Chapter 12: Ocean Sediments

Learning Objectives

After reading this chapter you should:

- know how sediments are classified based on physical characteristics (size, sorting etc.)
- identify the four main sources of marine sediments
- differentiate between organisms that produce different biogenous sediments
- understand the factors that determine the distribution of sediment types in the ocean
- know the different ways sediment samples can be obtained
- understand how biogenous sediments can be used to reconstruct past climate change

Let's be honest; for the majority of people with an interest in the oceans and oceanography it is not the allure of the sediments that first grabs their attention. At first glance the muddy seafloor may not seem that interesting, but the sediments play a vital role in marine ecosystems and our understanding of ocean and geological processes. The sediments provide habitat for a multitude of marine organisms, and they contain information about past climates, plate tectonics, ocean circulation patterns, and the timing of major extinctions, just to name a few. In this chapter we will examine the major types of sediments, and their distribution on the ocean floor.

12.1 Classifying Sediments

The term "sediment" refers to the tiny particles of rocks and other materials that sink to the ocean floor and eventually settle and accumulate on the bottom. All regions of the seafloor contain some form of sediment, although there are many different types of sediments from a variety of sources, and the amount of accumulated sediment can vary greatly from place to place. Globally, ocean sediments average about 1 km thick, but they can exceed 15 km thick in areas of high accumulation (Figure 12.1.1). These areas include regions near the mouths of rivers where there is high sediment discharge, and passive margins near the continents where the seafloor has had millions of years for sediment to accumulate. On the other hand, sediments are sparse along divergent plate boundaries where new oceanic crust is being formed, as the crust is too new for significant accumulation (see section 4.5), and in the central oceans that are far away from any significant sediment sources.

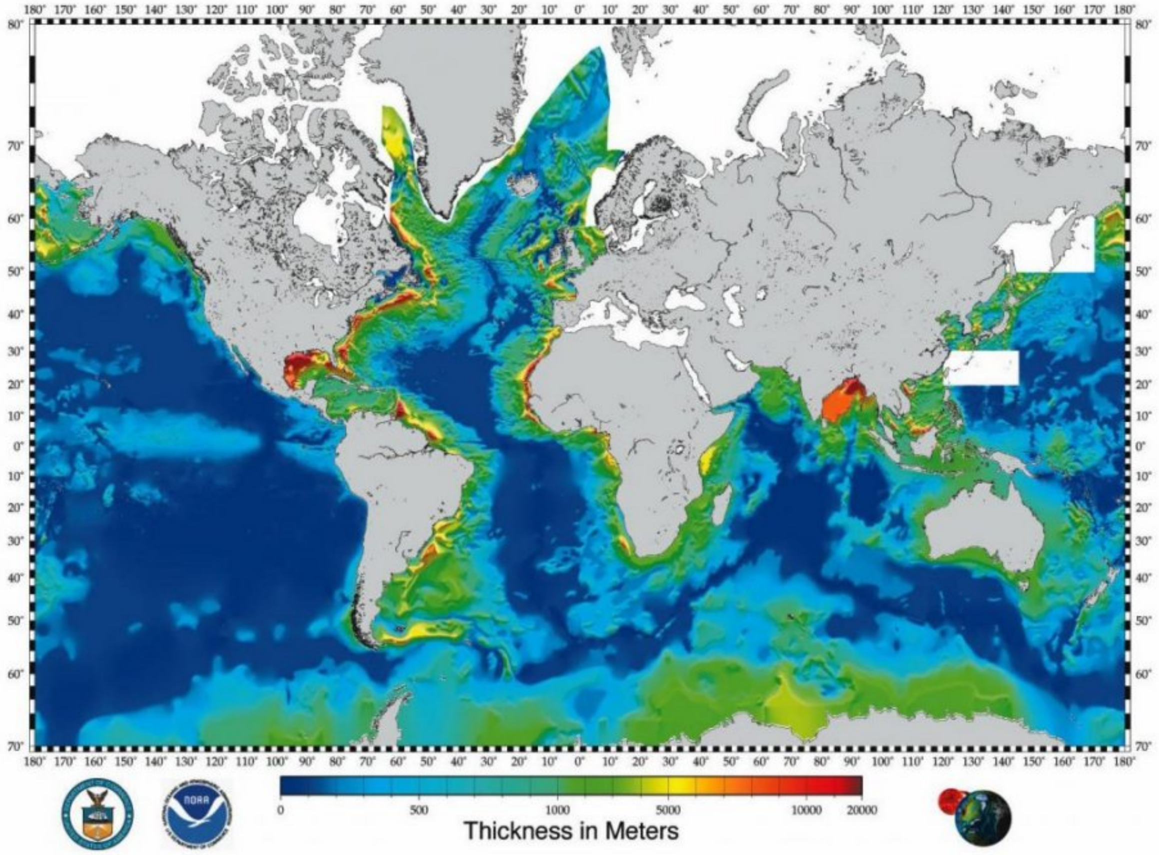

Figure 12.1.1 *Total sediment thickness of the world ocean (By Divins, D.L., NGDC Total Sediment Thickness of the World's Oceans & Marginal Seas, http://www.ngdc.noaa.gov/mgg/sedthick/sedthick.html (http://www.ngdc.noaa.gov/mgg/image/sedthick9.jpg) [Public domain], via Wikimedia Commons).*

As time passes over millions of years, these sediments can become **lithified** or turned into sedimentary rock. It has been estimated that over half of the exposed rock on the continents is sedimentary rock originally deposited in ancient oceans and uplifted by plate tectonics. Many tall mountains, including Mt. Everest, are composed of rock formations that contain fossils of marine creatures. These rocks were originally formed as ocean sediments which were then lithified and pushed upwards during the process of mountain formation.

Sediment Classification

There are a number of ways that we can classify ocean sediments, and some of the most common distinctions are based on the sediment texture, the sediment composition, and the sediment's origin.

Texture

Sediment texture can be examined through several variables. The first is **grain size**. Sediments are classified by particle size, ranging from the finest clays (diameter <0.004 mm) to the largest boulders (> 256 mm)(Figure 12.1.2). Among other things, grain size represents the conditions under which the sediment was deposited. High energy conditions, such as strong currents or waves, usually results in the deposition of only the larger particles as the finer ones will be carried away. Lower energy conditions will allow the smaller particles to settle out and form finer sediments.

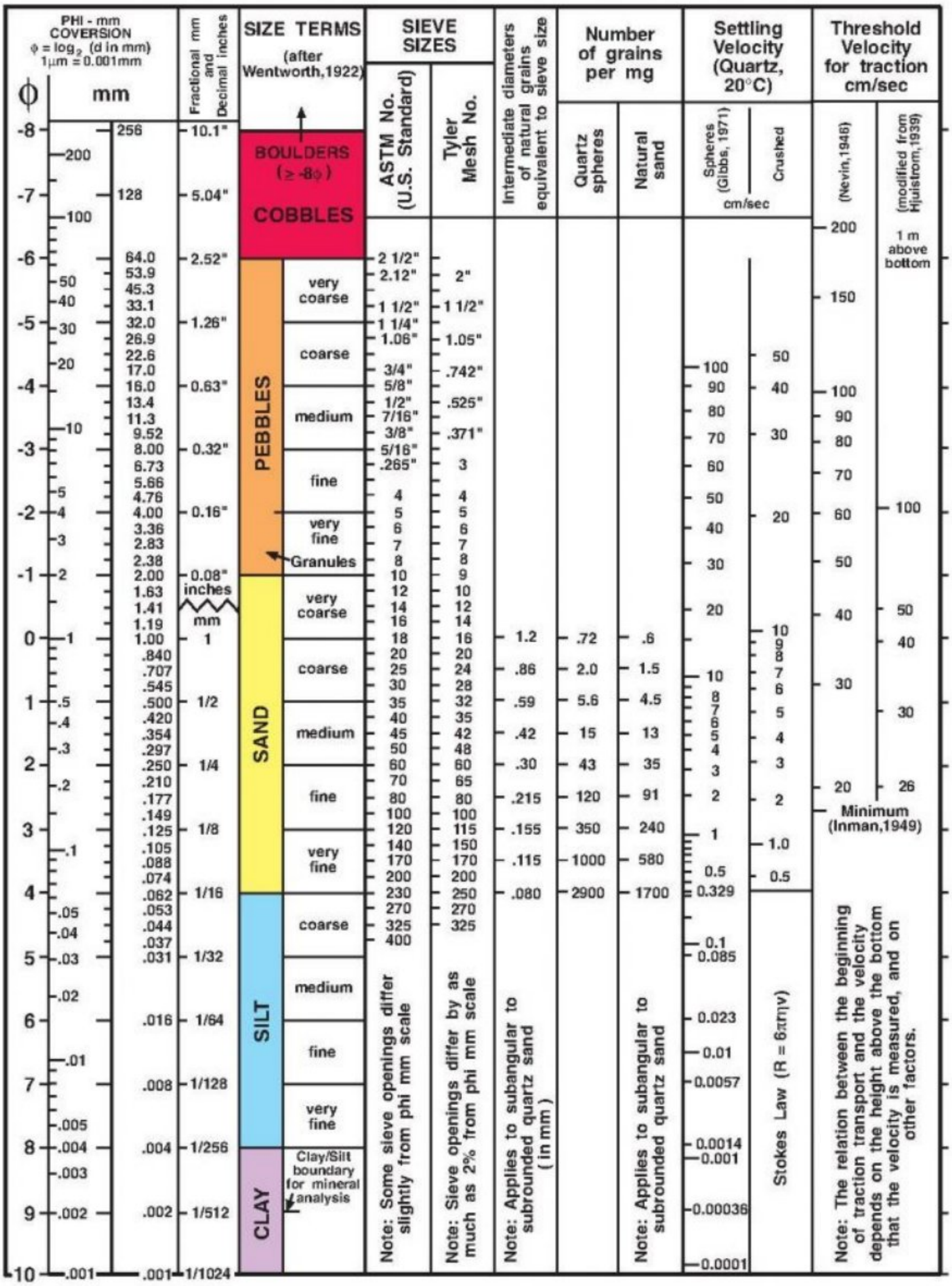

Figure 12.1.2 *Wentworth grain size chart for classifying sediments (By Jeffress Williams, Matthew A. Arsenault, Brian J. Buczkowski, Jane A. Reid, James G. Flocks, Mark A. Kulp, Shea Penland, and Chris J. Jenkins, USGS [Public domain], via*

Wikimedia Commons).

Sorting is another way to categorize sediments. Sorting refers to how uniform the particles are in terms of size (Figure 12.1.3). If all of the particles are of a similar size, such as in beach sand, the sediment is well-sorted. If the particles are of very different sizes, the sediment is poorly sorted, such as in glacial deposits.

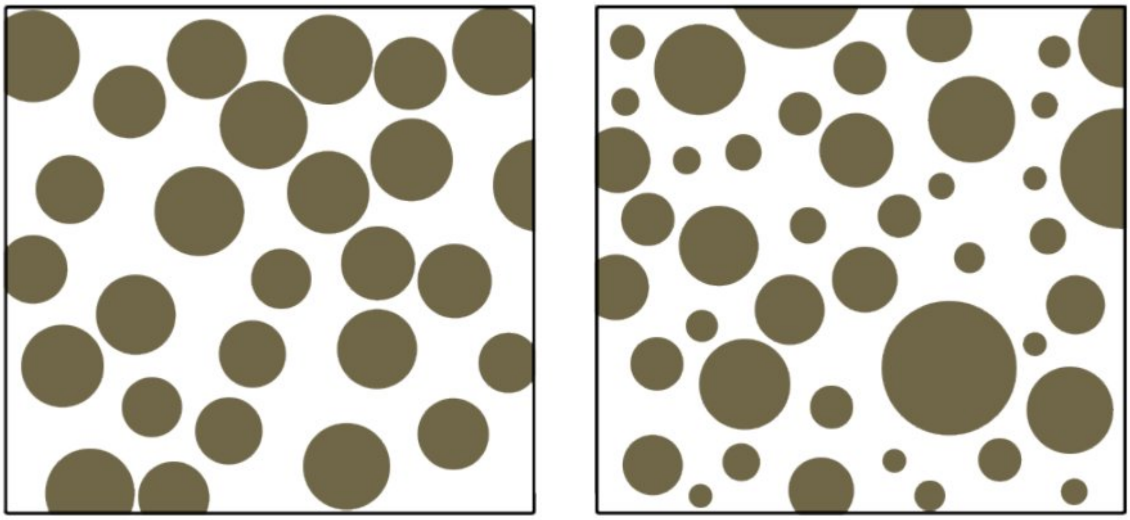

Figure 12.1.3 *Well-sorted sediments (left) have particles that are all of a similar size. Poorly sorted sediments (right) consist of particles of a wide range of sizes (Woudloper (Own work) [CC BY-SA 3.0], via Wikimedia Commons).*

Finally, sediment texture can be described based on its **maturity**, or how long the particles have been transported by water or other vectors. Maturity can be reflected by the degree of rounding of the particles, the amount of sorting, and the composition of the sediment. In the case of **rounding**, the more mature the sediment, the rounder the particles, as a result of the particles being abraded over time. A high degree of sorting indicates maturity, because over time the smaller particles will be washed away, and a given amount of energy will move particles of a similar size over the same distance. Lastly, the older and more mature a sediment the higher the quartz content, at least in sediments derived from rock particles. Quartz is a common mineral in terrestrial rocks, and it is very hard and resistant to abrasion. Over time, particles made from other materials are worn away, leaving only quartz behind. Beach sand is a very mature sediment; it is composed primarily of quartz, and the particles are rounded and of similar size (well-sorted).

Sources of sediments

Sediments are also classified based on their source of origin. There are four main categories for the origin of marine sediments:

- **Lithogenous** sediments are derived from preexisting rock. They are also called **terrigenous** sediments since most of it comes from the land masses and makes its way into the ocean.
- **Biogenous** sediments are composed of the remains of marine organisms.
- **Hydrogenous** sediments are formed when materials that are dissolved in water precipitate out and form solid particles.

- **Cosmogenous** sediments are derived from extraterrestrial sources.

The next few sections will address each of these sediment types in more detail.

12.2 Lithogenous Sediments

Lithogenous or terrigenous sediment is primarily composed of small fragments of preexisting rocks that have made their way into the ocean. These sediments can contain the entire range of particle sizes, from microscopic clays to large boulders, and they are found almost everywhere on the ocean floor. Lithogenous sediments are created on land through the process of weathering, where rocks and minerals are broken down into smaller particles through the action of wind, rain, water flow, temperature- or ice-induced cracking, and other erosive processes. These small eroded particles are then transported to the oceans through a variety of mechanisms:

- **Streams and rivers**: Various forms of runoff deposit large amounts of sediment into the oceans, mostly in the form of finer-grained particles (Figure 12.2.1). About 90% of the lithogenous sediment in the oceans is though to have come from river discharge, particularly from Asia. Most of this sediment, especially the larger particles, will be deposited and remain fairly close to the coastline, however, smaller clay particles may remain suspended in the water column for long periods of time and may be transported great distances from the source.

Figure 12.2.1 *River discharge in the Yukon Delta, Alaska. The pale color demonstrates the large amounts of sediment released into the ocean via the rivers (By Jesse Allen and Robert Simmon (NASA Earth Observatory) [Public domain], via Wikimedia Commons).*

- **Wind**: Wind-borne (**aeolian**) transport can take small particles of sand and dust and move them thousands of kilometers from the source. These small particles can fall into the ocean when the wind dies down, or can serve as the nuclei around which raindrops or snowflakes form. Aeolian transport is particularly important near desert areas (Figure 12.2.2).

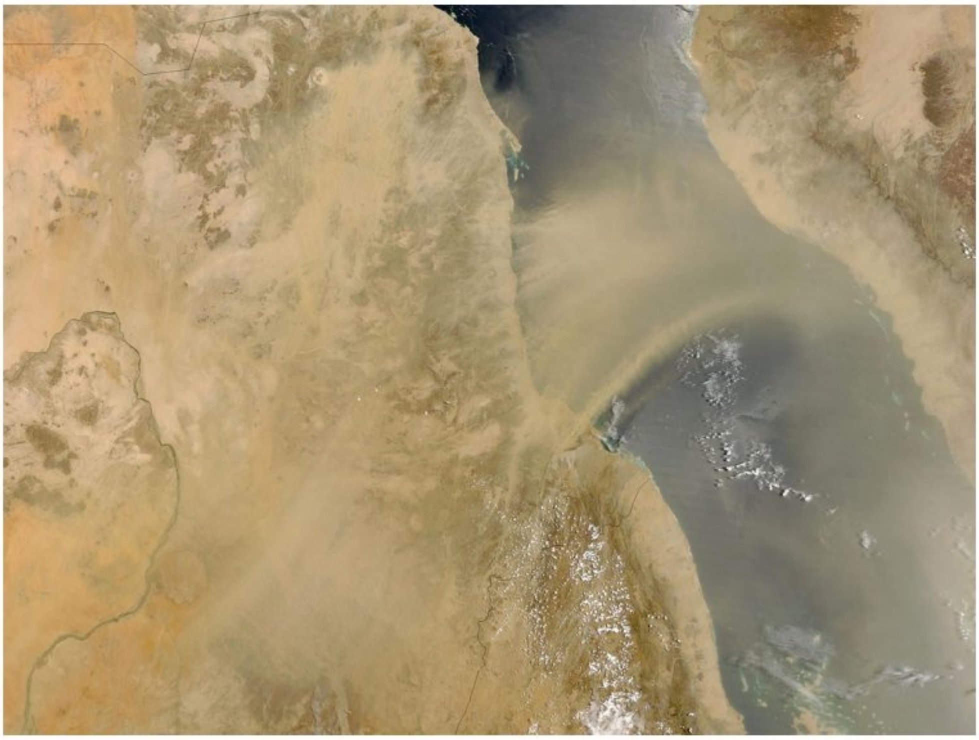

***Figure 12.2.2** A plume of wind-borne particles from Sudan (left) blow over the Red Sea (By NASA (http://visibleearth.nasa.gov/view_rec.php?id=5645) [Public domain], via Wikimedia Commons).*

- **Glaciers and ice rafting**: As glaciers grind their way over land, they pick up lots of soil and rock particles, including very large boulders, that get carried by the ice. When the glacier meets the ocean and begins to break apart or melt, these particles get deposited (Figure 12.2.3). Most of the deposition will happen close to where the glacier meets the water, but a small amount of material is also transported longer distances by rafting, where larger pieces of ice drift far from the glacier before releasing their sediment.

Figure 12.2.3 *When glaciers reach the sea they can break apart, depositing their sediments into the ocean, including very large pieces of rock (By Ianaré Sévi (Own work) [CC BY-SA 3.0], via Wikimedia Commons).*

- **Gravity**: Landslides, mudslides, avalanches, and other gravity-driven events can deposit large amounts of material into the ocean when they happen close to shore.
- **Waves**: Wave action along a coastline will erode rocks and will pull loose particles from beaches and shorelines into the water.
- **Volcanoes**: Volcanic eruptions emit vast amounts of ash and other debris into the atmosphere, where it can then be transported by wind to eventually get deposited in the oceans (Figure 12.2.4).

Figure 12.2.4 *Eruption of the Mayon Volcano, Philippines, in 1984. Much of the material spewed from a volcanic eruption may eventually make its way into the oceans (By C.G. Newhall [Public domain], via Wikimedia Commons).*

- **Gastroliths**: An interesting, although relatively minor avenue for the transport of lithogenous sediments to the ocean is in the form of gastroliths. Gastrolith means “stomach stones” and comes from the fact that many animals, including seabirds, pinnipeds, and some crocodiles will deliberately swallow stones in one area and regurgitate them in another. Often these stones swallowed on land will be regurgitated at sea. Why swallow stones? Possible explanations include using the stones to help grind up food in the stomach, to act as ballast to aid in buoyancy regulation, or to fill the stomach and reduce feelings of hunger during fasting periods on shore.

Most of these processes deposit lithogenous sediment fairly close to shore. The sediment particles can then be transported farther away by waves and currents, where they may eventually escape the continental shelf and reach the deep ocean floor.

Composition

Lithogenous sediments usually reflect the composition of whatever materials they were derived from, so they are dominated by the major minerals that make up most terrestrial rock. This includes quartz, feldspar, clay minerals, iron oxides, and terrestrial organic matter. Quartz (silicon dioxide, the main component of glass) is one of the most common minerals found in nearly all rocks, and it is very resistant to abrasion (see section 12.1), so it is a dominant component of lithogenous sediments, including sand.

12.3 Biogenous Sediments

Biogenous sediments come from the remains of living organisms that settle out as sediment when the organisms die. It is the "hard parts" of the organisms that contribute to the sediments; things like shells, teeth or skeletal elements, as these parts are usually mineralized and are more resistant to decomposition than the fleshy "soft parts" that rapidly deteriorate after death.

Macroscopic sediments contain large remains, such as skeletons, teeth, or shells of larger organisms. This type of sediment is fairly rare over most of the ocean, as large organisms don't die in enough of a concentrated abundance to allow these remains to accumulate. One exception is around coral reefs; here there is a great abundance of organisms that leave behind their remains, in particular the fragments of the stony skeletons of corals that make up a large percentage of tropical sand.

Microscopic sediment consists of the hard parts of microscopic organisms, particularly their shells, or **tests**. Although very small, these organisms are highly abundant and as they die by the billions every day their tests sink to the bottom to create biogenous sediments. Sediments composed of microscopic tests are far more abundant than sediments from macroscopic particles, and because of their small size they create fine-grained, mushy sediment layers. If the sediment layer consists of at least 30% microscopic biogenous material, it is classified as a biogenous **ooze**. The remainder of the sediment is often made up of clay.

The primary sources of microscopic biogenous sediments are unicellular algaes and protozoans (single-celled amoeba-like creatures) that secrete tests of either calcium carbonate ($CaCO_3$) or silica (SiO_2). Silica tests come from two main groups, the **diatoms** (algae) and the **radiolarians** (protozoans) (Figure 12.3.1).

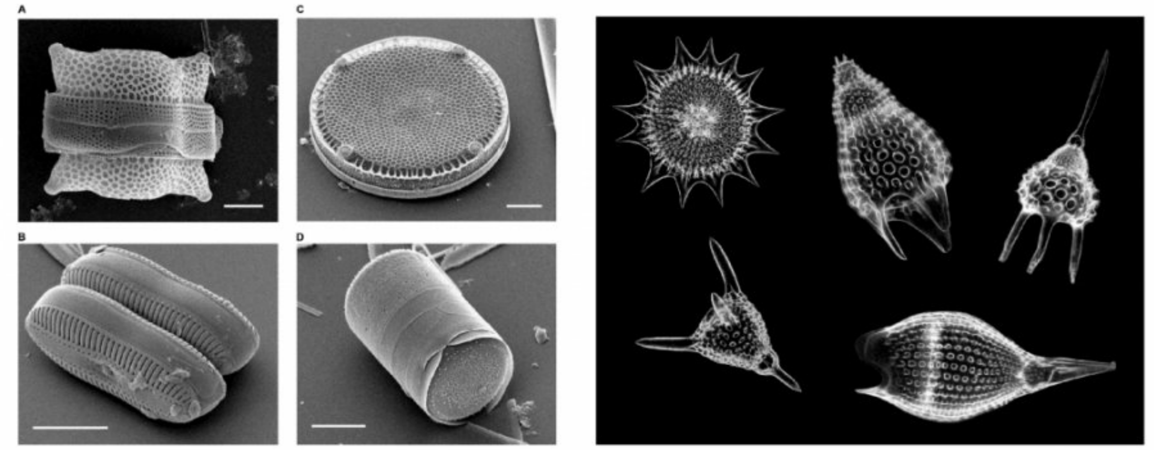

Figure 12.3.1 *Various diatom (left) and radiolarian (right) tests (Diatom images courtesy of Mary Ann Tiffany, San Diego State University [CC BY 2.5], via Wikimedia Commons; radiolarian images by Andreas Drews, https://pxhere.com/en/photo/239774, [CC by 2.0]).*

Diatoms are important members of the phytoplankton, the small, drifting algal photosynthesizers. A diatom consists of a single algal cell surrounded by an elaborate silica shell that it secretes for itself. Diatoms come in a range of shapes, from elongated, pennate forms, to round, or centric shapes that often have two halves, like a Petri dish (Figure 12.3.1 *left*). In areas where diatoms are abundant, the underlying sediment is rich in silica diatom tests, and is called **diatomaceous earth** (see box below).

What use are diatoms?

Diatoms are a vital piece of the global ecosystem for their role in oceanic primary production and the creation of much of the oxygen that organisms breathe. But diatoms are also important for many industrial and agricultural applications. Because of the very fine grain size, and the lattice-like structure of the diatom tests, diatomaceous earth has been used as a filtering agent in things like swimming pool filters and beer brewing. The microscopic tests have been added as an abrasive to toothpaste, facial cleansers and household cleaning agents. Alfred Nobel used diatomaceous earth to stabilize nitroglycerine in the production of dynamite. Diatomaceous earth also displays insecticide properties by stimulating dehydration in insects. It is marketed for this purpose in agriculture, as well as for household use to combat ants, cockroaches, and bedbugs. “Food grade” diatomaceous earth has also entered the market, with proponents touting a range of health benefits arising from its consumption. That’s a pretty impressive range of uses from a microscopic algae!

Radiolarians are planktonic protozoans (making them part of the zooplankton), that like diatoms, secrete a silica test. The test surrounds the cell and can include an array of small openings through which the radiolarian can extend an amoeba-like “arm” or pseudopod (Figure 12.3.1 *right*). Radiolarian tests often display a number of rays protruding from their shells which aid in buoyancy. Oozes that are dominated by diatom or radiolarian tests are called **siliceous oozes**.

Like the siliceous sediments, the calcium carbonate, or calcareous sediments are also produced from the tests of microscopic algae and protozoans; in this case the **coccolithophores** and **foraminiferans**. Coccolithophores are single-celled planktonic algae about 100 times smaller than diatoms. Their tests are composed of a number of interlocking $CaCO_3$ plates (coccoliths) that form a sphere surrounding the cell (Figure 12.3.2 *left*). When coccolithophores die the individual plates sink out and form an ooze. Over time, the coccolithophore ooze lithifies to becomes chalk. The famous White Cliffs of Dover in England are composed of coccolithophore-rich ooze that turned into chalk deposits (Figure 12.3.2 *right*).

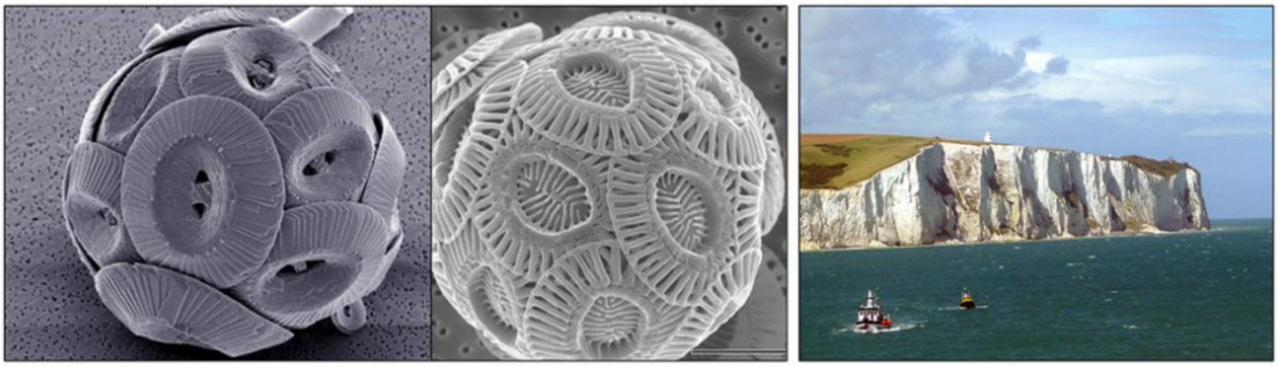

***Figure 12.3.2** (**Left**) coccolithophore tests (left; By Richard Lampitt, Jeremy Young, The Natural History Museum, London (http://planktonnet.awi.de/); center; by Alison R. Taylor (University of North Carolina Wilmington Microscopy Facility) (PLoS Biology, June 2011, Cover ([1])) [Both images CC BY 2.5], via Wikimedia Commons). (**Right**); the White Cliffs of Dover (Immanuel Giel (Own work) [CC BY-SA 3.0], via Wikimedia Commons).*

Foraminiferans (also referred to as "forams") are protozoans whose tests are often chambered, similar to the shells of snails. As the organism grows, is secretes new, larger chambers in which to reside. Most foraminiferans are benthic, living on or in the sediment, but there are some planktonic species living higher in the water column. When coccolithophores and foraminiferans die, they form **calcareous oozes**.

Figure 12.3.3 *Foraminifera tests collected from a beach in Myanmar (By Psammophile [GFDL (http://www.gnu.org/copyleft/fdl.html) or CC BY-SA 3.0], via Wikimedia Commons).*

Older calcareous sediment layers contain the remains of another type of organism, the **discoasters**; single-celled algae related to the coccolithophores that also produced calcium carbonate tests. Discoaster tests were star-shaped, and reached sizes of 5-40 μm across (Figure 13.3.4). Discoasters went extinct approximately 2 million years ago, but their tests remain in deep, tropical sediments that predate their extinction.

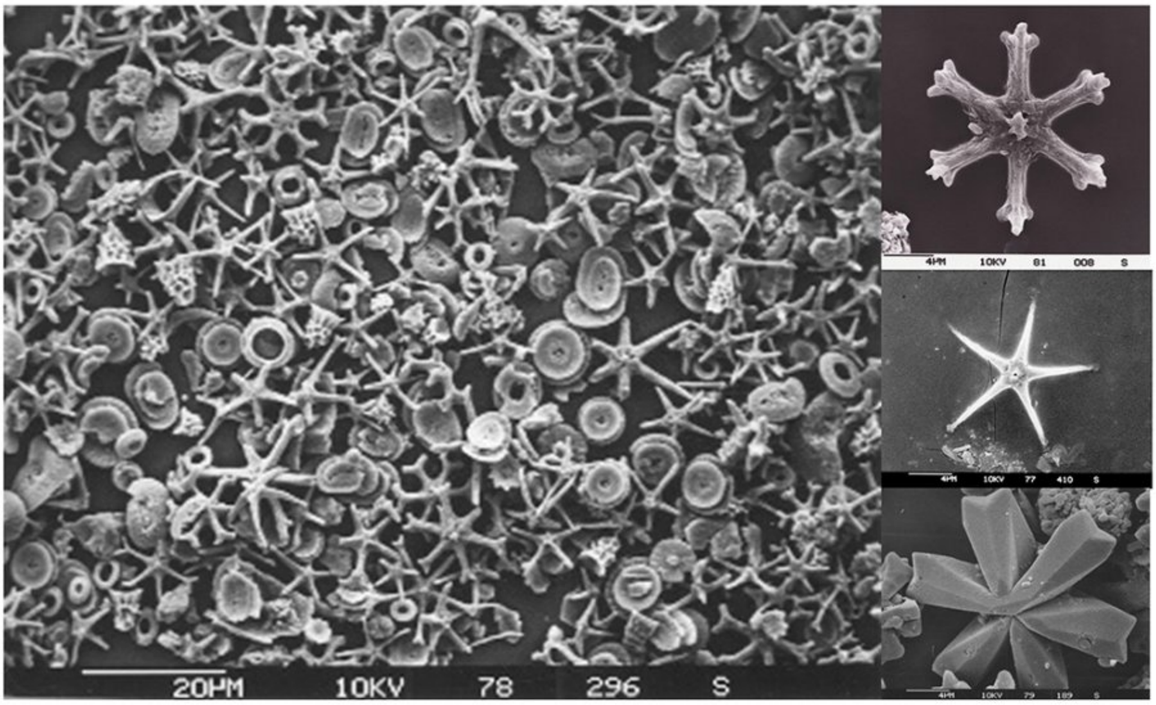

Figure 12.3.4 *Discoaster tests. Left: discoster tests with assorted coccoliths. Top right Discoaster surculus; center right: Discoaster pentaradiatus; bottom right: Discoaster surculus (All images by Hannes Grobe (Own work) [CC BY 3.0], via Wikimedia Commons).*

Because of their small size, these tests sink very slowly; a single microscopic test may take about 10-50 years to sink to the bottom! Given that slow descent, a current of only 1 cm/sec could carry the test as much as 15,000 km away from its point of origin before it reaches the bottom. Yet despite this, we find that the sediments in a particular location are well-matched to the types of organisms and degree of productivity that occurs in the water overhead. This means that the sediment particles must be sinking to the bottom at a much faster rate, so that they accumulate below their point of origin before the currents can disperse them. What is the mechanism for this increased sinking rate? Apparently most of the tests do not sink as individual particles; about 99% of them are first consumed by some other organism, and are then aggregated and expelled as large fecal pellets, which sink much more quickly and reach the ocean floor in only 10-15 days. This does not give the particles as much time to disperse, and the sediment below will reflect the production occurring near the surface. The increased rate of sinking through this mechanism is called the “fecal express.”

Reconstructing past climate through sediment analysis

As outlined in the opening to this chapter, examining marine sediments allows us to learn much about oceanographic and atmospheric processes, both past and present. Biogenous sediments are no exception, and they can allow us to reconstruct past climate history from oxygen isotope ratios.

Oxygen atoms exist in three forms, or isotopes, in ocean water: O^{16}, O^{17} and O^{18} (the number refers to the atomic masses of the isotopes). O^{16} is the most common form, followed by O^{18} (O^{17} is rare). O^{16} is lighter than O^{18}, so it evaporates more easily, leading to water vapor that has a higher proportion of O^{16}. During periods of cooler climate, water vapor condenses into rain and snow, which forms glacial ice that has a high proportion of O^{16}. The remaining seawater therefore has a relatively higher proportion of O^{18}. Marine organisms who incorporate dissolved oxygen into their shells as calcium carbonate will therefore have shells with a higher proportion of O^{18} isotope. In other words, the ratio of O^{16}:O^{18} in shells will be low during periods of colder climate.

When the climate warms, glacial ice melts, releasing O^{16} from the ice and returning it to the oceans, increasing the O^{16}:O^{18} ratio in the water. Now, when organisms incorporate oxygen into their shells, the shells will contain a higher O^{16}:O^{18} ratio. Scientists can therefore examine biogenous sediments, calculate the O^{16}:O^{18} ratios for samples of known ages, and from those ratios, infer the climate conditions under which those shells were formed. The same types of measurements can also be taken from ice cores; a decrease of 1 ppm O^{18} between ice samples represents a decrease in temperature of 1.5° C.

12.4 Hydrogenous Sediments

*Methane hydrate section modified from "Physical Geology" by Steven Earle**

As we saw in section 5.3 seawater contains many different dissolved substances. Occasionally chemical reactions occur that cause these substances to precipitate out as solid particles, which then accumulate as **hydrogenous sediment**. These reactions are usually triggered by a change in conditions, such as a change in temperature, pressure, or pH, which reduces the amount of a substance that can remain in a dissolved state. There is not a lot of hydrogenous sediment in the ocean compared to lithogenous or biogenous sediments, but there are some interesting forms.

Hydrothermal vents were discussed in section 4.11. Recall that in these systems, seawater percolates into the seafloor, where it becomes superheated by magma before being expelled by the vent. This superheated water contains many dissolved substances, and when it encounters the cold seawater after leaving the vent, these particles precipitate out, mostly as metal sulfides. These particles make up the “smoke” that flows from a vent, and may eventually settle on the bottom as hydrogenous sediment (Figure 12.4.1).

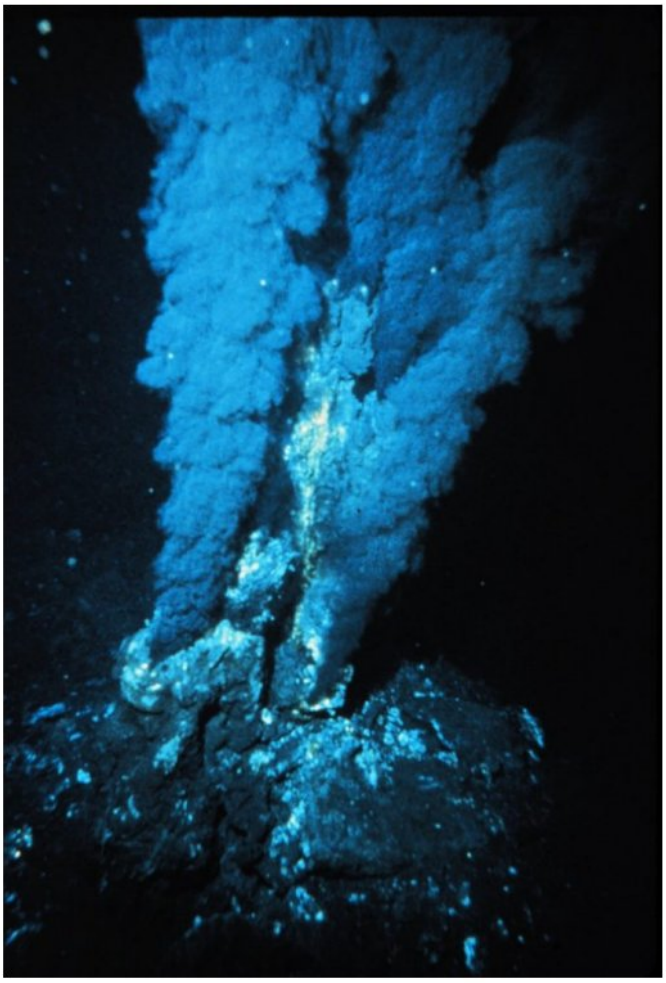

Figure 12.4.1 *A "black smoker" hydrothermal vent. The "smoke" consists of dissolved particles that precipitate into solids when exposed to colder water (NOAA, http://www.photolib.noaa.gov/htmls/nur04506.htm).*

Manganese nodules are rounded lumps of manganese and other metals that form on the seafloor, generally ranging between 3-10 cm in diameter, although they may sometimes reach up to 30 cm (Figure 12.4.2). The nodules form in a manner similar to pearls; there is a central object around which concentric layers are slowly deposited, causing the nodule to grow over time. The composition of the nodules can vary somewhat depending on their location and the conditions of their formation, but they are usually dominated by manganese- and iron oxides. They may also contain smaller amounts of other metals such as copper, nickel and cobalt. The precipitation of manganese nodules is one of the slowest geological processes known; they grow on the order of a few millimeters per million years. For that reason, they only form in areas where there are low rates of lithogenous or biogenous sediment accumulation, because any other sediment deposition would quickly cover the nodules and prevent further nodule growth. Therefore, manganese nodules are usually limited to areas in the central ocean, far from significant lithogenous or biogenous inputs, where they can sometimes accumulate in large numbers on the seafloor (Figure 12.4.2 *right*). Because the nodules contain a number of commercially valuable metals, there has been significant interest in mining the nodules over the last several decades, although most of the efforts have thus far remained at the exploratory stage. A number of factors have prevented large-scale extraction of nodules, including the high costs of deep sea mining operations, political issues over mining rights, and environmental concerns surrounding the extraction of these non-renewable resources.

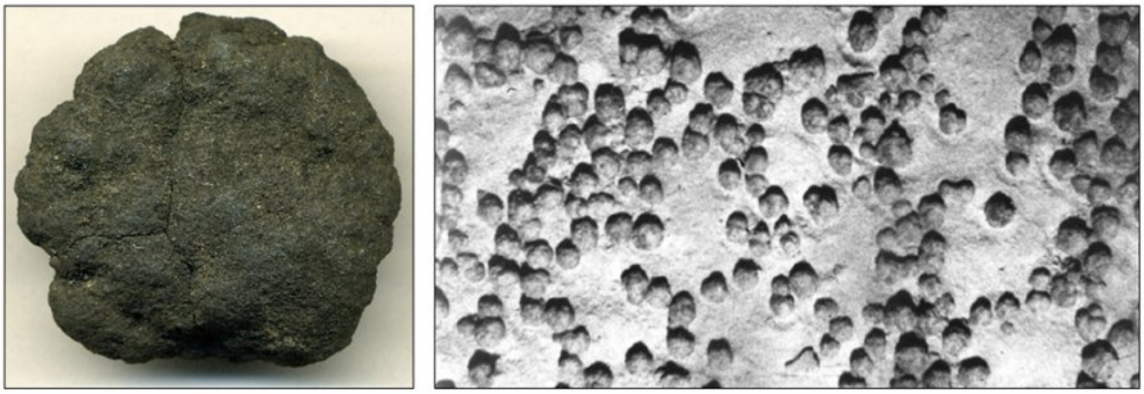

Figure 12.4.2 *(Left) manganese nodule from the subtropical eastern Pacific Ocean. The nodule is 4.1 cm in diameter (By James St. John, https://www.flickr.com/photos/jsjgeology/15139986302/in/photostream/ [CC-BY 2.0]). (Right) field of manganese nodules on the seafloor (By United States Geological Survey [Public domain], via Wikimedia Commons).*

Evaporites are hydrogenous sediments that form when seawater evaporates, leaving the dissolved materials to precipitate into solids, particularly halite (salt, NaCl). In fact, the evaporation of seawater is the oldest form of salt production for human use, and is still carried out today. Large deposits of halite evaporites exist in a number of places, including under the Mediterranean Sea. Beginning around 6 million years ago, tectonic processes closed off the Mediterranean Sea from the Atlantic, and the warm climate evaporated so much water that the Mediterranean was almost completely dried out, leaving large deposits of salt in its place (an event known as the Messinian Salinity Crisis). Eventually the Mediterranean re-flooded about 5.3 million years ago, and the halite deposits were covered by other sediments, but they still remain beneath the seafloor.

__Figure 12.4.3__ Salt farmers harvesting salt left behind from the evaporation of seawater, Pak Thale, Ban Laem, Phetchaburi, Thailand (By JJ Harrison (Own work) [CC BY-SA 3.0], via Wikimedia Commons).

Oolites are small, rounded grains formed from concentric layers of precipitation of material around a suspended particle. They are usually composed of calcium carbonate, but they may also from from phosphates and other materials. Accumulation of oolites results in oolitic sand, which is found in its greatest abundance in the Bahamas (Figure 12.4.4).

__Figure 12.4.4__ Oolites from a beach on Joulter's Cay, The Bahamas (By Wilson44691 (Own work) [Public domain], via Wikimedia Commons).

Methane hydrates are another type of hydrogenous deposit with a potential industrial application. All terrestrial erosion products include a small proportion of organic matter derived mostly from terrestrial plants. Tiny fragments of this material plus other organic matter from marine plants and animals accumulate in terrigenous sediments, especially within a few hundred kilometers of shore. As the sediments pile up, the deeper parts start to warm up (from geothermal heat), and bacteria get to work breaking down the contained organic matter. Because this is happening in the absence of oxygen (a.k.a. anaerobic conditions), the by-product of this metabolism is the gas methane (CH_4). Methane released by the bacteria slowly bubbles upward through the sediment toward the seafloor. At water depths of 500 m to 1,000 m, and at the low temperatures typical of the seafloor (close to 4°C), water and methane combine to create a substance known as methane hydrate. Within a few meters to hundreds of meters of the seafloor, the temperature is low enough for methane hydrate to be stable and hydrates accumulate within the sediment (Figure 12.4.5 *left*). Methane hydrate is flammable because when it is heated, the methane is released as a gas (Figure 12.4.5 *right*). The methane within seafloor sediments represents an enormous reservoir of fossil fuel energy. Although energy corporations and governments are anxious to develop ways to produce and sell this methane, anyone that understands the climate-change implications of its extraction and use can see that this would be folly.

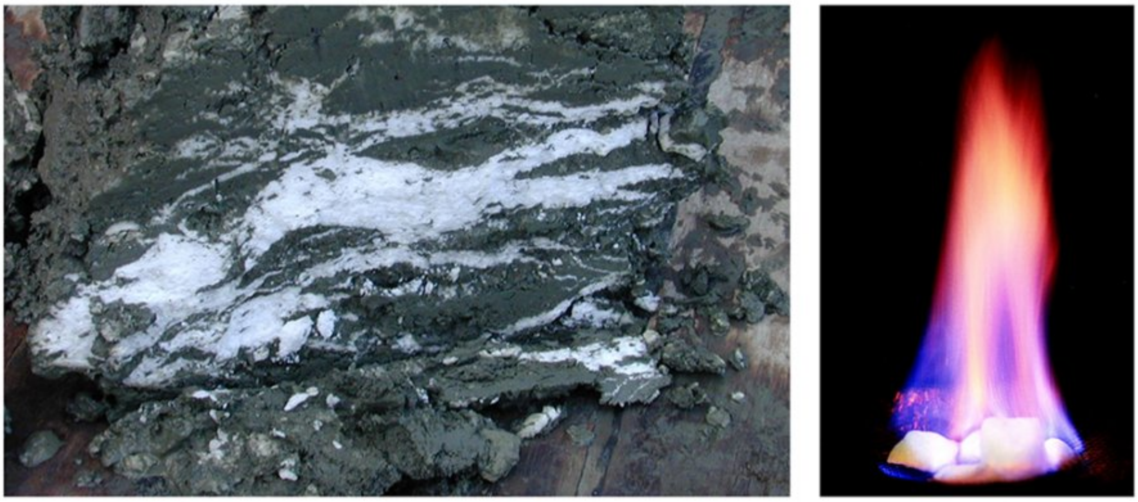

Figure 12.4.5 *(Left): Methane hydrate within muddy sea-floor sediment from an area offshore from Oregon (By Wusel007 (Own work) [GFDL (http://www.gnu.org/copyleft/fdl.html) or CC BY-SA 3.0], via Wikimedia Commons). (Right): Methane hydrate on fire (USGS, http://www.usgs.gov/blogs/features/files/2012/01/New-Image.jpg).*

12.5 Cosmogenous Sediments

Cosmogenous sediment is derived from extraterrestrial sources, and comes in two primary forms; microscopic spherules and larger meteor debris. Spherules are composed mostly of silica or iron and nickel, and are thought to be ejected as meteors burn up after entering the atmosphere. Meteor debris comes from collisions of meteorites with Earth. These high impact collisions eject particles into the atmosphere that eventually settle back down to Earth and contribute to the sediments. Like spherules, meteor debris is mostly silica or iron and nickel. One interesting form of debris from these collisions are tektites, which are small droplets of glass. They are likely composed of terrestrial silica that was ejected and melted during a meteorite impact, which then solidified as it cooled upon returning to the surface.

Figure 12.5.1 *Tektite-like glass found near the Volkov River in western Russia (By James St. John [CC BY 2.0], via Wikimedia Commons).*

Cosmogenous sediment is fairly rare in the ocean and it does not usually accumulate in large deposits. However, it is constantly being added to through space dust that continuously rains down on Earth. About 90% of incoming cosmogenous debris is vaporized as it enters the atmosphere, but it is estimated that 5 to 300 tons of space dust land on the Earth's surface each day!

12.6 Sediment Distribution

Now that we have an understanding of the types of sediments found in the ocean, we can turn our attention to the processes that cause different types of sediments to dominate in different locations. Sediment accumulation will depend on the the amount of material coming from the source, the distance from the source, the amount of time that sediment has had to accumulate, how well the sediments are preserved, and the amounts of other types of sediments that are also being added to the system.

Rates of sediment accumulation are relatively slow throughout most of the ocean, in many cases taking thousands of years for any significant deposits to form. Lithogenous sediment accumulates the fastest, on the order of 1 m or more per thousand years for coarser particles. However, sedimentation rates near the mouths of large rivers with high discharge can be orders of magnitude higher. Biogenous oozes accumulate at a rate of about 1 cm per thousand years, while small clay particles are deposited in the deep ocean at around 1 mm per thousand years. As described in section 12.4, manganese nodules have an incredibly slow rate of accumulation, gaining 0.001 mm per thousand years.

Marine sediments are thickest near the continental margins (refer to figure 12.1.1) where they can be over 10 km thick. This is because the crust near passive continental margins is often very old, allowing for a long period of accumulation, and because there is a large amount of terrigenous sediment input coming from the continents. Near mid-ocean ridge systems where new oceanic crust is being formed, sediments are thinner, as they have had less time to accumulate on the younger crust. As you move away from the ridge spreading center the sediments get progressively thicker (see section 4.5), increasing by approximately 100-200 m of sediment for every 1000 km distance from the ridge axis. With a seafloor spreading rate of about 20-40 km/million years, this represents a sediment accumulation rate of approximately 100-200 m every 25-50 million years.

Figure 12.6.1 shows the distribution of the major types of sediment on the ocean floor. Cosmogenous sediments could potentially end up in any part of the ocean, but they accumulate in such small abundances that they are overwhelmed by other sediment types and thus are not dominant in any location. Similarly, hydrogenous sediments can have high concentrations in specific locations, but these regions are very small on a global scale. So we will mostly ignore cosmogenous and hydrogenous sediments in the discussion of global sediment patterns.

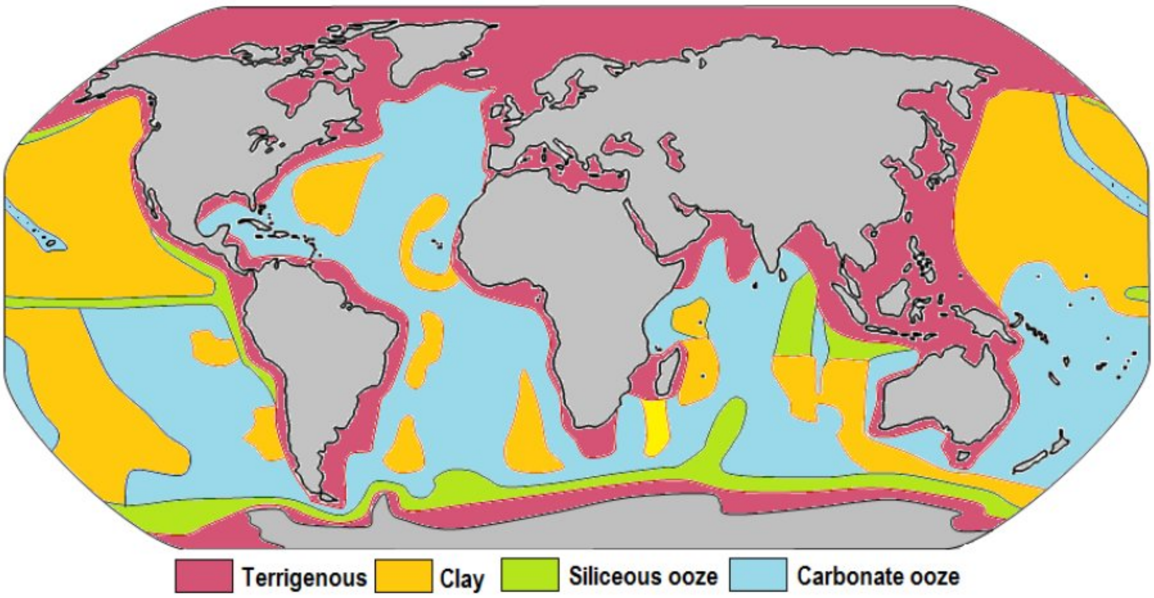

Figure 12.6.1 *The distribution of sediment types on the seafloor. Within each colored area, the type of material shown is what dominates, although other materials are also likely to be present (Steven Earle, "Physical Geology").*

Coarse lithogenous/terrigenous sediments are dominant near the continental margins as runoff, river discharge, and other processes deposit vast amounts of these materials on the continental shelf (section 12.2). Much of this sediment remains on or near the shelf, while turbidity currents can transport material down the continental slope to the deep ocean floor. Lithogenous sediment is also common at the poles where thick ice cover can limit primary production, and glacial breakup deposits sediments along the ice edge. Coarse lithogenous sediments are less common in the central ocean, as these areas are too far from the sources for these sediments to accumulate. Very small clay particles are the exception, and as described below, they can accumulate in areas that other lithogenous sediment will not reach.

The distribution of biogenous sediments depends on their rates of production, dissolution, and dilution by other sediments. We learned in section 7.4 that coastal areas display very high primary production, so we might expect to see abundant biogenous deposits in these regions. However, recall that sediment must be >30% biogenous to be considered a biogenous ooze, and even in productive coastal areas there is so much lithogenous input that it swamps the biogenous materials, and that 30% threshold is not reached. So coastal areas remain dominated by lithogenous sediment, and biogenous sediments will be more abundant in pelagic environments where there is little lithogenous input.

In order for biogenous sediments to accumulate their rate of production must be greater than the rate at which the tests dissolve. Silica is undersaturated throughout the ocean and will dissolve in seawater, but it dissolves more readily in warmer water and lower pressures; in other words, it dissolves faster near the surface than in deep water. Silica sediments will therefore only accumulate in cooler regions of high productivity where they accumulate faster than they dissolve. This includes upwelling regions near the equator and at high latitudes where there are abundant nutrients and cooler water. Oozes formed near the equatorial regions are usually dominated by radiolarians, while diatoms are more common in the polar oozes. Once the silica tests have settled on the bottom and are covered by subsequent layers, they are no longer subject to dissolution and the sediment will accumulate. Approximately 15% of the seafloor is covered by siliceous oozes.

Biogenous calcium carbonate sediments also require production to exceed dissolution for sediments to accumulate, but the processes involved are a little different than for silica. Calcium carbonate dissolves more readily in more acidic water. Cold seawater contains more dissolved CO_2 and is slightly more acidic than warmer water (section 5.5). Therefore calcium carbonate tests are more likely to dissolve in colder, deeper, polar water than in warmer, tropical, surface water. At the poles the water is uniformly cold, so calcium carbonate readily dissolves at all depths, and carbonate sediments do not accumulate. In temperate and tropical regions calcium carbonate dissolves more readily as it sinks into deeper water. The depth at which calcium carbonate dissolves as fast as it accumulates is called the **calcium carbonate compensation depth**, or **calcite compensation depth**, or simply the **CCD**. The **lysocline** represents the depths where the rate of calcium carbonate dissolution increases dramatically (similar to the thermocline and halocline). At depths shallower than the CCD carbonate accumulation will exceed the rate of dissolution, and carbonate sediments will be deposited. In areas deeper than the CCD, the rate of dissolution will exceed production, and no carbonate sediments can accumulate (Figure 12.6.2). The CCD is usually found at depths of 4 – 4.5 km, although it is much shallower at the poles where the surface water is cold. Thus calcareous oozes will mostly be found in tropical or temperate waters less than about 4 km deep, such as along the mid-ocean ridge systems and atop seamounts and plateaus. The CCD is deeper in the Atlantic than in the Pacific since the Pacific contains more CO_2, making the water more acidic and calcium carbonate more soluble. This, along with the fact that the Pacific is deeper, means that the Atlantic contains more calcareous sediment than the Pacific. All told, about 48% of the seafloor is dominated by calcareous oozes.

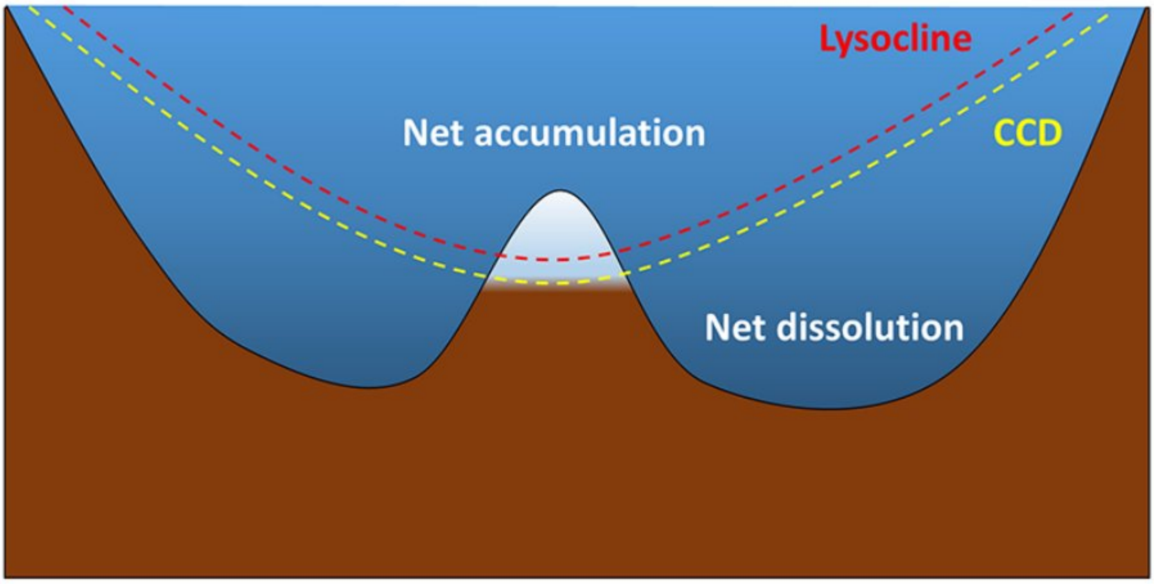

Figure 12.6.2 *Calcareous sediment can only accumulate in depths shallower than the calcium carbonate compensation depth (CCD). Below the CCD, calcareous sediments dissolve and will not accumulate. The lysocline represents the depths where the rate of dissolution increases dramatically (PW).*

Much of the rest of the deep ocean floor (about 38%) is dominated by abyssal clays. This is not so much a result of an abundance of clay formation, but rather the lack of any other types of sediment input. The clay particles are mostly of terrestrial origin, but because they are so small they are easily dispersed by wind and currents, and can reach areas inaccessible to other sediment types. Clays dominate in the central North Pacific, for example. This area is too far from land for coarse lithogenous sediment to reach, it is not productive enough for biogenous tests to accumulate, and it is too deep for calcareous materials to reach the bottom before dissolving. Because clay particles accumulate so slowly, the clay-dominated deep ocean floor is often home to hydrogenous sediments like manganese nodules. If any other type of sediment was produced here it would accumulate much more quickly and would bury the nodules before they had a chance to grow.