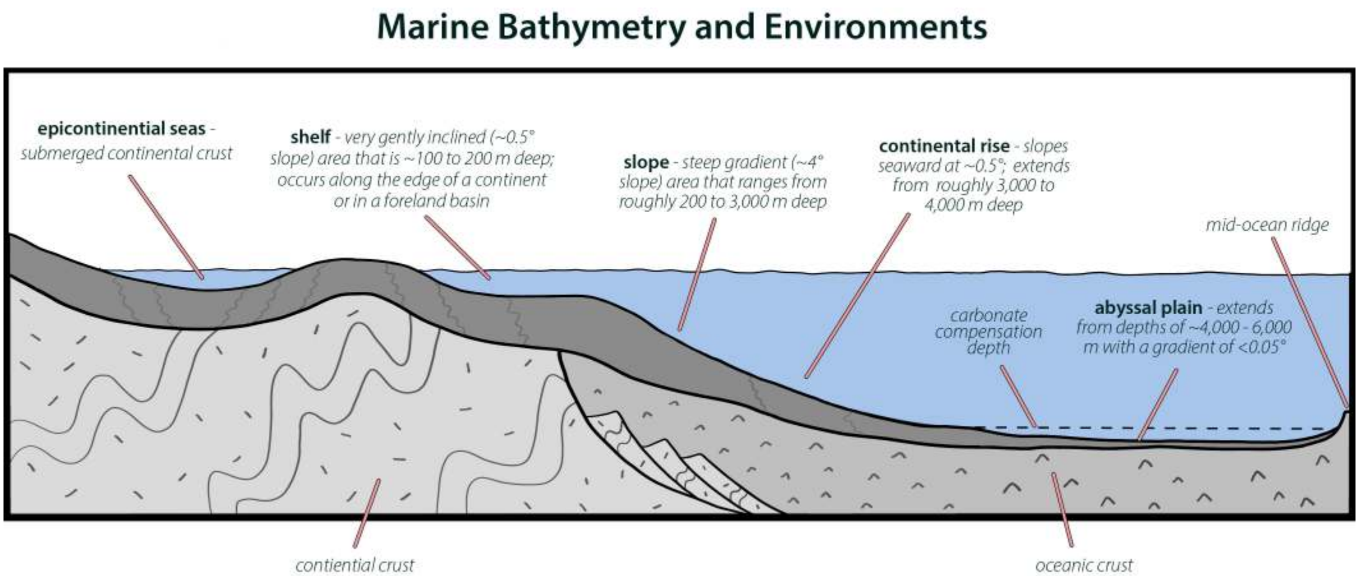

Figure 10.4.2: Cross-sectional sketch showing marine bathymetry and major environments (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

10.4.2: Shelf Environments

The shelf is a relatively shallow, very gently dipping area along the submerged margin of a continent or in a foreland basin. It extends from the boundary with the lower shoreface to the more steeply dipping shelf break. The width of the continental shelves varies greatly, ranging from a few kilometers to over 1,500 kilometers wide and ~100 to 200 m deep.

Figure 10.4.2: The gorge at Taughannock Falls, NY exposes a thick succession of Devonian-aged shelf mudrocks deposited as part of the Catskill clastic wedge (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0).

10.4.2.1: Inner Shelf

Waves are an important influence on sedimentation in many inner shelves. Fair-weather wave base is the depth at which waves influence the seafloor under normal conditions ... the depth is equal to 1/2 of the spacing between waves. Waves are larger during storms and their influence extends to greater depth (storm wave base). The offshore transition is the innermost portion of a wave-dominated inner shelf; it lies between fair weather wave base and storm wave base. It is characterized by admixed sand and mud, intense bioturbation and the resulting paucity of primary sedimentary structures. Hummocky cross-stratified (HCS) sandy layers deposited during storm events may be present locally if they were buried before they could be homogenized via bioturbation.

In relatively shallow seas with significant tidal currents sediment movement is more continuous and large, straight- to sinuous-crested sand ridges may be present. These features are elongate parallel to current direction and can be ornamented with bi-directional cross beds the reflect ebb and flood currents.

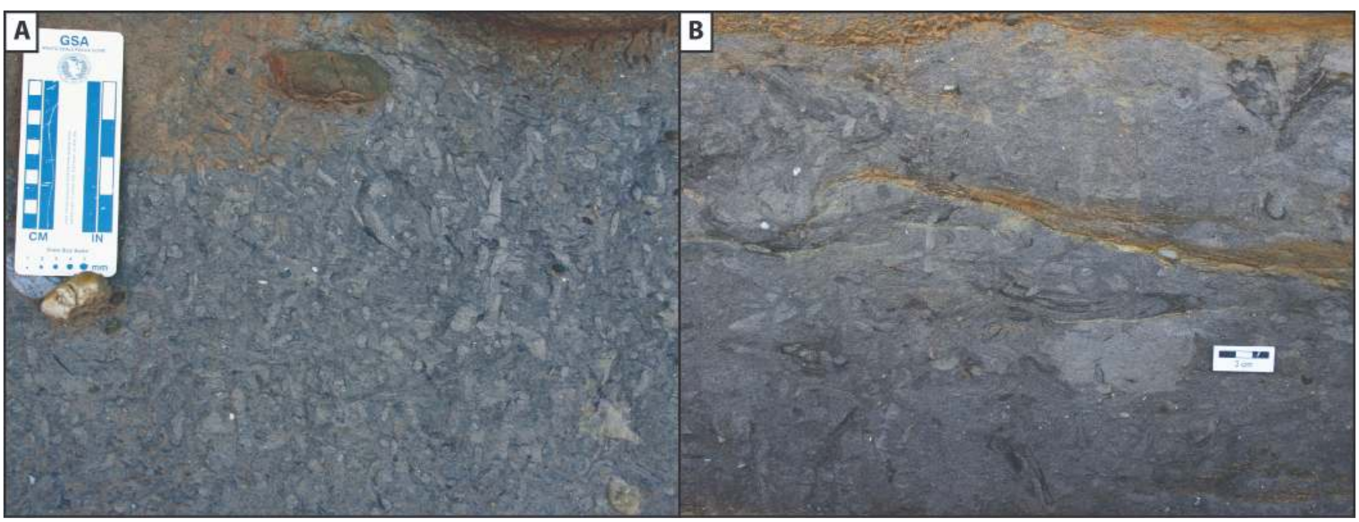

Figure 10.4.3: Intensely bioturbated sand and mud in deposits of the offshore transition (inner shelf). Images from Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0.

10.4.2.2: Outer Shelf

Beyond the inner shelf, the outer shelf lies at depths too great for wave and tidal influence. Here, variably bioturbated mudrock via suspension deposition is the most common lithology. Weak currents exist, mainly influenced by oceanic circulation or upwelling. Additionally, relict sand bodies may also be present - they represent remnants of coarser sediments deposited when sea levels were lower and sediment transport was more active. Carbonate deposition can also occur in deeper water portions of some clastic-dominated shelves.

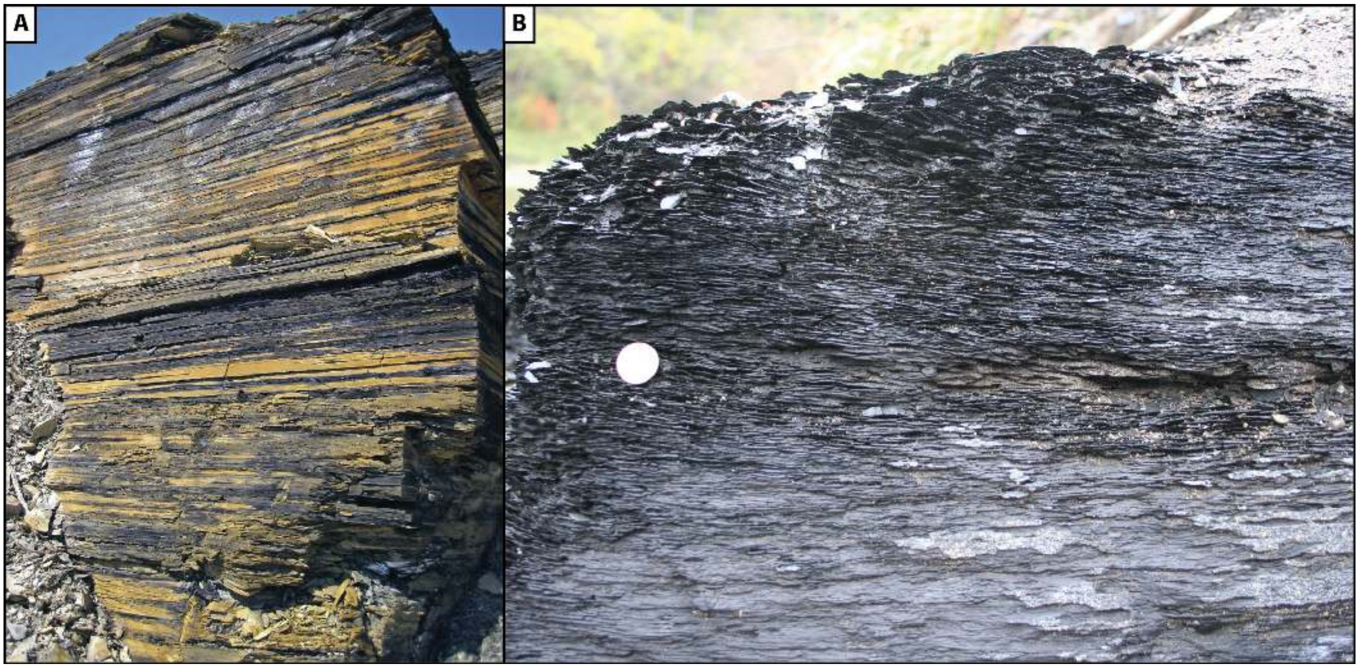

Figure 10.4.4: Organic-rich shales deposited in oxygen-poor outer shelf water depths. A) New Albany shale (Upper Devonian) at the McDonald Knob Outcrop, Bullitt County, Kentucky (James St. John via Flickr; CC BY 2.0) and B) Utica shale exposed along the NYS Thruway (Michael C. Rygel via Wikimedia Commons; CC BY-SA 3.0).

10.4.3: Deep Marine Environments

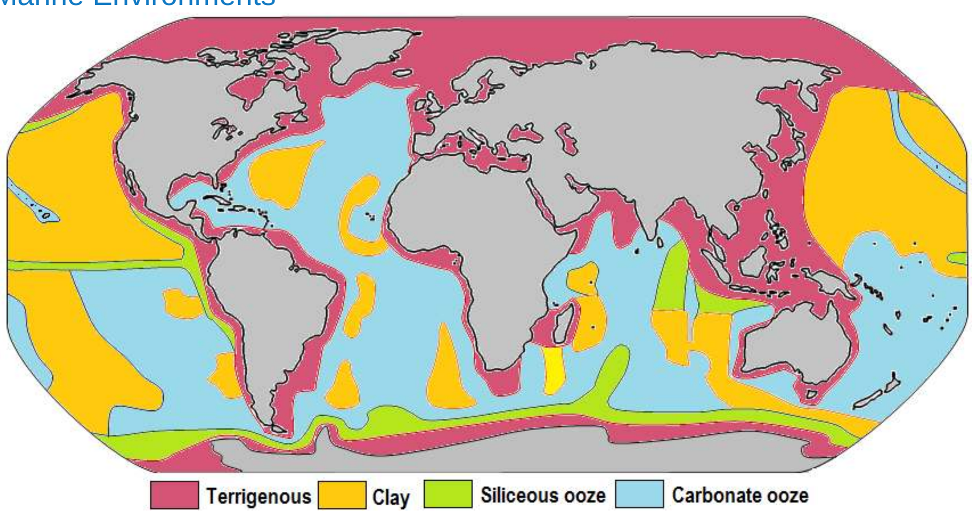

Figure 10.4.5: Distribution of sediment types on the seafloor; Within each colored area, the type of material shown is what dominates, although other materials are also likely to be present (Steven Earle, Physical Geology (2nd), CC BY 4.0).

10.4.3.1: Slope, Rise, and Abyssal Plain

Deep marine depositional settings include the continental slope, continental rise, and abyssal plain. The continental slope dips seaward at approximately 4 degrees and serves as a major conduit for sediment transport from the continental shelf to the deep ocean. The slope generally extends from depths of around 200 meters at the shelf break to approximately 3,000 meters where it transitions into the rise. This region also marks the transition from continental to oceanic crust. Sediment is delivered through suspension deposition and storm-driven transport (via gravity and exceptional currents), but the relatively steep gradient makes the slope prone to soft-sediment deformation, failure, and gravity mass movement. Slope failures lead to the development of turbidity currents (turbulent mixtures of sediment and water) which can carve submarine canyons and transport sediment downslope under the influence of gravity.

At the base of the continental slope, the continental rise consists of sediment transported by turbidity currents. This region, which slopes more gently at around 0.5 degrees, extends from approximately 3,000 to 4,000 meters in depth. These currents form channelized systems with levees, resembling subaerial floodplain channels. As turbidity currents lose energy, they deposit sediment in a characteristic sequence described by the Bouma model, which details the transition from high-energy turbidity currents to lower-energy traction currents and eventual suspension deposition. This results in a fining-upward sequence, with coarser material deposited first and finer sediments settling out as the current dissipates.

Figure 10.4.6: Stacked turbidite deposits (partial Bouma sequences) exposed in a road cut near Mill Creek, California (Dave Schumaker via Flickr; CC BY-NC-ND 2.0).

Beyond the continental rise lies the abyssal plain, one of the most extensive depositional environments on Earth (it covers nearly 50% of the planet's surface!). The abyssal plain extends from depths of about 4,000 to 6,000 meters and has an extremely low gradient, typically less than 0.05 degrees. Sedimentation here is almost exclusively pelagic, with fine-grained sediments settling out from suspension. These deep-sea sediments are primarily composed of siliceous or calcareous ooze and deep-sea clays.

10.4.3.2: Carbonate Compensation Depth

The carbonate compensation depth (CCD) plays a crucial role in determining the composition of abyssal sediments. At shallow depths, surface waters are supersaturated with calcite, allowing for relatively abundant carbonate deposition. However, as depth increases, cold temperatures and high pressure cause the water to become undersaturated, which contributes to the dissolution of calcareous material. At the CCD, the rate of carbonate dissolution exceeds the rate of supply, and below this depth carbonate sediment is not present and siliceous oozes or deep sea clays become dominant. Features like mid-ocean ridges and volcanic seamounts can create significant topographic relief in the deep ocean and the calcareous sediment deposited in these shallow waters may be eventually be buried with bedded cherts or shales as plate motion causes the plate to move, cool, and subside.

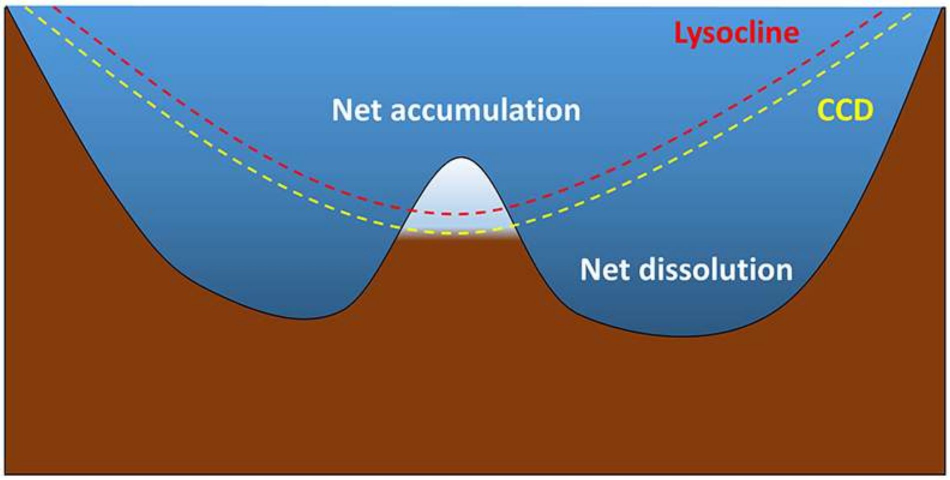

Figure 10.4.7: Calcareous sediment can only accumulate in depths shallower than the calcium carbonate compensation depth (CCD). Below the CCD, calcareous sediments dissolve and will not accumulate. The lysocline represents the depths where the rate of dissolution increases dramatically (Paul Webb via Introduction to Oceanography, CC BY 4.0).

10.5: Carbonate Environments

10.5.1: Distribution in Space and Time

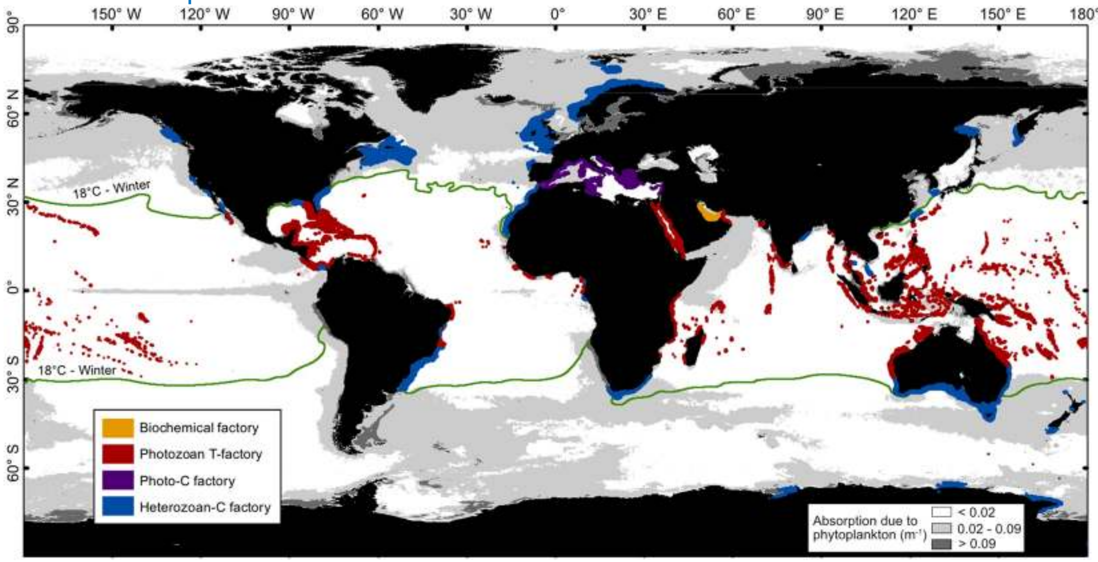

Figure 10.5.1: Distribution of modern shallow water carbonate environments. The different colors represent different "factories", each of which is a distinctive carbonate-forming ecosystems. They include biochemical factories (orange) dominated by lime mud, ooids, and/or stromatolites, photozoan - tropical (red) factories dominated by coral and green algae, Photozoan - cool water (purple) factories areas dominated by red algae and sea grass, and heterozoan - cool water (blue) factories dominated by bryozoans and mollusks. From Laugié et al, 2019; CC BY 4.0.

Figure 10.5.2: View of Egg Island, an 800 m^2 island in the Bahama Banks which is a large carbonate platform in the Caribbean Sea. Note the deep blue open ocean to the north of the reef that makes up the island (tropical photozoan factory) and the light-colored, tidally influenced oolitic sand bars south of the island (biochemical factory). From the European Space Agency via Wikimedia Commons; CC BY-SA 3.0 IGO.

Most of the world's coastlines today are dominated by clastic sediments, with only a few regions exhibiting significant carbonate production. These carbonate-producing areas are often in shallow, tropical areas that are removed from major rivers and other sources of clastic sediment. In environments where clastic sediments are present, carbonate production is often inhibited. Clastic sediments reduce light penetration, dilute carbonate material, and generally create unfavorable conditions for carbonate-producing organisms. While mixed clastic-carbonate systems do exist, it is often more practical to consider carbonate and clastic depositional environments as distinct end-members due to their differing sedimentary processes.

Carbonate deposition is most prevalent during warm climate phases and in shallow seas where sea levels are high. Although abiotic carbonate production can be locally important (oolitic limestones, whiting events, etc.), a large percentage of modern carbonate production is biological, derived from shelly organisms, calcareous algae, and calcareous plankton. Since most of these organisms thrive in the photic zone, carbonate deposition is generally restricted to shallow marine environments where light penetration is sufficient to support biological activity.

Carbonate rocks are significant in the geological record, but the organisms associated with their formation have changed dramatically over time. Unlike clastic sediments, carbonate sediments are intrabasinal, meaning they originate within the basin where they are deposited. The maximum grain size in carbonate sediments is controlled by the skeletal material of the organisms rather than by hydrodynamic processes. Additionally, the proportion of carbonate mud present in a deposit serves as a reliable proxy for assessing energy levels within the depositional environment.

Terminology

A stunningly vague, complex, and sometimes contradictory collection of terms has arisen to describe the spectrum of carbonate environments. In this section we adopt the following terminology based on our experiences and reading of the literature:

Basin - A deep marine environment well below storm wave base. Dominated by suspension deposition. Basinal deposits can be nearly identical to outer ramp or outer shelf deposits.

Lagoon - A relatively narrow body of water that sits between the mainland and some barrier with the open ocean (in this context, typically a shoal or reef). Communication with fully marine waters is restricted and oxygen and/or salinity stress are typical. Generally very low energy; some have peritidal areas around their margins.

Peritidal - Areas that are tidally influenced or immediately adjacent to them. Subenvironments include subtidal (below avg. low tide), intertidal (variously exposed and inundated by normal tides), and supratidal (above avg. high tide).

Platform - A nearly horizontal, shallow marine environment that is surrounded by marine waters. Flooded areas of large platforms can be considered shelves. Can be in direct communication with the open ocean (unrimmed) or can be bounded by a reef or shoal (rimmed).

Ramp - A moderately dipping, relatively shallow marine environment. Comparable to a shelf but a somewhat steeper slope. Can be in direct communication with the open ocean (unrimmed) or can be bounded by a reef or shoal (rimmed). Outer ramp deposits can be nearly identical of basinal facies.

Reef - An *in situ* accumulation of wave-resistant organisms (cf. shoal). Can form on the margins of platforms, ramps, or shelves.

Shelf - a very low gradient, relatively shallow marine environment in shallow epieric seas, continental margins, or atop platforms. Comparable to a ramp, but a shallower gradient. Can be in direct communication with the open ocean (unrimmed) or can be bounded by a reef or shoal (rimmed). Outer shelf deposits can be nearly identical of basinal facies.

Shoal - an accumulation of sand-sized carbonate grains that that builds up in higher energy, wave-dominated environments (cf. reef). Can form on the margins of platforms, ramps, or shelves.

Slope - A steeply dipping area that is transitional between the relatively shallow waters of a shoal, shelf, ramp, or reef and deep marine waters of the basin.

10.5.2: Major Depositional Environments

Carbonate sediments primarily form in relatively shallow water environments (meters to tens of meters) within the subtidal carbonate factory, where sediment is transported both basinward and shoreward. In addition to these regions, carbonate accumulation occurs in relatively shallow water reef and peritidal environments, while deepwater carbonate deposition can occur where calcareous skeletal material accumulates.

10.5.2.1: Peritidal Areas

Peritidal carbonates form in environments influenced by tides and the areas immediately adjacent to them. These include supratidal environments (only inundated during storm events), intertidal environments (regularly exposed and submerged by the tides), and shallow subtidal environments (just below the low tide level).

The most proximal (updip) areas experience extended periods of subaerial exposure and may consist of pebble- to boulder-sized carbonate breccias with a matrix that varies from (lime) mudstone to siltstone or sandstone. Supratidal to upper intertidal areas are more frequently inundated and may be composed of lime mudstone with variable amounts of interbedded skeletal wackstone to packstone. Lime mudstones may be planar laminated or have crinkly algal laminae; mudcracks and intraclasts can be present locally.

Intertidal environments represent low-energy, sheltered settings that occasionally experience high-energy storm events. Common facies include laminated mudstones with variable amounts of intraclastic, skeletal and/or peloidal wackestone to grainstone. Mudcracks, fenestrae, crinkly algal laminae, pisoids, peloids, and intraformational conglomerates may be abundant.

Shallow subtidal areas, which transition into lagoonal environments, can contain mudstone to packstone facies with rip-up clasts, peloids, and fossils present locally. Herringbone cross-beds and variably oriented ripple cross-laminae may be present and record reversing tidal currents. Thin quartz sandstones or arenaceous carbonates are present in some successions.

In all of these settings, dolomitization can be pervasive and may destroy or overprint primary textures.

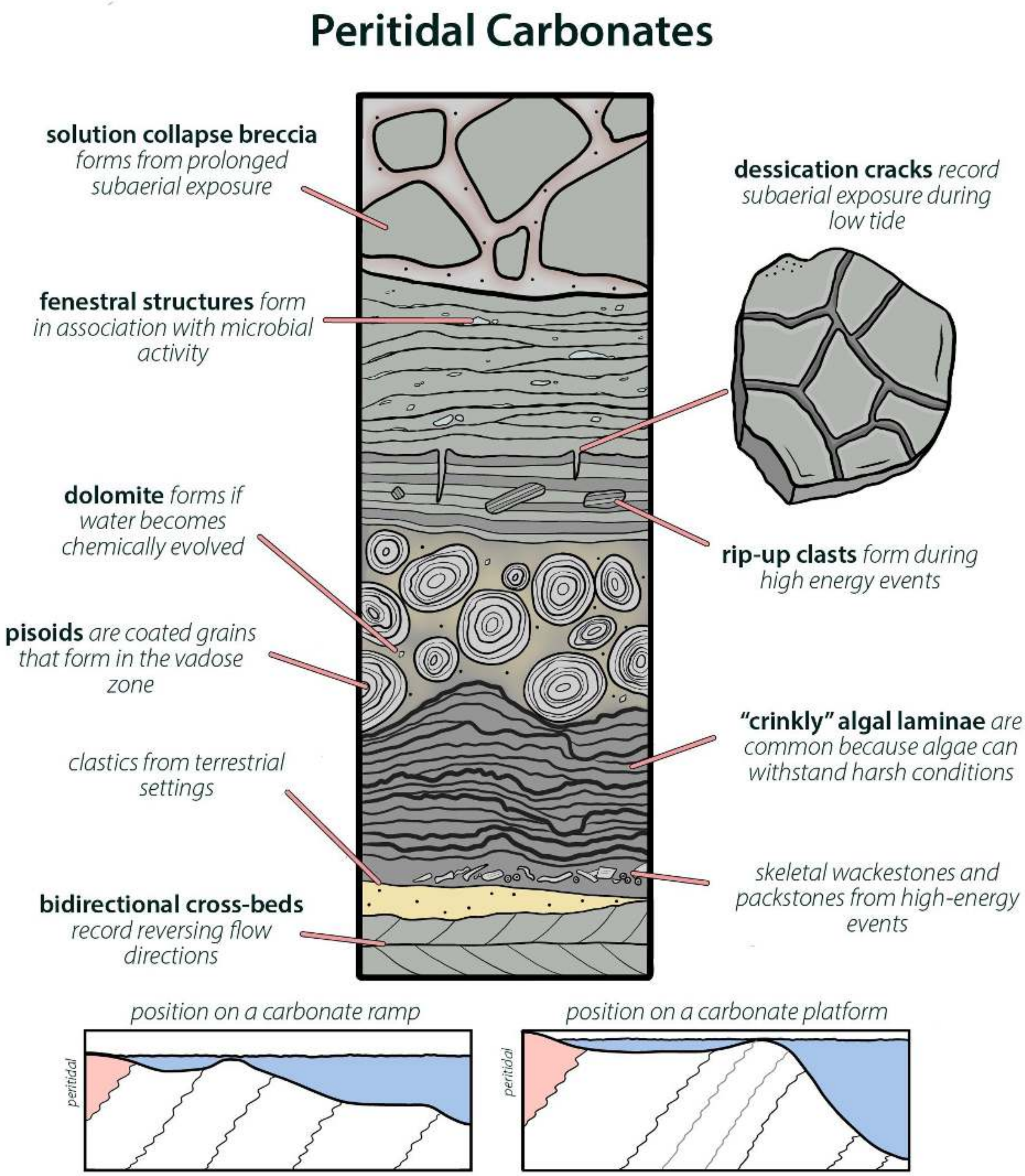

Figure 10.5.3: Sedimentology of peritidal carbonate environments (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

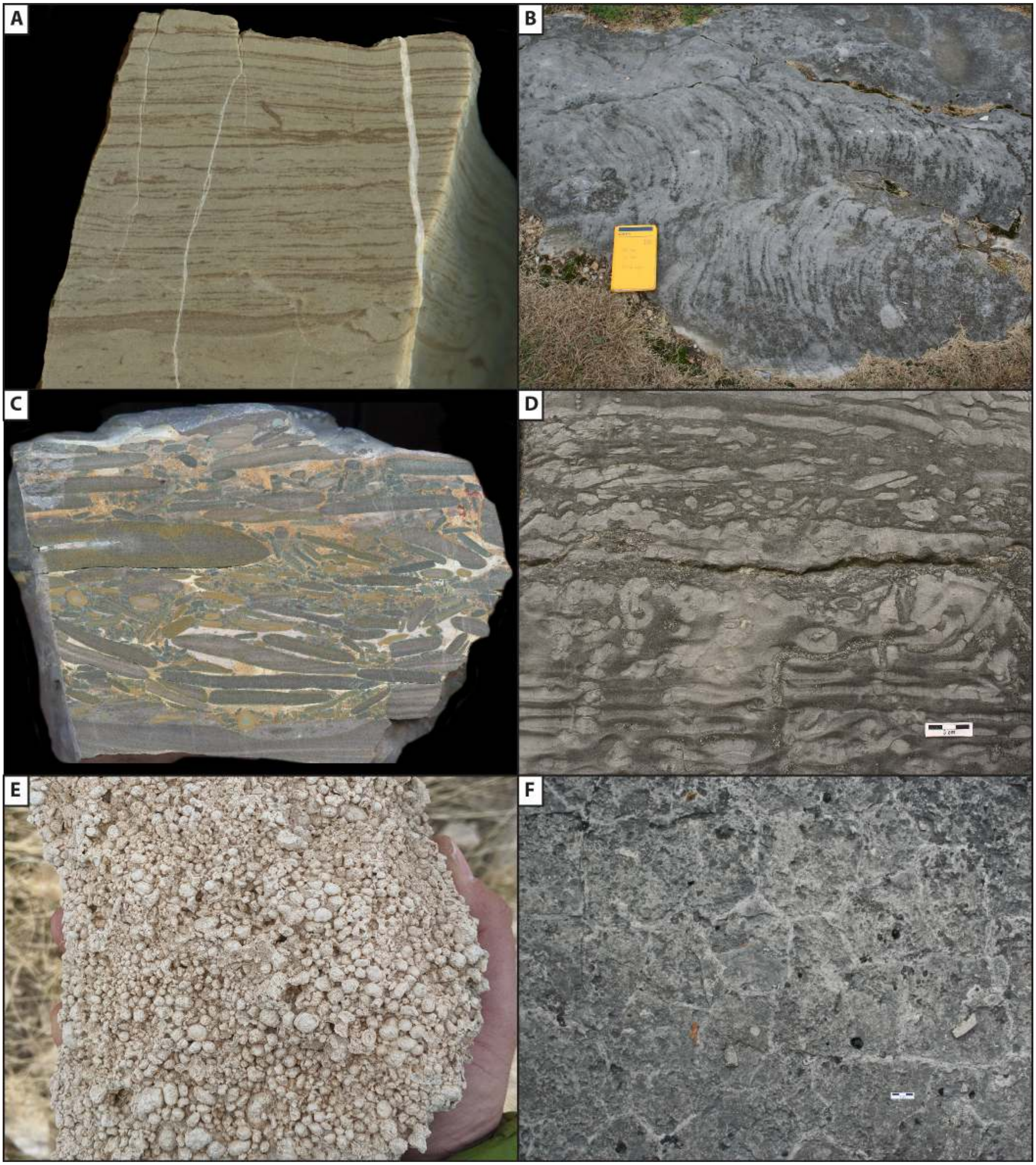

Figure 10.5.4: Peritidal carbonate deposits, Part 1. A) Heterolithic lime mudstone; sample is ~10 cm across. B) Bedding plane view of trough cross-beds recording flow in opposite directions. C) Imbricated lime mudstone intraclasts in a grainstone; the green color is from glauconite. Sample is 18 cm across. D) Intraclasts and oncoids in a packstone. E) Pisoid grainstone. F) Bedding plane view of mudcracks in a limestone. All images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0.

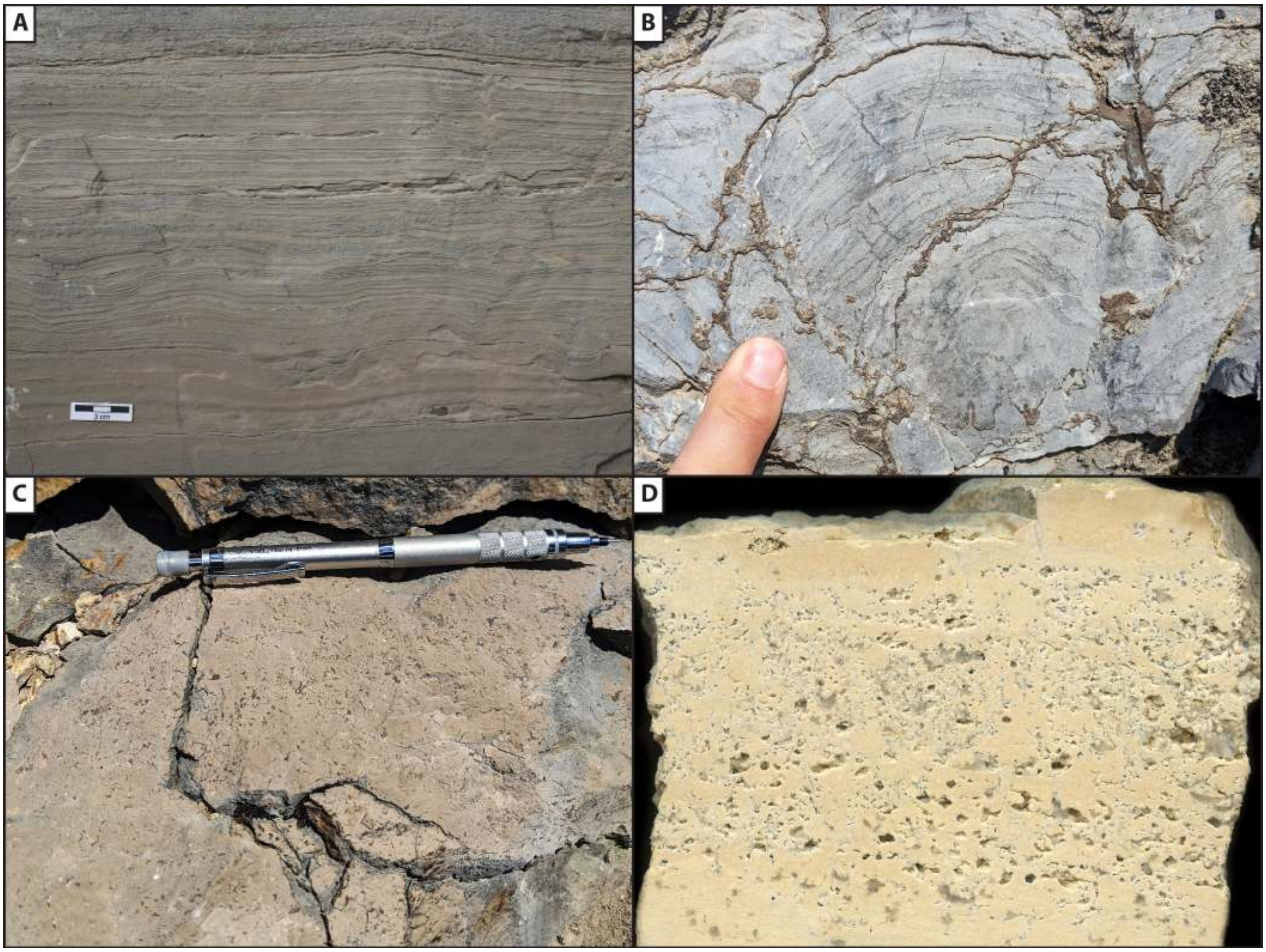

Figure 10.5.5: Peritidal carbonate deposits, Part 2. A) "Crinkly", likely algally-laminated lime mudstone B) Stromatolite in Glacier National Park, C) Fenestral limestone in the Mississippian Lodgepole Formation, and D) Fenestral limestone, Permian Reef complex, New Mexico. Field of view is ~15 cm. All images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0.

10.5.2.2: Lagoons

Lagoons are low-energy environments located on the landward side of reefs or shoals. Although entirely subaqueous, their limited connection to the open ocean makes them susceptible to oxygen and salinity stress. Deposition is primarily by suspension and organic-rich lime mudstones are very abundant, but peloidal grainstones can be abundant locally and intraclast and/or skeletal grainstones record washover events during storms.

Both trace and body fossil abundance and composition are highly variable, though low-diversity assemblages are most common. While scattered marine fossils may be present, organisms tolerant of stressed conditions (ostracods, calcispheres, etc.) tend to dominate. Algal laminae, stromatolites, and/or stromatoporoids may be present locally.

Overall, lagoons are typically preserved as organic-rich lime mudstones to wackestones, which are internally massive to faintly laminated. Bioturbation ranges from scattered to abundant. Dolomitization can be pervasive in lagoons that become chemically evolved.

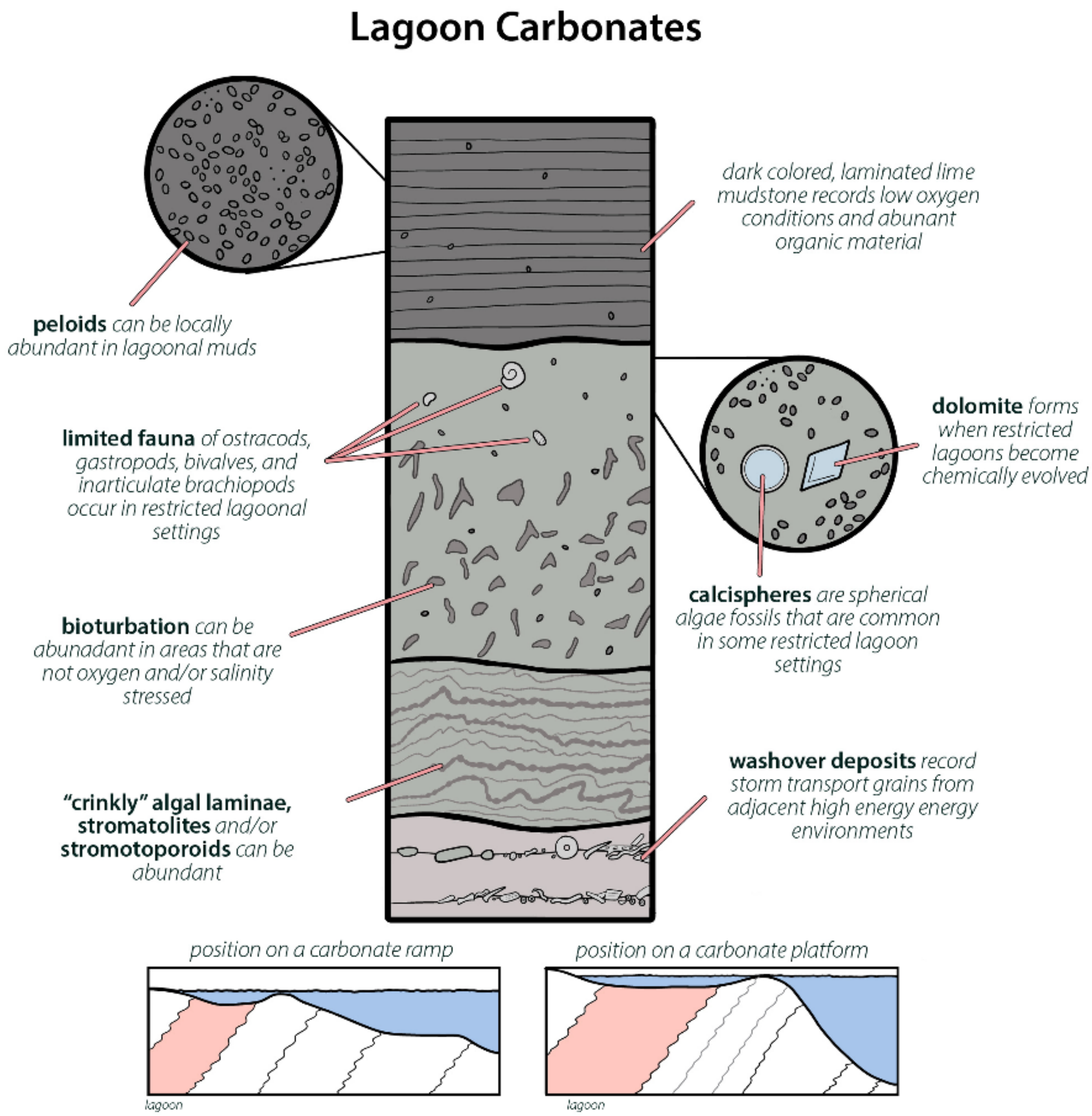

Figure 10.5.6: Sedimentology of lagoonal carbonate environments (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

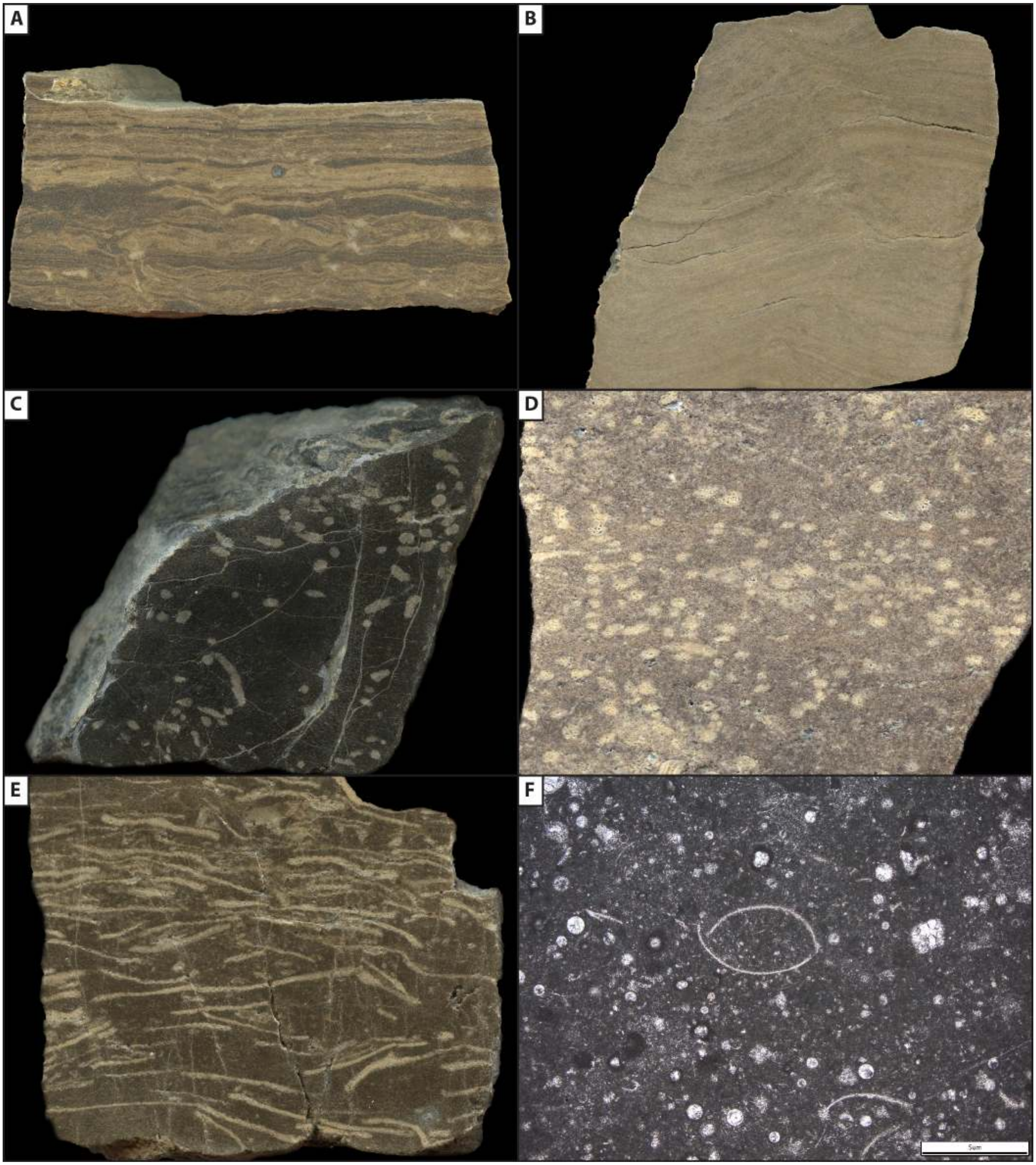

Figure 10.5.7: Photographs of carbonate facies deposited in lagoons. A) Heterollitic dolomudstone from the peritidal margin of a lagoon. B) "Crinkly", likely algally-laminated dolomudstone C) Possible burrows in an organic-rich, internally massive dolomudstone, D) Burrows in an internally massive dolomudstone, E) Amphiphora (stromatoporoids) in the Jefferson Formation (Devonian) near Sacagawea Peak, Bridger Range, Montana. F) Photomicrograph (PPL) showing ostracods and calcispheres in a lime mudstone from the Mississippian Lodgepole Formation in SW Montana. All images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0.

10.5.2.3: Shelf/Platform Interior

Shelves and platforms are both very low-gradient shallow marine environments; the margin of shelves and platforms can be in direct communication with the open ocean (unrimmed) or can be bounded by a reef or shoal (rimmed). Shelves can be adjacent to continental crust and represent a transition to the open ocean or form the interior of platforms. Platforms are isolated and surrounded by marine waters.

The spectrum of environments and processes on shelves can be extremely variable depending on local conditions. Carbonate deposition can occur seaward of, and pass laterally into, clastic-dominated areas on the margins of continental shelves. Common facies include lime mudstones, wackstones and packstones with largely *in situ* skeletal material, storm-generated wackestone to packstone beds with abraded skeletal material, as well as fossiliferous sandstones, siltstones, and shales.

Higher energy shelves have been described from epicontinental seas and the interior portions of isolated platforms. Common facies include grainstones that may contain a variety of transported grains, including skeletal fragments, ooids, pisoids, peloids, and/or intraclasts. Grainstones are commonly cross-bedded and may contain isolated larger grains transported during storms. In shallow, protected areas near the shore or behind shoals/reefs, fenestral-pisoid laminated facies and tepee complexes may be present. Scattered sandstones and/or siltstones may be present.

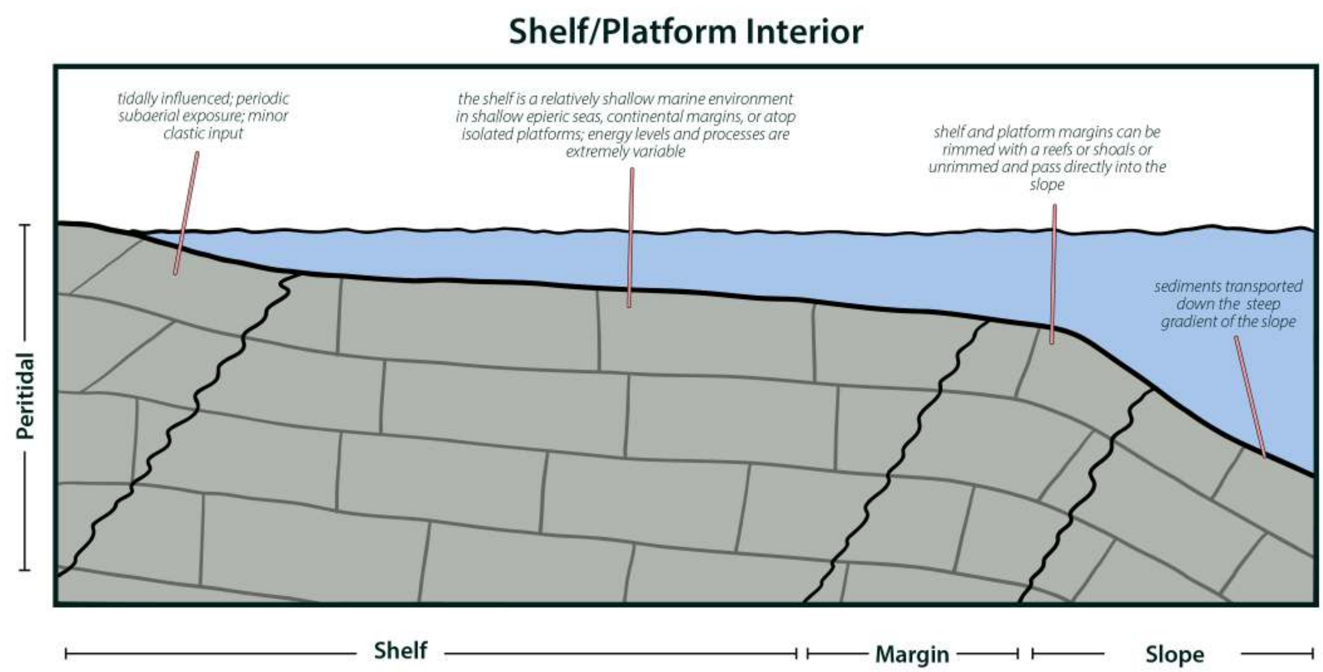

Figure 10.5.8: Schematic cross section of a very gently dipping shelf adjacent to a continental margin; comparable environments also occur in th einterior of some isolated carbonate platforms. From Page Quinton via Wikimedia Commons; CC BY-SA 4.0.

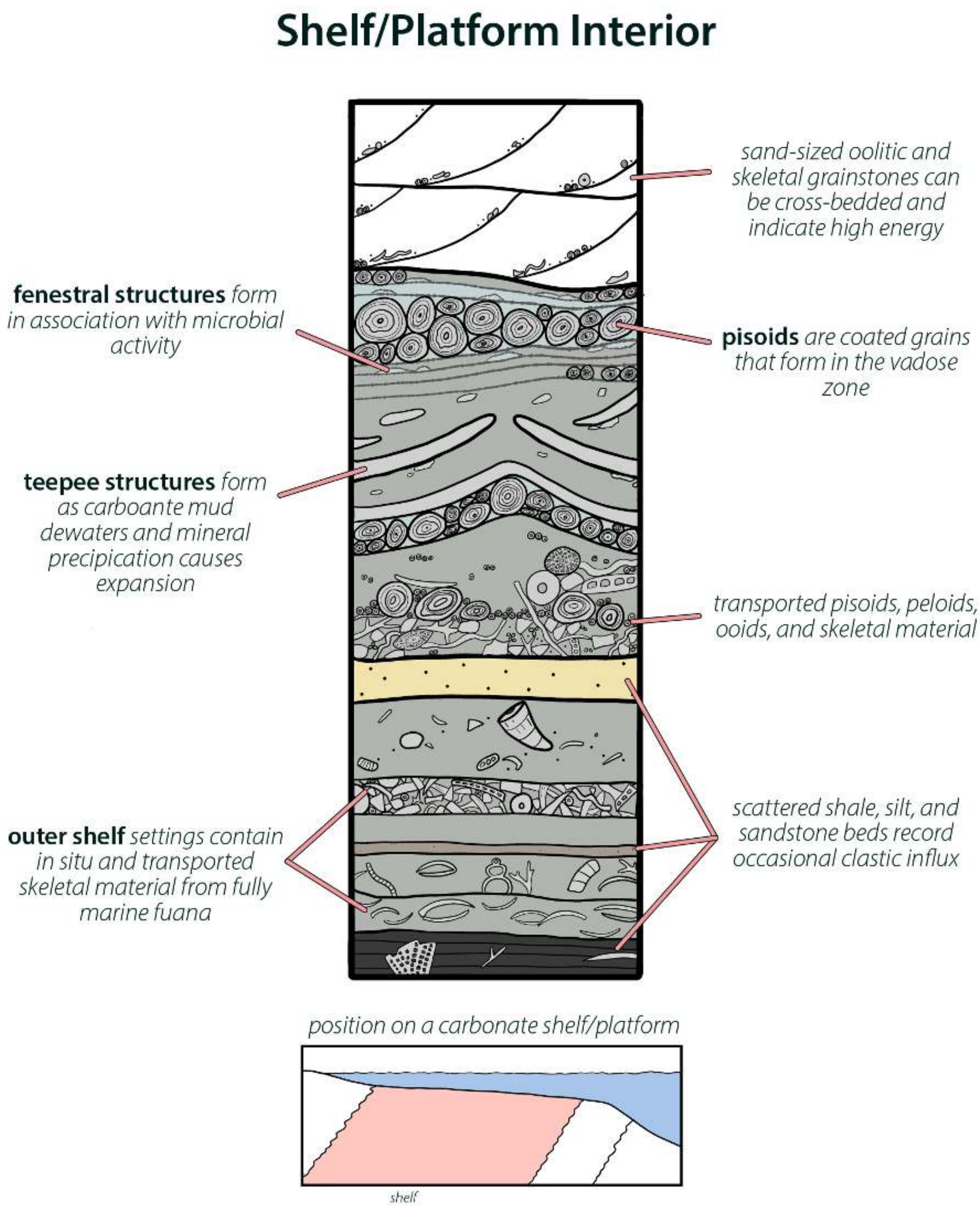

Figure 10.5.9: Sedimentology of carbonate shelf and platform interior areas (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

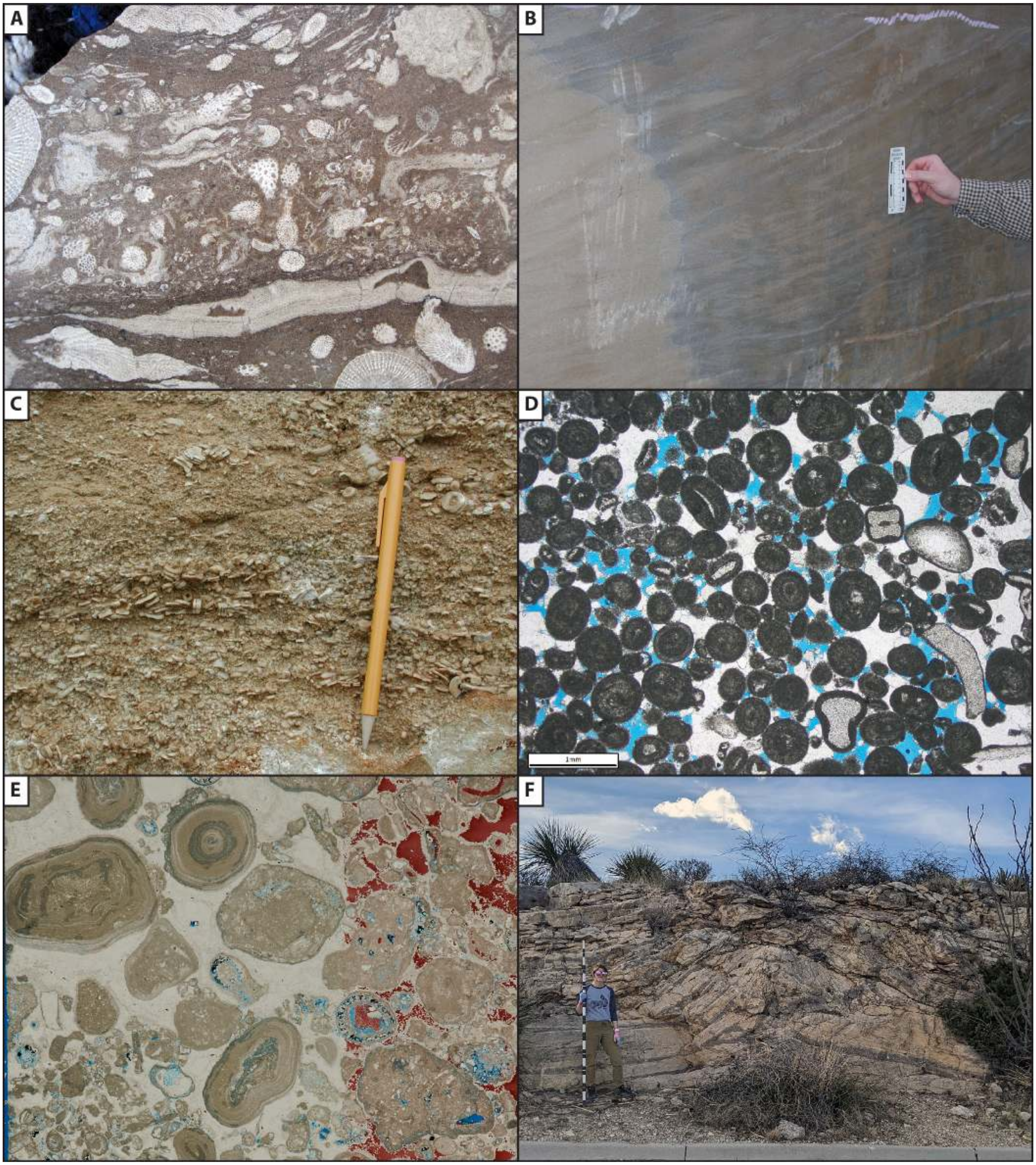

Figure 10.5.10: Photographs of carbonates deposited on shelves or platform interiors. A) Skeletal wackestone to packstone; unit and location uncertain but possibly from the Devonian of southern Indiana (James St. John via Wikimedia Commons; CC BY 2.0). B) Cross-bedded skeletal grainstone in the Salem Limestone, central Indiana. C) Detail of a crinoid-rich skeletal grainstone. D) Photomicrograph of A Pleistocene oolitic sandstone. E) Transported pisoids, intraclasts and ooids in a wackestone to packstone in the shelf crest of the Permian Reef complex. F) Tepee structure in shelf crest strata of the Permian Reef complex exposed in the parking lot of the Carlsbad Caverns National Park visitor center. All images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0.

10.5.2.4: Shoals

Carbonate shoals are high-energy environments where transported skeletal fragments and/or ooids accumulate above fair-weather wave base. While shoals and reefs occupy similar bathymetric positions, reefs consist of *in situ* accumulations of wave-resistant organisms, whereas shoals are composed of transported sediment. Both can form along broad shelf margins or isolated platforms, provided clastic input is minimal and fully marine conditions with abundant oxygen, energy, and warm temperatures are present. Although modern examples exist (e.g., the Bahama Banks), they were far more common in the geologic past when epeiric seas flooded continental crust, and global climates were warmer.

Shoals and reefs act as barriers, separating calm lagoons or shelf environments from more energetic open-marine conditions. Shoal development follows processes similar to those of a clastic shoreface. The upper shoal is well above fair-weather wave base, consists of shifting oolitic and/or skeletal sand (grainstone) with sedimentary structures such as cross-beds, ripple cross-laminae, low-angle laminae, cross-cutting erosion surfaces, and rip-up clasts. More distal foreshoal deposits, also above fair-weather wave base, contain more interbedded skeletal to oolitic packstones and wackestones. Bioturbation increases in abundance and diversity compared to the upper shoal. While sedimentary structures are similar, cross-beds become less common and may be replaced by hummocky cross-strata.

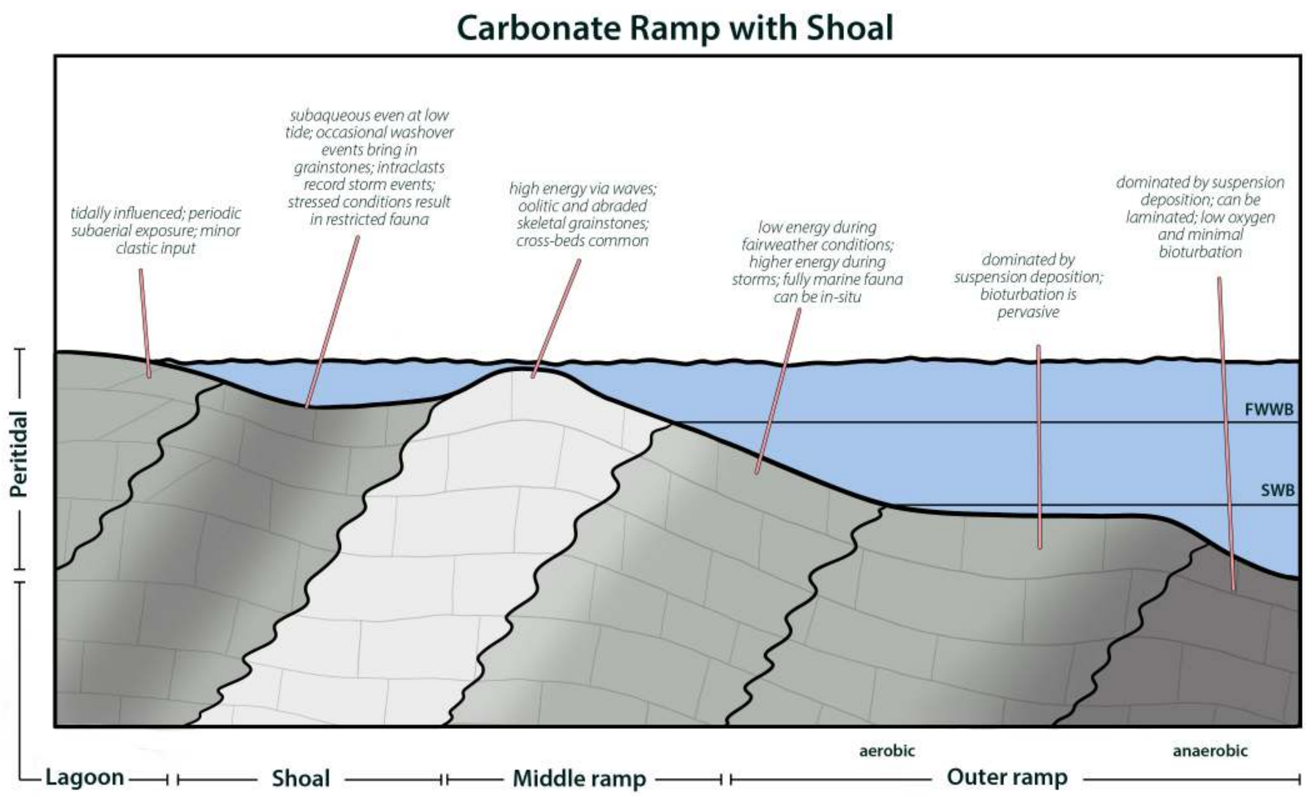

Figure 10.5.11: Schematic cross section of a relatively gently dipping carbonate ramp with a shoal developed above fair weather wave base (FWWB). From Page Quinton via Wikimedia Commons; CC BY-SA 4.0.

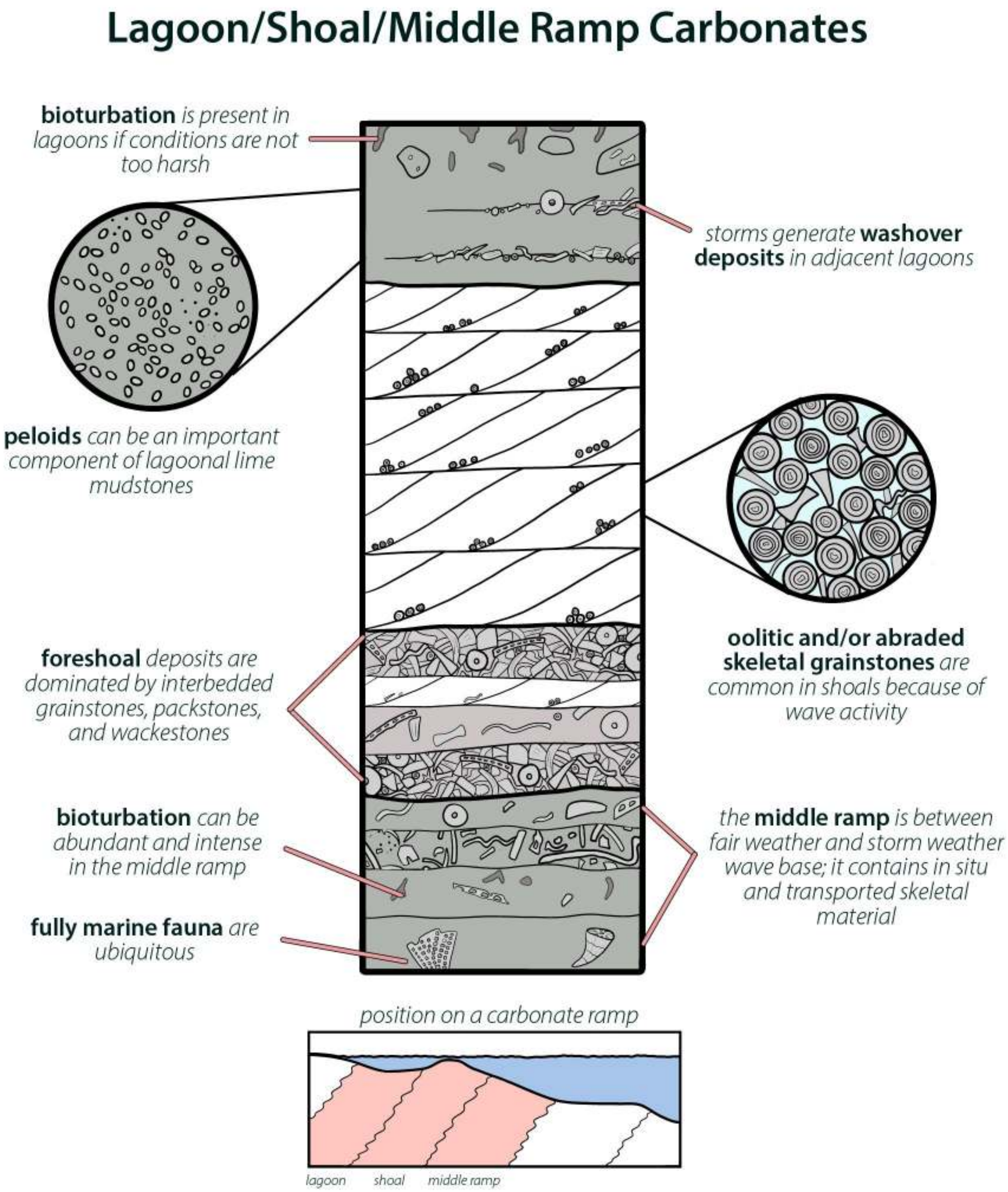

Figure 10.5.12: Sedimentology of a lagoon to shoal to middle ramp carbonate environments (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

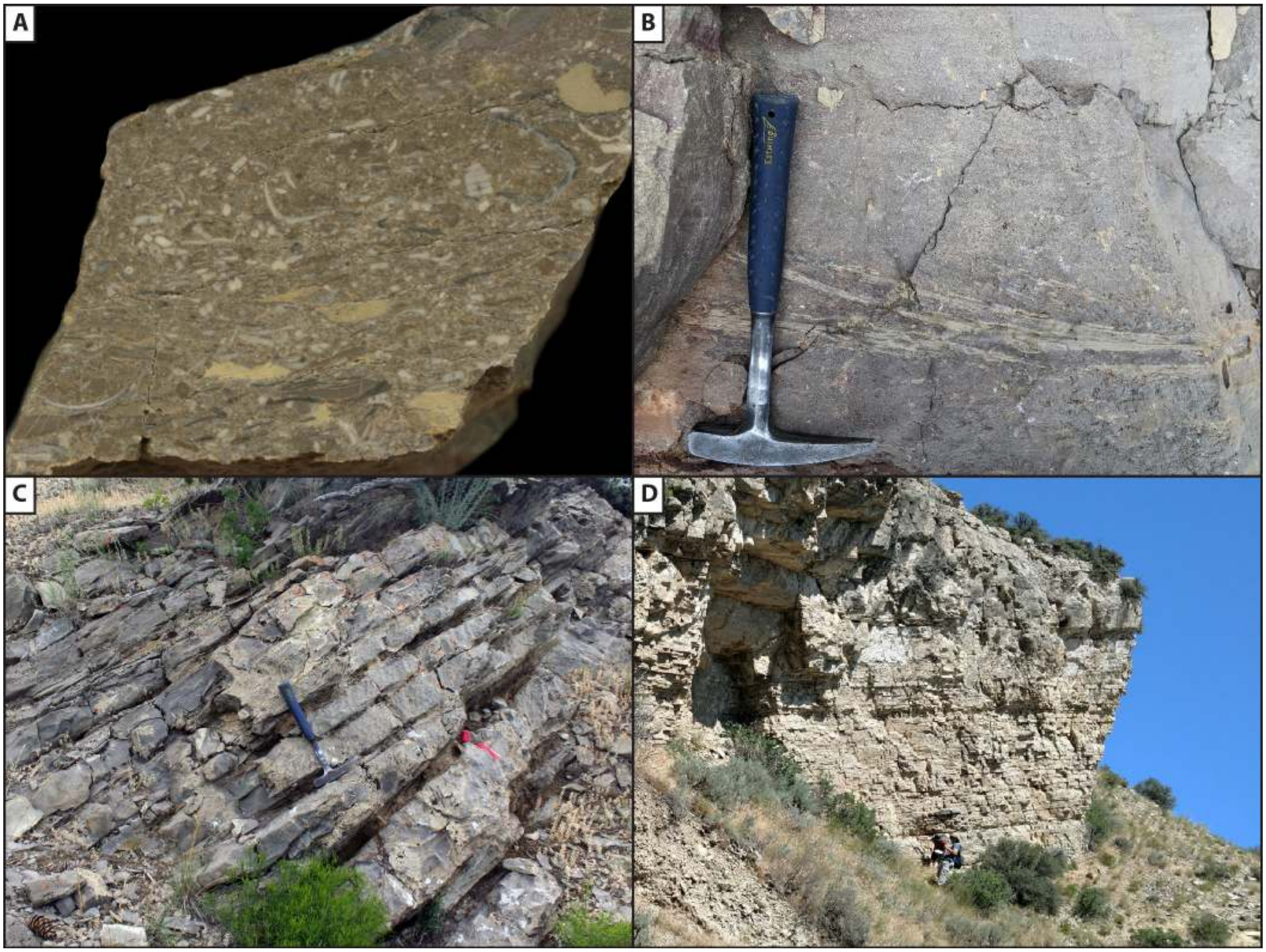

Figure 10.5.13: Photographs of carbonates deposited in shoal or middle ramp environments. A) Polished slab of a skeletal grainstone deposited in shoreface water depths on a shoal, Lodgepole Formation, Montana. B) Outcrop photograph of a cross-bedded skeletal grainstone. C) Interbedded wackestones, packstones, and grainstones deposited near the transition between a shoal and the middle ramp. D) Shallowing-upward parasequence through foreshoal and shoal deposits, Lodgepole Formation, Montana. All images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0.

10.5.2.5: Reefs

Reefs are wave-resistant, *in situ* accumulations of benthic organisms that thrive in high-energy, shallow-water environments. While they can endure crashing waves and grow rapidly enough to keep pace with sea-level rise, they generally require clear, warm, fully marine waters. The morphology of reef-building organisms is shaped by energy levels within the reef environment, with different organisms contributing in various ways—some form rigid, wave-resistant frameworks, others generate sediment, and some bind and stabilize the seafloor. Reef-building communities have varied over time, including stromatolites, archaeocyathids, corals, stromatoporoids, algae, sponges, and bivalves, among others. Major subenvironments of the reef are described below.

The back reef is a low-energy zone transitional between the reef and protected areas (ex: lagoon) behind it. It typically has a muddy substrate, sometimes with domal or robust branching organisms. A low-gradient reef flat, composed of storm-transported clasts cemented into a pavement, may separate it from the reef crest. The reef crest extends from the high tide mark to several meters below, this zone varies with energy levels. In extreme wave conditions, it consists of sheet-like encrusting organisms adapted to exposure and intense wave action; in calmer settings, robust branching forms dominate. The reef front contains a diverse collection of organisms that display a systematic progression of forms including massive corals near the reef crest, branching forms near wave base, and platy forms extending to the lower limit of the photic zone.

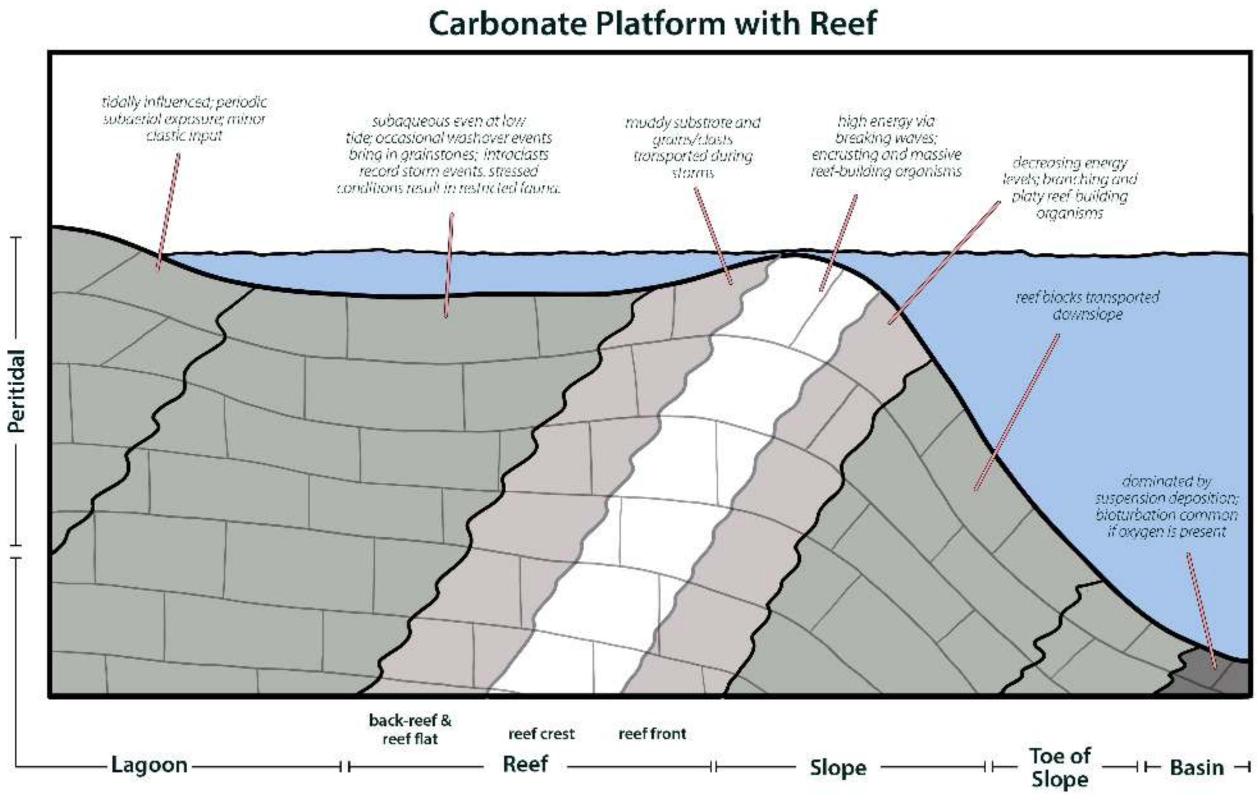

Figure 10.5.x14: Schematic cross section of a reef and a steeply dipping slope (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

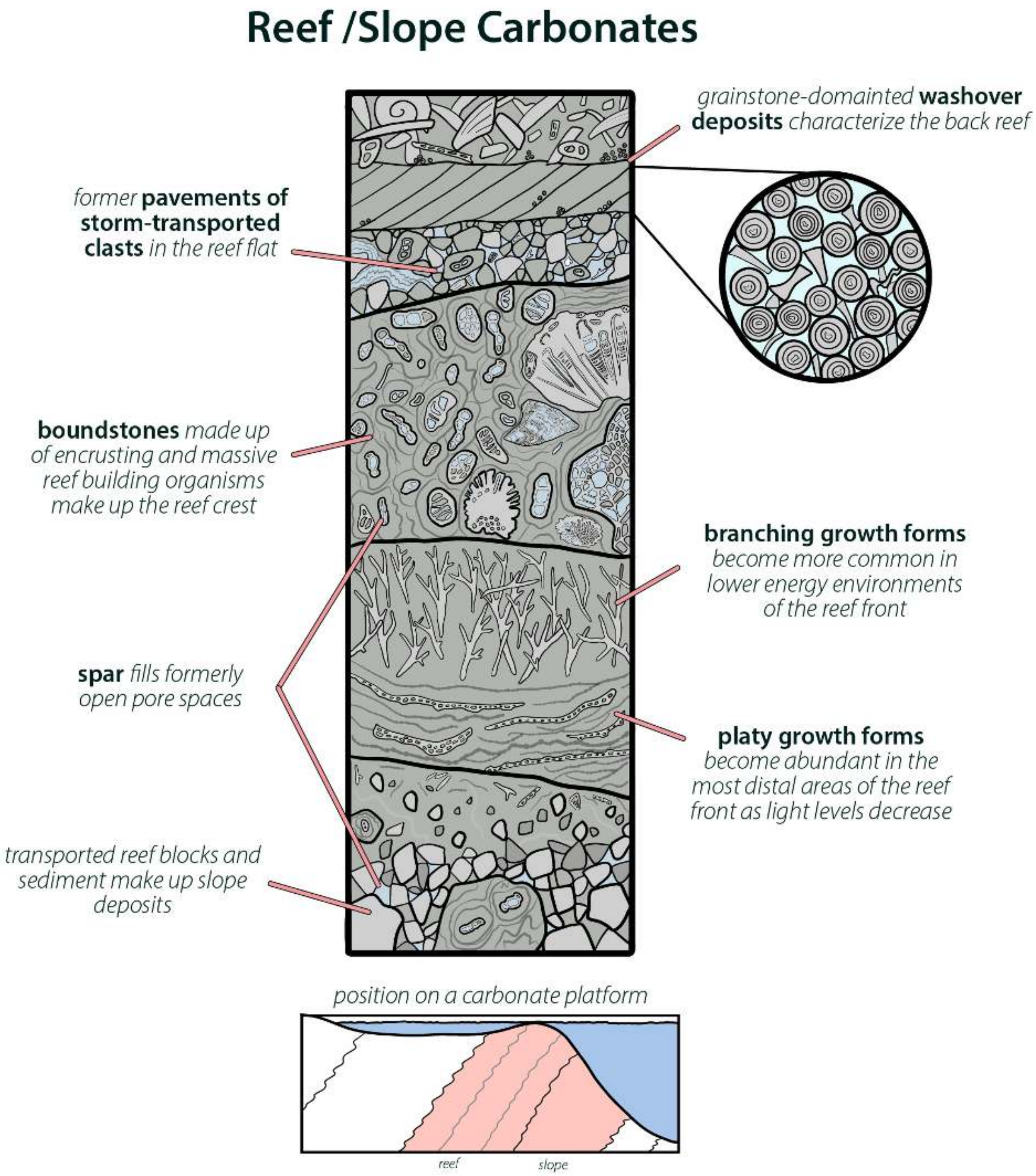

Figure 10.5.15: Sedimentology of reef and slope carbonate environments (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

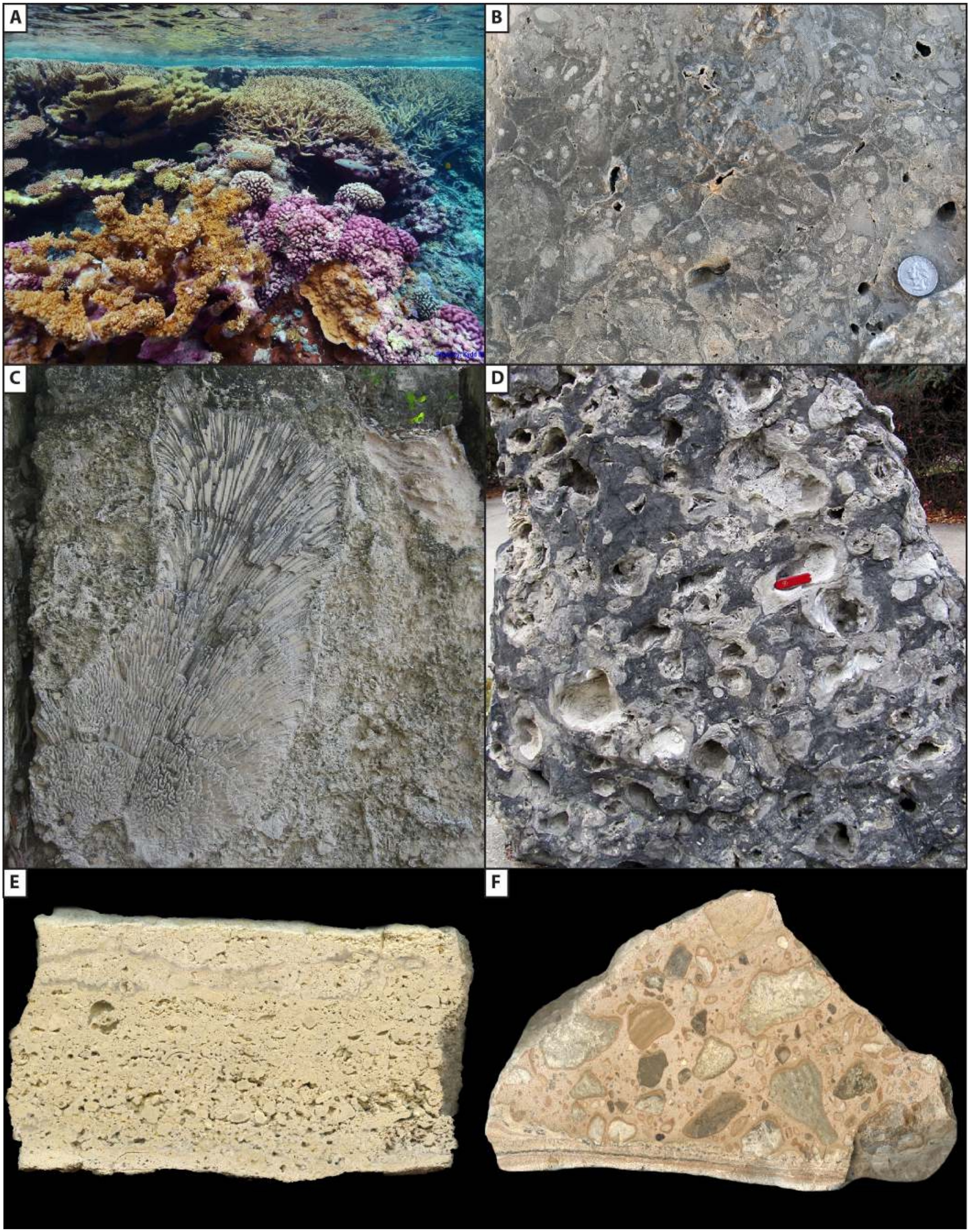
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Figure 10.5.16: Carbonate deposits associated with reefs. A) Modern reef (USFWS - Pacific Region via Flickr; CC BY-NC 2.0), B) Sponge boundstone block transported downslope from the Permian Reef Complex. The reef complex was entombed in a thick succession of evaporites when the basin became restricted and modern erosion exposes the amount of original relief on the platform. The Capitan Limestone is the core of this sponge-dominated reef and it prograded out over the slope and deep basinal deposits of the Cherry Canyon through Bone Spring Formations. The amount of topographic relief present today approximates the amount of bathymetric relief when the reef was active. C) Fossil brain coral (*Diploria*) at the Windley Key Fossil Reef Geological State Park. US Quarter near top for scale (Jstuby via Wikimedia Commons; public domain), D) Stromatoporoid reef from the Devonian Cairn Formation (Georgialh via Wikimedia Commons; CC BY-SA 4.0). E) Fenestral limestone and F) carbonate breccia deposited in a backreef setting, Permian Reef Complex, New Mexico (Michael Rygel via Wikimedia Commons; CC BY-SA 4.0)

10.5.2.6: Middle Ramp/Slope

Depending on geomorphic and plate tectonic settings, shoals and reefs transition seaward into either a middle ramp on gently dipping shelves or steeply dipping slopes along the margins of rimmed shelves or platforms.

The middle ramp extends from fair-weather wave base to storm wave base, comparable to the offshore transition (inner shelf) zone of clastic coasts. Common lithologies include lime mudstone interbedded with skeletal wackestones to packstones and thin grainstones. Skeletal material accumulates *in situ* in wackestones to packstones, while grainstones likely form from downslope sediment movement during storms. Fully marine faunas are present, and bioturbation is generally more intense and diverse than in shoals.

Steeply dipping slopes lie seaward of reefs or shoals on rimmed shelves or platforms, extending from the photic zone to depths of tens to hundreds of meters before transitioning to the low-gradient basin floor. These carbonate slopes contain fine-grained carbonate sediment that settled from suspension and coarser debris transported downslope from reefs or shoals. Common facies include interbedded lime mudstones, wackestones/packstones, and rudstones deposited by debris flows or turbidity currents, with occasional clastic sandstones or siltstones. Much of the skeletal material is transported downslope, and bioturbation varies from absent to intense. At the distal toe of slope, deep basin deposits alternate between lime mudstones formed by suspension deposition and rudstones from turbidity currents and debris flows.

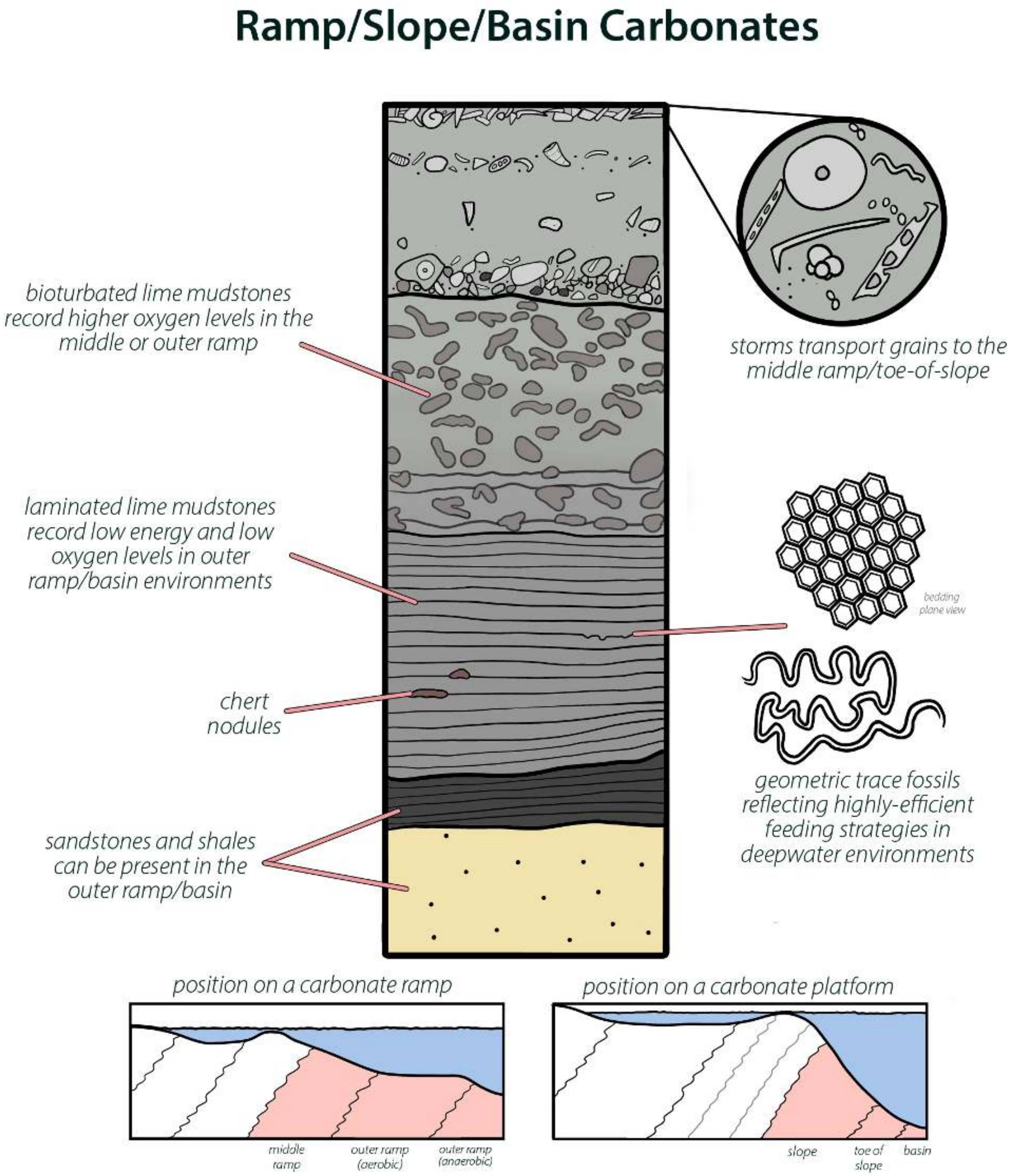


Figure 10.5.17: Sedimentology of middle ramp, outer ramp, toe of slope, and basin carbonate environments (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

Figure 10.5.18: Carbonate facies associated with the middle ramp (A-C) or slope (D-F). A) Interbedded shales, skeletal lime mudstone and skeletal wackestone. B) Skeletal packstone. C) Bedding plane expsoure of fossils in a skeletal packstone found beneath the rocks shown in A. A through C are all from the Mississippian Lodgepole Formation, Montana. D) Laminated and bioturbated lime mudstones, E) flow imbricated foraminifera, and F) transported fragments of reef boundstones. D through F are from the Permian Reef complex in west Texas. All images from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0.

10.5.2.7: Outer Ramp/Basin

As per the discussion in the abyssal plain and carbonate compensation depth sections in 10.4: Clastic Marine Environments, carbonate accumulation in deep marine outer ramp or basin environments primarily results from the suspension deposition of calcareous plankton skeletons (calcareous ooze), with occasional gravity-driven transport from the slope. Deposition of carbonate only occurs above the carbonate compensation depth. Common facies include laminated lime mudstones and/or chalk, often interbedded with fine-grained clastics and bedded cherts. Lime mudstones may contain microfossils, sparse bioturbation (typically *Zoophycos* or *Nereites* Ichnofacies), and peloids. Deep marine carbonates are often organic-rich and can serve as significant hydrocarbon source rocks.

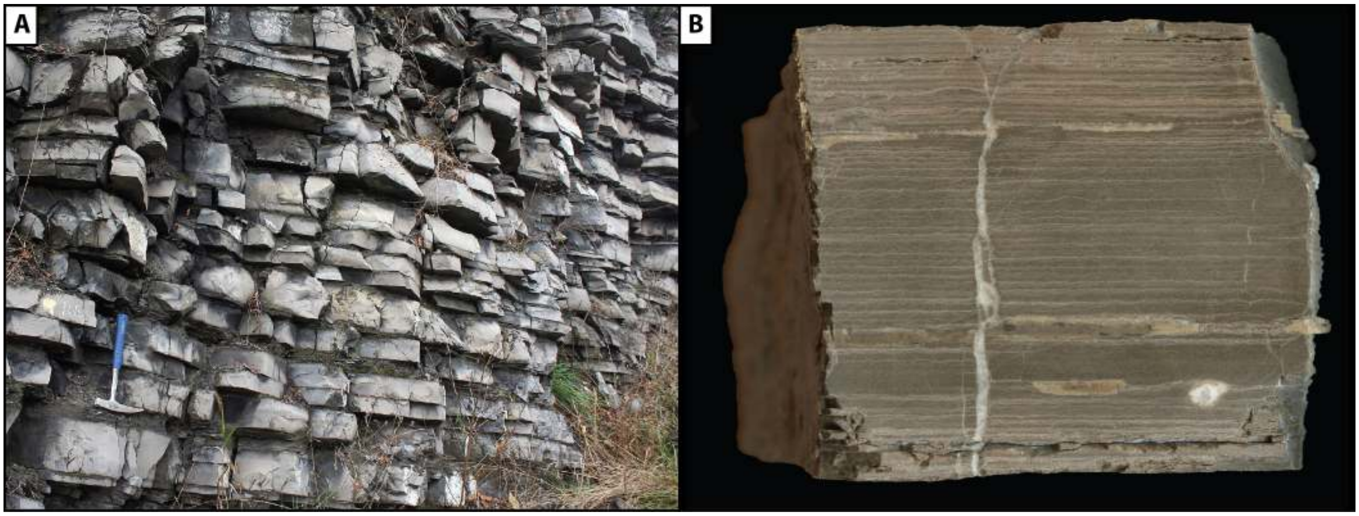


Figure 10.5.19: Carbonate facies associated with basin (deep water) environments. A) Internedded lime mudstone and black shale, Union Furnace section. B) Laminated lime mudstone from the Williams Ranch Road locality, west Texas. Both images from from Michael Rygel via Wikimedia Commons; CC BY-SA 4.0.

10.5.3: Readings and Resources

- Carbonate World website - https://carbonateworld.com/
- James, N.P. and Jones, B., 2015, Origin of Carbonate Sedimentary Rocks, Wiley, 464 p.
- Laugié, M., Michel, J., Pohl, A. et al. Global distribution of modern shallow-water marine carbonate factories: a spatial model based on environmental parameters. Sci Rep 9, 16432 (2019). https://doi.org/10.1038/s41598-019-52821-2; CC BY 4.0.

10.6: Glacial Environments

10.6.1: Glacial Influence

Glacial environments do not form a single, coherent depositional system but instead interact with and overprint existing systems such as fluvial, lacustrine, and eolian environments. While glacially-overprinted environments closely resemble warmer-climate versions of those same environments, diagnosis of glacial influence depends on recognizing the diagnostic features described below.

Figure 10.6.1: Glacial striations on bedrock in the Swiss Alps; lens cap for scale (Michael Rygel via Wikimedia Commons; CC BY-SA 3.0).

10.6.2: Glacial Indicators

One of the more widespread glacially influenced deposits is loess, composed of wind-blown silt. These fine-grained deposits accumulate downwind of glaciated regions, often forming extensive blankets of sediment. Once lithified, loess appears as a nondescript siltstone, making it difficult to distinguish from other silt-rich deposits unless it is found in association with other glacial indicators.

A more direct glacial deposit is glacial till, which lithifies into diamictite over time. This deposit consists of a poorly sorted mixture of mud, sand, and gravel, accumulating beneath or in front of a glacier. The lack of sorting and the mix of clast sizes distinguish it from other deposits, but care must be taken in interpretation, as similar textures can result from debris flows. A key diagnostic feature of glacial till is the presence of striated, faceted, or bullet-shaped clasts, which form due to abrasion within moving glacial ice, incorporation of faceted ventifacts, and dynamics within glacial ice, respectively.

Varves are annual couplets of sand and mud, recording alternating high-energy summer deposits and finer winter deposits when the lake is ice-covered. While varves provide a clear record of cold climate, similar rhythmic layering can occur in non-glacial settings such as tidal flats or fluvial systems with pulses of floodwater.

Ice-rafted debris serves as an additional indicator of glacial, or at least freezing conditions. As icebergs calve from sediment-laden glaciers, they can drift into open water and release debris as they melt. This process results in outsized clasts that are encased within fine-grained sediments. Mudrocks formed in cold climates might also contain glendonites, which are pseudomorphs of the mineral ikaite which forms only in near freezing conditions.

Ancient bedrock surfaces that were smoothed and polished by the motion of glacial ice and ornamented with striations and chatter marks created by transported clasts within the ice are common in modern glacial environments and have been reported from the ancient record.

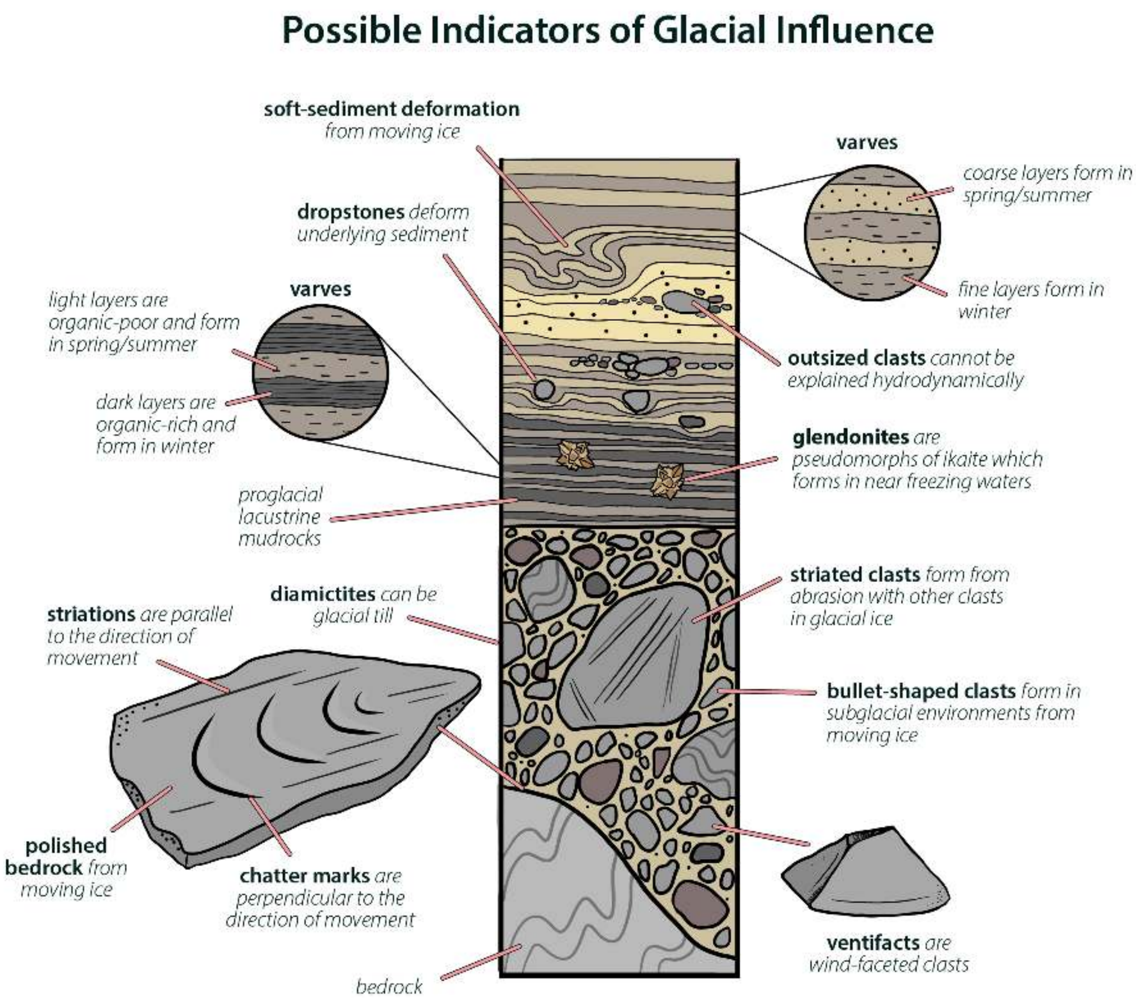


Figure 10.6.2: Possible indicators of glacial influence that might be preserved in the geologic record (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

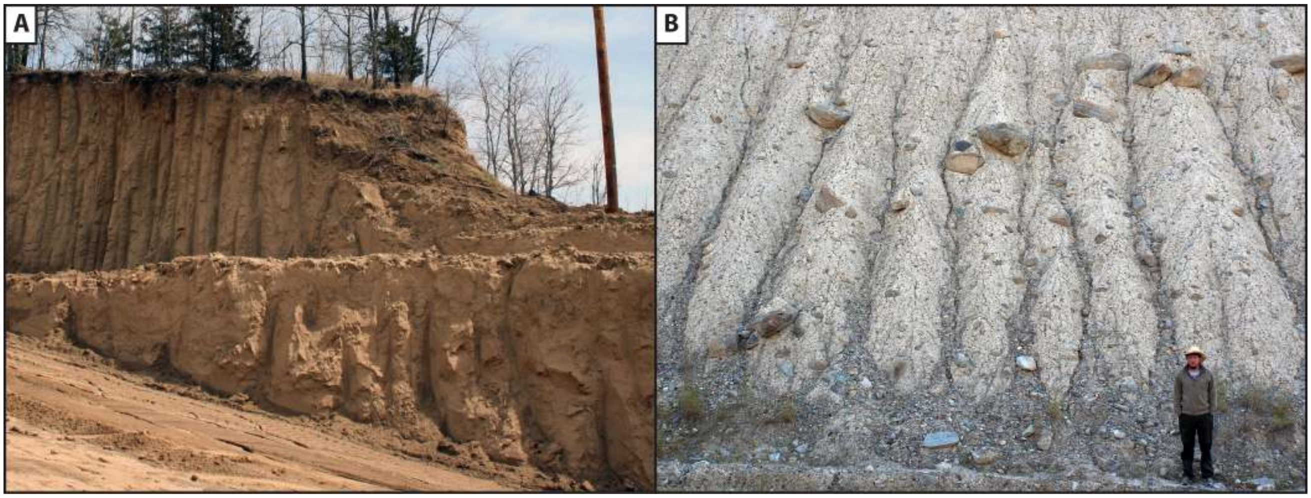


Figure 10.6.3: Deposits from the last glacial include loess from Nebraska (Michael Rygel via Wikimedia Commons; CC BY-SA 3.0) and till from northwest Montana (Michael Rygel via Wikimedia Commons; CC BY-SA 3.0).

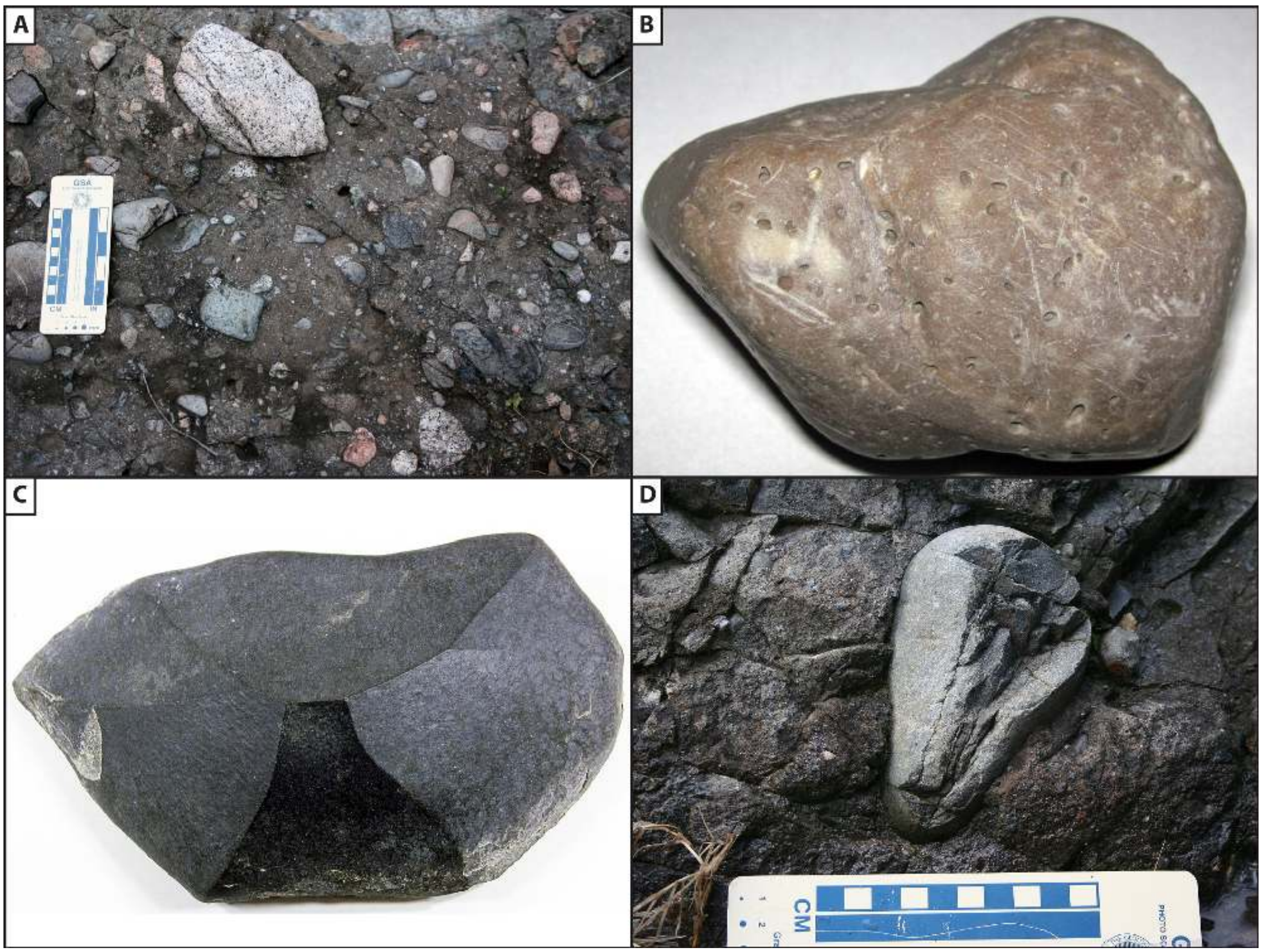


Figure 10.6.4: Diamictites and clasts that are associated with glaciers. A) Diamictite can reasonably be interpreted as glacial till if found in close association with other glacial features and if it contains scratched, faceted, or polished clasts. B) Striated clast (James St. John, Flickr, CC BY 2.0), C) ventifacts can be incorporated into till and preserved as faceted clasts (James St. John, Flickr, CC BY 2.0), D) Bullet-shaped clasts develop in subglacial environments from moving ice. A and D are from Michael Rygel via Wikimedia Commons; CC BY-SA 3.0.

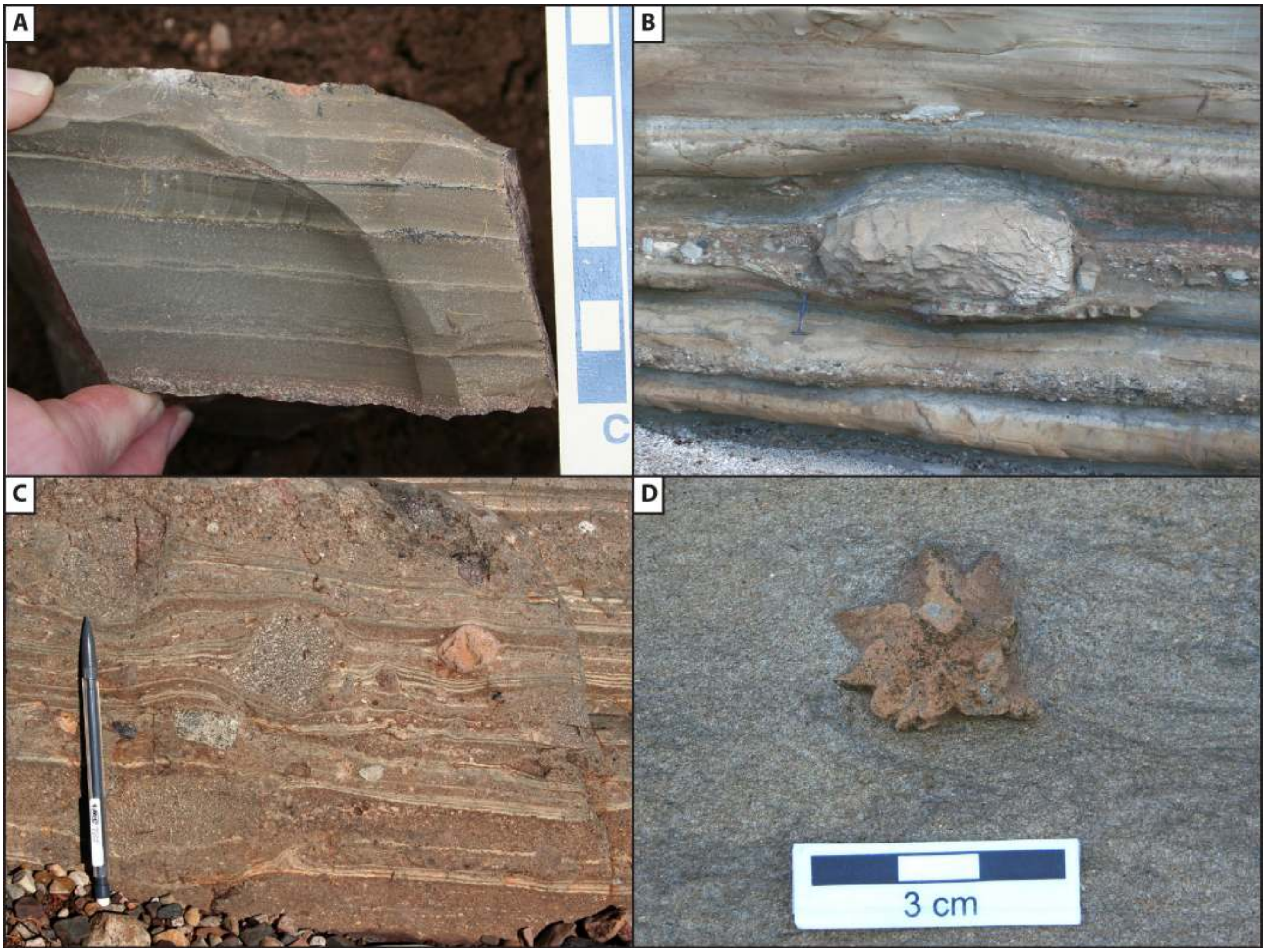


Figure 10.6.4: Glacial/cold climate indicators. A) Varves composed of coarse (light) and fine (greenish gray) couplets. B) Large quartzite boulder deposited as ice-rafted debris in shoreface sandstones. C) Varves and dropstones record ice-rafting into bodies of water that froze in the winter. D) Glendonites in mudstone indicate near freezing conditions. All images from Michael Rygel via Wikimedia Commons; CC BY-SA 3.0.

Figure 10.6.5: The Nooitgedacht Glacial Pavement is an ancient polished and striated bedrock surface formed in association with the Late Paleozoic Ice Age in what is now South Africa. This surface was later exhumed and used as a substrate for rock art during the late Stone Age. Image from Andrew Hall via Wikimedia Commons; Michael Rygel via Wikimedia Commons; CC BY-SA 3.0.

CHAPTER OVERVIEW

11: Sea Level

In the simplest sense, sea level is the elevation of the surface of the ocean. But when we think about it in the geologic sense, things quickly become more complicated as we must consider variations in space, changes in time, and the complex interactions between the surface of the ocean, the Earth's crust, and the rate at which sediment is supplied to an area. In this chapter we will briefly explore all of these topics and think about how they are preserved in the geologic record.

Learning Objectives

- Explain what happens during transgressions and regressions.
- Define water depth, eustatic sea level, and relative sea level.
- Explain how tectonism and eustasy interact to cause increases or decreases in relative sea level.

Chapter thumbnail shows a relative sea level curve (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

11.1: Transgressions and Regressions

11.1.1: Definitions

Changes in the relative sea-level history of an area are recorded in the geologic record by transgressions and regressions. If we had a time machine and were watching what happened in a region or specific location, we could define these terms based on the direction of shoreline movement (landward or seaward), changes in water depth (getting shallower or deeper). A transgression occurs when the shoreline moves in a landward direction and an overall deepening takes place. A regression occurs when the shoreline moves in a seaward direction and overall shallowing takes place.

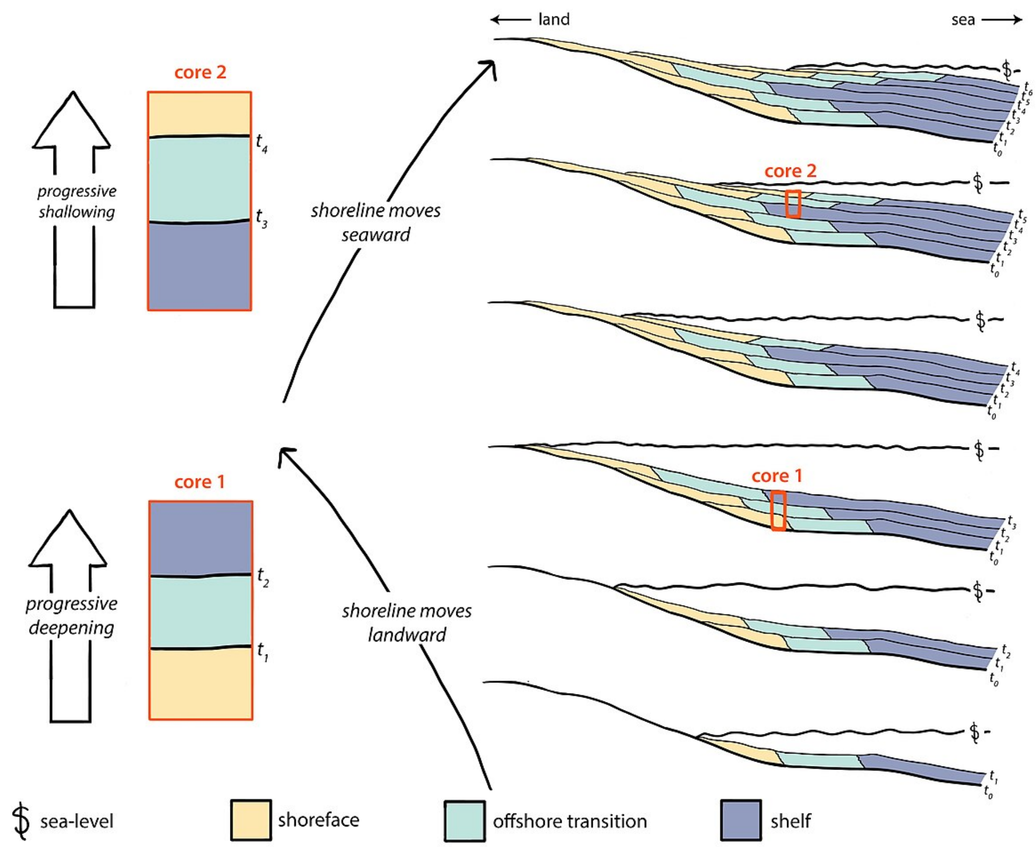


Figure 11.1.1: Overview of transgression (bottom three time slices) and regression (top three time slices) as recorded by changes in water depth, shoreline position, and changes in depositional environments through time (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

11.1.1.1: Walther's Law

Walther's Law states that "facies that occur in conformable vertical successions also occurred in laterally adjacent environments." In the diagram provided above, we have three laterally adjacent environments: the shoreface, offshore transition, and shelf - our studies of the modern confirm this sequence. If all you had was Core 1, you'd see that these facies are stacked atop one another and that things got progressively deeper through time. Given that, it would be reasonable to interpret this as a confirmable succession that records a transgression. In Core 2, we see exactly the opposite ... the facies succession records progressive shallowing which could be reasonably interpreted as a regression in a conformable succession.

Remember that the presence of unconformities violates Walter's Law and can place facies atop one another that were nowhere near each other in a lateral sense.

11.2: Sea Level Terminology

When talking about “sea level”, it's important to make sure that we define some important terms:

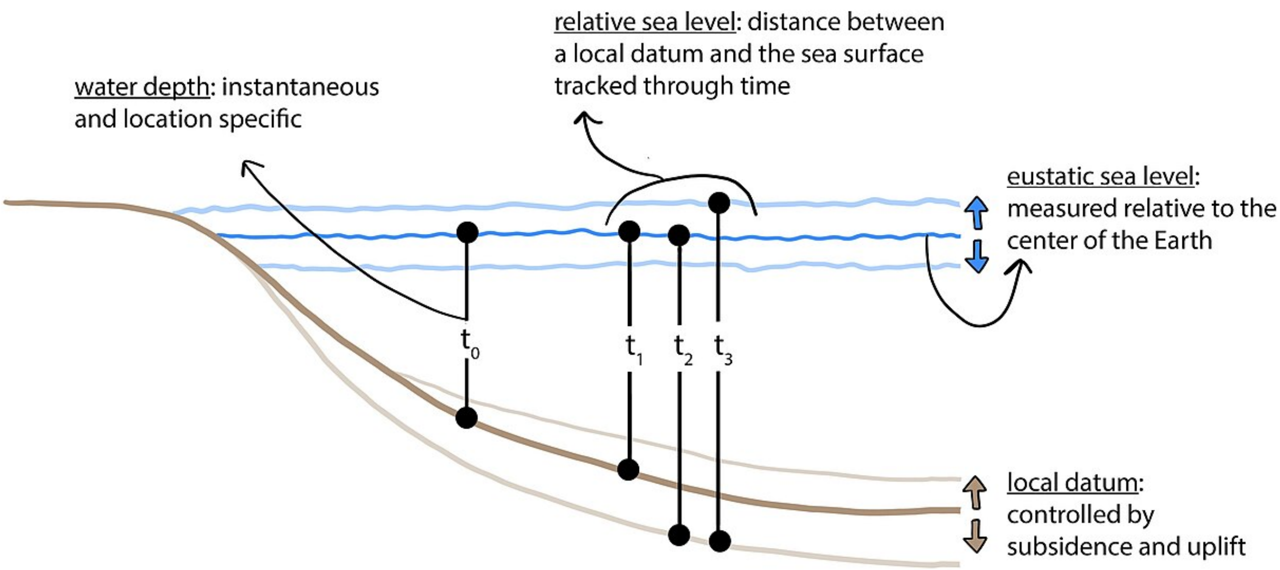


Figure 11.2.1: Overview of sea level nomenclature (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

11.2.1: Water Depth

Water depth is the distance between the sediment surfaces and the water surface. This is an instantaneous measurement that is location specific.

11.2.2: Eustatic Sea Level

Eustatic sea level is the distance between the center of the Earth and the sea surface. Given that the center of the Earth is a universal datum, this allows this value to be measured or calculated globally. You can change global sea level by changing the amount of water in the oceans or changing the amount of water the ocean basins can hold. Drivers of eustatic change include:

1. Thermohaline changes – changes in the volume of water in the ocean because of temperature changes. Over a scale of decades, changes in the temperature of the shallow ocean can change eustatic sea level by centimeters to tens of centimeters. Over centuries to millennia, circulation can warm the deep ocean as well and cause several meters of eustatic change.
2. Glacial ice volume – when large amounts of glacial ice accumulate on land, sea level falls. When glacial ice on land melts and runs back to the oceans, eustatic sea level rises. Because it's largely contained in the ocean already, the amount of floating sea ice has much less of an effect. Changes in the volume of glacial ice have the potential to cause very large changes in eustatic sea level over relatively short periods of time (many tens of meters over a few millennia).
3. Volume of terrestrial aquifers – this is the amount of groundwater stored on land. It has a relatively minor and gradual contribution, something on the order of a few meters over tens of millennia.
4. Volume of the ocean basins – Long term tectonic changes can change the amount of water held in the ocean basins. Although it is potentially a very large amount of water, these changes take a very long time (hundreds of meters over tens of millions of years).

11.2.3: Relative Sea Level (RSL)

Relative sea level is the distance between a local datum and the sea surface. It's a function of tectonism (subsidence or uplift) and eustasy and allows us to track the difference between them through time. You increase relative sea level when the distance between the two increases; you decrease relative sea level when the two get closer together.

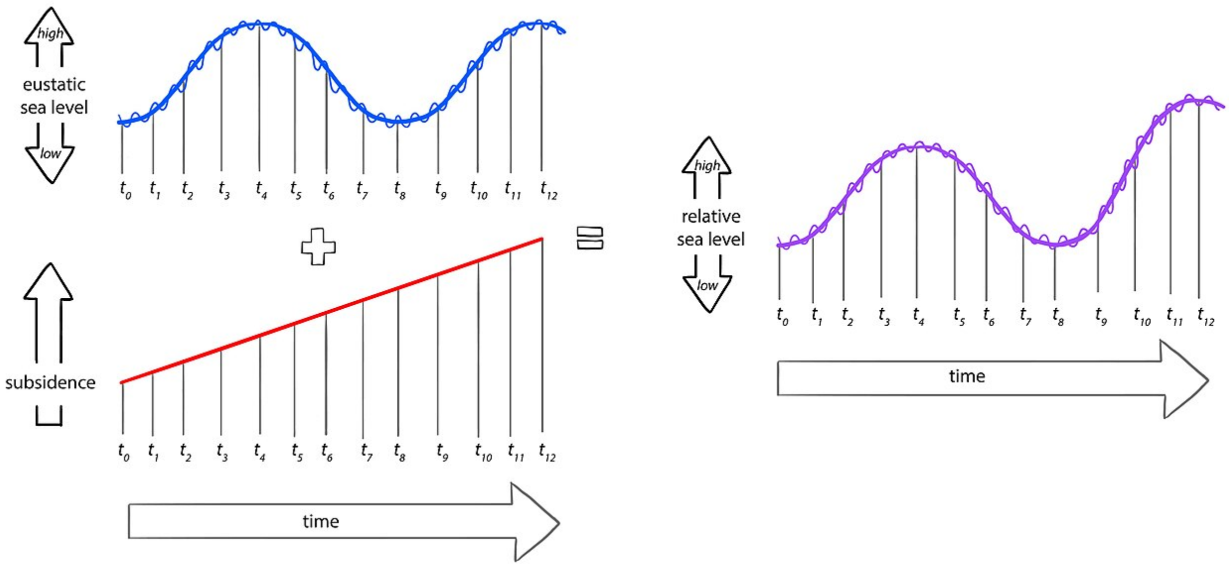


Figure 11.2.3: Relative sea-level (purple) is a function of eustatic sea level (blue) and tectonics (red). In this case, the tectonic component is subsidence which will enhance phases of rise, dampen falls, and result in the overall creation of space available for sediment to accumulate in (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

Relative sea level (RSL) is a useful concept to consider with regards to transgressions because it determines "accommodation (A)", which is the room available for sediment to accumulate. An increase in RSL increases accommodation and gives us the potential to preserve sediment. An overall decrease in RSL gives us less (or removes) accommodation because there is less space for sediment to accumulate. In terms of what causes transgressions and regressions, accommodation is a useful concept, but it alone is not enough to control transgression versus regression. We must also consider sediment supply (S) which is the amount of sediment that is delivered to an area (which is a function of climate, tectonism, etc.).

Given all of that, we can revisit the definitions of transgression and regression that we talked about at the beginning of this section to think about how the balance between sediment supply and accommodation controls transgressions and regressions. In cases where the the sediment supply (S) is greater than accommodation (A), a regression occurs and the shoreline is pushed in a seaward direction (overall shallowing). In cases where the sediment supply is less than accommodation, a transgression occurs and the shoreline migrates in a landward direction (progressive deepening).

CHAPTER OVERVIEW

12: Stratigraphy

Stratigraphy is the study of the geometric relationships, composition, origin, age, and relationships between stratified rocks deposited at the surface of the Earth. We will begin with an overview of different types of unconformities and geologic contacts and then move on to a discussion of the most common approaches to subdivide and classify sedimentary rocks. These schemes are based on lithology (lithostratigraphy), depositional trends (sequence stratigraphy), or seismic reflection data (seismic stratigraphy).

Learning Objectives

- Identify different types of contacts and unconformities and explain their implications for reconstructing the geologic history of an area.
- Identify and describe the different types of lithostratigraphic contacts and units.
- Describe a complete sea level cycle using sequence stratigraphic terminology and the concepts of accommodation and sediment supply.

Chapter thumbnail shows a relative sea level curve (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

12.1: Review of unconformities and other types of contacts

Although we are likely dealing only with disconformities when talking about sedimentary rocks and stratigraphic relationships, its worth remembering that there are three types of unconformities:

Angular unconformity – occurs when there is a marked discordance between layers caused by a period of uplift and erosion during unconformity development.

Nonconformity – occurs when you have sedimentary rocks deposited atop crystalline rocks that formed at depth.

Disconformity – occurs when you have sedimentary rocks atop sedimentary rocks with no obvious angular discordance. The unconformity surface can have significant amounts of erosional relief.

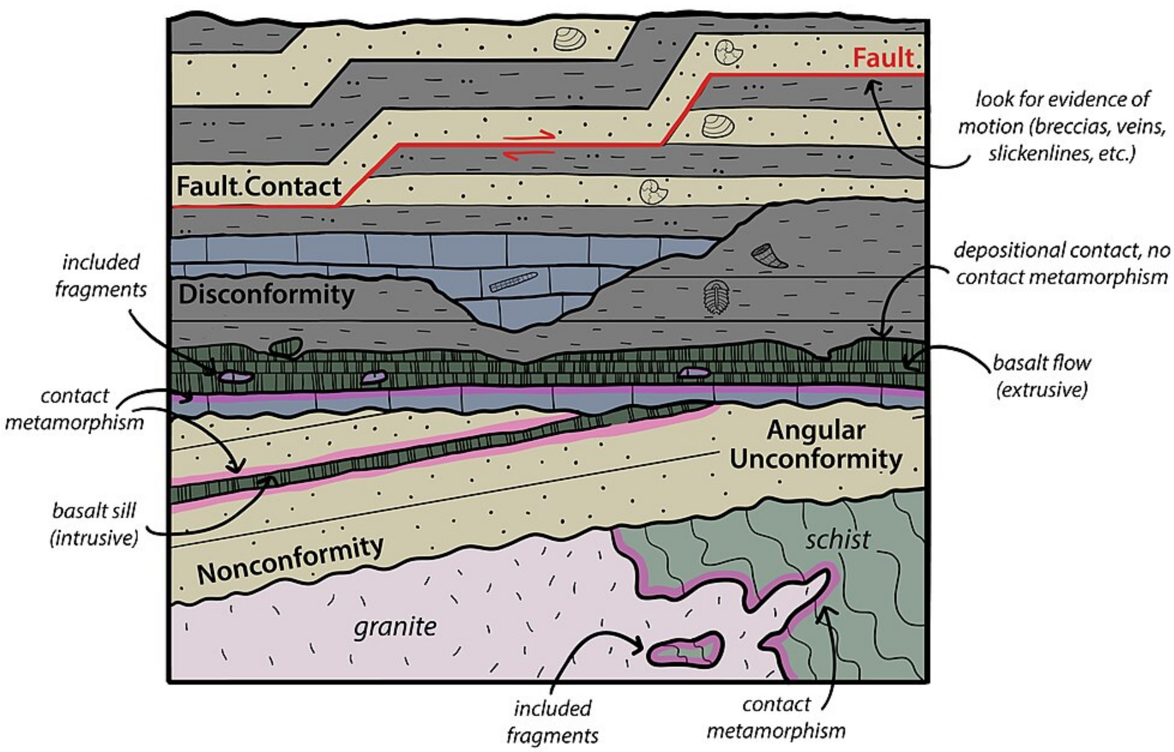


Figure 12.1.1: Diagram showing the three types of unconformities as well as different types of igneous contacts and a variety of expressions of a fault contact (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

12.2: Lithostratigraphy

Lithostratigraphy is a particular type of stratigraphy that uses the lithologic character of the rock as the basis for subdividing and classifying the rocks. Our ability to do lithostratigraphy is based on the fundamental concepts outlines by Steno's Principles and the concepts of relative and absolute dating.

12.2.1: Lithostratigraphic Relationships

Vertical contacts are most commonly described as sharp, gradational, or interbedded. Lateral transitions are most often described as gradational, intertonguing, or pinching out.

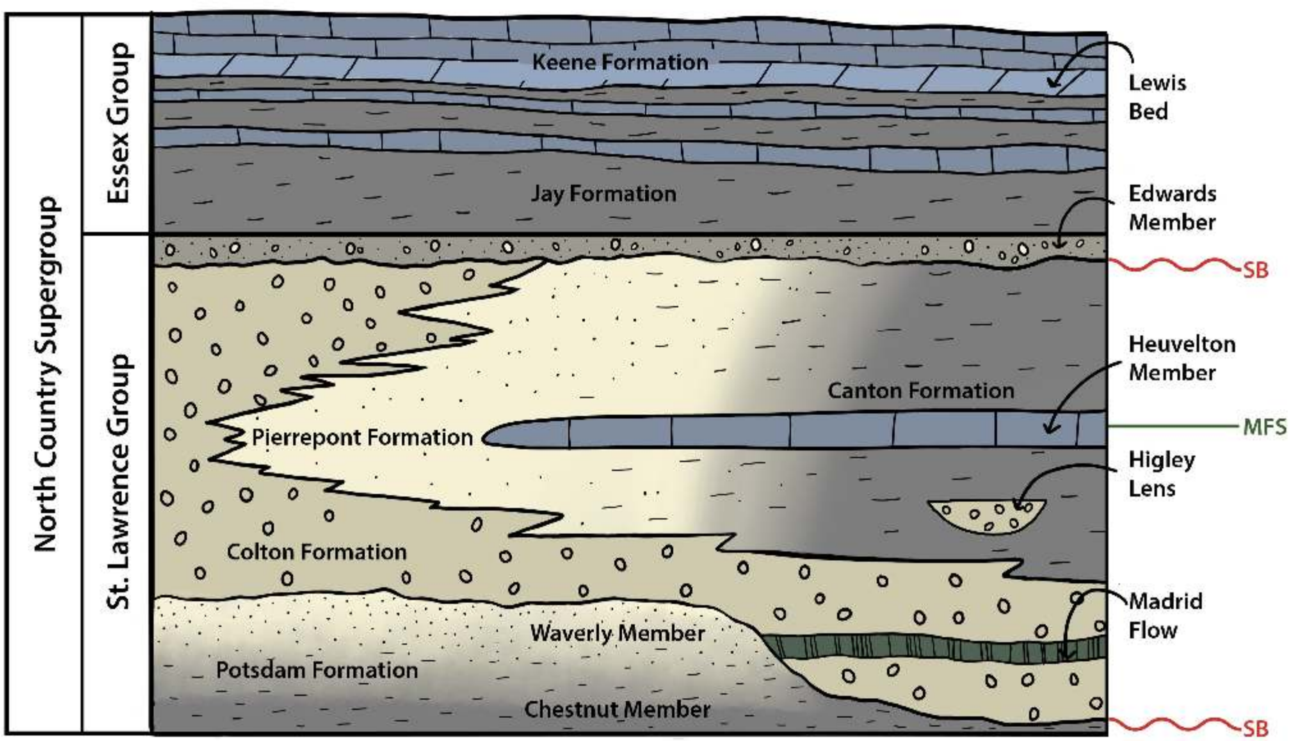


Figure 12.2.1: Diagram showing the major types of lateral and vertical lithologic contacts and lithostratigraphic units; a sequence stratigraphic subdivision of the same units is provided along the right margin of the diagram (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

12.2.2: Lithostratigraphic Units

We are governed by the North American Stratigraphic Code which lays out a comprehensive set of rules for naming and recognizing units. In order to have a unit be formally named (they are proper nouns) an author must provide all of the required information (inc. justification, boundaries, type sections, history, etc.) in a peer-reviewed publication that is widely available. The type section for a named lithostratigraphic unit must be accessible, well-exposed, and characteristic of the unit.

Lithostratigraphic units are given a two part name; the first part refers to a place where the unit is well exposed and the second part tells the reader what the rank of the unit is. This, the Mauch Chunk Formation occurs near and is a lithostratigraphic formation named for the town of Mauch Chunk (now renamed Jim Thorpe, after one of the greatest athletes of the last century). In descending order, the different ranks of lithostratigraphic units include:

Supergroup – is the largest formal lithostratigraphic unit. They are composed of collections of named formations and or groups.

Group – the unit above formation, it may be composed entirely or partially of named formations.

Formation – the fundamental unit of lithostratigraphy. They must be mappable at the surface or in the subsurface, usually across an area that is at least the size of a 1:24k USGS topographic map.

Member – the lithostratigraphic unit below a formation in rank. All members must occur within formations. Formations need not be divided (in whole or in part) into members. Members can extend laterally from one formation into another. Geographically restricted members are often referred to as members or tongues.

Bed (or Flow) – the smallest formal lithostratigraphic unit. It is a discrete and distinctive unit that is easily recognized.

12.3: Sequence Stratigraphy

Sequence stratigraphy is based on the recognition and correlation of stratigraphic surfaces that represent changes in depositional trends (Embry, 2001). Remember that this is different from lithostratigraphy which is when we subdivide rocks into lithologic formations based on their physical characteristics.

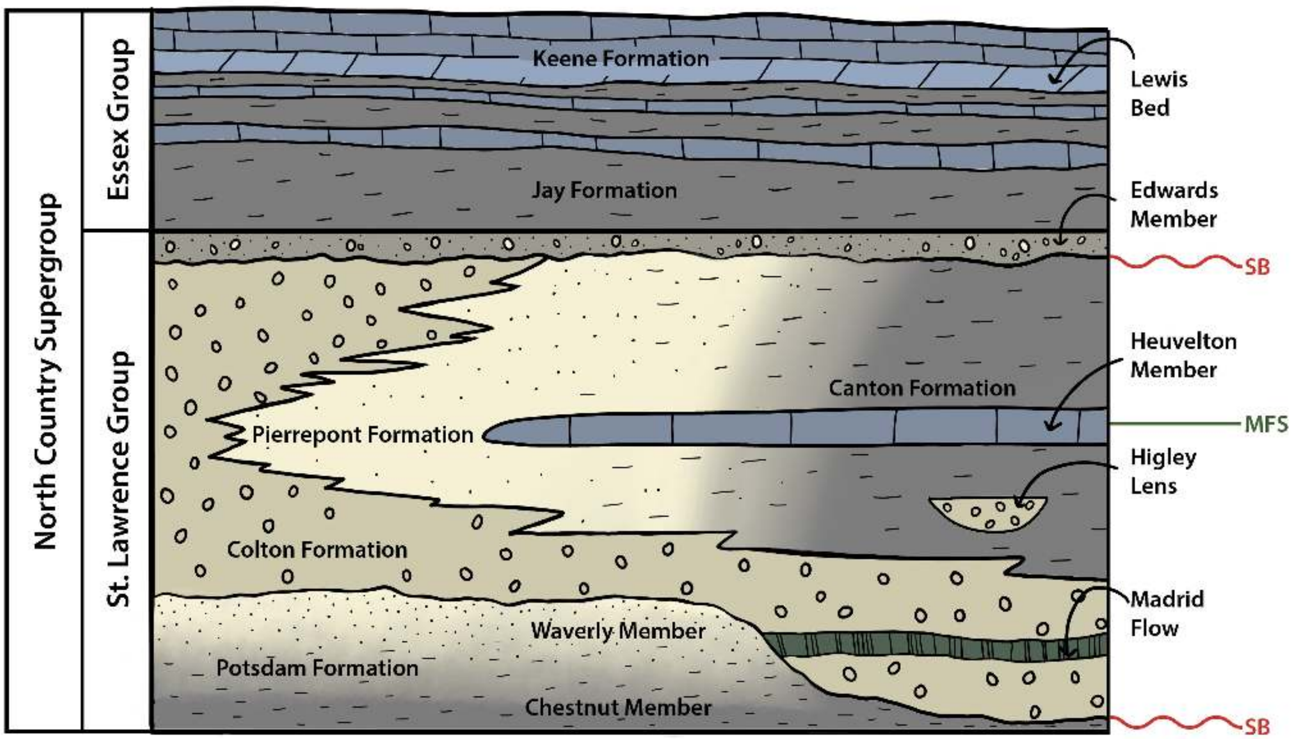


Figure 12.3.1: Diagram showing the major types of lateral and vertical lithologic contacts and lithostratigraphic units; a sequence stratigraphic subdivision of the same units is provided along the right margin of the diagram (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

12.3.1: Parasequences

Relative relative sea-level (RSL) curves are the combined product of eustatic and tectonic curves. The RSL curve that we made is a smooth theoretical one; in reality these curves are very complex and composed of a series of small-amplitude, short term oscillations in RSL and sediment supply (caused by things like delta-lobe switching, etc).

Parasequences are the sedimentary record of the small amplitude changes described above. They are small scale, shallowing upward succession of beds bounded by flooding surfaces (abrupt, nearly instantaneous increases in water depth). The flooding surfaces are overlain by deposits that record progressive shallowing. Parasequence thickness is extremely variable and can range from a few tens of centimeters to several meters. Laterally, parasequeces are often tracable over tens to thousands of square kilometers.

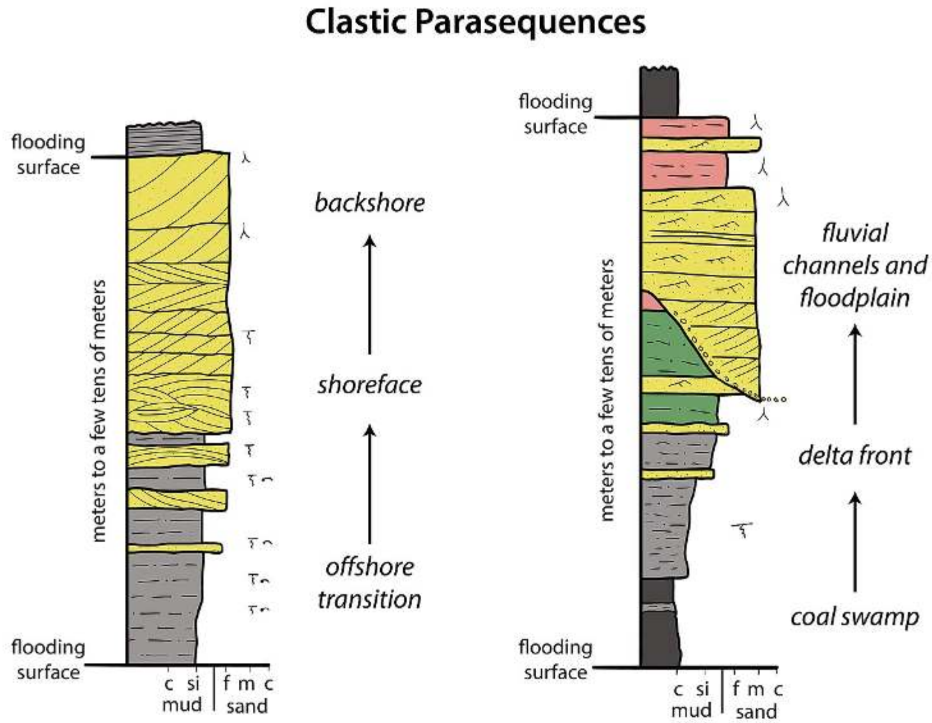


Figure 12.3.2: Typical facies successions in common types of clastic parasequences (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

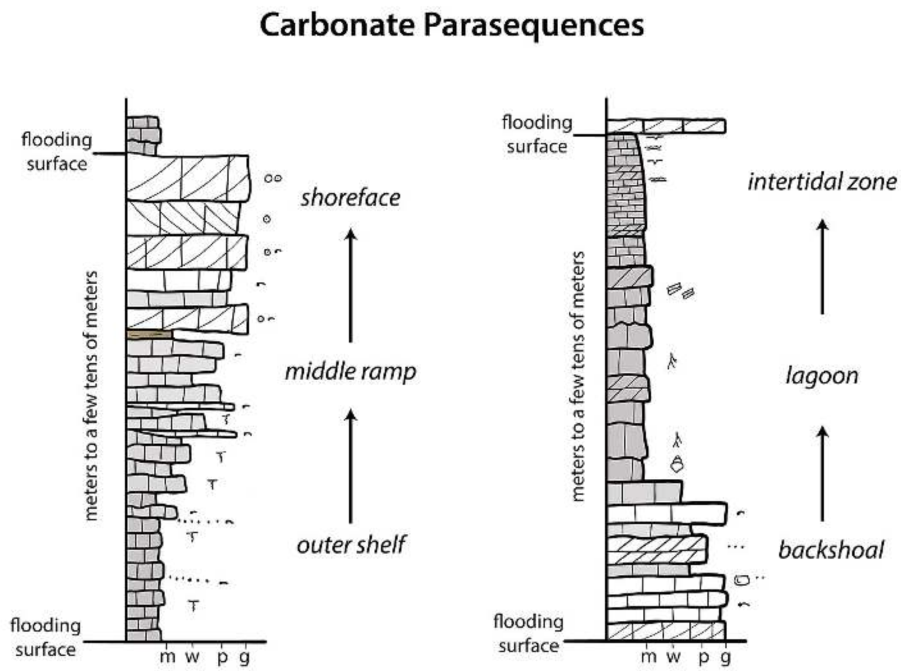


Figure 12.3.3: Typical facies successions in common types of carbonate parasequences (Page Quinton via Wikimedia Commons; CC BY-SA 4.0).

12.3.2: Parasequence Sets

When stacked atop one another, parasequences collectively show long-term changes in shoreline position. We can recognize three different parasequence stacking patterns (parasequence sets).

Progradational parasequence sets – form when sedimentation rates outpace the creation of accommodation which results in each parasequence being shallower than the previous one, progressive shallowing, and an overall seaward shift in shoreline position (regression).

Aggradational parasequence sets – happen when there is no net shift in shoreline position or average water depth.

Retrogradational parasequence sets – form when accommodation rates are outpacing sediment supply rates. The result is that each parasequence represents deepening relative to the previous one and an overall landward shift in shoreline position (transgression).

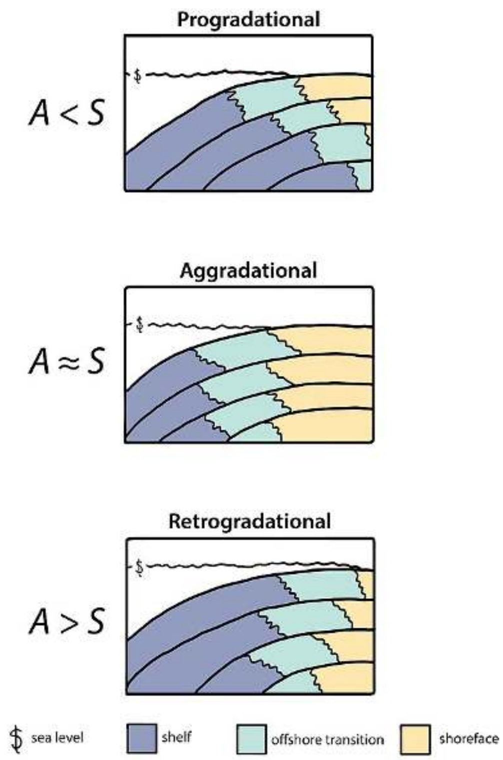


Figure 12.3.4: Parasequence set architecture (Page Quinton via Wikimedia Commons; CC BY-SA 4.0 - after Figure 4.7 in Coe and Church, 2003).

12.3.3: Systems Tracts, Bounding Surfaces, and Sequences

Sequences are the fundamental unit of sequence stratigraphy. They are composed of a package of genetically-related strata deposited during one cycle of sea-level rise and fall. In many basins, they are bounded by unconformities and their correlative conformities.

Systems tracts are parts of a sequence that are deposited during a given phase of the sea-level cycle; each has a distinctive parasequence architecture. There are a variety of models for sequence architecture; the simplest schemes have just two systems tracts and the most complex ones have four. Each systems tract is separated from the next via a stratigraphically important surface.

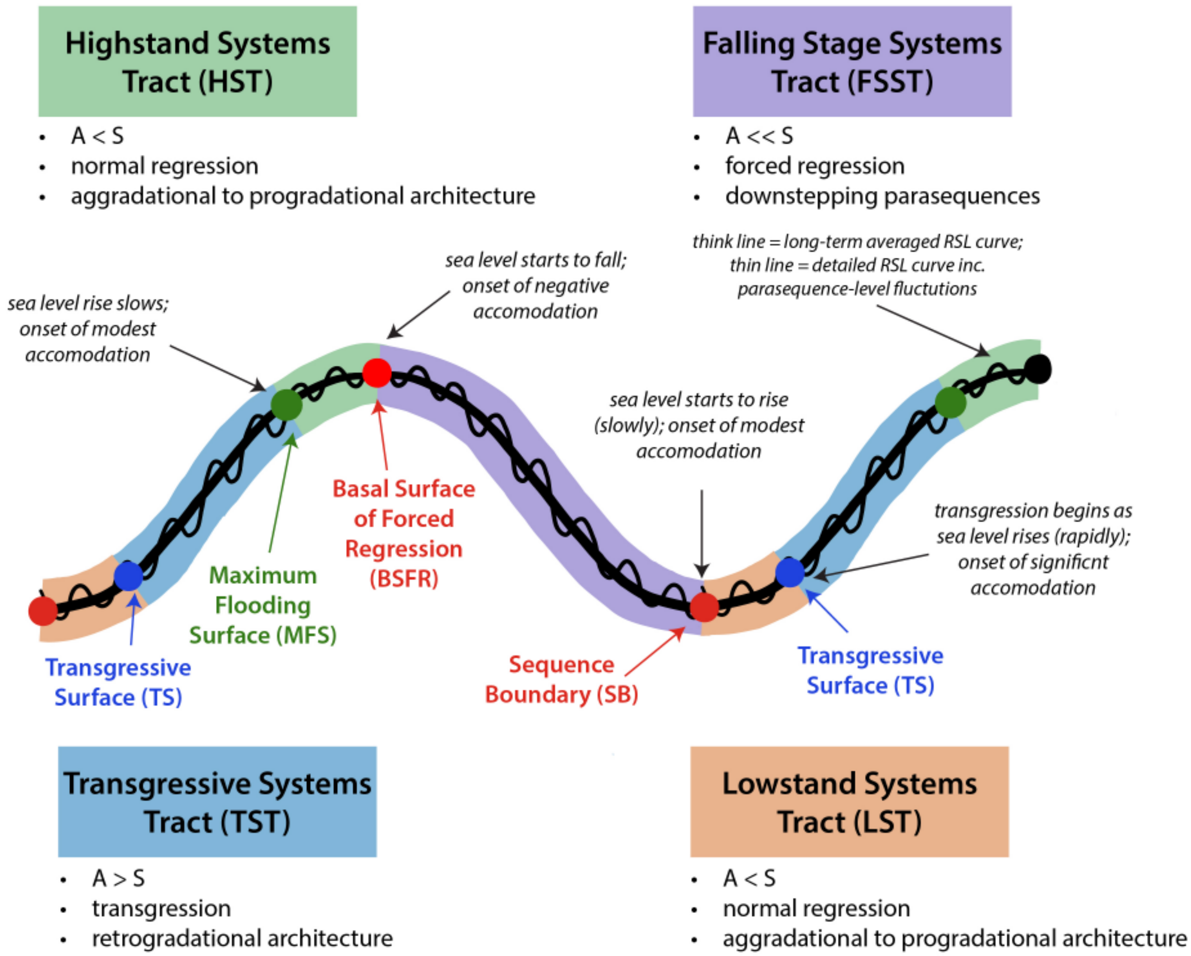


Figure 12.3.5: Overview of sequence stratigraphic nomenclature relative to a sea level cycle (Page Quinton via Wikimedia Commons; CC BY-SA 4.0). Systems tracts and placement of sequence stratigraphic surfaces is after the Depositional Sequence IV approach in Catuneanu (2022) which is based on Helland-Hanson and Gjelberg (1994), Hunt and Tucker (1995, 1996), and Plint and Nummedal (2000).

12.3.3.1: Highstand systems tract (HST)

The highstand systems tract forms when sea-level is still rising, but rising slowly enough that it is being overpowered by sediment supply. The result is an aggradational to progradational parasequence set architecture, overall (normal) regression, and the position of the shoreline to build both up and out to sea. It is composed of progradational to aggradational parasequence sets. It is bounded below by the maximum flooding surface and above by the basal surface of forced regression and/or the start of unconformity development.

12.3.3.2: Basal Surface of Forced Regression (BSFR)

The basal surface of forced regression occurs at the moment that sea-level starts to fall and negative accommodation is generated. As the name implies, the resulting regression is forced because there is no option but for the shoreline to shift both seaward and topographically down. That stands in contrast to a normal regression which can happen under a positive accommodation regime any time that sediment supply outpaces the generation of accommodation.

12.3.3.3: Falling Stage Systems Tract (FSST)

The falling stage systems tract is only preserved in basins that are experiencing a high subsidence rate and/or in deepwater parts of basins that are not subaerially exposed. Sea level is falling throughout and the basin experiences a massive forced regression under a negative accommodation regime. It is composed of downstepping progradational parasequnce sets. The shoreline is progressively moving out and down and exposed sediments are eroded or pedogenically overprinted as unconformity development occurs across an ever increasing percentage of the basin. It is underlain by the BSFR and overlain by the sequence boundary or its correlative conformity.

12.3.3.4: Sequence Boundary (SB)

The sequence boundary represents the maximum extent of the unconformity across the sedimentary basin. In parts of the basin that were never subaerially exposed, it passes laterally into the correlative conformity. Remember that the unconformity is a diachronous event, it starts sooner, ends later, and lasts longer in updip parts of the basin.

12.3.3.5: Lowstand Systems Tract (LST)

The lowstand systems tract forms in the early phases of sea-level rise while sediment supply is still outpacing the generation of accommodation. Regression continues, but its no longer forced and although the position of the shoreline continues to build out into the basin it can also start to build topographically upward. It is composed of aggradational to progradational parasequence sets.

The LST is underlain by the sequence boundary and overlain by the transgressive surface.

12.3.3.6: Transgressive surface (TS)

The transgressive surface marks the start of transgression and occurs at the moment when transgression begins because accommodation overtakes sediment supply. Some authors will refer to this surface as the maximum regressive surface. At this time the shoreline reaches its most basinward position.

12.3.3.7: Transgressive Systems Tract (TST)

The transgressive systems tract occurs during the most rapid phase of sea level rise and is the only systems tract where A>S which causes a transgression. It is composed of retrogradational parasequence sets formed as the shoreline moves up and in a landward direction.

It is underlain by the transgressive surface and overlain by the maximum flooding surface.

12.3.3.8: Maximum flooding surface (MFS)

The maximum flooding surface marks the end of transgression, the most landward position of the shoreline, and the deepest water deposition in the basin.

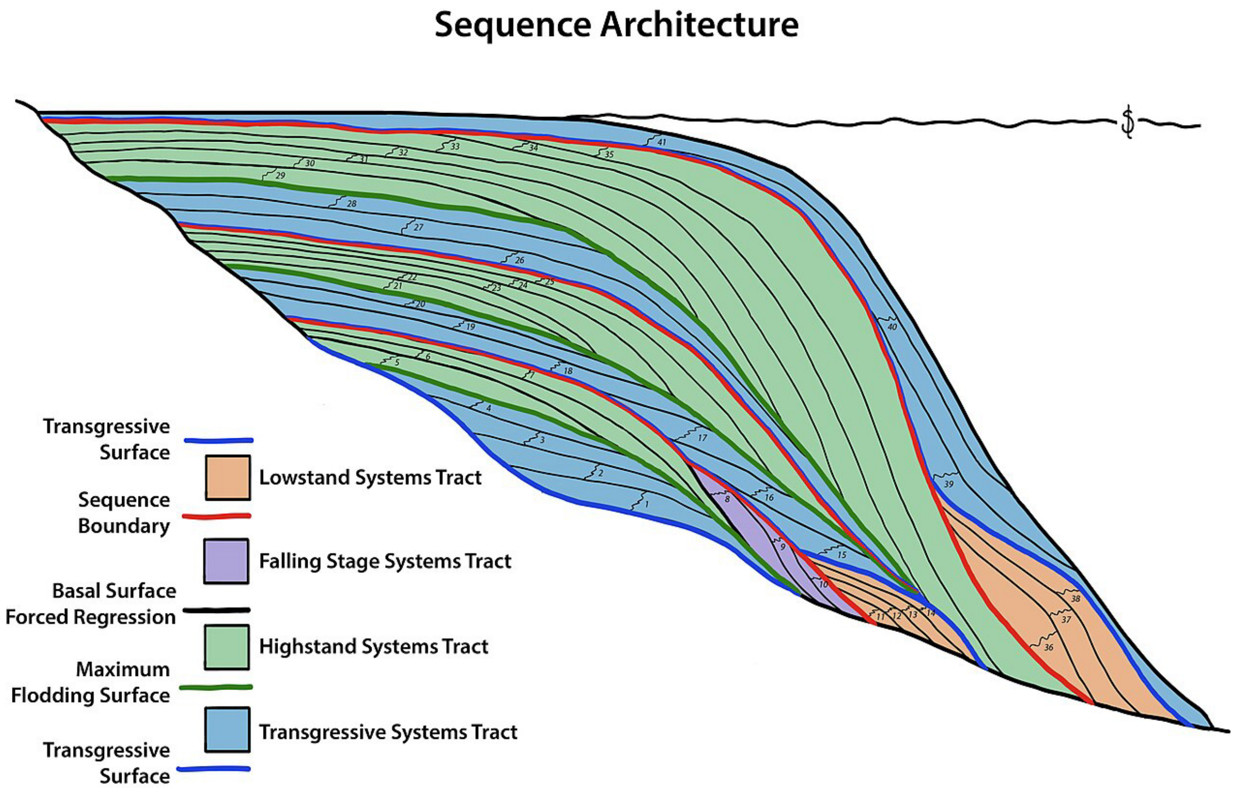


Figure 12.3.6: Cross-sections sketch of a shelf edge showing sequence stratigraphic stacking patterns (Page Quinton via Wikimedia Commons; CC BY-SA 4.0). Systems tracts and placement of sequence stratigraphic surfaces is after the Depositional Sequence IV approach in Catuneanu (2022) which is based on Helland-Hanson and Gjelberg (1994), Hunt and Tucker (1995, 1996), and Plint and Nummedal (2000).

12.3.4: Lithostratigraphy vs. Sequence Stratigraphy

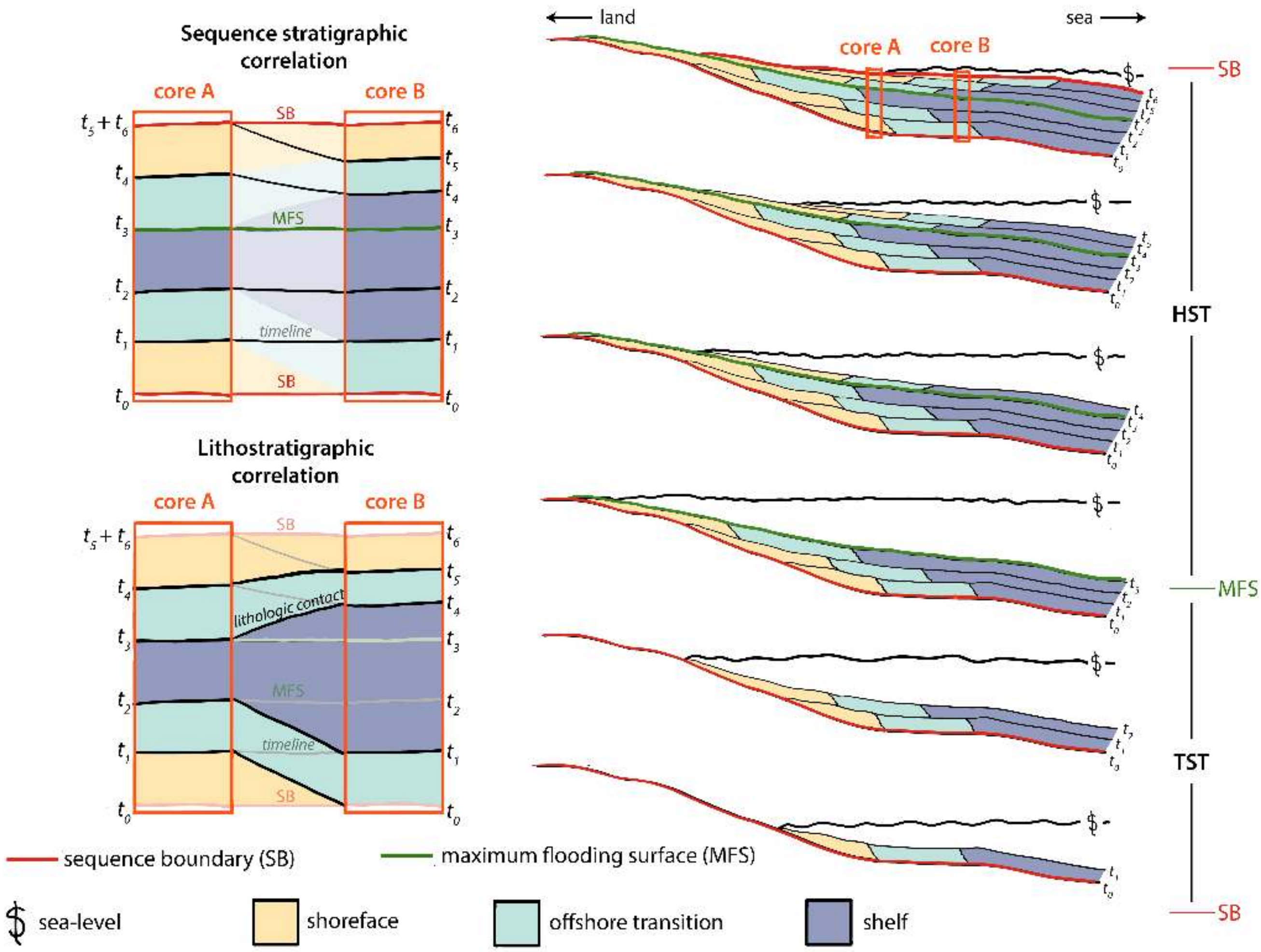


Figure 12.3.6: Comparison of sequence stratigraphic correlation versus lithostratigraphic correlation (Page Quinton via Wikimedia Commons; CC BY-SA 4.0)

12.3.5: Alternate Sequence Architectures

The sequence stratigraphic model described above represents the most complex and complete expression of a sequence. Its important to remember that not all systems tracts are going to be preserved at all locations and/or in all sedimentary basins.

ExxonMobil invented sequence stratigraphy and their scheme does not identify the FSST, instead they lump these sediments together with lowstand deposits.

In the “Genetic sequence” model, it’s the maximum flooding surface that defines the start and end of a sequence. This model works well in rapidly subsiding basins where the flooding event is widespread/well preserved and the regression and sequence boundary is suppressed.

12.3.6: Readings and Resources

- Catuneanu, O., 2022, Principles of Sequence Stratigraphy, Elsevier, Oxford, 375 p. https://doi.org/10.1016/C2009-0-19362-5
- Coe, A.L. and Church, K.D., 2003, Sequence Stratigraphy, *in* Coe A.L. (ed), The Sedimentary Record of Sea-Level Change, Cambridge University Press, Cambridge U.K., p. 57-98. ISBN 9780521538428. Publisher URL.
- Hunt, D. and Tucker, M.E., 1995, Stranded parasequences and the forced regressive wedge systems tract: deposition during base-level fall—reply, Sedimentary Geology,
- Volume 95, Issues 1–2, p. 147-160, https://doi.org/10.1016/0037-0738(94)00123-C.
- North American Stratigraphic Code - https://ngmdb.usgs.gov/Info/NACSN/05_1547.pdf
- Plint, A.G., Nummedal, D., 2000, The falling stage systems tract: recognition and importance in sequence stratigraphic analysis. In: Hunt, D., Gawthorpe, R.L. (Eds.), Sedimentary Response to forced regression, Geol. Soc. London Special Publication 172, p. 1–17.

12.4: Seismic Stratigraphy

12.4.1: Seismic Reflection

Seismic reflection is a technique used to image structures and stratigraphy in the subsurface. The general idea is broadly similar to sonar, in that you are producing an image by sending energy into material and creating images from reflected energy. Advances in computing power in the last few decades has dramatically improved data quality and we can now resolve horizons that are <5 m thick thousands of meters down in the subsurface.

Video 12.4.1: A general overview of seismic data collection and what is it used for.

12.4.2: Data Collection

Seismic reflection data can be collected on land or at sea. Data collection begins when a seismic source (airguns at sea; small explosions or truck-mounted vibrating plates on land) generate seismic waves that travel into the Earth and pass through layers of sediment and rock. Different layers have different physical properties which result in different seismic velocities as the waves pass through them (generally 2-8 km/s). Some of the energy reflects off of boundaries between layers, unconformities, and faults and reflects back to the surface. Sensors called geophones or hydrophones detect the reflected seismic waves and produce traces that show the two-way travel time from the energy source to the boundary and up to the sensor.

In the greatly simplified portrayal in Figure 1, we show what a single geophone would record the energy from a single source; the source and the geophone are moved to get a second trace. In practice, geophones/hydrophones are deployed in long arrays with hundreds or thousands of evenly-spaced sensors; each release of seismic energy is recorded by the entire array. Having large arrays allows the reflections to be picked up from multiple perspectives which removes noise from the system. The data is collected, processed, and most commonly displayed in two-dimensional lines or three-dimensional cubes that can be examined from a variety of perspectives.

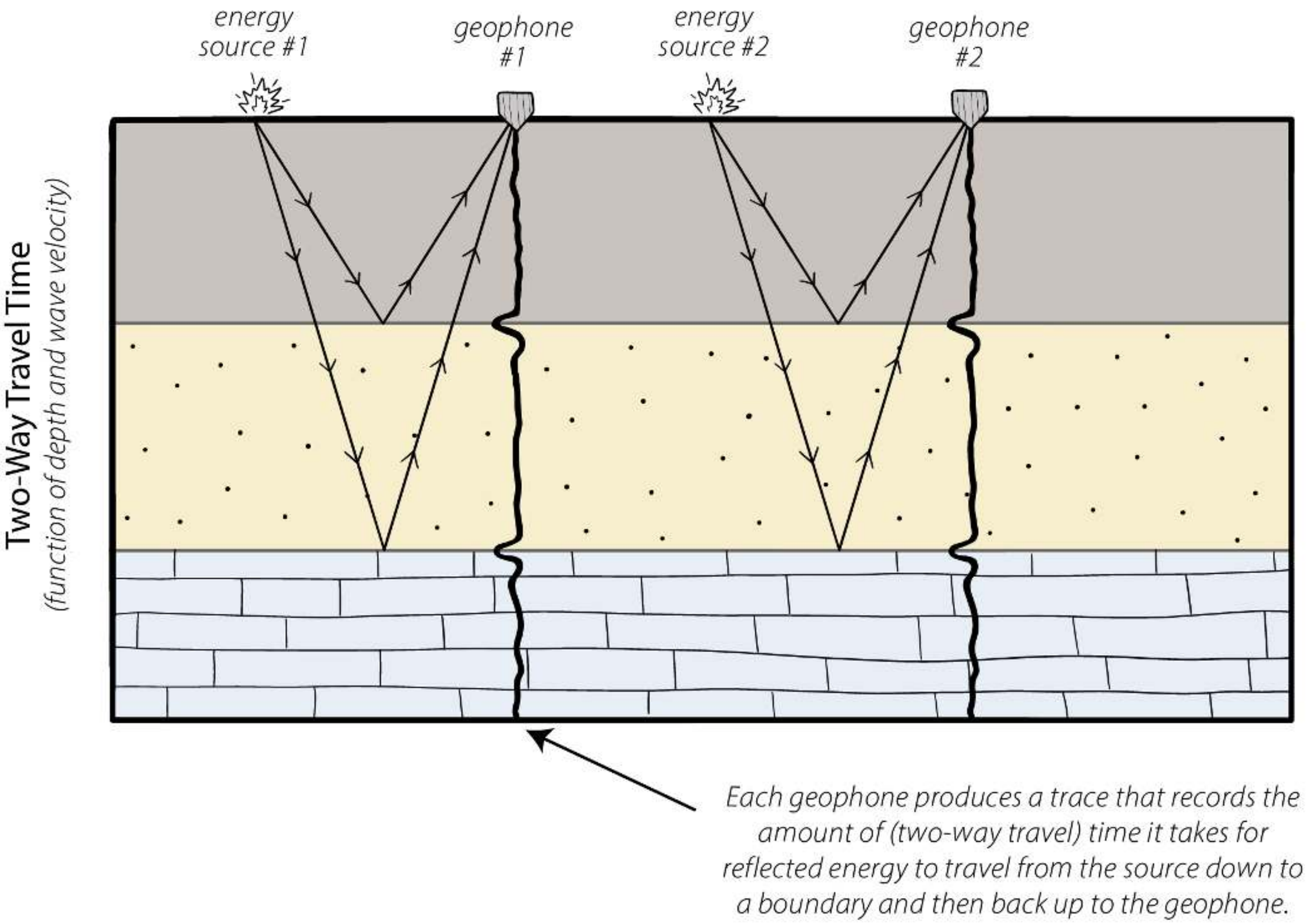


Figure 12.4.1: Simplified illustration of seismic reflection data collection. Energy is released from a source and travels into the subsurface. Different layers have different physical properties that impact seismic velocities and some energy reflects back to the surface when the waves hit the coundaries between layers. Geophones pick up the reflections and make a trace that shows two way travel time for each reflection. By combining multiple releases of energy and deploying extensive arrays of geophones, a detailed picture of the subsurface can be created (Page Quinton, *via Wikimedia Commons;* CC BY-SA 4.0)

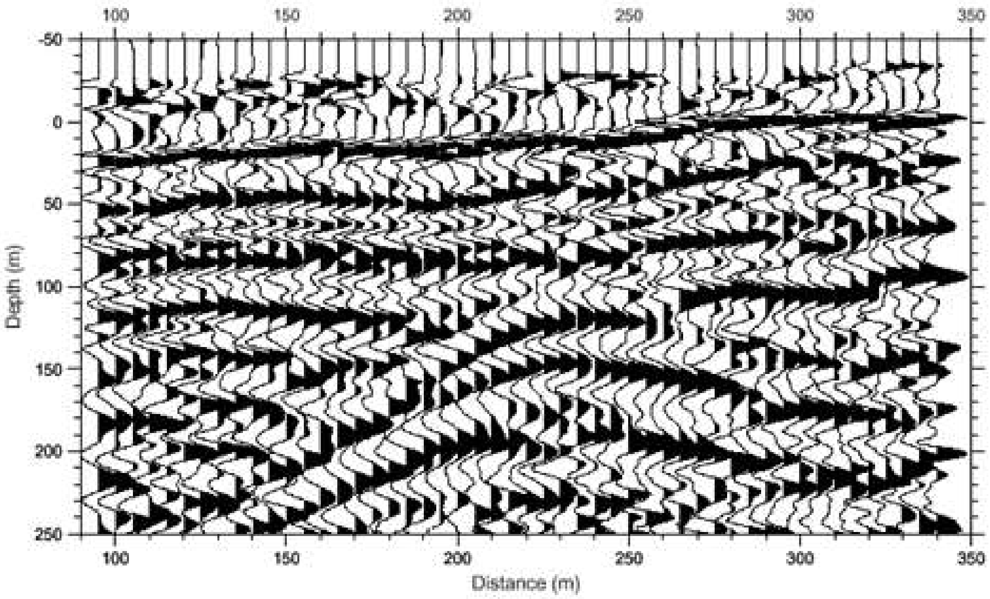


Figure 12.4.2: Simple two-dimension seismic reflection line. The vertical axis shows depth (converted from two-way travel time) and the horizontal axis shows regularly spaced traces along the line. Troughs on individual traces are colored black to enhance the appearance of reflectors in the subsurface (United States Geological Survey, via Wikimedia Commons; public domain).

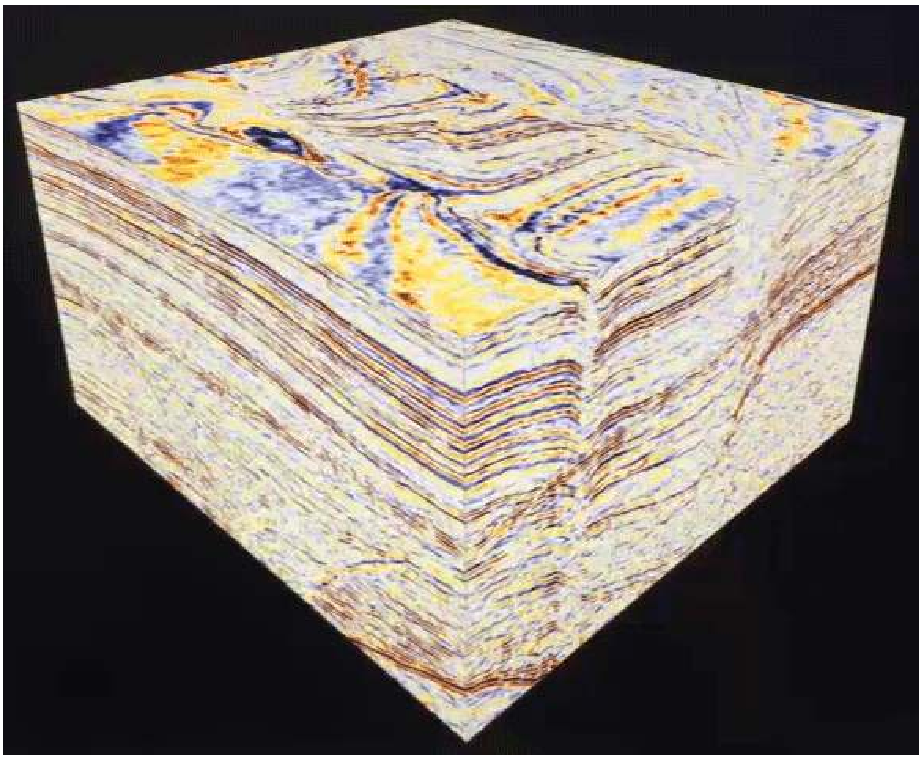


Figure 12.4.3: Animation of a 3D seismic cube. Peaks and troughs are colored red and blue to enhance the appearance of reflectors in the subsurface. 3D cubes can be sliced in a variety of ways to enable visualization and correlation (Pacific Coastal and Marine Science Center via USGS; public domain).

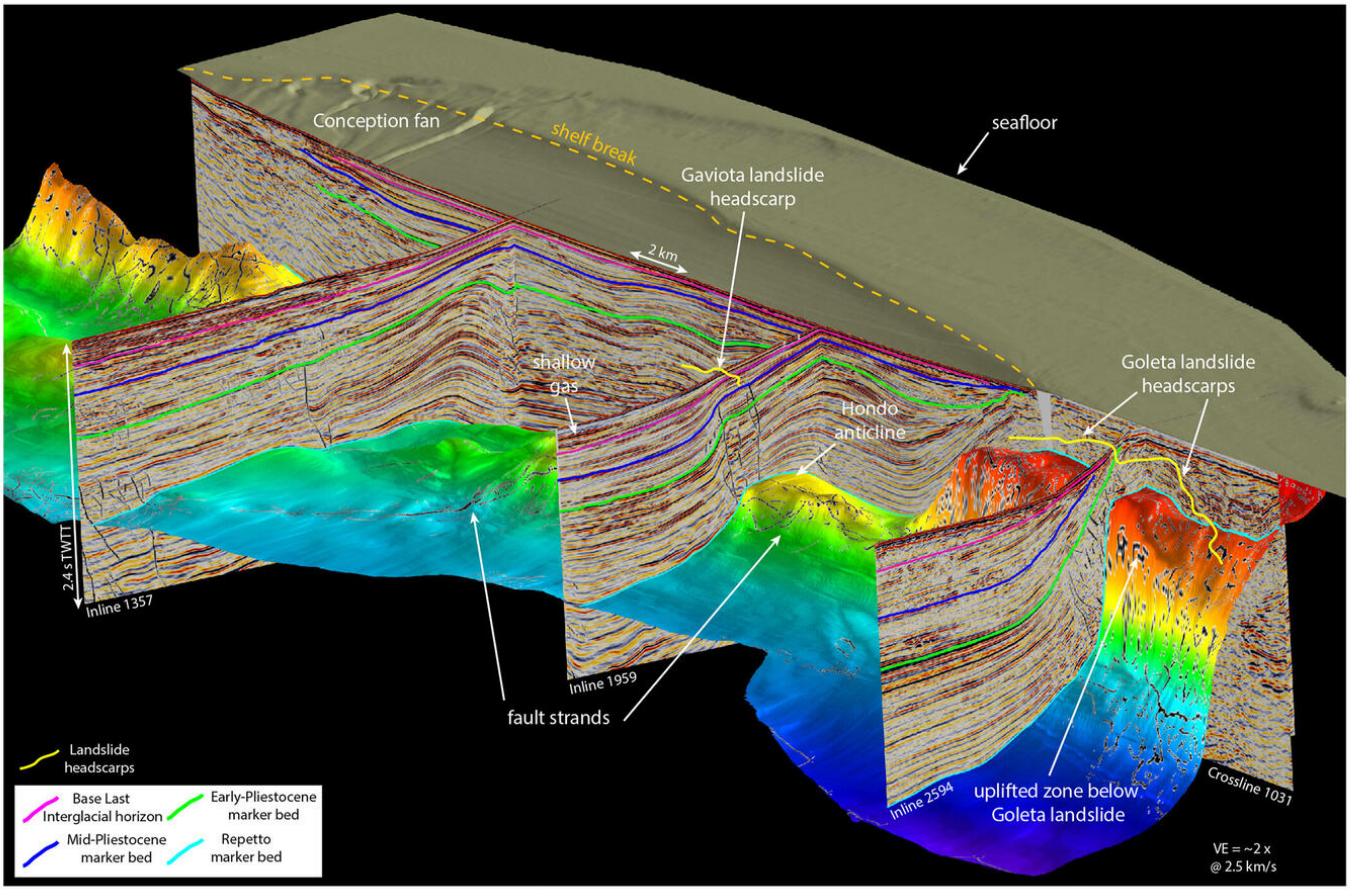


Figure 12.4.4: Bedding contacts, unconformities, and faults can be mapped in great detail using high quality seismic datasets. In this example dataset from the Santa Barbara Channel, offshore southern California faults are shown in black, landslide headscarps are shown in yellow, and various marker beds are shown in blues, green, and purple (Pacific Coastal and Marine Science Center via USGS; public domain).

12.4.3: Sequence Stratigraphic Analysis of Seismic Lines

As discussed above, seismic reflectors can represent bedding surfaces or unconformities. Each one roughly approximates a timeline that marks the depositional surface at a moment in time. Improved processing and interpretation techniques can allow for some interpretation of the properties of the material between reflectors, but generally speaking lithologic facies changes do not image well. Careful examination of seismic lines shows that in many areas, distinctive patterns in reflector terminations and geometries are present. Major termination types include:

- onlap - progressive termination against a surface with a greater dip
- downlap - progressive termination against a surface with a lower dip
- toplap - very low angle termination against an overlying surface
- erosional truncation - termination by an erosional surface (often somewhat irregular)

A seismic sequence is a package of genetically-related strata deposited during one sea-level cycle. In the simplest case, this would be an unconformity-bound package where a package of retrogradational reflectors (TST) onlap the basal sequence boundary and a progradational package of reflectors (HST or a "regressive systems tract") downlap the maximum flooding surface.

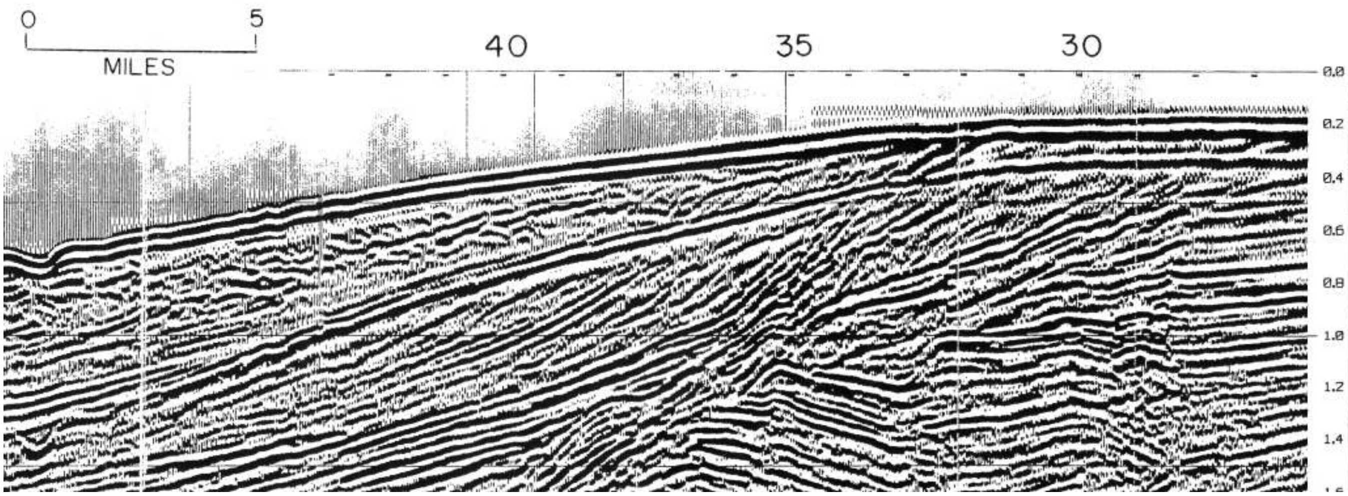


Figure 12.4.5: Portion of an uninterpreted 2D seismic line showing many of the distinctive patterns in reflector terminations illustrated below (Seismological Facility for the Advancement of Geoscience via Seismic Sequences; CC BY 4.0).

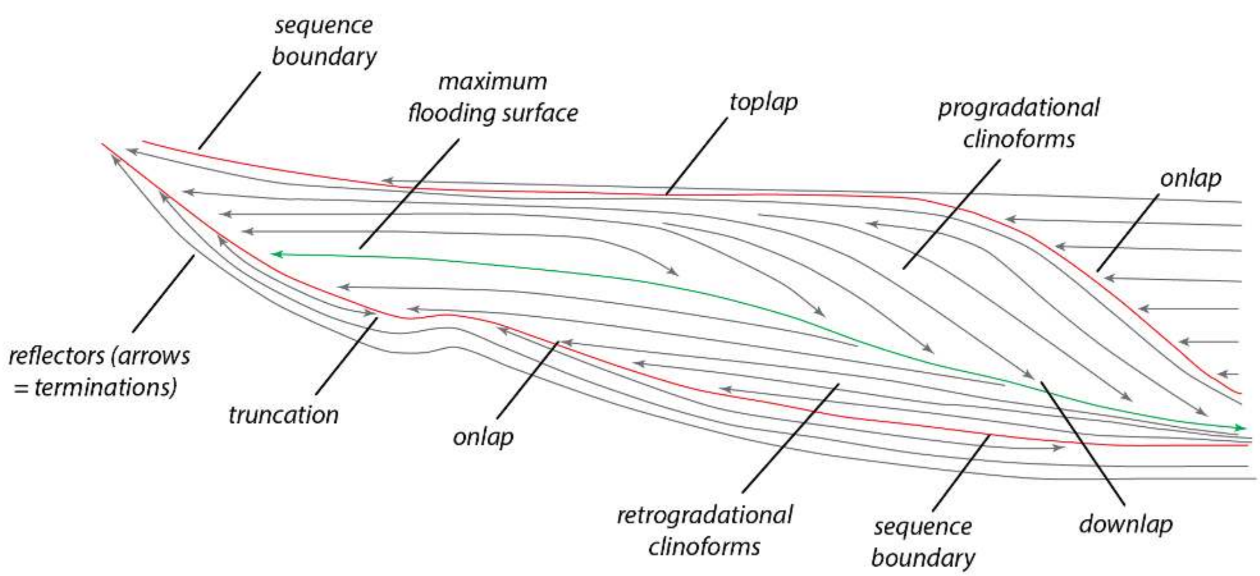


Figure 12.4.6: Diagram showing main types of reflector terminations, sequence stratigraphic surfaces, and clinoform stacking patterns (Page Quinton, via Wikimedia Commons; CC BY-SA 4.0; diagram is after Vail, 1987)

12.4.4: Seismic Facies

Within, or in addition to, a sequence stratigraphic framework, reflector terminations, patterns, and properties can be used to delineate seismic facies, which are mappable packages of reflectors that are distinct from surrounding ones. Seismic facies can be used to match seismic attributes to drilling data and to provide information about lateral changes in lithology and/or fluid types.

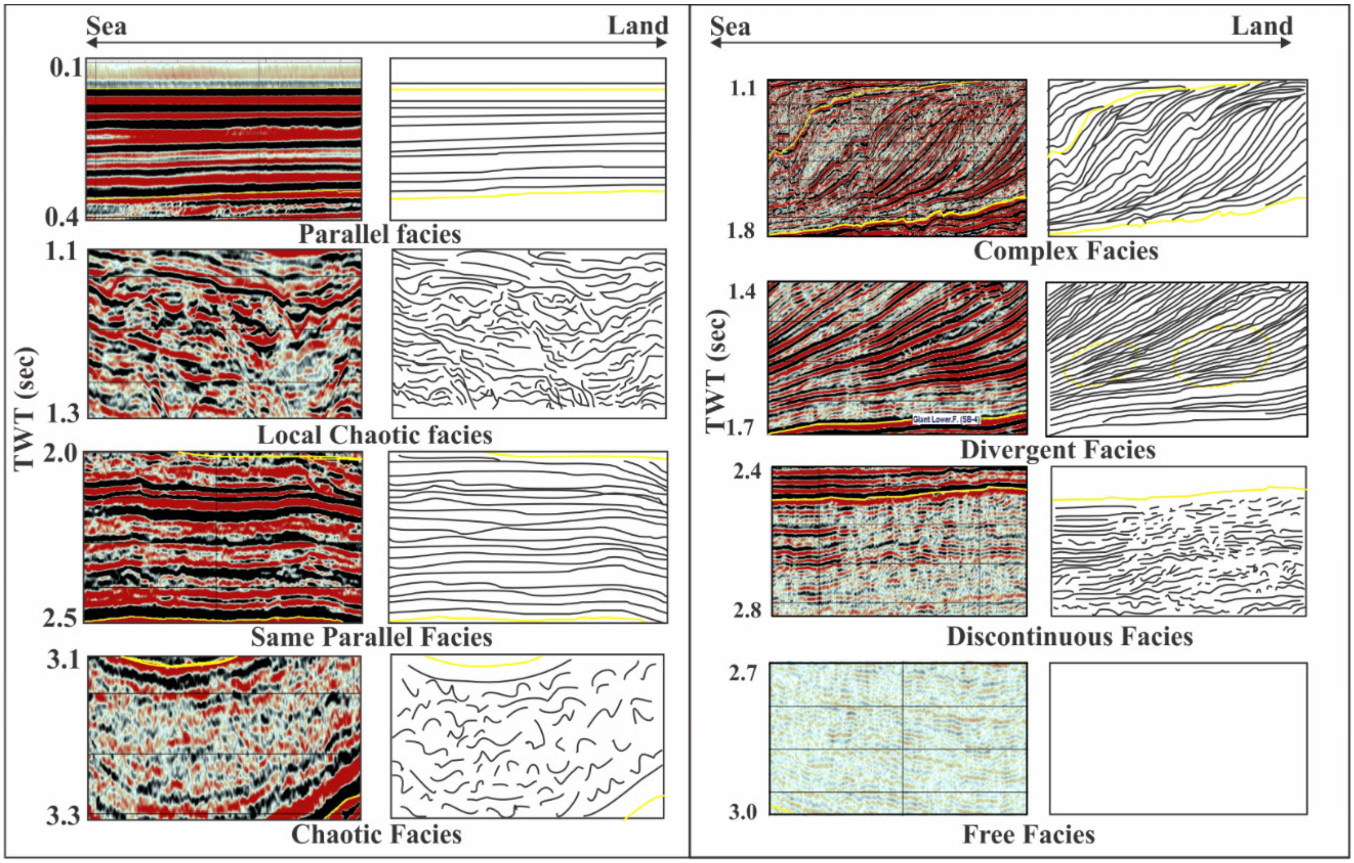


Figure 12.4.5: Example seismic facies from a seismic reflection dataset (Al-Masgari et al, 2021 via Seismic Sequence Stratigraphic Sub-Division Using Well Logs and Seismic Data of Taranaki Basin, New Zealand; CC BY 4.0).

12.4.5: Resources and Readings

- https://www.mdpi.com/2076-3417/11/3/1226#
- Seismic Facies Analysis on AAPG Wiki
- Vail, P.R., 1987, Seismic Stratigraphy Interpretation Using Sequence Stratigraphy: Part 1: Seismic Stratigraphy Interpretation Procedure, in Bally, A.W. (ed), AAPG Studies in Geology #27, volume 1: Atlas of Seismic Stratigraphy, p. 1-10.

13.1: Review of Plate Tectonics

The text provided below is reused from 4.4: Plates, Plate Motions, and Plate-Boundary Processes in Karla Panchuk's Physical Geology textbook on Libretexts.

Shifting the Paradigm to Shifting Plates

Critics of continental drift shared a mental picture of Earth's outer layer that made it very difficult for them to imagine how continents could move. They envisioned Earth as having a solid one-piece outer shell, represented by the rocks making up the ocean floor. They thought of continents as large blocks of rock that would have to slide across or through the ocean floor in order to move. On the cusp of the plate-tectonics paradigm shift In the 1960s, geologists were faced with increasingly difficult-to-deny evidence that continents had moved, and they also knew more than ever before about the shape of the ocean floor. They did the best they could to reconcile what they knew with an unworkable model of Earth's structure: some even hypothesized about how the newly discovered ocean floor structures might actually be the deformation in the wake of a continent plowing through ocean crust. Imagine their plight as a fly bonk-bonk-bonking against a window, trying to get outside, but working within a conceptual framework that has no place for the existence of glass.

Now we know from the shape of the ocean floor and the distribution of earthquakes that Earth's outer layer is broken into many fragments, called tectonic plates. This is why the term *continental drift* doesn't match up very well with our present-day understanding of plate tectonics. It isn't just continents that are moving, but entire slabs of lithosphere that can include both continental and oceanic crust attached to the uppermost part of the mantle (Figure 4.23). The lithospheric plates not only move, but can grow and also split into pieces. They crash into each other, break, and fold. Some are being swallowed up by the Earth as you read this. "Continental drift" is far too peaceful a term to describe the behaviour of Earth's lithosphere.

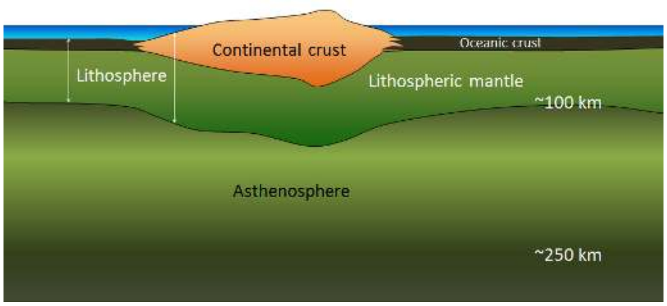


Figure 4.23 The crust and upper mantle. Tectonic plates consist of lithosphere, which includes the crust and the lithospheric (rigid) part of the mantle. Source: Steven Earle (2015), CC BY 4.0. View source.

While lithospheric plates are rigid, the asthenosphere immediately below the lithosphere is not. It contains very small amounts of melted rock, and this makes it weak. Lithospheric plates can move because the weak asthenosphere deforms by flowing. The behavior of the asthenosphere is exactly what some geologists thought the ocean crust would have to behave like in order for continents to drift through it.

A Map of Moving Plates

The ideas of "continental drift" and sea-floor spreading became widely accepted by 1965, and more geologists started thinking in these terms. By the end of 1967, Earth's surface had been mapped into a series of plates (Figure 4.24). The major plates are Eurasian, Pacific, Indian, Australian, North American, South American, African, and Antarctic plates. There are also numerous small plates (e.g., Juan de Fuca, Nazca, Scotia, Philippine, Caribbean), and many very small plates or sub-plates. The Juan de Fuca Plate is actually three separate plates (Gorda, Juan de Fuca, and Explorer), all moving in the same general direction but at slightly different rates.

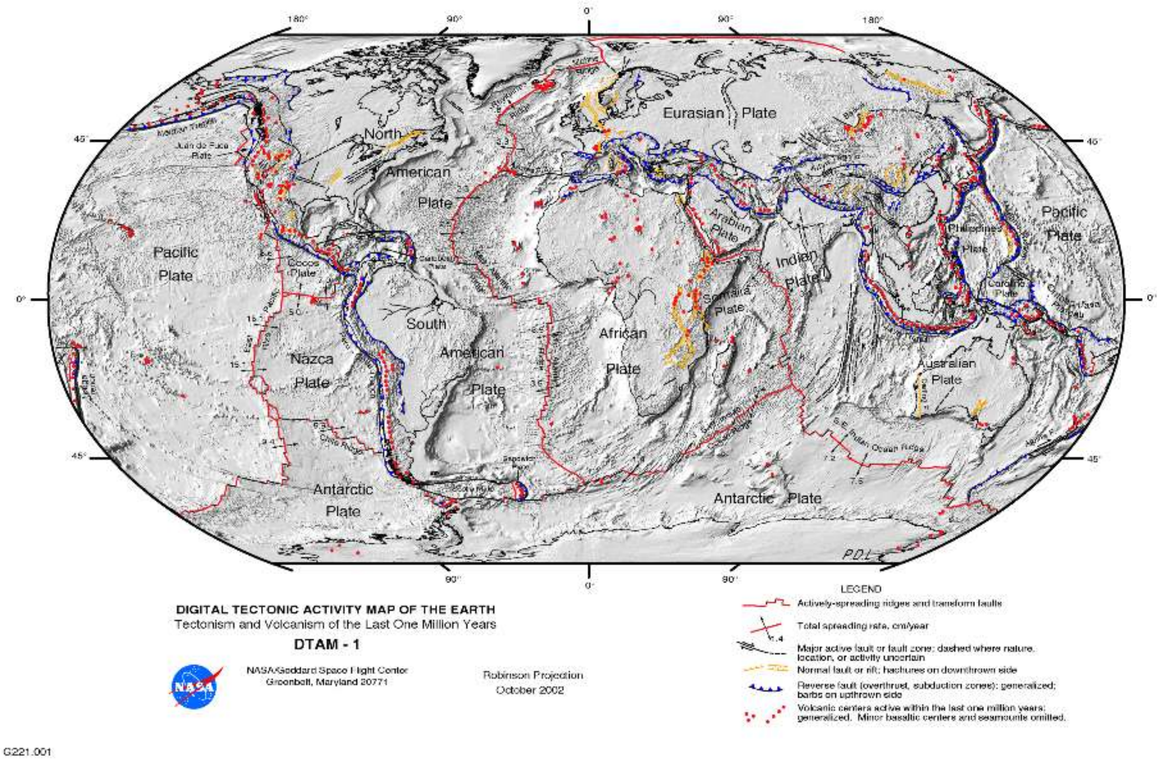


Figure 4.24 A detailed map of Earth's tectonic plates. Source: Paul Lowman and Jacob Yates, NASA Goddard Space Flight Center (2002), Public Domain. View source.

Plate motions can be tracked using Global Positioning System (GPS) data from different locations on Earth's surface. Rates of motions of the major plates range from less than 1 cm/y to more than 10 cm/y. The Pacific Plate is the fastest, moving at more than 10 cm/y in some areas, followed by the Australian and Nazca Plates. The North American Plate is one of the slowest, averaging ~1 cm/y in the south up to almost 4 cm/y in the north.

Plates move as rigid bodies, so it may seem surprising that the North American Plate can be moving at different rates in different places. The explanation is that plates can rotate as they move. The North American Plate rotates counter-clockwise, while the Eurasian Plate rotates clockwise.

The fact that plates include both crustal material and lithospheric mantle material makes it possible for a single plate to be include both oceanic and continental crust. Notice in Figure 4.24 how the North American Plate includes most of North America, plus half of the northern Atlantic Ocean. Similarly the South American Plate extends across the western part of the southern Atlantic Ocean, while the European and African plates each include part of the eastern Atlantic Ocean. The Pacific Plate is almost entirely oceanic, but it does include the part of California west of the San Andreas Fault.

Types of Plate Boundaries

Boundaries between the plates are of three types: **divergent** (moving apart), **convergent** (moving together), and **transform** (moving side by side). Although the plates are in constant motion, and move in different directions, there is never a significant amount of space between them.

Practice with Plate Boundary Types

Divergent Boundaries

Most divergent boundaries are spreading centres within oceans, where magma from partially melted mantle rock rises up and freezes to form new oceanic crust (Figure 4.25). Normally the pressure is too high in Earth's mantle to allow melting, but spreading centres are places where mantle convection is moving rocks upward, thus decreasing the pressure on them. The decrease is enough to trigger partial melting of the rock, meaning that some of the minerals in the mantle rock can begin to melt.

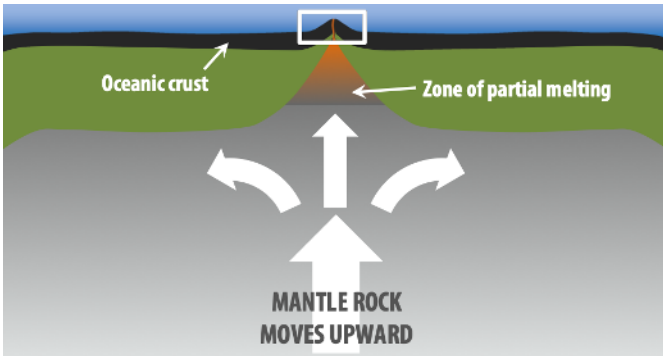


Figure 4.25 Mantle upwelling taking place along an oceanic divergent plate margin. Source: Karla Panchuk (2021), CC BY-NC-SA. Adapted from Steven Earle (2015), CC BY 4.0. View source.

The triangular zone of partial melting near the ridge crest is approximately 60 km thick and the proportion of magma is about 10% of the rock volume. This produces crust that's about 6 km thick once the melt rises up from the rock in which it formed. Crustal material created from mantle partial melts at a spreading boundary is always oceanic in character; in other words, it's mafic igneous rock (basalt or gabbro, with minerals rich in iron and magnesium).

Spreading rates vary considerably, from 1 cm/y to 3 cm/y in the Atlantic, to between 6 cm/y and 10 cm/y in the Pacific. Some of the processes taking place in this setting include (Figure 4.26):

- Melted rock (magma) from the mantle rising up to fill the voids left by divergence of the two plates
- **Pillow lavas** forming where melted rock emerges on the ocean floor and is cooled by seawater (inset)
- Vertical sheeted dykes intruding into cracks resulting from the spreading
- Magma cooling more slowly in the lower part of the new crust, forming bodies of gabbro

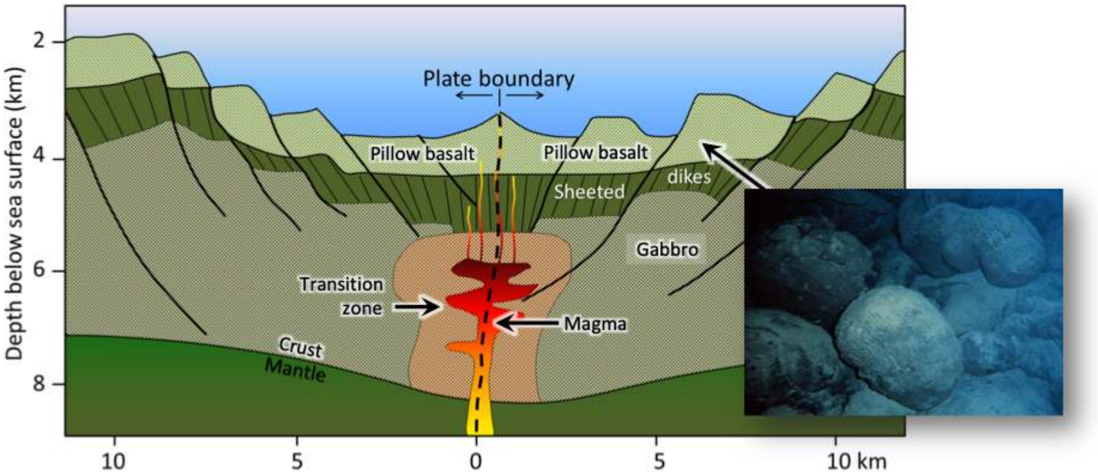


Figure 4.26 Expanded view of the white box in Figure 4.25 showing divergent boundary processes and materials. Inset: Pillow basalts from the ocean floor of Hawai'i. Source: Karla Panchuk (2021), CC BY-NC-SA 4.0. Adapted from Steven Earle (2015), CC BY 4.0. View source. (Modified after Sinton and Detrick (1992)). Inset: NOAA (1988), Public Domain. View source.

Spreading is thought to start with lithosphere being warped upward into a dome by buoyant material from an underlying mantle plume or series of mantle plumes. The buoyancy of the mantle plume causes the dome to fracture in a radial pattern, with three arms spaced at approximately 120° (Figure 4.27).

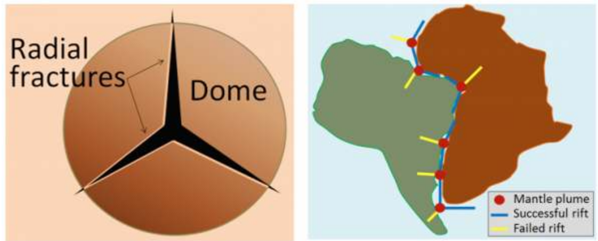


Figure 4.27 Dome and three-part rift formation (left) and continental rifting (right) between the African and South American parts of Pangea at around 200 Ma. Source: Steven Earle (2015), CC BY 4.0. View source.

When a series of mantle plumes exists beneath a large continent, the resulting rifts may align and lead to the formation of a rift valley, such as the present-day Great Rift Valley in eastern Africa (Figure 4.28). This type of valley may eventually develop into a linear sea (such as the present-day Red Sea), and finally into an ocean (such as the Atlantic). It's likely that as many as 20 mantle plumes—many of which still exist—were responsible for the initiation of the rifting of Pangea along what is now the Mid-Atlantic Ridge.

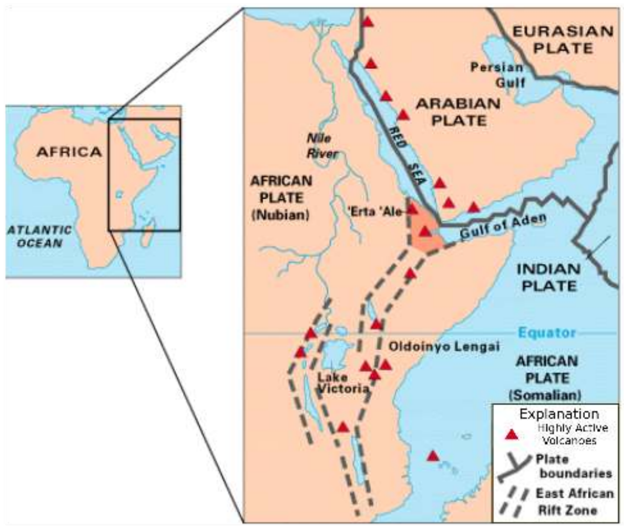


Figure 4.28 The East African Continental Rift Zone and spreading centres in the Red Sea and Gulf of Aden. If spreading continues along continental rift zones, ocean basins form. See this region in Google Earth. Source: U. S. Geological Survey (1999), Public Domain. View source.

Convergent Boundaries

Convergent boundaries are where two plates are colliding with each other. There are three types, classified according to whether ocean or continental crust is present on either side of the boundary. These are ocean-ocean, ocean-continent, and continent-continent convergent boundaries.

Ocean-Ocean Convergent Boundaries

At an ocean-ocean convergent boundary, a plate margin consisting of oceanic crust and lithospheric mantle is **subducted**, or travels beneath, the margin of the plate it's colliding with (Figure 4.29). Often it's the older and colder plate that is denser and subducts beneath the younger and hotter plate. Ocean trenches commonly form along these boundaries.

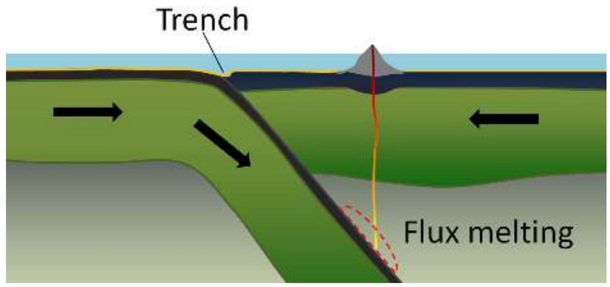


Figure 4.29 Configuration and processes of an ocean-ocean convergent boundary Source: Steven Earle (2015), CC BY 4.0. View source.

As the subducting crust is heated and the pressure increases, water is released from within the subducting material. This water comes primarily from alteration of the minerals pyroxene and olivine to serpentine near the spreading ridge shortly after the rock's formation. The water mixes with the overlying mantle, which lowers the melting point of mantle rocks, causing magma to form. This process is called **flux melting** or **fluid-induced melting**.

The newly produced magma rises through the mantle and sometimes through the overlying oceanic crust to the ocean floor where it creates a chain of volcanic islands known as an **island arc**. A mature island arc develops into a chain of relatively large islands (such as Japan or Indonesia) as more and more volcanic material is extruded and sedimentary rocks accumulate around the islands. The largest earthquakes occur near the surface where the subducting plate is still cold and strong.

Examples of ocean-ocean convergent zones are subduction of the Pacific Plate south of Alaska (Aleutian Islands, view in Google Earth) and west of the Philippines, subduction of the Indian Plate south of Indonesia, and subduction of the Atlantic Plate beneath the Caribbean Plate.

Ocean-Continent Convergent Boundaries

Subduction of an oceanic plate also happens at an ocean-continent convergent boundary, except this time the overriding plate carries continental crust. Rocks and sediment on the continental slope are thrust up into an **accretionary wedge**, and compression leads to faults forming within the continental plate (Figure 4.30). The mafic magma produced adjacent to the subduction zone rises to the base of the continental crust and leads to partial melting of the crustal rock. The resulting magma ascends through the crust, producing a mountain chain with many volcanoes.

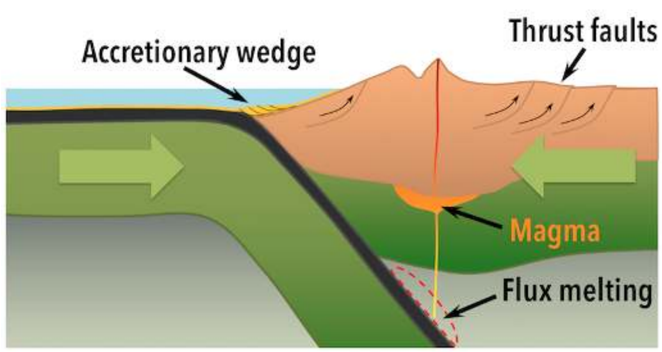


Figure 4.30 Configuration and processes of an ocean-continent convergent boundary Source: Karla Panchuk (2018), CC BY 4.0. Modified after Steven Earle (2015), CC BY 4.0. View source.

Examples of ocean-continent convergent boundaries are subduction of the Nazca Plate under South America (which has created the Andes Range) and subduction of the Juan de Fuca Plate under North America (creating the mountains Garibaldi, Baker, St. Helens, Rainier, Hood, and Shasta, collectively known as the Cascade Range).

Continent-Continent Convergent Boundary

A continent-continent collision occurs when a continent or large island that has been moving along with subducting oceanic crust collides with another continent (Figure 4.31). Prior to the continent-continent collision, the situation would have looked like the ocean-continent collision shown in Figure 4.30, except imagine that continental lithosphere is attached to the plate of oceanic lithosphere just to the left of the area shown in the image. Continent-continent collisions build on to the edges of existing continents, and can even merge two continents into a single larger one.

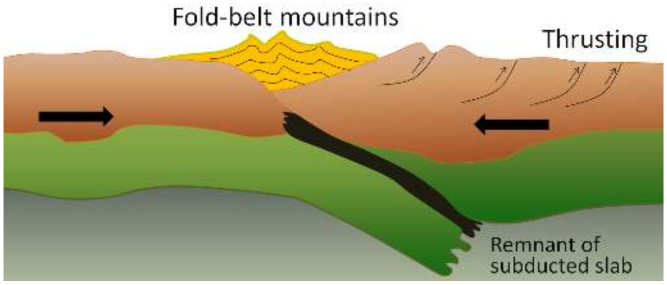


Figure 4.31 Configuration and processes of a continent-continent convergent boundary Source: Steven Earle (2015), CC BY 4.0. View source.

Continental lithosphere is too low in density to be forced into the mantle the way that oceanic lithosphere is, so subduction doesn't happen. Pre-existing continental rocks are deformed into giant mountain belts, as are any sediments that accumulated along the shores of both continental masses. Some ocean crust and upper mantle material may also be included. These mountains are not volcanic, because not only is the pressure on the mantle increased, water is no longer being added by a subduction zone.

Eventually, the edge of the ocean plate that was subducted (before the two masses of continental lithosphere collided) will break off and sink into the mantle. When this happens, the weight of rock in the collision zone is suddenly reduced, and the mountain belt can spring upward and float higher in the mantle (like your air mattress in a swimming pool once you push your friend off).

Examples of continent-continent convergent boundaries are the collision of the India Plate with the Eurasian Plate, creating the Himalaya mountain range (view in Google Earth), and the collision of the African Plate with the Eurasian Plate, creating the series of ranges extending from the Alps in Europe (view in Google Earth) to the Zagros Mountains in Iran.

When a subduction zone is jammed shut by a continent-continent collision, plate tectonic stresses that are still present can sometimes cause a new subduction zone to develop outboard of the colliding plate.

Transform Boundaries

Transform boundaries exist where—in an ideal scenario—one plate slides past another without producing or destroying crust. In situations where the transform boundary has bends and jogs, however, there will be small-scale collisions and divergence as the jogs crash into the bends, or open up small windows to deeper crust.

Most transform faults connect segments of mid-ocean ridges and are thus ocean-ocean plate boundaries. Notice where the red segments in Figure 4.32 offset the black segments marking mid-ocean ridges. Some transform faults connect continental parts of plates. The San Andreas Fault connects the southern end of the Juan de Fuca Ridge with the northern end of the East Pacific Rise (a ridge) in the Gulf of California. The part of California west of the San Andreas Fault and all of Baja California are on the Pacific Plate (Figure 4.33). But transform faults don't just connect divergent boundaries; the Queen Charlotte Fault connects the north end of the Juan de Fuca Ridge, starting at the north end of Vancouver Island, to the Aleutian subduction zone.

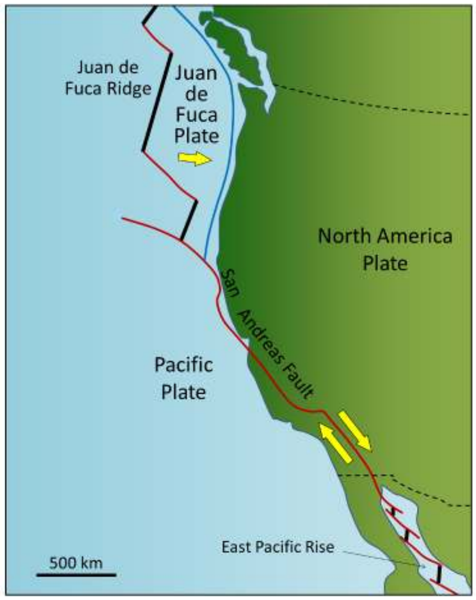


Figure 4.32 The San Andreas Fault extends from the north end of the East Pacific Rise in the Gulf of California to the southern end of the Juan de Fuca Ridge. All of the red lines on this map are transform faults. Black lines are divergent margins, and the blue line is a convergent margin. Source: Steven Earle (2015), CC BY 4.0. View source.

Figure 4.33 The San Andreas Fault at Parkfield in central California. The person with the orange shirt is standing on the Pacific Plate and the person at the far side of the bridge is on the North American Plate. The bridge is designed to slide on its foundation. Source: Steven Earle (2015), CC BY 4.0. View source.

Practice with Plate Boundary Features

Plate Tectonics and Supercontinent Cycles

The present continents were once all part of a supercontinent that Alfred Wegener named **Pangea** (*all land*). More recent studies of continental matchups and the magnetic ages of ocean-floor rocks have enabled us to reconstruct the history of the break-up of Pangea.

Pangea began to rift apart along a line between Africa and Asia and between North America and South America at around 200 Ma (Figure 4.34). During the same period the Atlantic Ocean began to open up between northern Africa and North America, and India broke away from Antarctica. Between 200 and 150 Ma, rifting started between South America and Africa and between North America and Europe, and India moved north toward Asia. By 80 Ma, Africa had separated from South America, and most of Europe had separated from North America. By 50 Ma, Australia had separated from Antarctica, and shortly after that, India collided with Asia.

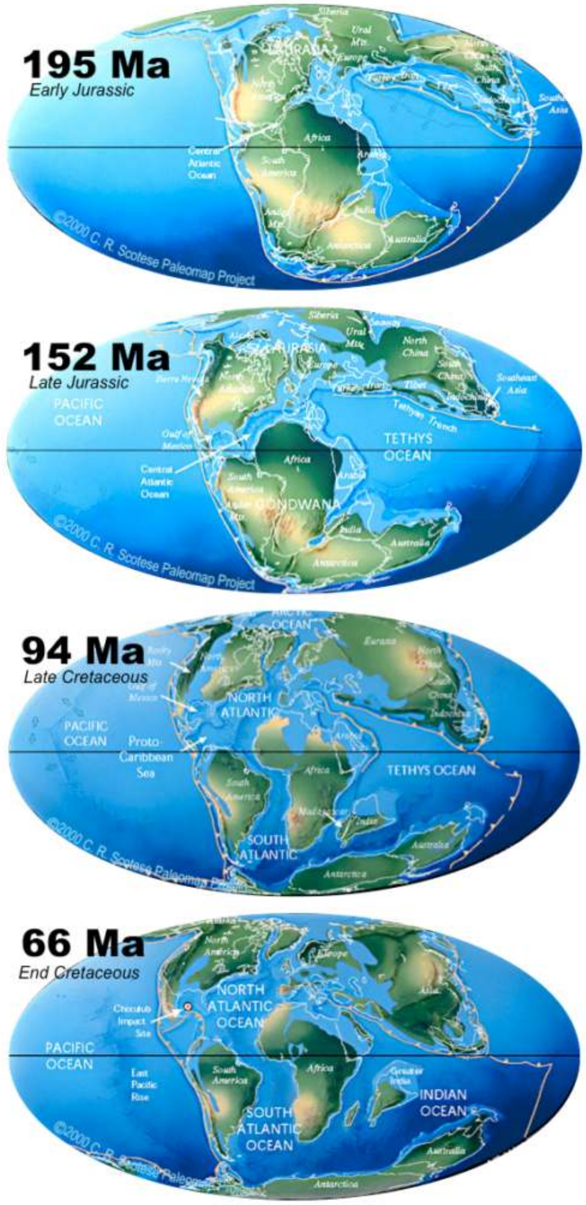


Figure 4.34 Sequence of paleogeographic reconstructions showing the breakup of Pangea. Source: Karla Panchuk (2017) CC BY-NC-SA 4.0. Maps from C. R. Scotese, PALEOMAP Project (www.scotese.com). Click for map sources and terms of use.

Within the past few million years, rifting has occurred in the Gulf of Aden and the Red Sea, and also within the Gulf of California. Incipient rifting has begun along the Great Rift Valley of eastern Africa, extending from Ethiopia and Djibouti on the Gulf of Aden (Red Sea) all the way south to Malawi.

Pangea was not the first supercontinent. It was preceded by Pannotia (600 to 540 Ma), Rodinia (1,100 to 750 Ma), and by others before that. In fact, in 1966, Tuzo Wilson proposed that supercontinents are part of an on-going cycle, which we now refer to as a **Wilson cycle**. In a Wilson cycle, continents break up, and fragments drift apart only to collide again and make a new continent.

At present we are in the stages of a Wilson cycle where fragments are drifting and changing their configuration. North and South America, Europe, and Africa are moving with their respective portions of the Atlantic Ocean. The eastern margins of North and South America and the western margins of Europe and Africa are called passive margins because there is no subduction taking place along them. Because the oceanic crust formed by spreading along the mid-Atlantic ridge is not currently being subducted (except in the Caribbean), the Atlantic Ocean is slowly getting bigger, and the Pacific Ocean is getting smaller.

This situation may not continue for too much longer, however. As the Atlantic Ocean floor gets weighed down around its margins by great thickness of continental sediments, it will be pushed farther and farther into the mantle, and eventually the oceanic lithosphere may break away from the continental lithosphere and begin to subduct (Figure 4.35).

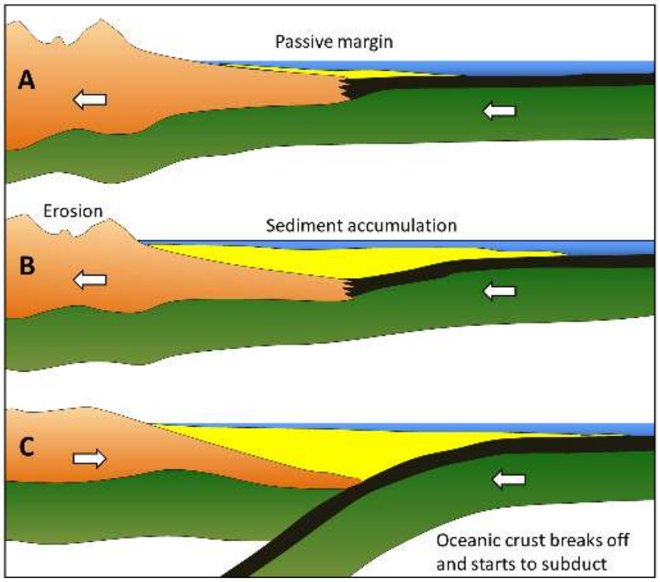


Figure 4.35 Development of a subduction zone at a passive margin. Times A, B, and C are separated by tens of millions of years. Once the oceanic crust breaks off and starts to subduct, the continental crust (North America in this case) may no longer be pushed to the west and could start to move east because the rate of spreading in the Pacific basin is faster than along the Mid-Atlantic Ridge. Source: Steven Earle (2015), CC BY 4.0. View source.

A subduction zone will develop, and the oceanic plate will begin to descend under the continent. Once this happens, the continents will no longer continue to move apart because the spreading at the mid-Atlantic ridge will be taken up by subduction. If spreading along the mid-Atlantic ridge continues to be slower than spreading within the Pacific Ocean, the Atlantic Ocean will start to close up, and eventually (in a 100 million years or more) North and South America will collide again with Europe and Africa. If this continues without changing for another few hundred million years, we will be back to where we started, with one supercontinent (Figure 4.36).

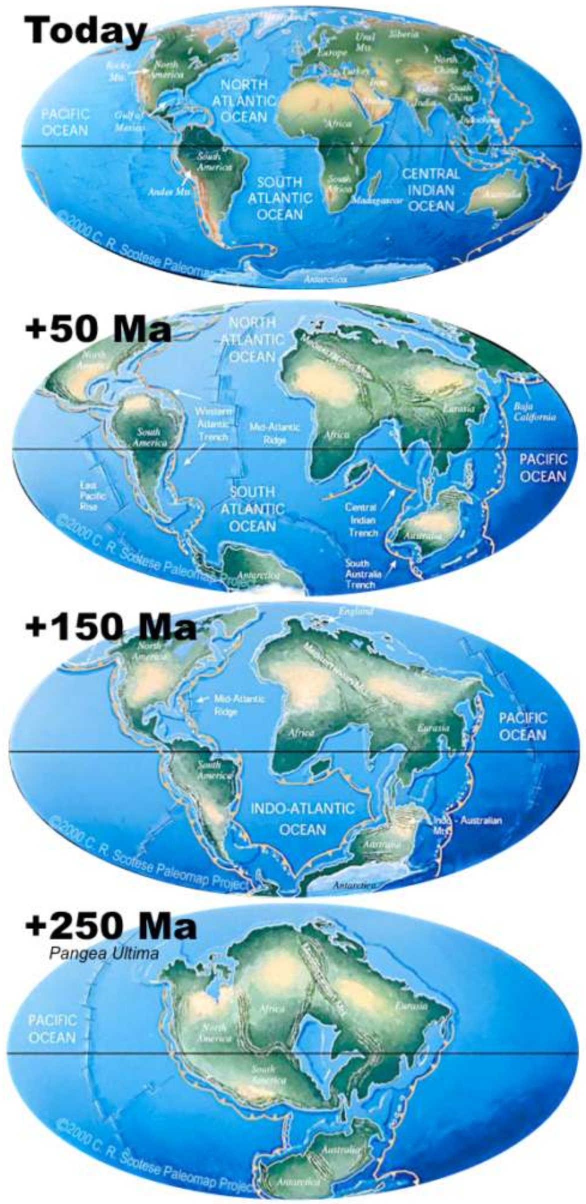


Figure 4.36 Sequence of reconstructions showing the possible future configuration of land masses on Earth at 50, 150, and 250 million years from now. Movements culminate in the formation of a new supercontinent called Pangea Ultima. Source: Karla Panchuk (2017) CC BY-NC-SA 4.0. Maps from C. R. Scotese, PALEOMAP Project (www.scotese.com). Click for map sources and terms of use.

There is strong evidence around the margins of the Atlantic Ocean that this process has taken place before. There are roots of ancient mountain belts along the eastern margin of North America, the western margin of Europe, and the north-western margin of Africa, which show that these landmasses once collided with each other to form a mountain chain. The mountain chain might have been as big as the Himalayas.

The apparent line of collision runs between Norway and Sweden, between Scotland and England, through Ireland, through Newfoundland and the Maritimes, through the north-eastern and eastern states, and across the northern end of Florida. When rifting of Pangea started at approximately 200 Ma, the fissuring was along a different line from the line of the earlier collision. This is why some of the mountain chains formed during the earlier collision can be traced from Europe to North America and from Europe to Africa.

It is probably no coincidence that the Atlantic Ocean rift may have occurred in approximately the same place during two separate events several hundred million years apart. The series of hot spots that has been identified in the Atlantic Ocean may also have existed for several hundred million years, and thus may have contributed to rifting in roughly the same place on at least two separate occasions (Figure 4.37).

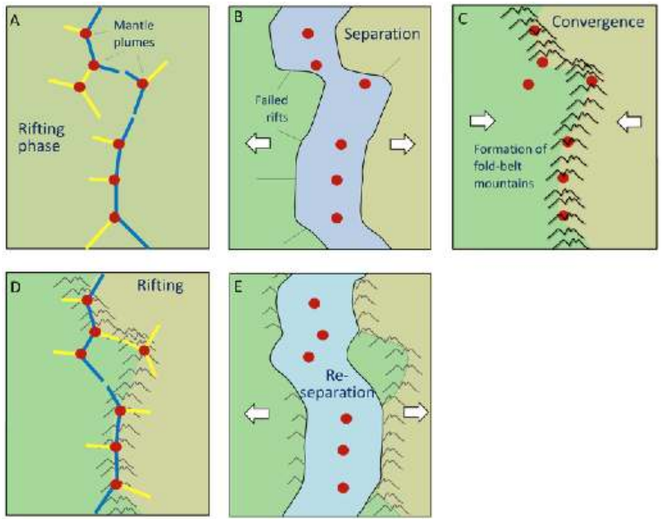


Figure 4.37 A scenario for the Wilson cycle. The cycle starts with continental rifting above a series of mantle plumes (red dots, A). The continents separate (B), and then re-converge some time later, forming a fold-belt mountain chain. Eventually rifting is repeated, possibly because of the same set of mantle plumes (D), but this time the rift is in a different place. Source: Steven Earle (2015), CC BY 4.0. View source.

References

Sinton, J. M., and Detrick, R. S. (1992). Mid-Ocean Ridge Magma Chambers. *Journal of Geophysical Research 97*(B1), 197-216.

13.2: Basins Formed in Extensional Settings

13.3: Basins Caused by Crustal Loading

13.4: Other Areas of Sediment Accumulation

CHAPTER OVERVIEW

14: Appendices

Topic hierarchy

Chapter thumbnail shows An outcrop-scale sketch (1878) and photograph (2021) of the unconformity between Cretaceous sedimentary rocks and much younger volcanic rocks in Yellowstone National Park (Geology of the unconformity on Mount Everts in Yellowstone National Park via USGS; public domain).

14.1: Geologic Sketches

14.1.1: The importance of sketching

Sketches are an effective way for geologists to quickly and effectively convey key geologic observations and relationships. They began out of necessity at a time when the only way to convey what you were seeing was a sketch or a written description. Although technology now allows us to take photographs and even make three dimensional models, sketches remain an important way to capture information because they allow the creator to highlight things that are important, de-emphasize things that are not, and convey what it all means. The example in Figure 1 shows an outcrop-scale sketch; other common types/scales including: views through a microscope (see 5: Siliciclastic Sedimentary Rocks and 6: Carbonate Sedimentary Rocks), hand samples (Figure 2), general notes (Figure 3), sketch maps (Figure 4), and measured sections (Figure 5).

Sketching is a skill that can be learned and does not require someone to be a gifted artist; it just requires neatness and a plan. In fact, a good sketch does not require photo-realism, but rather a simplified portrayal that conveys geologic realism. In many cases, annotations on sketches are of greater or equal value than the artwork itself.

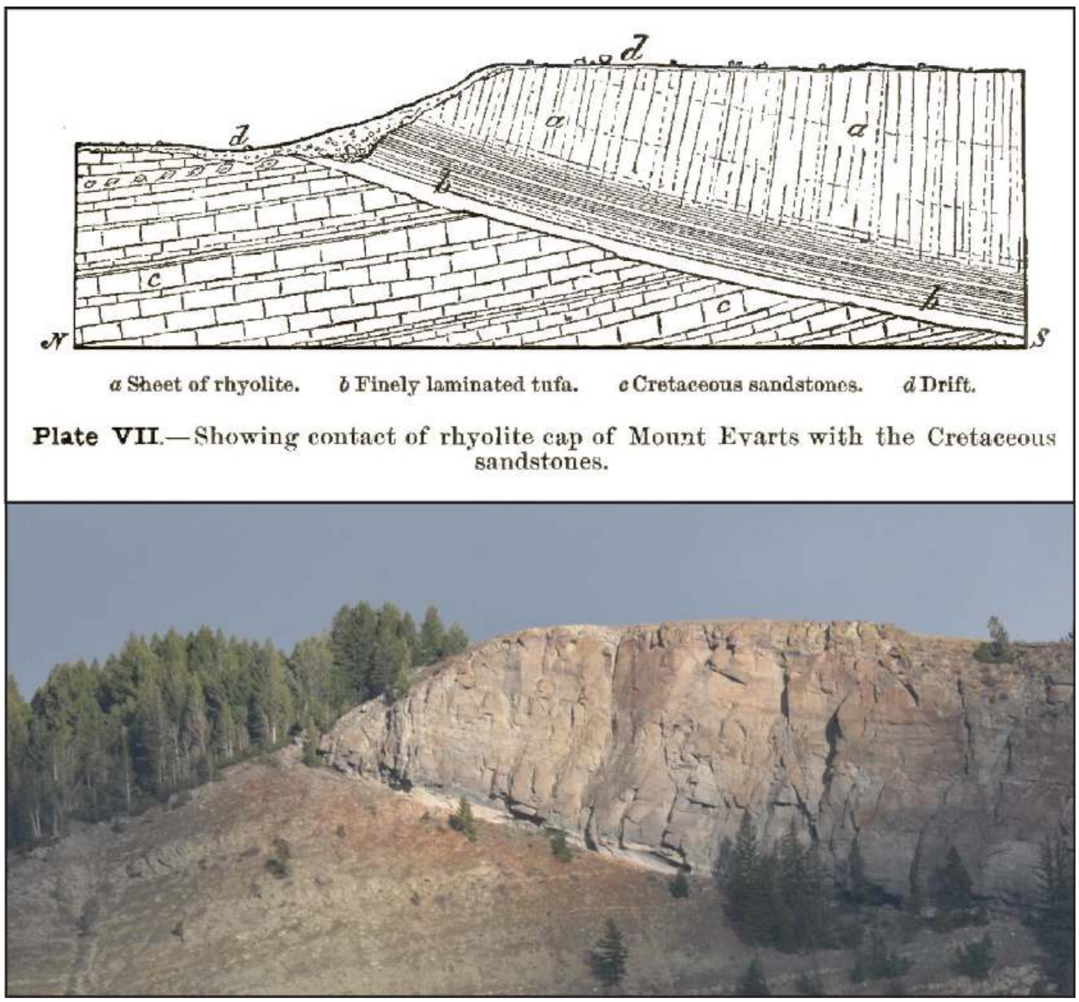


Figure 14.1.1: An outcrop-scale sketch (1878) and photograph (2021) of the unconformity between Cretaceous sedimentary rocks and much younger volcanic rocks in Yellowstone National Park. In many ways, the sketch is the more valuable tool for conveying geologic information because it emphasizes things that are most important and allows the use of artistic license to show things that might not be obvious from a distance (Geology of the unconformity on Mount Everts in Yellowstone National Park via USGS; public domain).

14.1.2: Notebook best practices

A geologic sketch is just one part of the geologist's notebook. No matter whether you are in the field, in a lab, or looking through a microscope, a notebook is an extremely important tool for geologic data collection. The details will vary based on the task, but in general, one's notebook entry for a day/project should include:

- A descriptive heading
- Day, date, time
- Location - as appropriate this could include coordinates, sketch map, access details, etc.
- Overview of goals/objectives
- Names of field party or collaborators
- Notes on weather, challenges, or anything extraordinary that might help you recall the specifics of that day.
- Pencil is generally much better for making detailed sketches because it allows you to erase.

14.1.3: How to make a sketch

1. Identify the subject of your sketch and think about what is most important about it.
2. Define the limits of the subject and orient your notebook in a way that matches with the layout of the subject. Paper is relatively inexpensive and its typically worth it to center the drawing in the page and give yourself ample margins for annotations.
3. Frame the subject by using a few simple strokes to outline its basic shape in the area you've allotted for the drawing. Don't start with detailed lines, that is a great way to run out of room or end up with inaccurate proportions. Instead, begin with a few simple strokes and then come back and modify them if needed. Things like edges of a feature, the skyline, and/or the boundaries with the background and foreground are good things to start with.
4. Once you've framed the edges of the subject, you should start making lines that represent the most important geologic features. You can and should use artistic license to show things that might not be obvious given the perspective that you are drawing from. Most commonly the major geologic lines are going to be things like:
    - Stratification - to shows changes in orientation and boundaries between units
    - Edges of grains, clasts, fossils to show things like size, shape, orientation, and changes in these characteristics
    - Structural features like folds, faults, and cleavage - these features are often easier to draw if done in combination the things listed above.
    - Remember that you should show only the most important features.
5. Remember that if you can't show everything that you need to with a sketch of a given scale that you can use things like pop-out windows to illustrate details.
6. Use subtle labels or lines/shading to show things like talus and vegetation to help the viewer link the drawing to the view. You can also use geologically-appropriate fill patterns to help show lithology, texture, etc. Avoid cross-hatching and other bold geometric patterns because they can confuse the reader.
7. Annotate your drawing to make it obvious to yourself and others what you are drawing and why. If you can show everything that you need to, remember that you can have windows that show enlarged areas.
8. Give the drawing a title, a vertical and horizontal scale bar, and a label to indicate perspective (facing north, map view, cross-sectional view, etc.). Be sure to record structural data if appropriate

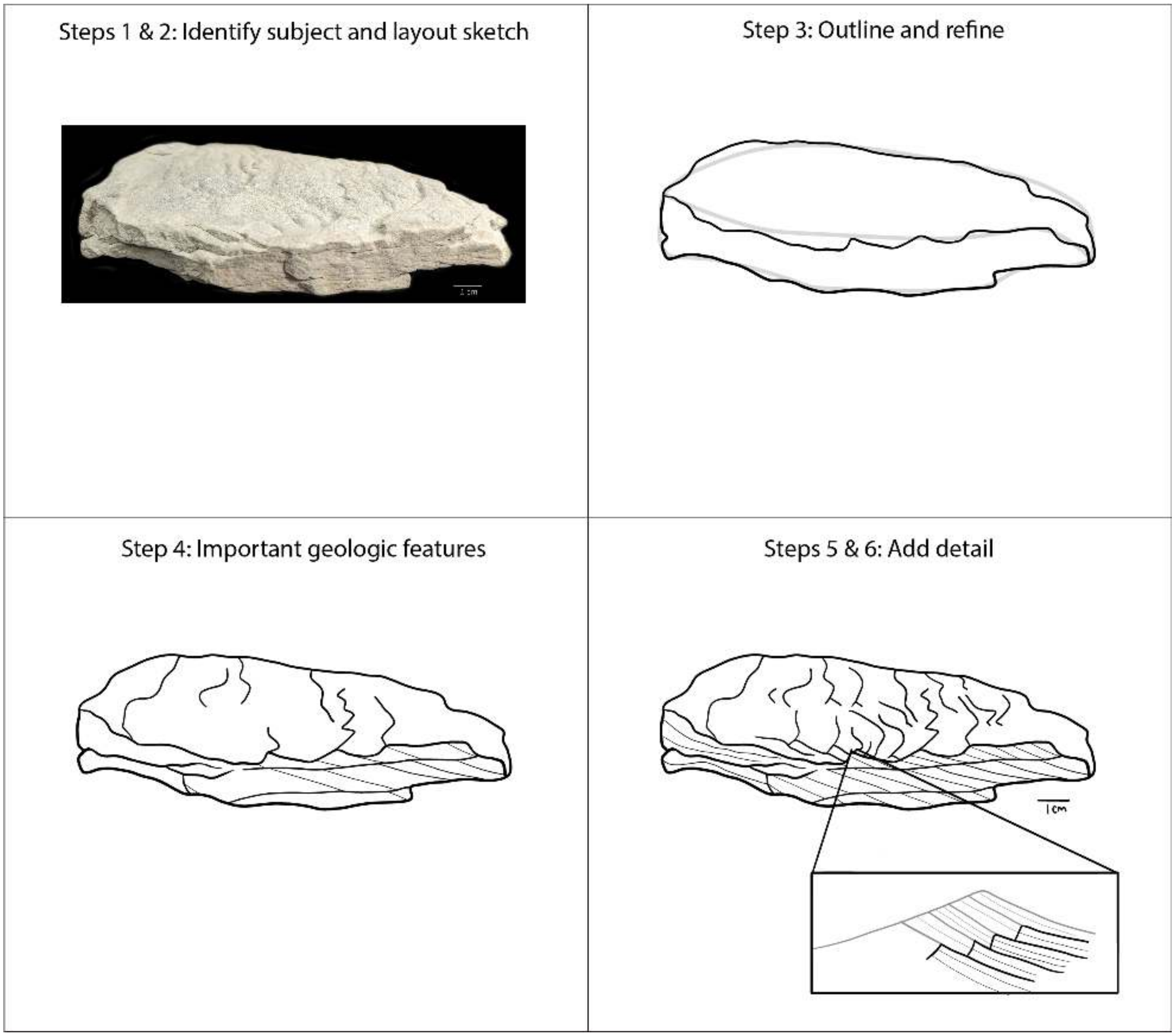


Figure $14.1.2a$: Initial steps in creating a geologic sketch

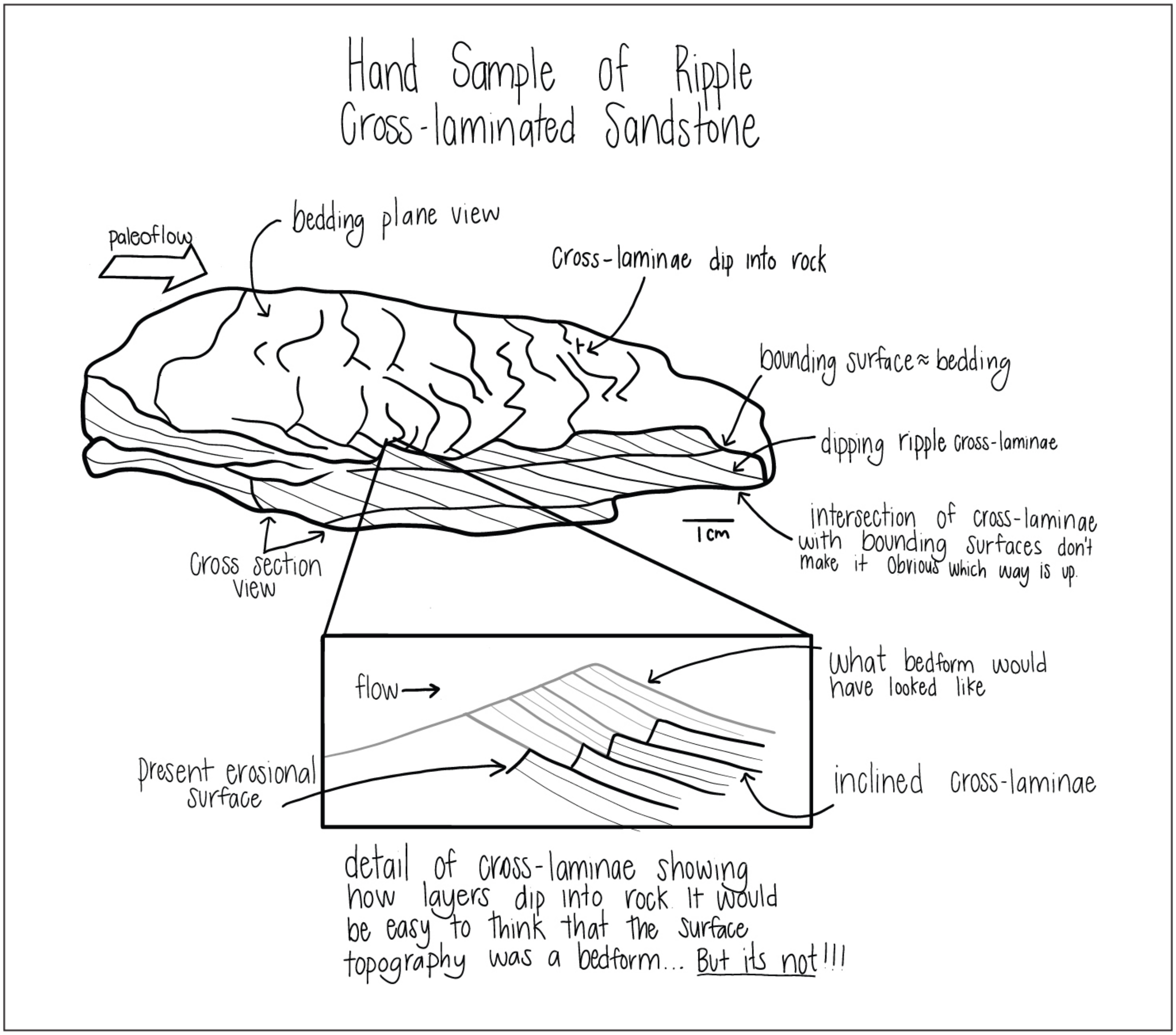


Figure 14.1.2*b*: Finished geologic sketch. The addition of annotations adds value to the sketch by recording interpretive infromation and enhancing the visuals.

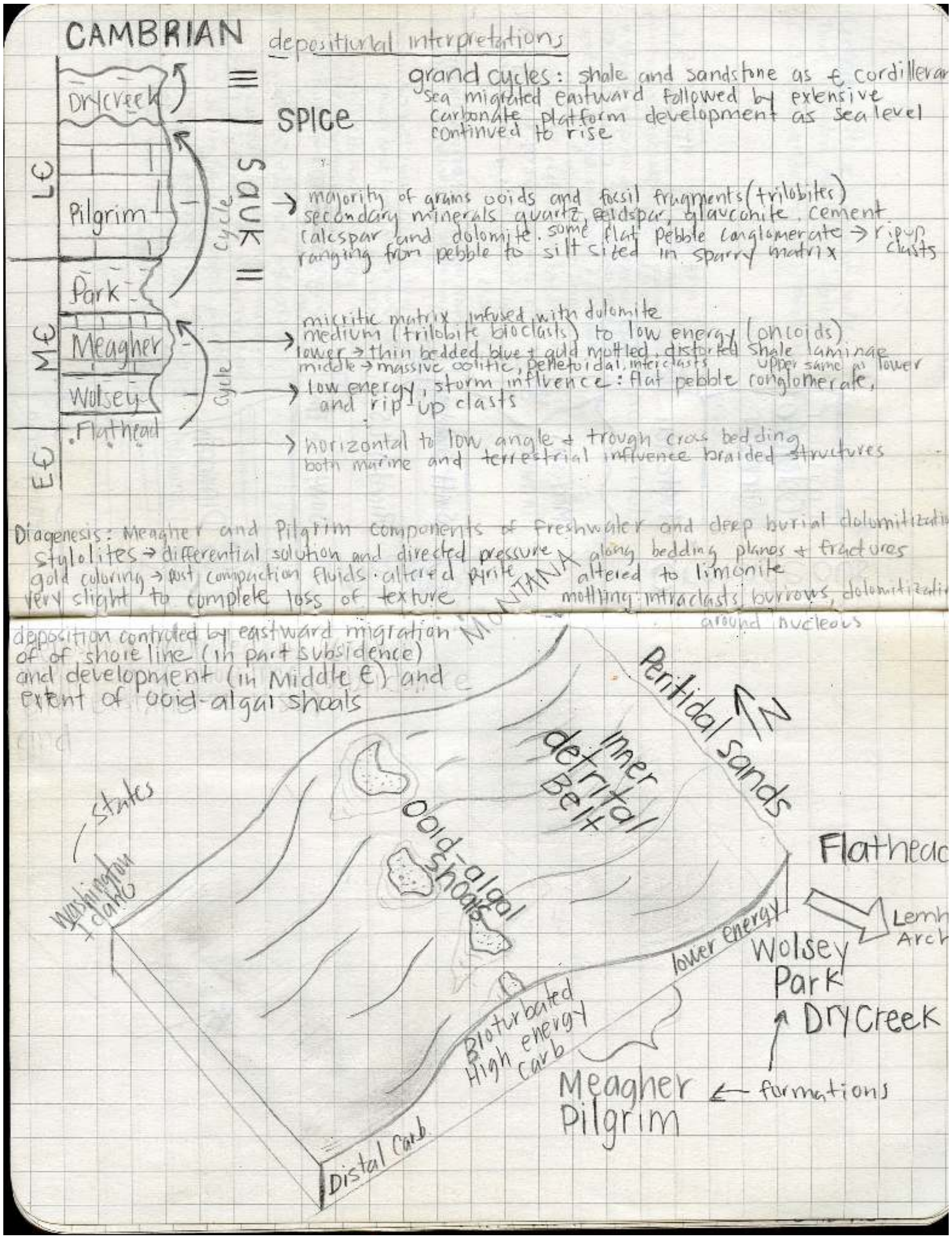


Figure 14.1.3: General notes and sketches summarizing the early Paleozoic stratigraphy of the northern Rocky Mountains.

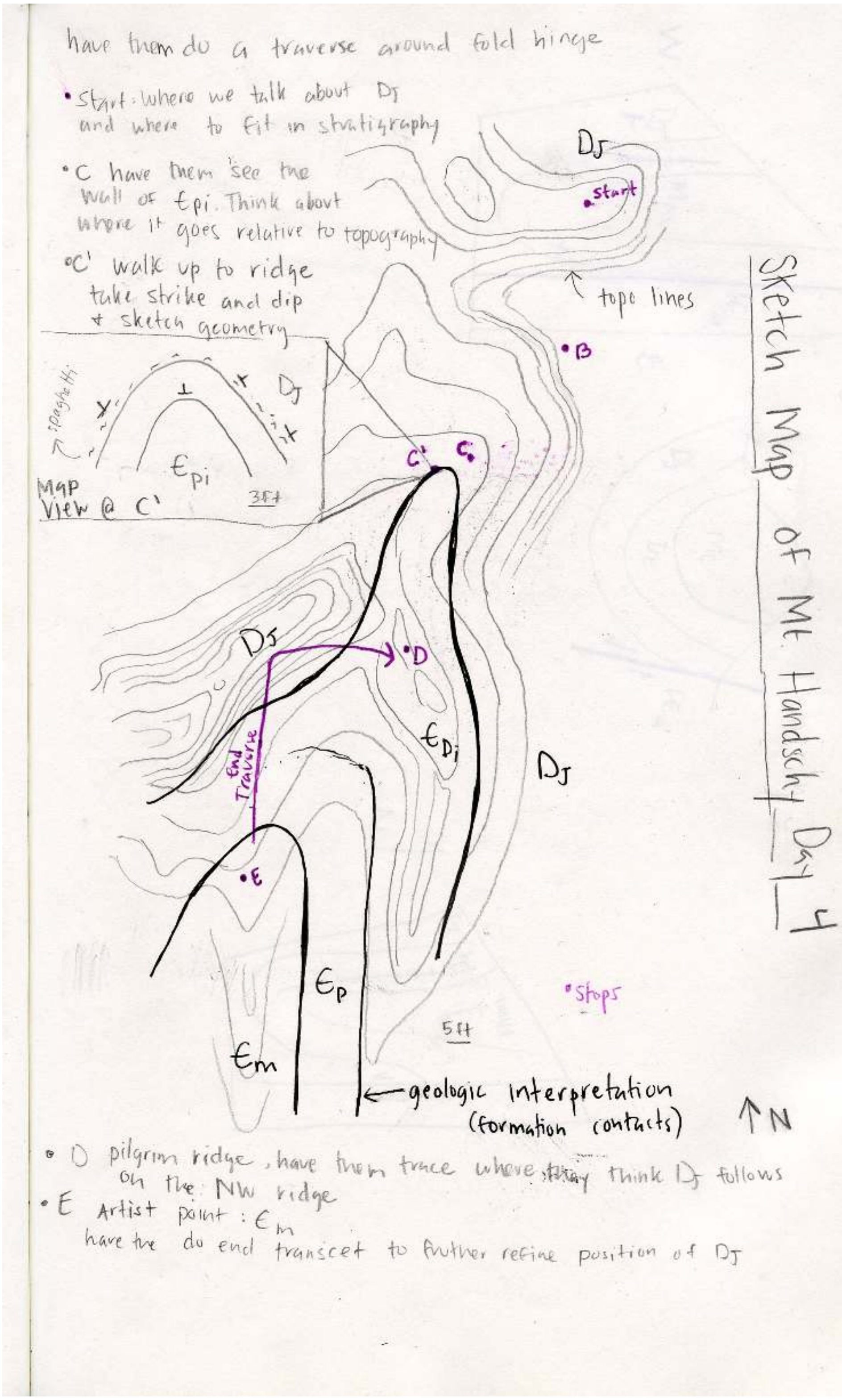


Figure 14.1.4: Sketch map summarizing a day of geologic mapping. Details include route, topography, and geology.

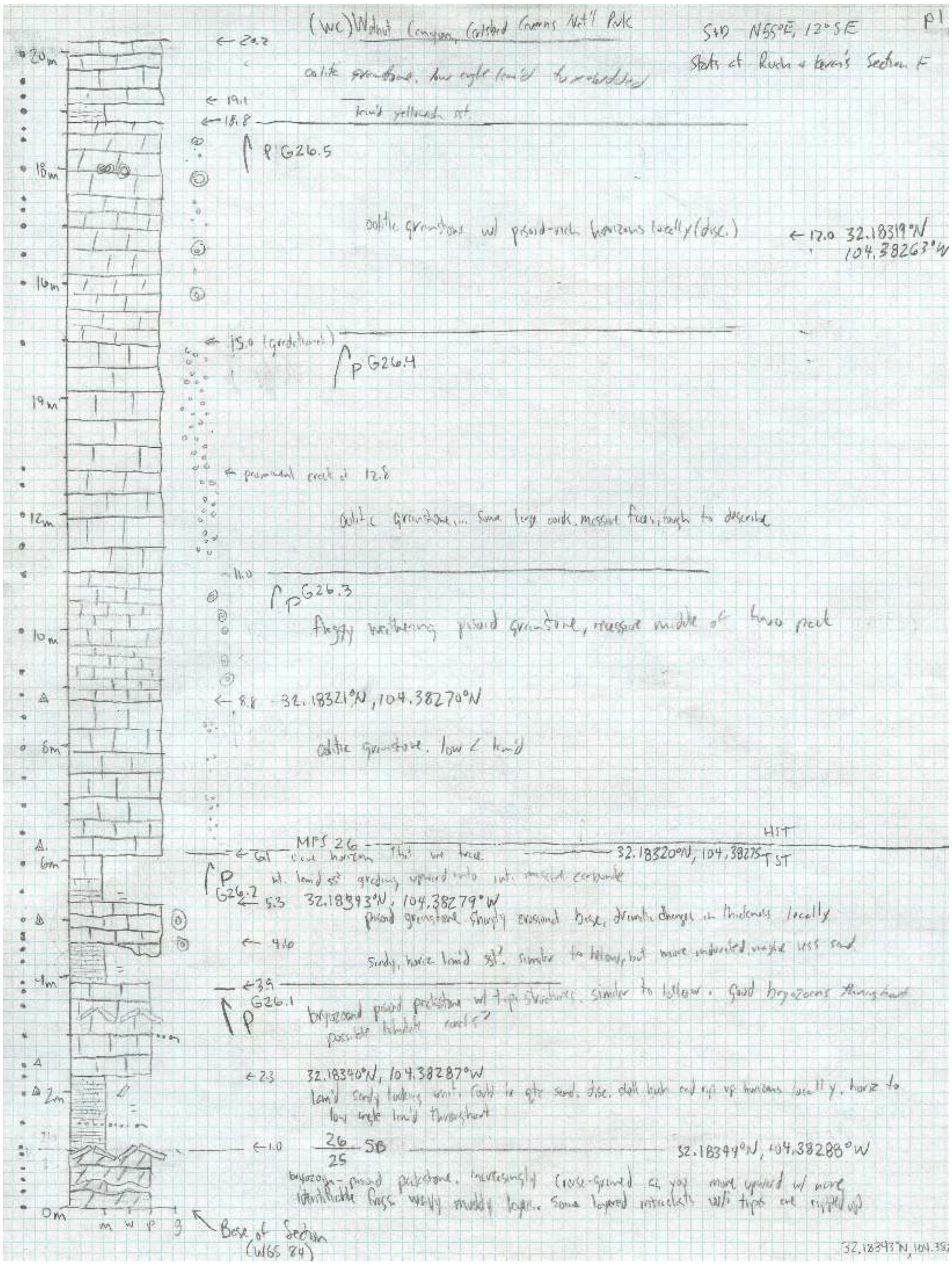
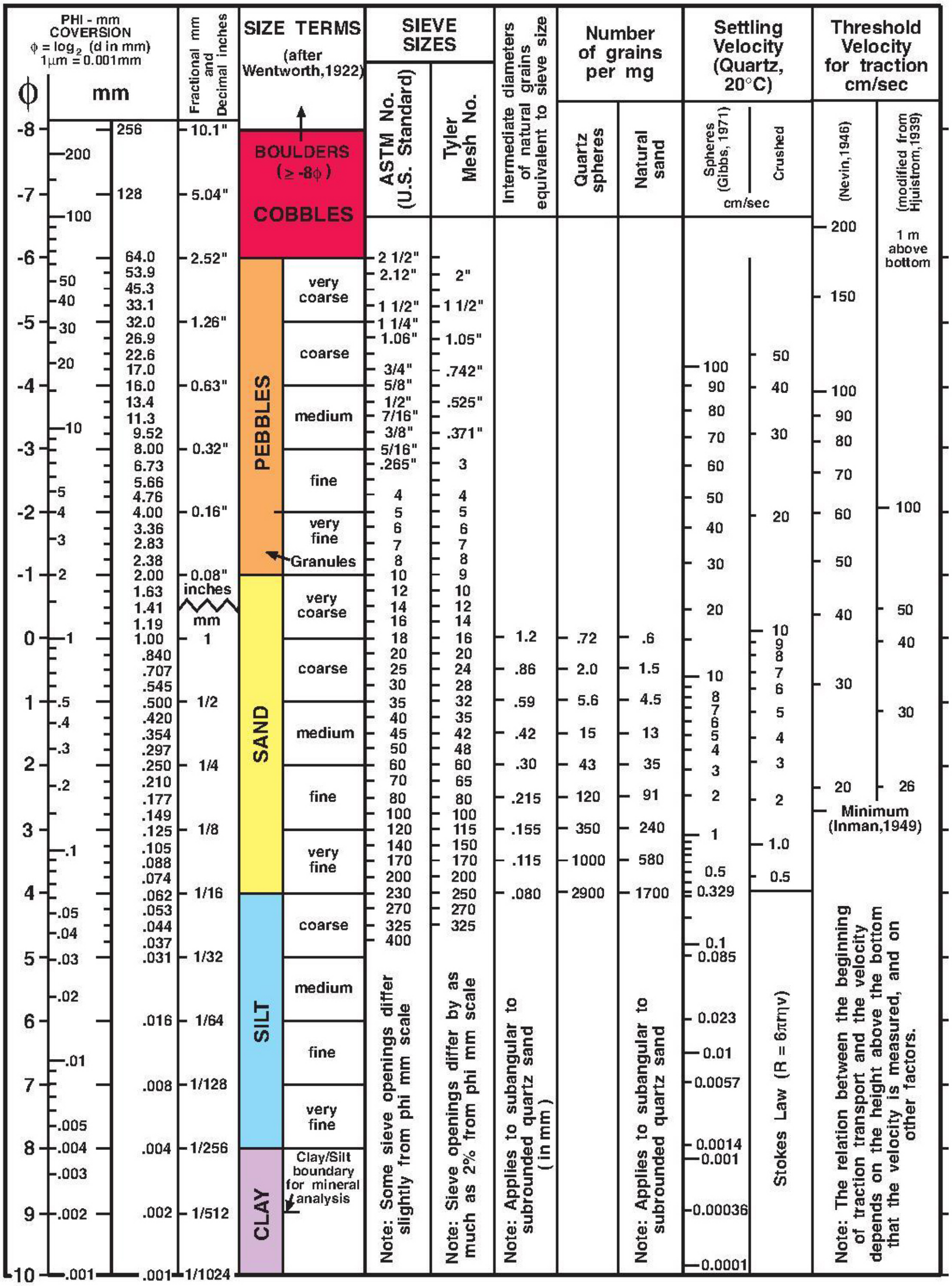


Figure 14.1.4: A graphic measured section showing unit thicknesses, rock types, fossils, and sedimentary structures.

14.2: Wentworth Grain Size Scale



Wentworth grain size chart modified to include particle diameters in inches and phi units as well as standard sieve sizes and other useful parameters. Diagram is from Williams, S.J., Arsenault, M.A., Buczkowski, B.J., Reid, J.A., Flocks, J.G., Kulp, M.A., Penland, S., and Jenkins, C.J., 2006, Surficial sediment character of the Louisiana offshore Continental Shelf region: a GIS Compilation, U.S. Geological Survey Open-File Report 2006-1195, online at http://pubs.usgs.gov/of/2006/1195/index.htm and content is in the public domain.

14.3: Well Log Interpretation

14.4: Book Content Mapped to ASBOG's Content Domain C

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