CHAPTER 8
COASTS

1. INTRODUCTION

1.1 The term *coastal studies*, which is in common use for a great variety of approaches to coastlines, covers a large area of endeavor. What do I mean by the term *coastal*? There are various definitions or interpretations, depending on how much or how little is included seaward and landward of the shoreline. The ***shoreline*** is fairly generally taken to be *the line or trace where the sea meets the land*, and this is fairly well defined except in areas with estuaries, tidal flats, etc., unless you want to quibble about where the high-water line is.

1.2 In this course I'm going to interpret the term ***coast*** in a fairly broad sense to mean *a marine area extending from the shoreline out onto the continental shelf for some distance, together with some land area immediately landward of the shoreline*. This admittedly sloppy definition succeeds in generally including areas both seaward and landward of the shoreline in which a great variety of processes operate that most people would categorize as coastal processes. I intend the definition to include only the innermost part of the ***continental shelf***, the part that's most strongly affected by the adjacent shoreline. (The continental shelf is the broad belt of shallow water adjacent to the coastline. In many areas of the world it is as much as a few hundred kilometers wide, and water depths at the shelf edge are no more than about two hundred meters.) In many coastal areas there's a well defined ***coastal plain*** that may be well over a hundred kilometers wide; only the part nearest the shoreline, directly affected by modern coastal processes, is included in my definition.

1.3 Coastal environments are unusually varied. If you let your mind go blank and I say the word "coast" to you, what image do you first conjure up? Rugged rocky sea cliffs? Wide sandy beaches? Shining coral reefs bathed in transparently blue water? These are only some of the many important coastal environments in the world today. How can I possibly deal with all these environments in one small part of this course, you might ask? Obviously I can't, so I'll concentrate on just a few kinds of coastlines and the most important processes that operate around them.

1.4 There's a lot of coastline in the world today: by one estimate, the coastlines of the world are almost 450,000 km long. One difficulty with an estimate like that is that the closer you look at a given stretch of coastline, the longer it comes out to be. On a large scale, that's because maps always involve a certain degree of generalization, depending upon the scale of the map. But the effect is still there even when you're looking at a small segment of the coastline right at your feet, because how do you take account of the outlines of the individual little sand grains?

1.5 Here's a list of the important factors that govern the nature of a coast, together with a few initial comments about each. Keep in mind that there's a rather strong interdependence among these factors.

- **sediment supply from land.** Clearly this is important in determining whether there is any sediment for the ocean to shape into beaches, deltas, barrier islands, etc. And the *size* of the sediment is important too: if only mud is supplied to the shoreline (as in certain low-latitude areas with hot and humid climate), you shouldn't expect to have sandy beaches!
- **climate.** The climate of the land area inland from the coast has various indirect effects on the nature of the coast: *size and supply of sediment* (see above), *river runoff*, and in some cases *glaciation*.
- **sea-level history.** Sea level hasn't stayed the same relative to the land: it rises and falls at rates ranging from something of the order of a millimeter per century, as a representative value throughout much of geologic time, to as much as a meter per century during rapid continental deglaciation, as happened in very recently in geologic time (within the last 20,000 years!). The nature of the coastline is not much affected by very slow changes in sea level, because coastal processes have plenty of time to equilibrate, but rapid changes in sea level are known to have strong effects on the nature of the coastline.
- **hydrodynamic setting.** By this vague term I mean the picture of *waves, tides, and currents* that affect the coastline. The relative importance of these three kinds of water motions varies greatly, and the nature of the coastline depends strongly on their relative importance.
- **tectonic setting.** In many areas the Earth's crust is unstable, and undergoes both vertical and lateral movements that can be very rapid on geologic time scales and can be substantial even on the time scales of long-term coastal processes. You have to think about the effect of tectonism on the climate and sediment supply landward of the coastline and on the submarine topography offshore, as well as the more direct effects on sea level at the coastline.

2. CLASSIFICATION

2.1 What are coasts like generally, and how can they be classified? I won't try to develop any formal or exhaustive classification here, but there are several important varieties I want you to be aware of and thinking about.

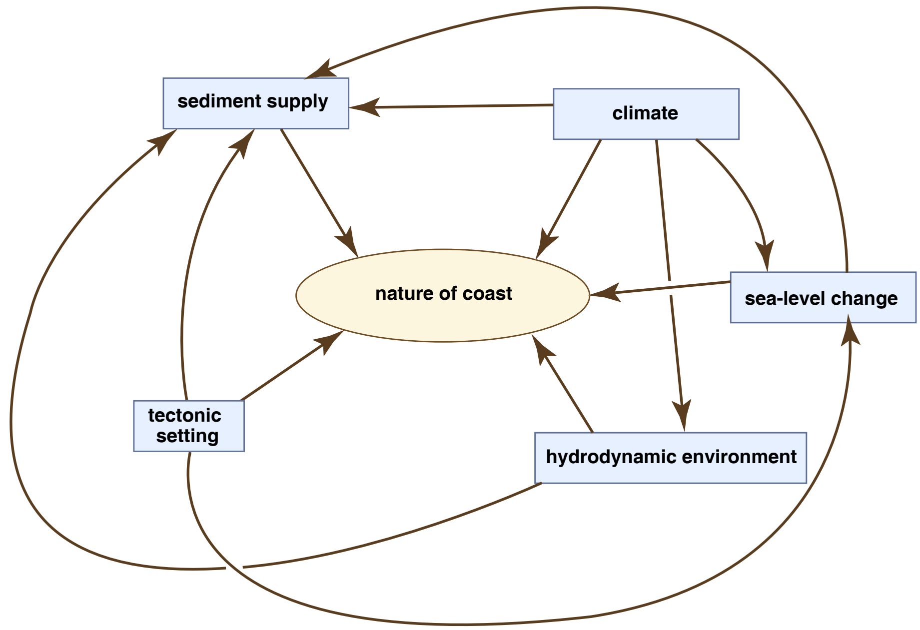

Figure by MIT OCW.

Figure 8-1 An attempt to show, in a simple diagram, the important relationships among the factors that govern the nature of coasts.

2.2 Look at the present shorelines of the world. What do you see?

Beaches. Beaches are perhaps the most common kind of coastline. It's easy to define a ***beach***: *an accumulation of loose sediment at the shoreline, shaped by the action of shoaling and breaking waves*. Sediment size ranges from fine sand to coarse gravel. Beaches range in extent from little pockets only meters to tens of meters long, to many hundreds of kilometers without a break.

Rocky or cliffy coasts. Along such coasts there's little or no loose sediment above mean low tide, so the bedrock of the continent is exposed to slow marine erosion. (The bedrock is usually resistant, or else the forces of the sea would produce enough sediment to form a sedimentary coastal environment like a beach.) There are all gradations from rocky coasts to beachy coasts.

Tidal flats. These are usually found in protected embayments or along coasts where waves are not important. They typically show an interlacing network of tidal distributary channels, with coarse sediment in the channels, passing into broad areas of fine sediment away from the channels.

Estuaries. An ***estuary*** is *a body of water with salinity lower than the open ocean, with restricted access to the ocean, where fresh water mixes with salt water*. There are several varieties, the main kinds these days being drowned-river estuaries and barrier-system estuaries. They seem to be unrepresentatively common today, presumably because of the effects of the most recent Pleistocene glaciation and deglaciation. Estuaries are often closely associated with tidal flats.

Deltas. A ***delta*** is *a body of sediment delivered to the coastline by a river, and built out into a body of water*, a lake or the ocean. Deltas show a great variety of geometries, and they can exist in combination with many of the other kinds of coasts.

Reefs. A ***reef*** is *a marine structure, built by organisms, that withstands the erosive action of waves*. Reefs are common in low-latitude areas with perennially warm water and low supply of siliciclastic sediment from land. They are often found in combination with carbonate-sand beaches.

Muddy open shorelines. These are not as common, and lie well outside the coastal experiences of most of us (including me). They are found mainly in the tropics in areas where substantial fine siliciclastic sediment is supplied from land but no carbonate sediment is produced near the shoreline, by reason of either too-cold water or swamping by the siliciclastics. (Why are they more common in the tropics? Presumably because of the nature of terrestrial weathering.) *Mangrove swamps*, important along some low-latitude shorelines, should be included here as well.

2.3 Coasts can also be classified into three kinds on the basis of *the effect of sea-level change*:

Emergent. Sea level *falling* relative to the land. These exist today, but they are not common. They are found only in areas where tectonic uplift has more than offset the postglacial sea-level rise.

Submergent. Sea level *rising* relative to the land. These are very common now, because of the recent rise in sea level.

Stable: These are not common now, but they must have been throughout much of geologic time.

Keep in mind that this is an atypical time, because of the large changes in sea level worldwide since the disappearance of the last great Pleistocene continental ice sheets. Sea level rose to nearly its present level from minus 120–130 m between 20,000 yr BP (years before the present) and now, and most of that was between 18,000 yr BP and 6,000 yr BP!

2.4 Coasts can be usefully classified on the basis of their *plate-tectonic setting* into leading-edge coasts and tailing-edge coasts:

Trailing-edge coasts. These are tectonically undifferentiated, once sedimentation has become established well enough to mask the complexities of initial rifting; slow subsidence, relatively low sediment supply, wide shelves; the east coast of North America is a good example.

Leading-edge coasts. These show varied uplift and subsidence; topographically differentiated (basins and ridges both offshore and onshore); relatively high sediment supply, narrow shelves; the west coast of North America and the west coast of South America are two different examples.

2.5 Finally, coasts are naturally classified on the basis of *hydrodynamic setting* into four basic groups, obviously with intergradations among them:

Tide-dominated coasts. Large tidal range; day-to-day sediment movement by strong tidal currents overshadows the effects of important but infrequent storms.

Wave-dominated coasts. Large breaking waves from distant storms overshadow the effects of important but infrequent nearby storms.

Storm-dominated coasts. Waves and currents produced by large coastal storms overshadow other effects.

Current-dominated coasts. Throughgoing strong ocean currents impinging from the adjacent deep ocean dominate sediment movement and shaping of the coast. (These are the least common of the four kinds.)

3. TIDES AND TIDAL CURRENTS

3.1 Introduction

3.1.1 Go to any seacoast and build a tide gauge to obtain a record of sea level as a function of time (over a period of weeks or months) by somehow damping out or averaging over or filtering out the effects of waves on time scales of seconds and storms on time scales of days to weeks. What would you observe? In most places, you would find regular and systematic fluctuations in water level with dominant “periods” of about half a day or about one day, together with more subtle longer-term patterns on time scales of months and even years.

3.1.2 And if you then set up a similar tide gauge in a different place, you would find the same general effect (and the same frequency) but with differences in details of the shape of the time record, and also differences in timing of highs and lows.

3.1.3 As you all know, these effects you’re observing are the tides. It doesn’t take much brilliance on anybody’s part to relate the tides to the moon, because of identity of tidal periods and lunar periods. This was known in ancient times, and probably in even more distant prehistoric times. (It might trouble you, though, that the times of high and low tides are not coincident with the times when the moon is overhead; more on that later.)

3.1.4 Here's the plan of action for this section on the tides:

- Forces involved in generating the tides, and the origin of the tidal bulge
- How the Earth's rotation under the tidal bulges accounts for the observed tidal variations
- Forced-wave nature of the tides, which accounts for the phase lags observed
- Complicating effects needed to account for real tides: standing waves and Coriolis effects
- Tidal currents: periodic motions of the water in conjunction with the rise and fall of the tides

3.2 How Forces in the Earth–Moon System Cause the Tidal Bulge

3.2.1 First, here are some comments on the Earth–Moon system, as an astronomical entity governed by the laws of mechanics. Usually one thinks of the Moon revolving about the Earth once in about each month. But it's better to think of the Earth and the Moon as revolving around a single point, the center of mass of the Earth-Moon system. The Earth–Moon system together forms a mass unit, and it's a well known principle in dynamics that the motions of this unit with respect to external forces can be analyzed in terms of an equal mass concentrated at the center of mass of the system. This is not the place for details, but here's a homey example: throw a skinny dumbbell up in the air, and watch the way the center of mass follows a parabolic trajectory, just like a concentrated mass point, even though the dumbbell is twirling wildly. Because the Earth is much more massive than the Moon, the center of mass of the Earth–Moon system is actually *inside the Earth*, about 4600 km from the center (Figure 8-2)!

3.2.3 How do the Earth and the Moon stay in equilibrium? The rate of revolution of the two bodies about each other is such that the overall gravitational force of attraction is just in balance with the overall centrifugal force that tends to make the bodies fly apart from one another.

3.2.4 Given this *overall* balance, however, the *local* balance between centrifugal forces and gravitational forces differs from point to point within the Earth, depending upon the position the point relative to Earth–Moon axis. It's this differing balance that causes the tides, by virtue of creating *a bidirectional tidal bulge* that draws the waters of the ocean out both in the direction of the Moon and in the direction away from the Moon, relative to directions normal to the Earth–Moon axis (Figure 8-3). (The solid Earth undergoes the same effect, as what are called earth tides, but because the solid Earth is a lot more rigid than the waters of the oceans the Earth tide is far smaller, only of the order of ten centimeters, than the ocean tide, which is a few meters, more or less.)

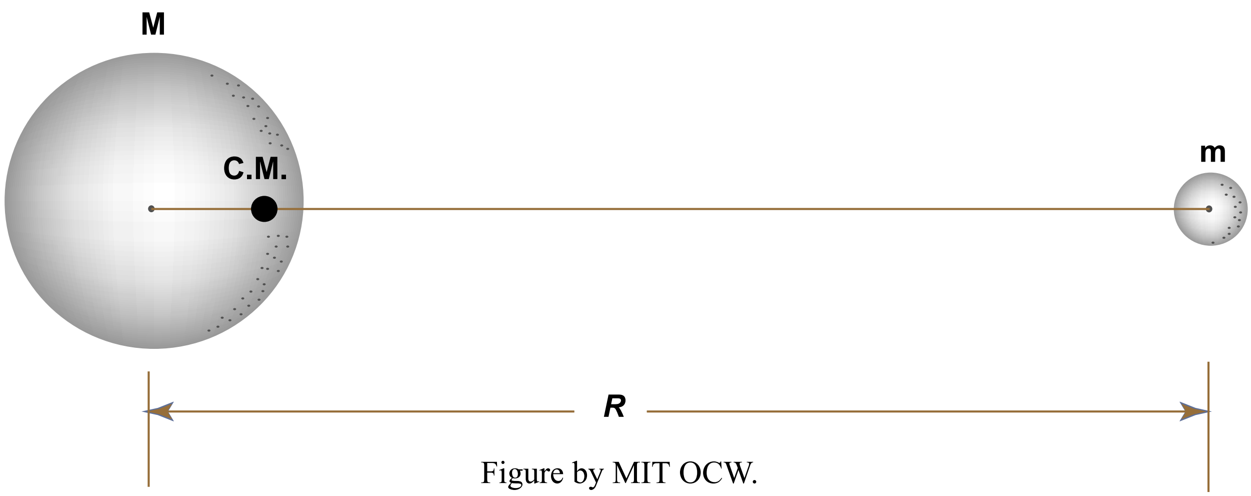

Figure 8-2. The Earth–Moon system.

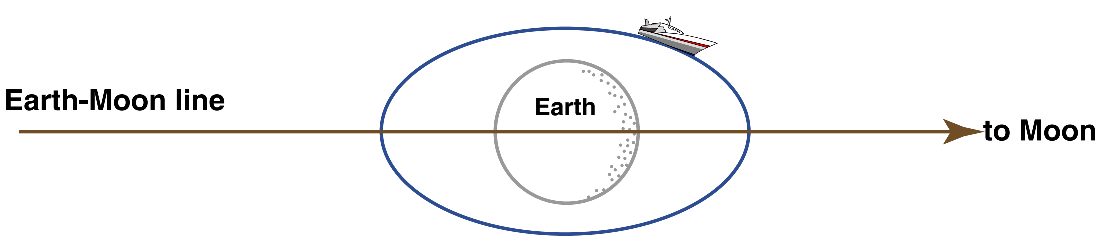

Figure by MIT OCW.

Figure 8-3. The double tidal bulge.

3.2.5 I imagine that the existence of this double-sided tidal bulge seems counterintuitive to you: *why not just in the direction of the Moon?* The following paragraphs attempt to provide a qualitative explanation, which itself will probably not seem immediately obvious to you; if you would like to see a more rigorous explanation, see the "advanced topic" section below.

3.2.6 The basic idea has to do with the relative magnitudes of two forces: the gravitational attraction of the Moon, and the centrifugal force on the Earth. As noted above, overall these two forces have to be in balance, or else the Earth–Moon system would not be in stable equilibrium with each other. The fundamental point is that, despite this overall balance, the relative magnitudes of the two forces vary from point to point within the Earth. The force of gravitational attraction is straightforward, but the centrifugal force needs some careful explanation. To help you understand it, try simulating the Earth–Moon system with your fists, as follows.

3.2.7 Make two fists. Hold the left fist pointing upward, and the right fist about a foot away, pointing downward. Your left fist represents the Earth, and your right fist represents the Moon. You can easily make your right fist revolve around your left fist in a circular orbit. But to simulate the motion of the Earth–Moon system, you need to move your left fist at the same time in a much smaller circular motion in the same sense of revolution, in such a way that the center of mass of the Earth–Moon system, which lies just inside your left fist, stays in the same place, relative to the floor below you. (It's easier to do than to describe or to illustrate; I'll demonstrate in class.)

3.2.8 Important: ignore the Earth's rotation, because that's not relevant to the origin of the tides. Concentrate on the circular revolution of the Earth (your left fist). Every point in your left fist is experiencing a centrifugal force, because of the circular motion, and that centrifugal force is the same at every point in your fist. Moreover, that centrifugal forces is always directed opposite to your right fist. (To understand that, you have to be doing the movements right.)

3.2.9 At the same time, every point in your left fist is experiencing a gravitational attraction from your right fist. That gravitational attraction is always directed toward your right fist. So at every point in your left fist there are two opposing forces, which are almost in balance. In contrast to the constant centrifugal force, the gravitational attraction of your right fist is slightly greater on the side of your left fist nearest your right fist and slightly smaller on the side of your left fist farthest from your right fist.

3.2.10 Now think about the relative magnitudes of the centrifugal force and the gravitational force. On the side nearest your right fist, the gravitational force is slightly larger than the centrifugal force, resulting in a net force directed outward from the surface of your left fist. On the side farthest from your right fist, the gravitational force is lightly smaller than the centrifugal force, also resulting in a net force outward from the surface of your left fist. These upward forces on opposite sides of your left fist are the forces that raise the tidal bulge on the two sides of the Earth!

ADVANCED TOPIC: QUANTITATIVE TREATMENT OF THE DOUBLE TIDAL BULGE

1. I mentioned above that the Earth–Moon system is characterized by a balance between gravitational force and centrifugal force. By Newton's law of universal gravitation, the gravitational force is

$$F = \frac{Gm_1m_2}{r^2} \qquad (1)$$

where, given two bodies, F is the gravitational force of attraction between the two bodies, M_1 is the mass of body 1, M_2 is the mass of body 2, r is the distance between the centers of mass of the bodies, and G is the gravitational constant, equal to 6.667 x 10^{-8} cgs units. (Important note: this G is ***not*** the same as the acceleration of gravity g on Earth!). So, for the Earth–Moon system,

$$\text{centrifugal force} = \frac{GMm}{R^2} \qquad (2)$$

where M is the mass of the Earth, m is the mass of the moon, and R is the distance between the centers of mass of the Earth and the Moon.

2. But at any point on or in the Earth the local gravitational force and the local centrifugal force are not necessarily in balance. In terms of a unit mass located anywhere in the Earth, the centrifugal force (per unit mass) is the same everywhere, Gm/R^2, found by dividing the overall centrifugal force by the mass of the Earth, but the gravitational force per unit mass varies from point to point because the distance from that point to the center of the Moon varies.

3. *Why is the centrifugal force the same everywhere?* (This is tricky to think about; at first thought, it seems as though it should vary from place to place.) I think the key here is to neglect the rotation of the Earth around its own axis; that's a *different* centrifugal-force effect, not related in any way to analysis of the tides in terms of the revolution of the Earth and the Moon around their common center of mass.

4. If we assume that the orbits of the Earth and the Moon about their common center of mass are circles (actually they're ellipses with very small eccentricity), a little thought should convince you that every point on the (nonrotating) Moon describes a circular path in space, and (important) the radii of all the circles are equal—although the circles are nonconcentric. And the same holds true for all points of the Earth, even though a nonrotating Earth revolves around a point actually located inside the Earth.

5. But how about the effect of the Earth's rotation? That simply changes the equilibrium shape of the Earth to be a prolate spheroid instead of a sphere; it's a matter of a balance between the "pancake" effect of the spin and the "glob" effect of the Earth's own gravity field. The picture of tidal forces, based on considerations of gravitational versus centrifugal forces for a nonrotating Earth, is *superimposed* on this equilibrium shape.

6. Now think about the local balance of gravitational and centrifugal forces at two special points on the Earth: the point on the Earth's surface that lies along the Earth–Moon axis on the side *nearest* the Moon (called the ***sublunar point***), and the point on the Earth's surface that lies along the Earth–Moon axis on the side *farthest* from the Moon (called the ***nadir point***). Set up a coordinate system as shown in Figure 8-4. What I'm going to do is write the sum of the local gravitational force per unit mass and the local centrifugal force per unit mass at each of these special points.

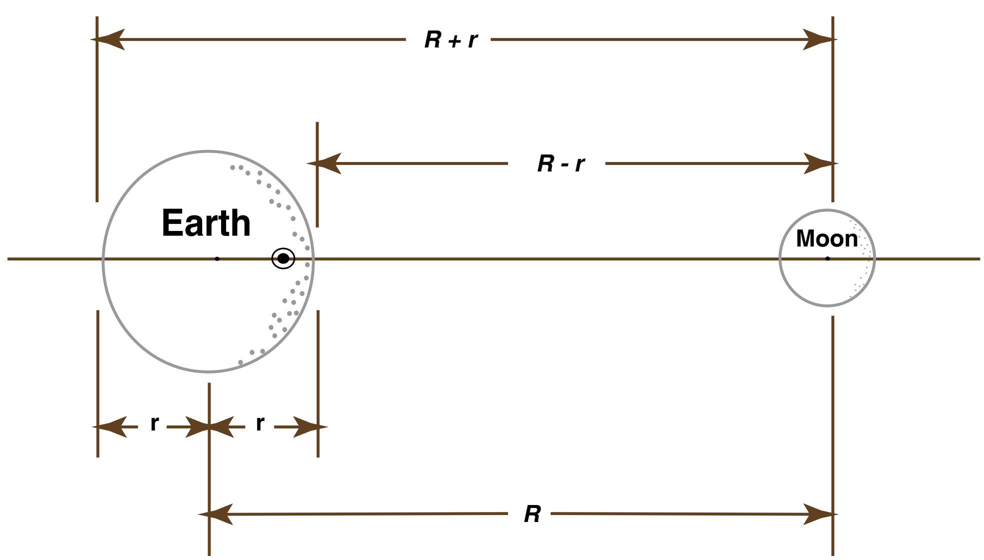

Figure by MIT OCW.

Figure 8-4. Definition sketch for developing the equilibrium theory of the tides.

7. At the sublunar point the local gravitational force per unit mass is $+Gm/(R-r)^2$, where $R-r$ is the distance from the sublunar point to the center of mass of the Moon. The local centrifugal force is $-Gm/R^2$, as discussed above; the minus sign is there because the centrifugal force acts in the direction opposite the gravitational force. So the *sum* of these two forces, which I'll call F_T, the net force per unit mass on Earth material located at the sublunar point, is

$$
\begin{aligned}
F_T &= -\frac{Gm}{R^2} + \frac{Gm}{(R-r)^2} \\
&= \frac{-Gm(R-r)^2 + GmR^2}{R^2(R-r)^2} \\
&= Gm\left[\frac{R^2-(R-r)^2}{R^2(R-r)^2}\right] \\
&= Gm\left[\frac{2rR-r^2}{R^2(R^2-2rR+r^2)}\right] \qquad (3)
\end{aligned}
$$

This looks formidable, but since $r << R$, then $r^2 << rR$ and $rR << R^2$, so to a very close approximation

$$
\begin{aligned}
F_T &= Gm\left[\frac{2Rr}{R^2(R^2)}\right] \\
&= \frac{2Gmr}{R^3} \qquad (4)
\end{aligned}
$$

8. On the opposite side of the Earth, at the nadir point, the same considerations on the net force per unit mass lead to

$$
\begin{aligned}
F_T &= -\frac{Gm}{R^2} + \frac{Gm}{(R+r)^2} \\
&= Gm\left[\frac{-(R+r)^2+R^2}{R^2(R+r)^2}\right] \\
&= Gm\left[\frac{-2rR-r^2}{R^2(R^2+2rR+r^2)}\right] \\
&= Gm\left[\frac{-2rR}{R^2(R^2)}\right] \\
&= -\frac{2Gmr}{R^3} \qquad (5)
\end{aligned}
$$

9. Wow! The net force per unit mass on a little drop of water in the ocean on the side of the Earth facing *toward* the Moon is *toward* the Moon, as you might have expected, but the net force per unit mass on a little drop of water in the ocean on the side of the Earth facing *away* from the Moon is *away* from the Moon (Figure 8-5). A deliciously counterintuitive result, no?

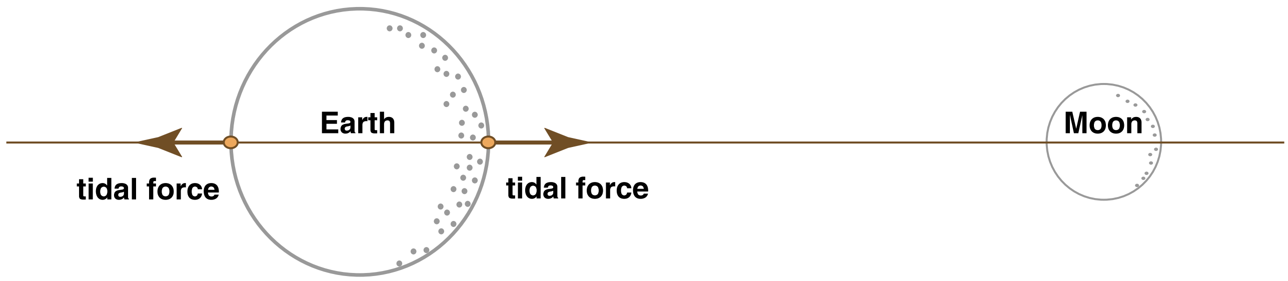

Figure by MIT OCW.

Figure 8-5. Tidal forces at the sublunar point and the nadir point.

3.2.11 At this point we should expand our consideration from one dimension to two dimensions and think about all the points on the intersection of the Earth's surface with any plane that passes through the Earth–Moon axis. That's more complicated mathematically, because we have an angle to worry about, but it's straightforward. I won't do it here. The result is shown qualitatively in Figure 8-6, which shows for a number of representative points the gravitational force, the centrifugal force, and the resultant of the two, and in Figure 8-7, which shows just the resultant.

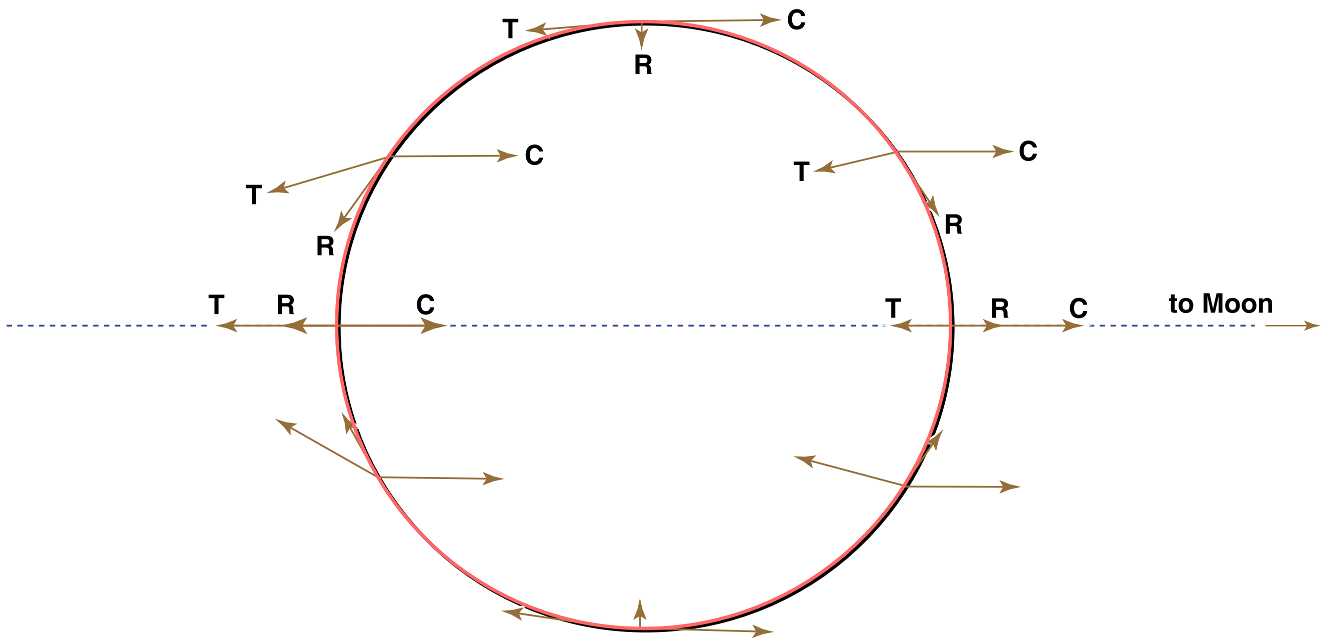

Figure by MIT OCW.

Figure 8-6. Tidal forces. T = gravitational force, C = centrifugal force, R = resultant force.

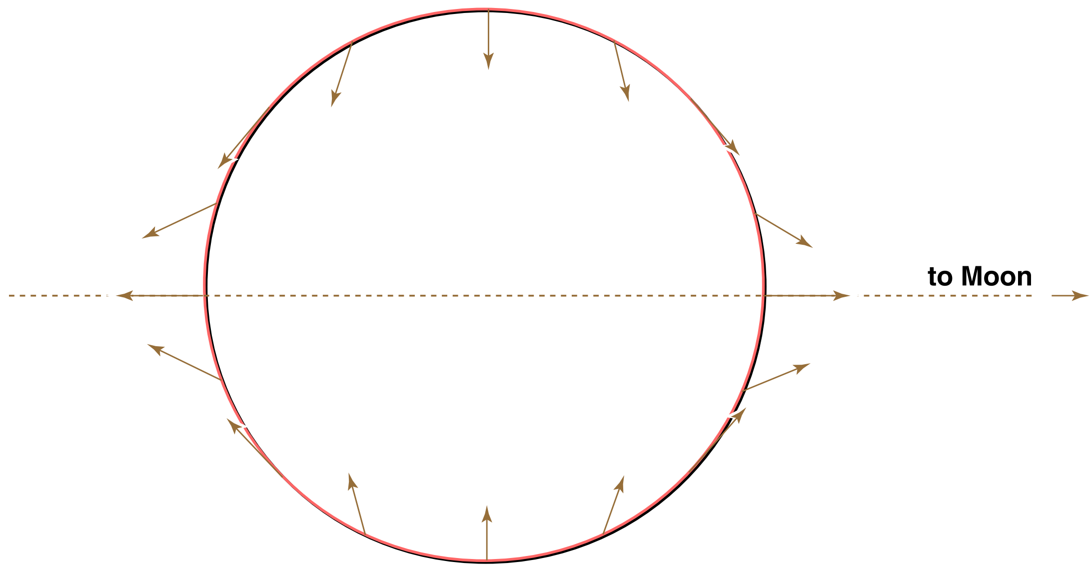

Figure by MIT OCW.

Figure 8-7. Resultant tidal forces.

3.2.12 The forces shown in Figure 8-7 can themselves be resolved into components along the local vertical and the local horizontal. If you plug in some numbers, you would find that the local vertical component of F_T is only about 10^{-7} that of the Earth's gravitational pull on an object. So the vertical component of the net force is has a negligible effect on things like ocean water at the Earth's surface. But the *horizontal* component of this net force is of about the same magnitude as the horizontal pressure-gradient forces that drive the water motions in the ocean. So the horizontal forces are important, and they can't be ignored. They are what cause the tides.

3.2.13 Figure 8-8 shows just the horizontal components of the net force F_T at various points around the Earth. They're zero at the sublunar and nadir points and at all points around the great circle that's normal to the Earth–Moon axis, and they're at a maximum around two small circles normal to the Earth–Moon axis and at 45° to that axis.

3.2.15 It's these horizontal forces that cause the tides. They pile up ocean water in two symmetrical bulges at the sublunar point and the nadir point and lower the water level in the vicinity of the great circle normal to the Earth–Moon axis.

3.2.16 The Sun produces a tidal effect of the same kind as that of the Moon, but the effect is smaller. Although the Sun has far greater mass than the Moon, it's also much farther away from the Earth. The tidal effect of the Sun is about 46% of that of the Moon. You can show this by forming the ratio

$$\frac{2Gm_{moon}\, r/R_{moon}{}^3}{2Gm_{sun}\, r/R_{sun}{}^3}$$

using the term on the right-hand side of Equation 4 or Equation 5 and substituting the appropriate values.

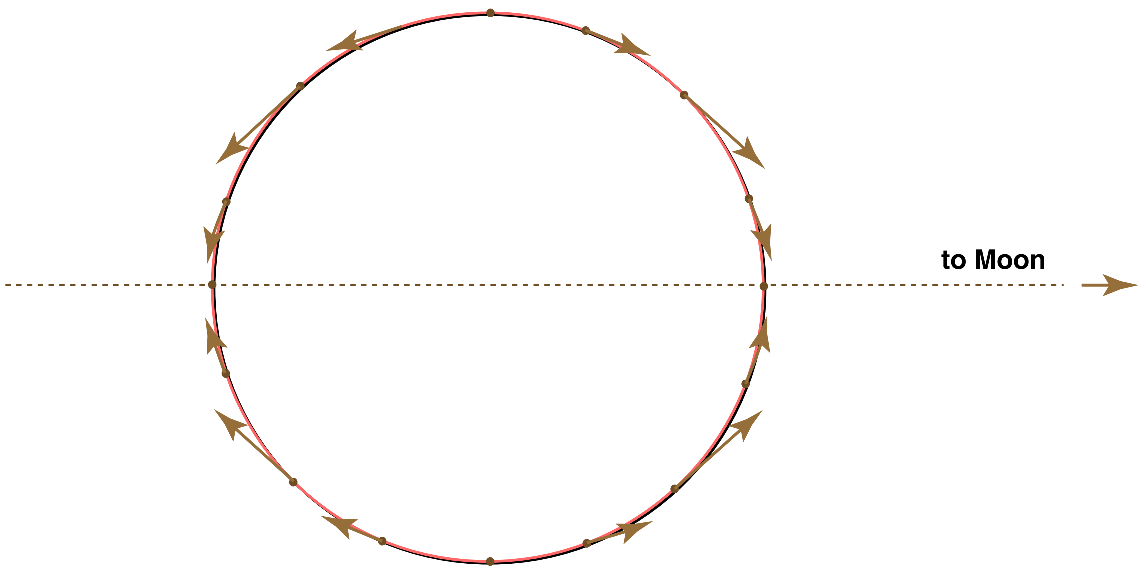

Figure by MIT OCW.

Figure 8-8. The horizontal component of the resultant tidal forces.

3.2.17 At this point you have to visualize that the Earth rotates on its axis under the tidal bulge, which, remember, is locked into position relative to the Moon. To you, the observer on the Earth, a tidal bulge seems to speed around the Earth twice a day, causing the rise and fall of the tides at your station. The period is the same as half the time it takes for the moon the reoccupy the meridian on successive days (24 hr 50.47 min, to be exact).

3.3 Some Complexities of the Equilibrium Tide

3.3.1 What I've presented above is basically what's called ***the equilibrium theory of the tides.*** Newton laid the foundation by developing this concept, and the details were well worked out by over a hundred years ago. But this is really only the beginning of the story of the equilibrium tide, because we would have to get more deeply involved in analysis of the relative motions of the Earth, the Moon, and the Sun to take account of all the variations. Below are just two examples of such matters—but they are the two most important ones.

3.3.2 In general the Earth's rotation axis is not perpendicular to the Earth–Moon axis. So when a point not on the Equator rotates under the two-ended tidal bulge the high tide at that point alternates between a higher high tide and a lower high tide (Figure 8-9). In Figure 8-9, a station at x_1 has a higher high tide now, but about 12 hours later, when that same station experiences another high tide, that high tide is a lower high tide, because now the station is at x_2, off the center of the tidal bulge. This is one cause of what's called ***diurnal inequality***, whereby *one of the two semidiurnal tides is larger than the other at a given station.*

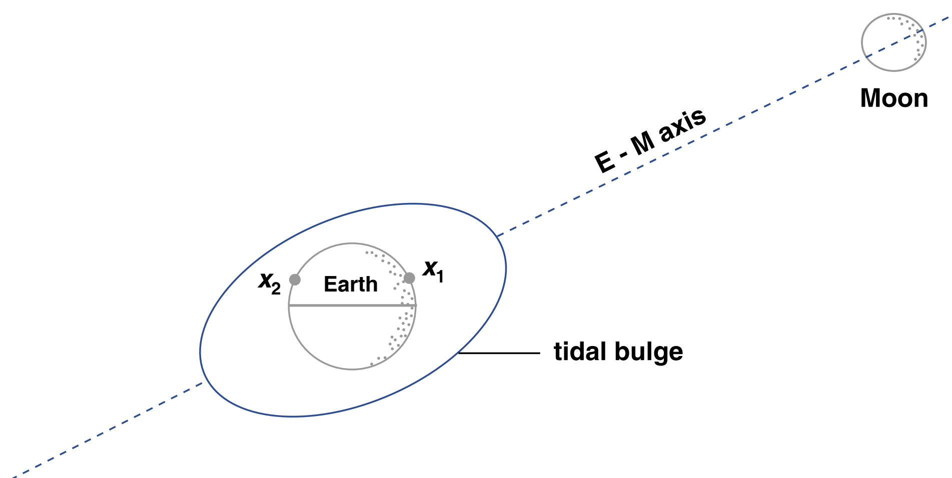

Figure by MIT OCW.

Figure 8-9. The orientation of the tidal bulge relative to the Earth's axis of rotation.

3.3.3 A striking worldwide effect of the tides is an alternation between greater and smaller tidal ranges on a time period of one lunar month. Tides with the *greater* tidal ranges are called ***spring tides*** (no relation to the spring season of the year!), and tides with the *smaller* tidal ranges are called ***neap tides***. The explanation of the spring–neap cycle is simple, and it has the side benefit of explaining also the phases of the Moon—something humankind in modern society has largely lost touch with, except for a swarm of boaters, smaller numbers of campers and fishermen, a rapidly shrinking group of old-fashioned gardeners and farmers who still insist on planting by the Moon, and a tiny band of scientists keeping knowledge alive.

3.3.4 In the course of one lunar month, the Moon makes one full revolution around the Earth. The Earth is at the same time revolving about the Sun, but it travels only a small fraction of a revolution in the time it takes for the Moon to go once around the Earth. Figure 8-10 shows the Moon at various positions around the Earth during one of its revolutions. The sense of rotation of the Earth around its own axis is shown also. The view in Figure 8-10 is perpendicular to the plane of revolution of the Earth around the Sun (called ***the plane of the ecliptic***), and looking down onto the Northern Hemisphere of the Earth.

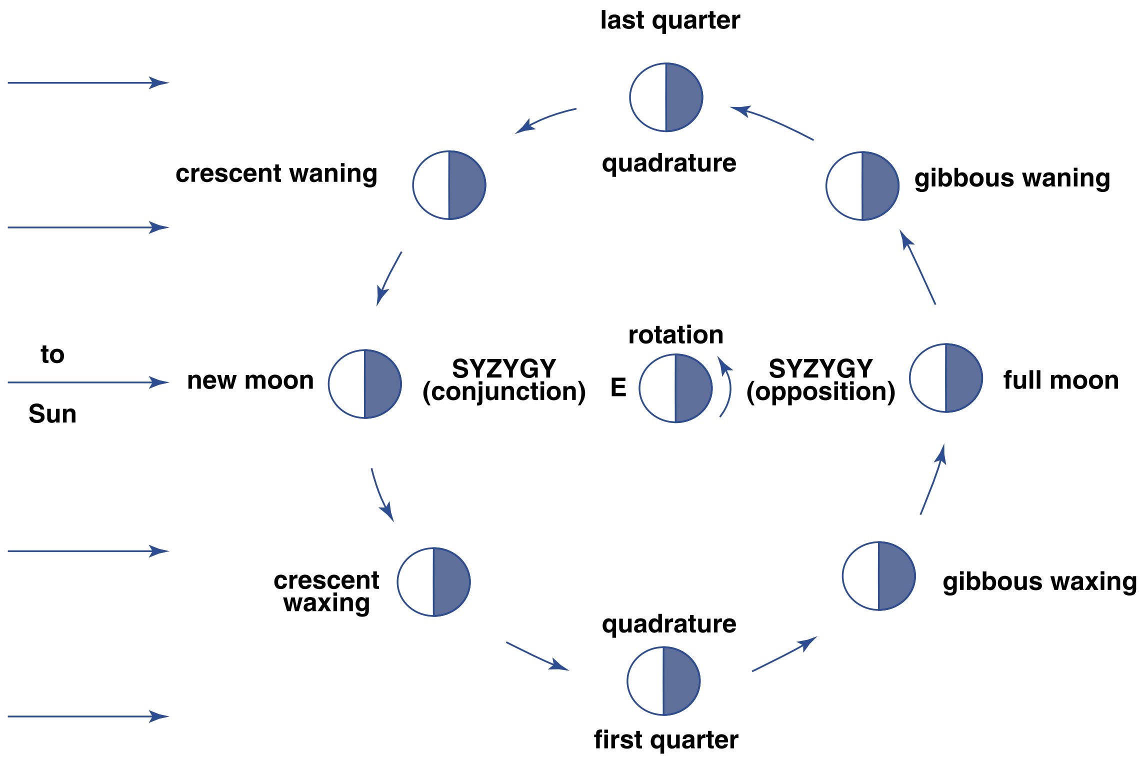

Figure by MIT OCW.

Figure 8-10. The phases of the moon

3.3.5 Twice during the month the Sun, the Earth, and the Moon are approximately in alignment. These are called ***syzygies*** (singular: ***syzygy***). The syzygy for which the Moon is toward the Sun is called ***conjunction***, and that for which the Moon is away from the Sun is called ***opposition***. When the line between the Moon and the Earth makes a right angle with the line between the Sun and the Earth, the Moon is said to be in ***quadrature***.

3.3.6 Obviously, the Sun and the Moon reinforce each other in their tidal effect at the time of syzygy, giving rise to spring tides, and they tend to cancel each other's effects at the time of quadrature, giving rise to neap tides.

3.3.7 You can also see in Figure 8-10 the explanation for the phases of the Moon. At the time of conjunction, you can't see the Moon; that's called the ***new moon***. At the time of opposition, you see the ***full moon***. At the times of quadrature you see a ***half moon***, officially called the ***first quarter*** and the ***last quarter***. The full set of terms is given in Figure 8-10. If you think about it just a little, you can also see from Figure 8-10 why the Moon rises a little later each day throughout the lunar month; it's because the sense of rotation of the Earth on its axis is the same as the sense of revolution of the Moon around the Earth, both being as shown in the figure.

3.3.8 There are many other such astronomical effects on the tides, having to do with

- the relative positions of the Earth, the Moon, and the Sun,
- the relative distance of the Moon from the Earth as a function of time during one lunar month, and
- the relative distance of the Sun from the Earth as a function of time during one year.

Accordingly, the tide varies complexly with time. These variations can be sorted out mathematically into what are called ***tidal constituents***: *regular periodic variations, with different amplitudes and periods, which are related directly or indirectly to the astronomical variations*. These constituents are given names and code designations. Table 8-1 shows the most important ones. The principal lunar tide, M_2, is the most important; it's the one derived in the "advanced topic" section above.

3.4 Tides on the Real Earth

3.4.1 That something is disastrously wrong with the equilibrium theory of the tide should be apparent to anybody who makes intelligent observations of the tides: the time of high tide is not the same as the time when the Moon is at its highest point in the sky. There's *a time lag between the time the Moon is highest in the sky and the time of high tide*. That time lag is called the ***lunitidal interval***. It's different at different points on the Earth, but it's always the same at any given point.

3.4.2 The reason for this effect is tied up with the fact that *the tide is a wave*: it has two crests and two troughs around the world, and it moves as a progressive wave from east to west. As the tide wave passes, the water oscillates, just as when a much smaller wind-generated waves passes, as described in Chapter 1, but the water undergoes no net translation over one tidal cycle. (Actually that's often not true in coastal environments with complex shoreline geometry and sea-bed topography, but it's certainly true as a broad spatial average.)

3.4.3 In the terminology of Chapter 1, the tide wave is a grossly *shallow-water wave*: the ratio of water depth (a few kilometers) to wavelength (basically halfway around the Earth) is extremely small.

Principal Lunar	M_2	12.42	semidiurnal
Principal Solar	S_2	12.00	
Larger Lunar elliptic	N_2	12.66	
Lunisoloar	K_2	11.97	
Lunisolar	K_1	23.93	diurnal
Principal Lunar	O_1	26.87	
Principal Lunar	P_1	24.07	
Lunar Fortnightly	M_f	327.86	spring - neap
Lunar Monthly	M_m	661.30	
Solar Semiannual	S_{sa}	2191.43	

Table 8-1. Tidal constituents. Figure by MIT OCW.

3.4.4 At this point you have to make a clear distinction between *free waves* and *forced waves*. A ***free wave*** propagates at a speed determined by its own wavelength and by the depth of water in which it's propagating. A ***forced wave***, on the other hand, is forced by external constraints to travel at some speed, slower or faster than its free-wave speed. The tide wave is clearly a forced wave, because it's constrained, whether it likes to or not (and it *doesn't*; see below), to travel around the Earth once a day. Whether such a forced wave lags behind its constraint or rides out in front of it depends on whether the ratio of the free-wave speed to the forced-wave speed is less than one or greater than one.

3.4.5 It's well known from the theory of water waves that the speed of propagation of a shallow-water wave is equal to $\sqrt{gh}$, where g is the acceleration of gravity and h is the water depth. You could easily compute from this that the free-wave speed of the tide wave is much less than the speed needed to get it around the Earth in one day. (Use 9.8 m/s^2 for g, and assume a representative water depth of 4000 m in the oceans. The number of seconds in the lunar day of 24 hr 50.47 min is about 89,428.) So *the tide wave lags behind the apparent motion of the Moon around the Earth* (Figure 8-11).

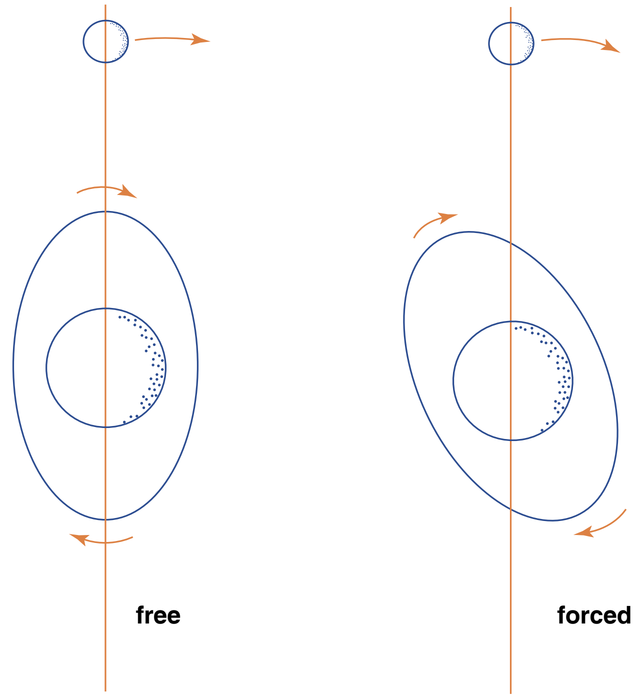

Figure by MIT OCW.
Figure 8-11. A forced tide wave vs. a free tide wave.

3.4.6 We're still not close to describing the true state of affairs in the real oceans. Only around the Southern Ocean, encircling a broad area in the Southern Hemisphere around Antarctica in the present configuration of the continents, can the tides do their thing as a wave that travels uninterruptedly all around the globe. All other areas of the oceans today are almost-closed basins, large or small, open only at their northern and/or southern ends. The critical question is: *How do the tides behave in such basins?*

3.4.7 The North Atlantic ocean basin is a good close-to-home example. If you ignore the minor effect of the narrow opening on the north into the Arctic Ocean, the North Atlantic is open only at its southern end. The basin feels a regularly rising and falling water level at its southern end. The effect of this rise and fall on the water levels and water motions in the basin is something that's not quite within your everyday experience, but it's something you could almost observe in your own bathtub. Here's a very simple home experiment you can make to study the effects:

3.4.8 Cut a sheet of plywood into long strips and nail the strips together to form a simple long trough closed at both ends (Figure 8-12). On the bottom of one end of the trough, place a separate short board of wood that stretches almost from one side of the trough to the other. Nail a vertical stick to this little board so that you can move it up and down regularly, to oscillate the water level at that end of the tank. That simulates the periodic rise and fall of the tide wave at the southern end of the ocean basin.

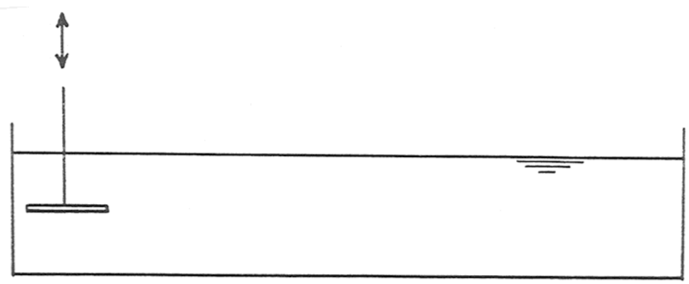

Figure 8-12. Making waves in a wave tank.

3.4.9 The up-and-down oscillation of the board at the end of your trough makes waves, and if your trough were endless you would just see a nice propagating train of waves, as described in Chapter 1. But the waves you make are reflected from the far wall of your trough. Those reflected waves travel back along the trough at the same speed as the oncoming waves, and they are of about the same amplitude (if you neglect the slight loss in energy when they're reflected at the vertical wall). The result is a beautiful pattern of ***standing waves,*** with ***nodes*** where the water level stays the same and ***antinodes*** where the oscillation in water level is at its maximum (Figure 8-13). The number of nodes depends on the length of your tank and the period of the oscillation you impose on it. So although waves are actually moving in both directions, all you see is their sum, a standing wave.

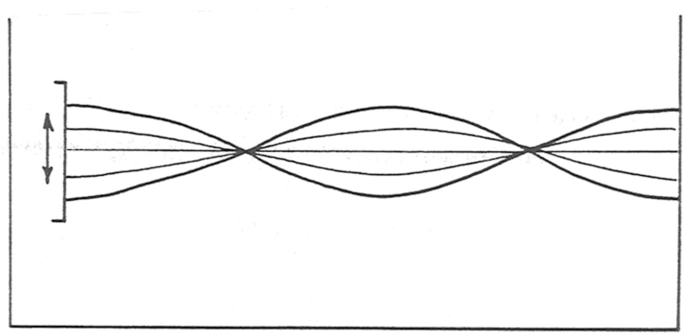

Figure 8-13. Standing waves in a wave tank. Each curve represents the water surface at a different time.

3.4.10 I used to have a wall phone in my kitchen, with one of those long spiraling lines to the handset. During particularly boring phone calls, I played at jiggling my end of the wire up and down to make standing waves on the wire. The faster I jiggled the wire, the more nodes I got. This is closely analogous to the behavior of the water in your trough.

3.4.11 Figure 8-14 shows the standing wave in your trough as a time-sequence cartoon for the simple case of only one node, in the middle of the tank. Given the size of real ocean basins, and the speed and period of the tide wave, this is what you're likely to find in the oceans. Note that the horizontal movement of the water is at its maximum at the nodes and is zero at the ends of the trough. This is just the kind of water motion observed in relatively long and narrow ocean basins.

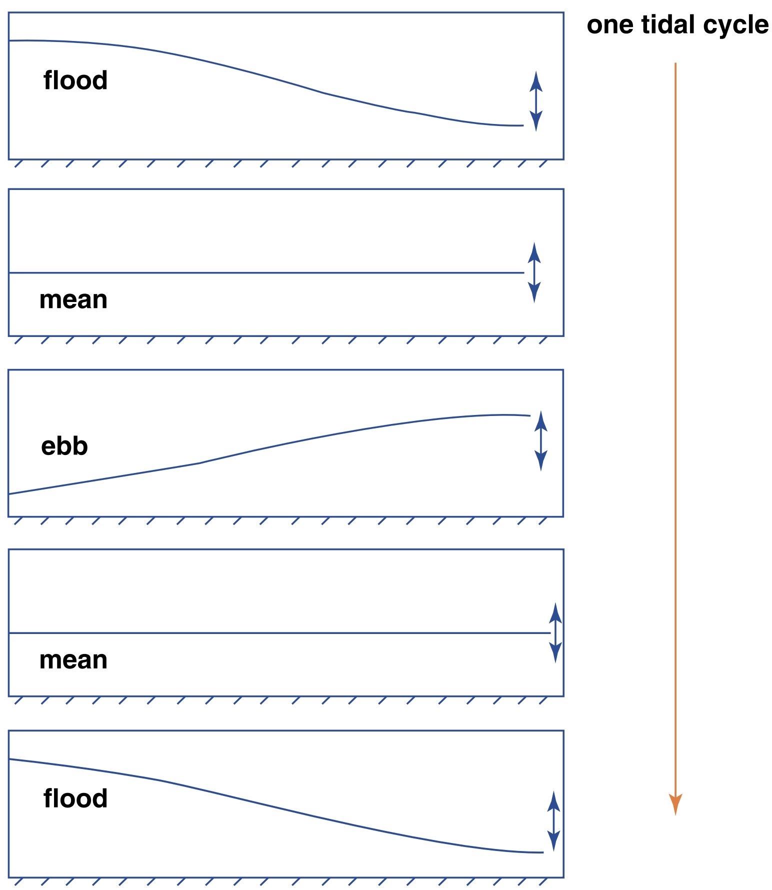

Figure by MIT OCW.
Figure 8-14. Time-sequence cartoon of a first-mode standing wave.

3.4.12 The Bay of Fundy is a good example of tidal standing waves. You probably know that the tides at the head of the Bay of Fundy are the highest in the world. Why? Because the size and depth of the Bay of Fundy are just about such that its natural frequency is close to the natural period of oscillation of seiching. (A ***seiche*** is *a periodic back-and-forth sloshing motion of water in an elongated lake or basin*. Once a difference in water level between the two ends of the water body is set up—by the wind, for example—the sloshing continues until it dies out by friction.) The amplitude of the forced oscillation is well known to be greatest when the forcing period is close to the natural period; this condition is called ***resonance***. You've all had lots of everyday experiences with resonance, whether you know it or not. Pumping on a swing is a good childhood example. Pushing your car in a rocking motion to get it unstuck on an icy road is a more grown-up example, as is sloshing the water in the roasting pan when you're trying to scrub the pan in the sink.

3.4.13 But this still isn't the end of the story! Unless the basin is very narrow, so that the water is constrained to move only back and forth, the Coriolis force has an important effect, unless you're near the Equator. (See the following background section to learn something about the extremely important, but also extremely counterintuitive, Coriolis effect.) Figure 8-15 shows how the Coriolis force affects the water motions in the basin. Keep in mind that in the Northern Hemisphere the Coriolis force acts to the right of the direction of motion. So when the water is moving from the back end of the basin toward the front end, the Coriolis force banks it up against the left wall, until the downslope component of gravity, acting away from the left wall, balances the Coriolis force. Then, when the water is moving toward the back of the basin, the Coriolis force banks the water against the right wall. The result is *a kind of circular sloshing motion* that's not easy to describe in a few words. This is called ***amphidromic motion***.

BACKGROUND: THE CORIOLIS EFFECT

1. In a large unobstructed indoor area (a gymnasium or a warehouse would be best, but a big room in your house would suffice), build a giant flat horizontal turntable—just a disk mounted at its center point on a vertical rotating shaft (Figure 8-16). You can rotate the whole disk at any desired constant rotation rate. It would be best if you painted the surface of the disk a flat black, the better to observe the motions of the brilliantly white marker spheres you're going to roll around on the surface. To make things really exciting, be sure to coat the white marker spheres with a thick chalky coating of some sort that tracks off evenly onto the surface of the turntable as they roll about.

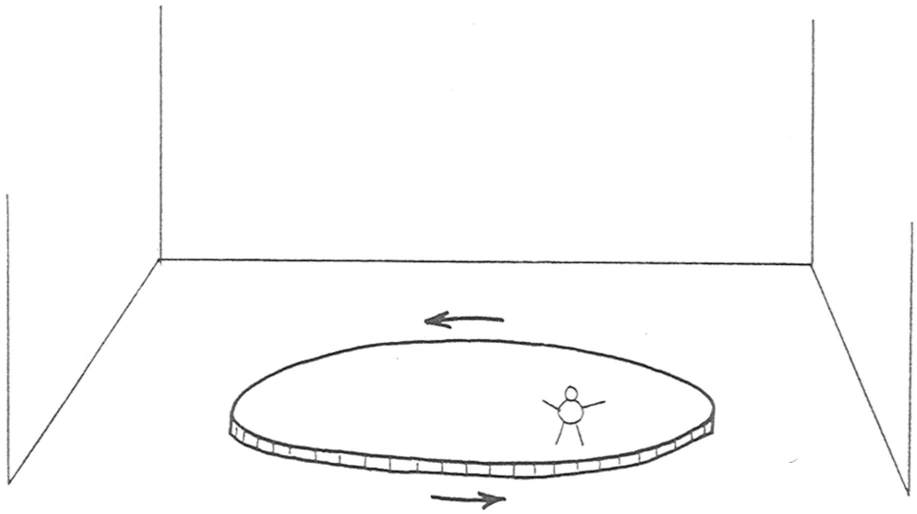

Figure 8-16. A turntable for investigating the effect of the earth's rotation.

2. To do things right, you're also going to need an observation perch above the turntable. This perch should be easily movable from place to place above the surface, but (and this is important) stationary relative to the floor of the room while in use. One of those mechanized cherry-picker seats, extending horizontally from the margin of the disk, would do nicely, if you can afford it.

3. Set the rotation rate, get into your perch, occupy a point just over the turntable, and roll one of your marked spheres onto the table, just as if you were at a bowling alley. Here's the big question: *What would the track of the sphere look like* on the turntable? (You're going to have to assume that the turntable exerts no substantial force on the rolling ball. That's not really true, but the effects are small enough that you can safely ignore them for the purposes of this demonstration. If you don't like that assumption, you can always imagine using a magic air-hockey puck that scoots frictionlessly over the turntable, leaving a powdery white trail behind it.)

4. The big jump that your powers of deduction or imagination have to take here is to see that *the track left by the ball on the table would be curved* (Figure 8-17). (And, once you're comfortable with that idea, it might naturally occur to you to think about whether that curved track is a circular arc. The answer turns out to be NO, although the reasons are a little too intricate to deal with at the moment.)

5 If you can't afford the time and money to build the turntable, but you still want to get some useful results, here's a much simpler and cheaper way of demonstrating the phenomenon (Figure 8-18). Pin a big piece of posterboard to the wall so that it can be rotated about its center point, and have an assistant stand to one side and rotate the posterboard in a hand-over-hand motion as steadily as possible. Stand in front of the posterboard with a marker pen, and draw a line on the posterboard in such a way that *the tip of the pen is moving in uniform rectilinear motion relative to the underlying wall* (that is, a straight line at constant speed). That's tough to do, because you have to try to ignore the surface of the posterboard and the mark that's coming out onto it and instead concentrate on the imaginary path of the pen point on the motionless wall behind. You would find (Figure 8-19) that no matter where you start on the posterboard, and no matter which direction you choose for your line, *the mark on the posterboard is a curving arc!*

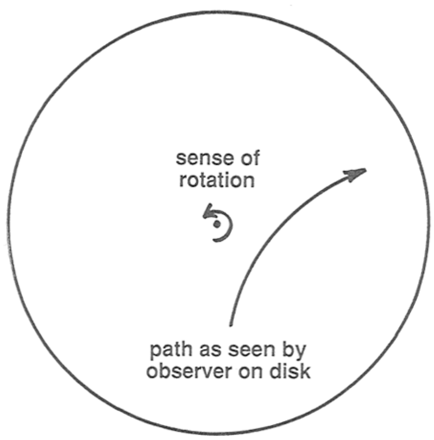

Figure 8-17. The curved track left by a ball rolling on a rotating turntable.

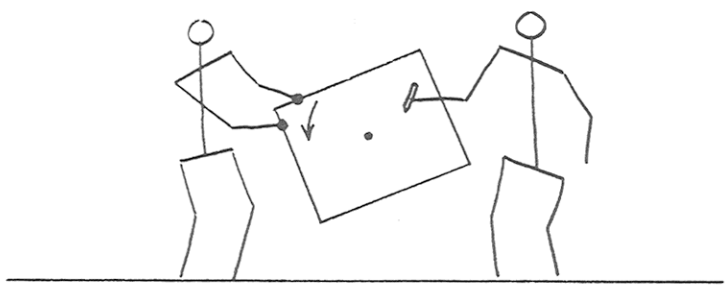

Figure 8-18. A simple way of demonstrating the Coriolis effect.

6. The next thing to do is have your assistant roll a marker sphere onto the turntable while you're riding on the turntable. Watch the ball as it rolls and leaves its circular-arc track. It will look to you as though some mysterious sideways force is continuously acting on the ball normal to its path to push it off its course.

Something seems to be wrong with Newton's first law, which tells you that the ball should be moving in a straight line at constant speed. You know what the problem is, of course: the fictitious side force is an artifact of your observing the ball from the standpoint of the rotating turntable. If you reoccupied your perch and rolled a clean, chalkless ball onto the dimly lit black surface of the turntable, you'd see the ball roll in a nice straight line! The fictitious side force that seems to act on moving bodies in a rotating environment is called the ***Coriolis force***, after the nineteenth-century French engineer–scientist Gaspard Gustave de Coriolis (1792–1843), who first studied the effect. And the apparent acceleration of the sphere (it's a radial acceleration, not a tangential acceleration, in that only the direction changes, not the speed) is called the ***Coriolis acceleration***. The entire effect is called the ***Coriolis effect***.

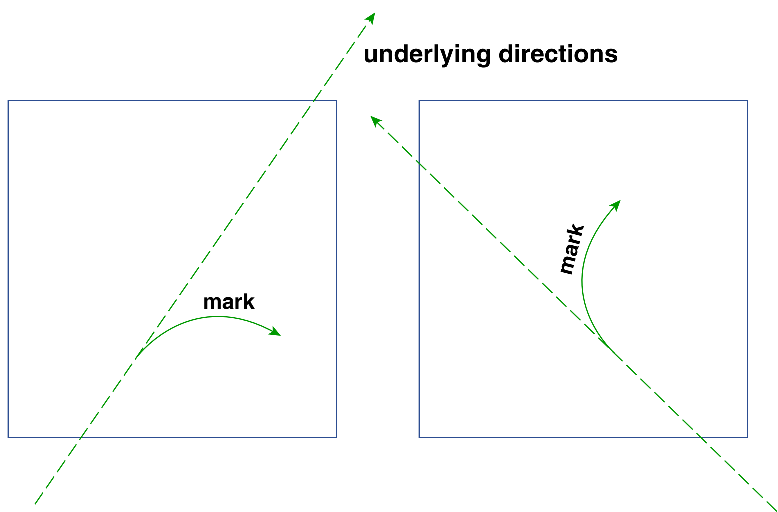

Figure by MIT OCW.
Figure 8-19. The results of the posterboard experiment.

3.4.14 In basins like this (Figure 8-20), there's some *point in the center of the basin where the tidal change in water elevation is zero*. That's called an ***amphidromic point***. Around the amphidromic point you can draw *closed contours that give the loci of equal tidal range*. These lines are called ***co-range lines***. Finally, you can also draw spoke-like *lines radiating outward from the amphidromic point toward the margins of the basin that show the locus of points at which the tide reaches a maximum at the same given time*. These lines are called ***cotidal lines***. A basin that has one amphidromic point with its associated cotidal lines and co-range lines is called an ***amphidromic system***. Figure 8-21 shows the North Atlantic as an amphidromic system. The complex geometry of the bottom topography and shoreline trace makes the details of the amphidromic system irregular, but the essential features are clearly there.

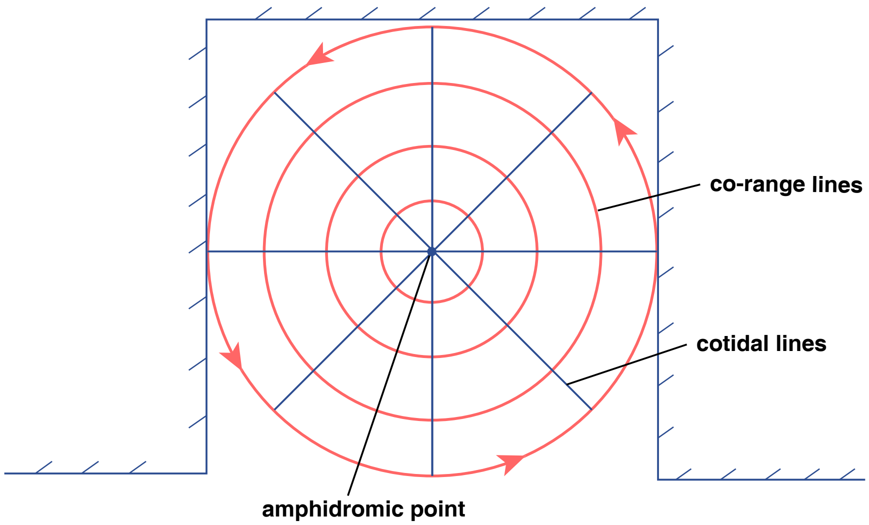

Figure by MIT OCW.

Figure 8-20. Elements of an amphidromic system.

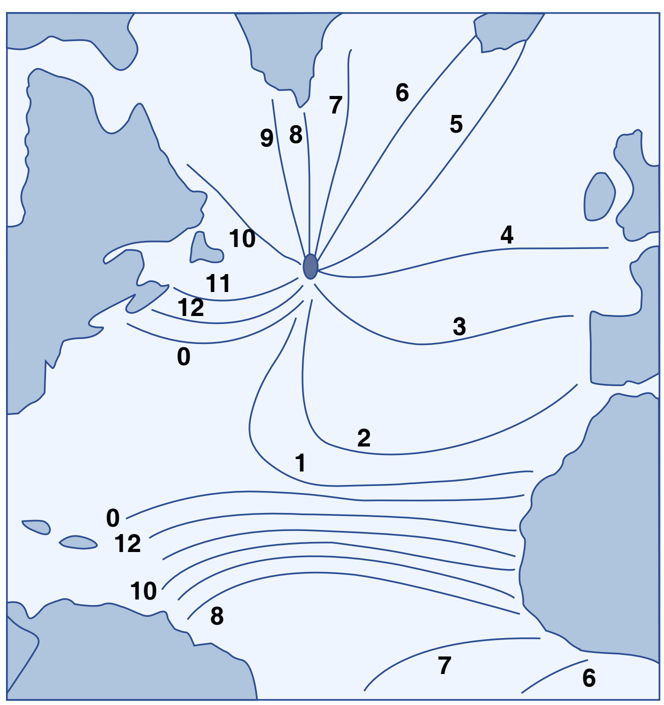

Figure by MIT OCW.

Figure 8-21. The amphidromic tidal system of the Atlantic Ocean.

3.4.15 Large amphidromic systems can act as drivers or forcers for smaller amphidromic systems. Smaller basins like large gulfs or bays around the margins of a large ocean basin have their own amphidromic systems, which are forced by the periodic tidal rise and fall of water level at their mouths as the tide sweeps amphidromically around the larger basin. But whatever the size of the basin, the foregoing exposition serves to explain nicely the progressive difference in times of low tide and high tide along a given stretch of coastline. Figure 8-21 shows why the times of low and high tides get later and later from north to south along the east coast of North America.

3.5 Tidal Currents

3.5.1 Tidal currents are the horizontal water movements associated with the tidal rise and fall of the sea surface. If you occupy a station in the ocean, you observe not only a systematic change in water level but also a systematic change in current velocity.

3.5.2 The connection between tidal water-surface elevation and tidal-current velocity is an indirect one, though. Because of the complexity of the actual tidal wave, depending on where you are there can be different relationships between the tide and the tidal current, and there can even be places with *tides but no tidal currents*, or *tidal currents but no tides*. But this shouldn't be too surprising, in light of our examination of standing tide waves in the preceding section. For the same reasons, there's no direct relationship between the times of high and low tides and the times of slack water, when no tidal currents are running.

3.5.3 Another thing: you have to average over a time long enough compared to the tidal cycle to factor out currents caused by random effects like winds and storms—but with the added complication that there may be nonzero average currents at a locality, caused by currents of other kinds in the oceans. Also, don't assume that the tidal currents themselves at a given locality have to average out to zero.

3.5.4 Why are there tidal currents? Tidal currents are simply the horizontal water motions associated with the passage of the tide wave, whatever the nature of that tide wave. Remember that tidal waves are very long compared to the water depth, so they are shallow-water waves. The water motion associated with such waves is a back-and-forth oscillation, actually an extremely flat ellipse, because the amplitude is so small relative to the wavelength (Figure 8-22). Even the sketch in Figure 8-22 has great vertical exaggeration, and also exaggeration of the length of the orbit.

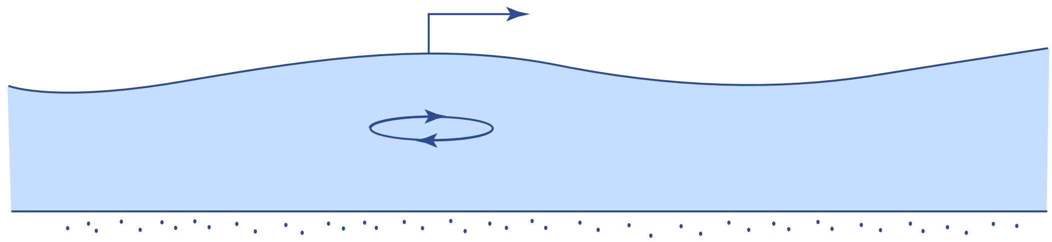

Figure by MIT OCW.
Figure 8-22. Water motions associated with the passage of the tide wave. (Vertically exaggerated.)

3.5.5 Some definitions:

flood current: the current that runs while the tide is rising

ebb current: the current that runs while the tide is falling

slack water: the period of no tidal current while the tide reverses

(Note: the concept of slack water is relevant only to areas with nearly bidirectional tidal currents.)

3.5.6 Complications in tidal currents arise because there's usually more than one progressive tidal wave, because of reflection, and commonly there are standing-wave systems as discussed above. So there's no general relationship between high and low tide, on the one hand, and the time of slack water and the speed of the tidal currents, on the other hand. That relationship varies from area to area. The time of slack water may or may not be close to the times of high and low tides.

3.5.7 Another complication is that there's no necessary relationship between the speed of the tidal current and the tidal range at a given place. In a very general way, of course, the speed of the tidal currents increases with the tidal range. But the speed of the tidal current is a matter of hydraulics, given the tidal range, and it depends on the volume of water moved versus the size of the passage through which the water has to move. For example, in the Gulf of Maine there's a large tidal range but weak tidal currents, whereas in Nantucket Sound there's a small tidal range but strong tidal currents. At a given station, however, the speed of the tidal currents is proportional to the tidal range, so currents are strongest during spring tides and weakest during neap tides.

3.5.8 How does one measure tidal currents? It's more difficult than measuring the tidal range. Like any other measurement of current, it requires a station that's fixed relative to the bottom, and a current meter of some kind. The picture of tidal currents is well known only for populated coastal areas.

3.5.9 What are the typical magnitudes of tidal currents? They're highly variable. Strong tidal currents are 2–3 m/s at the surface (or 75–80% of this when depth-averaged). And 1–2 m/s tidal currents are very common. Currents like that can move a lot of sediment—if the sediment is there to move.

3.5.10 How does one represent tidal currents? The best way is to plot the tidal-current velocity vector as a function of time over a complete tidal cycle, with all the vectors having a common origin. That kind of plot is called a ***tidal-current rose***. Figure 8-23 is an example from the Nantucket Shoals lightship. The following are some notes on Figure 8-23:

- The angular spacing of the vectors is not uniform, meaning that the turning of the current is not uniform in time.
- Low tide and high tide are not spaced equally in time.
- The maximum currents are not symmetrical in time.
- A floating object would make a closed traverse with the same shape (although not the same size!) as the curve connecting the tips of the vectors (and because the velocity vectors are turning clockwise, the object would move clockwise also).
- A current rose like that shown in Figure 8-23 is often approximately elliptical. If that's the case, then it's called a ***tidal ellipse***.
- Current roses can be open, as in Figure 8-23, or more closed, as in Figure 8-24. As the current rose becomes flatter, the tidal currents become bidirectional tidal currents.

4. BEACHES

4.1 Shoaling Waves

4.1.1 First go back to Chapter 1 and review the material on surface gravity waves. Now it's time to think about what happens to a train of waves propagating from the deep ocean to the shoreline. The waves start as deep-water waves, but as they shoal the begin to "feel the bottom", in the sense that they make the water move at the bottom. This bottom friction causes some loss of wave energy. Also, if the orbital velocity of the water at the bottom becomes great enough to move sand, ***oscillation ripples*** (a kind of oscillatory-flow bed configuration) develop (Figure 8-25).

4.1.2 One of the most important effects connected with shoaling waves is that they *slow down*. It turns out that for shallow-water waves the wave speed c is proportional to the square root of the water depth. Because the wave period itself doesn't change (the waves keep on being supplied from offshore at the same period), by the relation $L = cT$ *the wavelength decreases*. Wave height H also increases, for reasons connected with conservation of wave energy that are too complicated to discuss here, and the combined effect leads to *increase in wave steepness*. This can't go on indefinitely (the limiting steepness H/L turns out to be one-seventh), so eventually the waves fall over on themselves, or in other words they *break*.

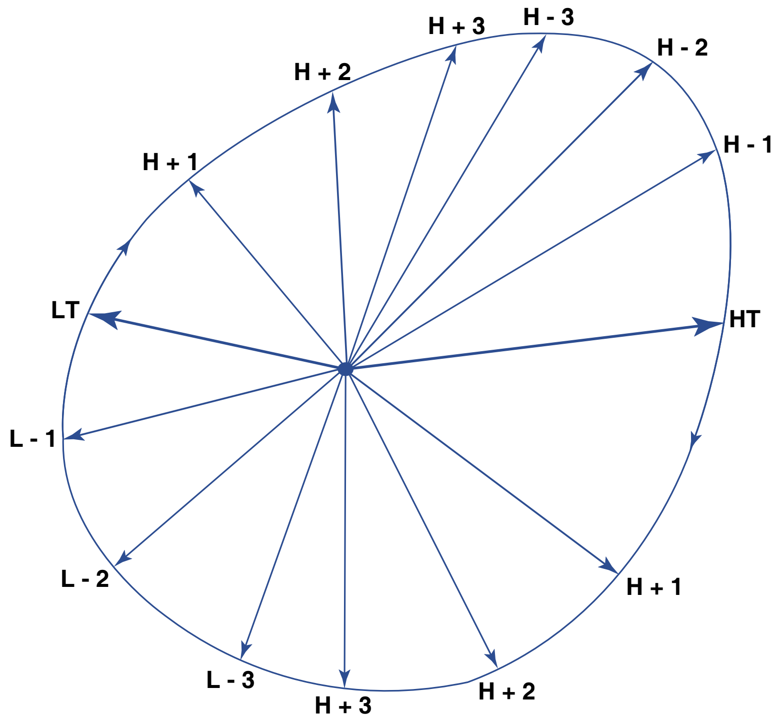

Figure by MIT OCW.

Figure 8-23. The tidal-current rose measured at the Nantucket Lightship.

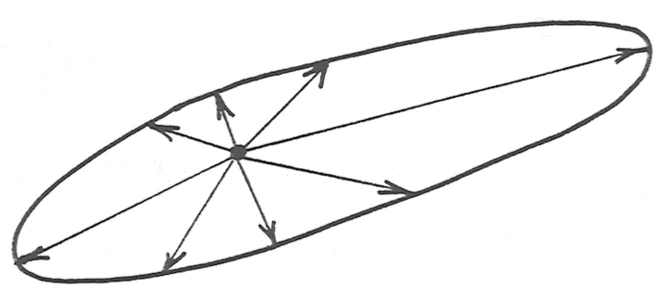

Figure 8-24. An example of a more elongated tidal-current rose.

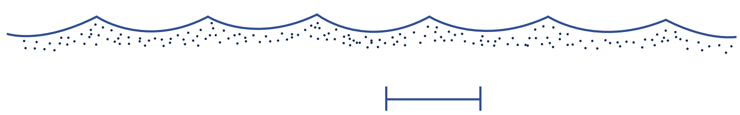

Figure by MIT OCW.
Figure 8-25. Oscillation ripples.

4.1.3 The form taken by the breaking waves depends on the rate of change in the steepness as the waves shoal. Three types of breakers are generally recognized (but keep in mind that there's a gradation among them):

spilling breakers: the wave peaks up slowly, the crest becomes unstable and spills down the front surface of the wave, and the wave height slowly decreases (Figure 8-26A). These tend to occur with gentle bottom slope and originally steep waves.

plunging breakers: the wave peaks up rapidly, and the crest becomes a thin vertical wall that curls over forward and then plunges forward and downward, causing a catastrophic decrease in wave height (Figure 8-26B). These tend to occur with steep bottom slope and originally intermediate-steepness waves.

surging breakers: the wave peaks as if to plunge, then the base of the wave surges up the beach face, and the crest collapses and disappears (Figure 8-26C). These tend to occur with very steep beaches and originally low-steepness waves.

4.1.4 After the wave has broken, the water rushes up the beach as a moving mass, which can be thought of as a *wave of translation* (in contrast to former nature of the wave, wherein the water had no net translation). This called the ***swash***. The water carried up the beach as swash returns under the downslope pull of gravity as ***backwash***.

4.1.5 Another effect of the slowing of waves as they shoal is ***refraction***. You may have noticed that waves moving toward shore tend to curve around so as to be more nearly parallel to the shore (Figure 8-27). This is an inevitable consequence of the slowing of the waves as they move into shallower water. Here's an easily understandable analogy. The drill sergeant is putting a squad of marching persons through their exercises. At one point he/she tells the people on the left to take baby steps and the people on the right to take giant steps. You can easily envision what will happen: the direction of movement of the squad will curve around to the left. This is exactly the same refractive effect that makes the shoaling waves swing around to be more nearly parallel to the shoreline.

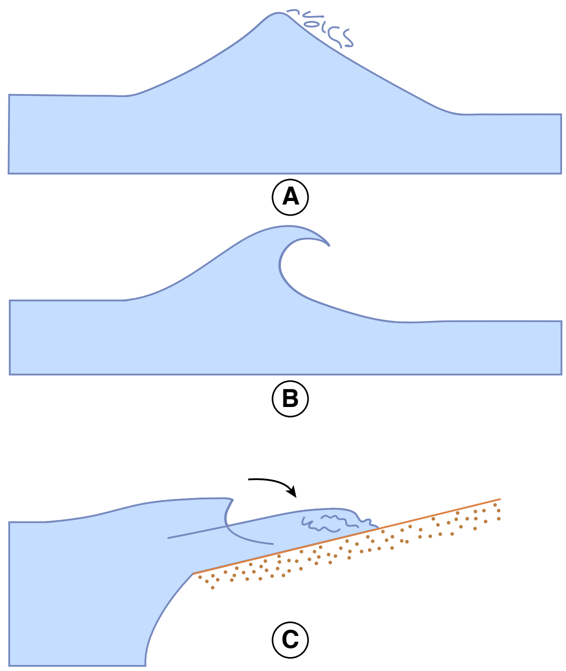

Figure by MIT OCW.
Figure 8-26. Varieties of breaking waves. A) Spilling breakers. B) Plunging breakers. C) Surging breakers.

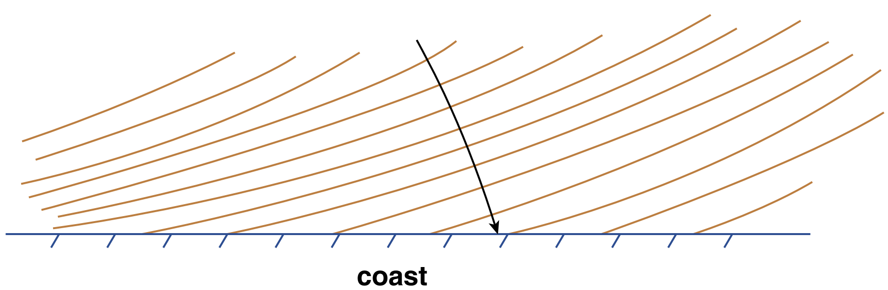

Figure 8-27. Refraction of waves as they shoal toward a shoreline.

4.2 Beach Profiles

4.2.1 It's easy to define a beach. A ***beach*** is *a mass of noncohesive sediment along a shoreline that's molded by wave action*. The sediment size of beaches ranges from the finest sand to coarse gravel. The best way to approach the dynamics of beaches is to look at a vertical cross section normal to the shoreline; such a cross section is called the ***beach profile***.

4.2.2 To establish some terminology, Figure 8-28 is a sketch of a representative beach profile. I want to emphasize that beach profiles vary a lot in their geometry and in the kinds of features they show—both from beach to beach and at the same beach as a function of time. The sketch in Figure 8-28 is a typical snapshot, but it's not representative of all beaches.

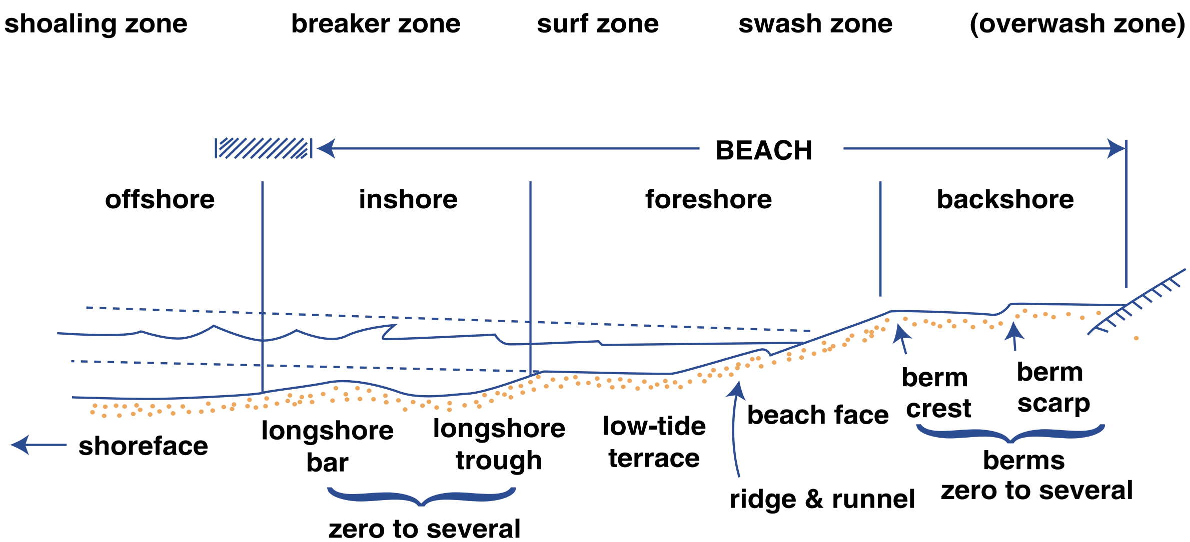

Figure by MIT OCW.

Figure 8-28. Shore-normal vertical profile through a beach.

4.2.3 (The *seaward limit* of the beach is a matter of definition; depending upon who's doing the defining, it can range from mean low tide to the seaward limit of breaking waves, which can be much farther offshore.)

4.2.4 Some general points about this profile:

• As discussed in the section on shoaling waves above, the strength and nature of bottom-water movements associated with breaking waves vary greatly and systematically from deep water to land.

• The beach profile is always changing with time, in an effort to keep up with the ever-changing wave state. We naturally assume that for each given wave state and beach sediment there's an equilibrium profile, in the sense that, whatever the initial geometry of the beach, the final equilibrium geometry attained by the beach after prolonged exposure to that wave state is the same. But because in the real world the wave state changes all the time, the beach is always adjusting toward a new equilibrium.

• The nature of sediment movement on the beach varies not only with the waves but also with the sediment size, which itself depends in a complex way on the long-term average wave state and on the nature of sediment supply to the beach.

4.3 Onshore–Offshore Transport

4.3.1 The main feature of sediment movement on the beach is the back-and-forth movement of the sand in response to the swash and backwash of the waves. Sand is carried up the beach as both bed load and suspended load by the relatively deep and strongly turbulent swash, and it's carried back down the beach mostly as bed load by the thinner and less turbulent backwash.

4.3.2 For a given beach sand, the vertical profile of the beach is very sensitive to wave conditions. At any given time there's nearly a balance between sand moved up the beach by the swash and sand moved down the beach by the backwash. The volume of backwash is always smaller than the volume of swash, because water percolates into the porous beach sand. To maintain the balance in upslope and downslope sand movement, the beach takes on a particular slope so that the force of gravity hinders upslope transport by the swash but aids downslope transport by the backwash.

4.3.3 What's therefore important in governing the slope of the upper part of the beach is *the ratio of backwash volume to swash volume*. As wave height increases, the volume of water moving up and down the beach increases, but because the volume lost by percolation doesn't increase very greatly the percentage of backwash compared to swash is greater than for waves with low steepness. So more material is moved down the beach, and the slope on the upper part of the beach decreases. The sand thus eroded from the upper part of the beach is parked farther offshore, although still within the beach system. During periods of good weather, with smaller waves, the ratio of backwash volume to swash volume decreases, and the slope of the upper part of the beach increases again, as sand is supplied from offshore. Figure 8-29 shows beach two profiles, one for small waves and one for large waves.

4.3.4 Ordinarily the sand that's parked offshore during periods of strong waves isn't lost to the beach system, and it's moved back onto the upper part of the beach—the part we see and lie on and play on—during periods of weaker waves. Only during unusually large storms can there sometimes be mobilization of great masses of sand and transport farther out to sea.

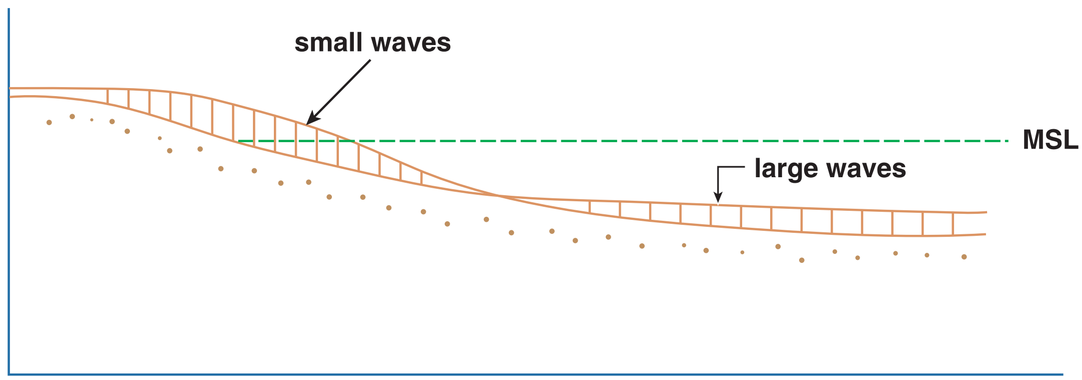

Figure by MIT OCW.
Figure 8-29. Beach profiles for times with small waves and times with large waves.

4.3.5 By much the same line of reasoning, *the slope of the beach varies directly with sand size*: coarser beaches admit of greater percolation, so for a given wave state the beach has to be steeper to strike the balance between upslope and downslope sediment movement. Beaches with fine sand have very gentle slopes, sometimes of only a few degrees. Such beaches can be hundreds of meters wide. On the other hand, beaches formed of coarse gravel in areas exposed to very strong waves can be as steep as the angle of repose of the gravel, well over 30°. (The ***angle of repose*** is *the slope angle that's assumed by a pile of loose sediment that's formed by pouring the sediment from a point above the pile*.) On such beaches there's no backwash: the water of the swash percolates so rapidly into the beach that there's none left to flow back down the beach as backwash! Then the downslope pull of gravity is the only thing that counteracts the upward transport by the swash.

4.4 Longshore Transport

4.4.1 So far I've talked only about sediment movement in the onshore–offshore direction. Sand is also moved *along* the beach, in various ways. Such movement is called ***longshore transport***, or ***littoral drift***, or ***longshore drift***.

4.4.2 How is longshore transport observed? It's notoriously difficult to make direct measurements of sediment transport rate, especially in complicated situations like beaches, where various kinds of water movements are superimposed upon one another, but two lines of indirect evidence should be readily understandable:

- piling up of sand in front of groins and jetties, and depletion behind them (Figure 8-30);
- building of sand spits at headlands (Figure 8-31).

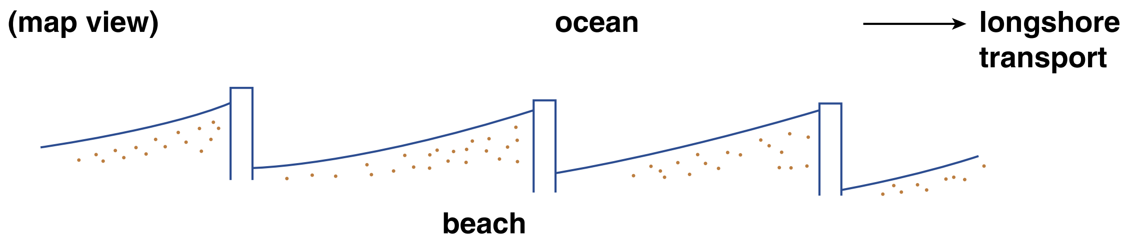

Figure by MIT OCW.
Figure 8-30. How sand piles up a the updrift sides of groins.

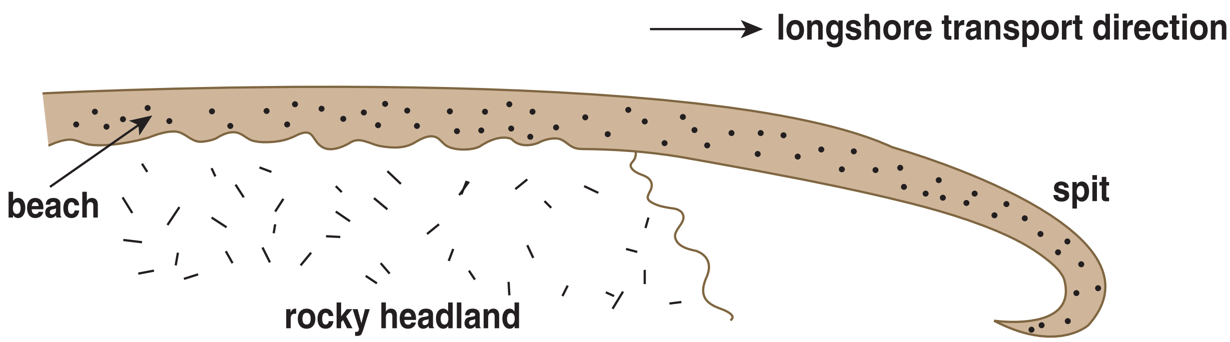

Figure by MIT OCW.
Figure 8-31. A sand spit at a headland.

4.4.3 Another good line of evidence, less obvious to the casual observer, is the presence of distinctive sediment along the beach downdrift from the mouth of a river (Figure 8-32).

4.4.4 Finally, you can imagine marking or tagging a small volume of sediment, with paint or even with a radioactive coating, and then looking for it in the downdrift direction after some period of time. That's an excellent way of getting qualitative information on longshore transport. But getting actual rates of movement, in terms of mass of sand moved per unit time, is much more difficult, largely because of problems connected with mixing of the marked sand downward into the underlying sand.

4.4.5 Longshore transport is complicated, because several different mechanisms contribute to it. One kind of longshore transport, called ***beach***

drifting, is easy to understand: it comes about because the waves approach the beach obliquely rather than straight on. (It's generally the case that waves don't approach the beach straight on.)

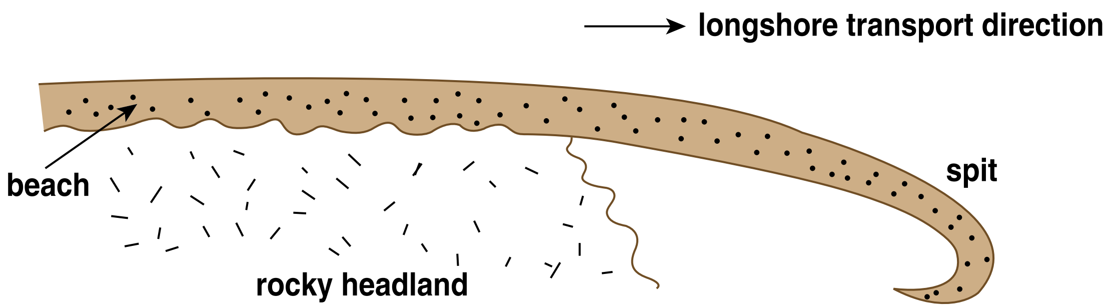

Figure by MIT OCW.
Figure 8-32. Downdrift movement of a tracer introduced to the coastline by a stream.

4.4.6 Despite the tendency for wave refraction to make the waves more nearly parallel to shore, waves still tend to be slightly oblique to shore when they break (Figure 8-27). So the swash (the wave of translation produced by breaking) runs not straight up the beach but diagonally, moving sand diagonally with it. As the swash loses its momentum and is pulled back down the beach by gravity, it flows more nearly straight down the beach, again carrying some sand with it. The net effect on the sand is to be displaced slightly down the beach in sawtooth fashion, in the direction of wave approach to shore (Figure 8-33).

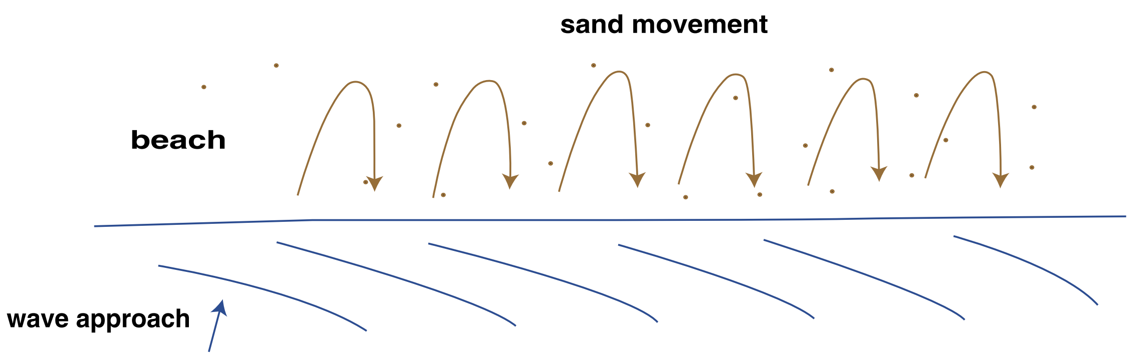

Figure by MIT OCW.
Figure 8-33. Beach drifting.

4.4.7 The other important cause of longshore transport is the presence, immediately adjacent to the shoreline, of a steady current flowing parallel to the beach. Such a current is called a ***longshore current***. The causes of longshore currents are varied. Here are three possibilities:

- the sawtooth movement of swash and backwash by obliquely approaching waves, described above;
- currents caused by complex processes associated with the oblique approach of the waves themselves;
- currents present in the nearshore zone but not caused by wave shoaling.

4.4.8 Whatever their origin, longshore currents can transport sand just like currents in a river. In fact, the ability of longshore currents to transport sand is enhanced by the suspension of sand by breaking waves in the breaker zone: there can be appreciable longshore transport by longshore currents even though the currents are not strong enough to move the sand by themselves.

4.4.9 Beach erosion is much in the news, especially in southern New England and on Long Island. The best way to think about beach erosion is in terms of the budget of sand involved in longshore transport, because we've seen already that little sand is lost to the beach by direct offshore movement. In this view, beach erosion in one locality is offset by beach deposition somewhere else. In areas where less sand is supplied from upstream than is transported downstream there's beach *erosion*, and the position of the beach moves *landward*; in areas where more sand is supplied from upstream than is transported downstream, there's beach *deposition*, and the position of the beach moves *seaward*.

4.4.10 In the context of Cape Cod or Long Island, you have to keep in mind that they are basically just gigantic piles of coarse sediment that were parked there while the terminus of the last continental ice sheet stayed in about the same place for a long time at the height of advance of the ice sheet, and after sea level rose to about its present position those piles of sediment have been exposed to wave action and longshore transport without any new source of sand, as from a major river system. So the only source of sand for the inevitable longshore transport comes from the beach itself, and the sediment cliffs behind the beach. On the other hand, lots of new beach is being built by longshore transport to the "downstream ends" of the beaches—Provincetown and Monomoy, in the case of Cape Cod.

4.4.11 Figure 8-34 is a sketch of outer Cape Cod, showing the direction of approach of the dominant sand-moving waves and the net longshore transport directions. South of a point about in the middle of the face of the outer Cape, net longshore transport is to the south, and north of that point, net longshore transport is to the north. The dividing point is called a ***null point***. For a long distance north and south of the null point there's net erosion of the beach. But to the north and south of that zone of net erosion, there's net deposition, as the spits extend themselves ever farther.

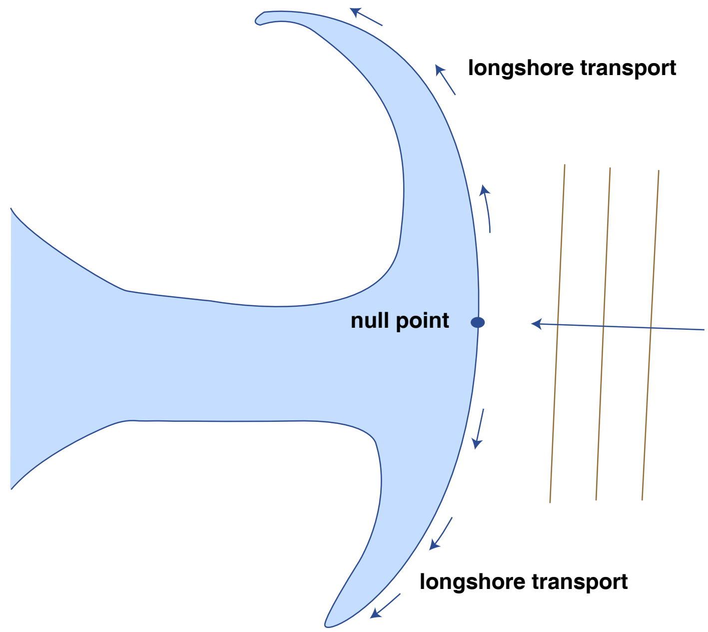

Figure by MIT OCW.
Figure 8-34. Longshore transport along the outer face of Cape Cod.

5. DELTAS

5.1 Introduction

5.1.1 A ***delta*** is *a body of sediment deposited at a point along a body of water where a sediment-transporting channelized flow of water enters the water body*. Deltas range in size from those little decimeter-scale bodies that you can see forming when a rivulet of rainwater enters a puddle during and right after a rain, to giant bodies at the mouths of major rivers like the Mississippi. The reason deltas form—and I suppose that this is obvious—is that as the channel flow enters the water body it spreads out and thereby decelerates, dropping much of its sediment load at or not far from its mouth.

5.1.2 The concept of a delta is simple, but the geometry of large deltas in the real world is rather complicated and highly varied, owing to a variety of factors. More on that later; first, some material on the hydrodynamics of deltas.

5.2 The Hydrodynamics of Deltas

5.2.1 To understand deltas, you need to know something about jets. In fluid dynamics, a ***jet*** is *a fluid motion created where a high-speed flow in a pipe or other conduit enters a large body of relatively still fluid*. It's easy to visualize the classic jet (Figure 8-35). A pipe enters a large tank of still water. When the water flowing in the pipe enters the tank, it no longer feels the force (the downstream pressure gradient) that was driving the flow in the pipe. Its momentum carries it out into the water of the tank, but friction with the surrounding fluid robs it of its momentum, and it eventually slows to a stop. Most such jets are turbulent—unless they are very small, very slow, and/or consist of very viscous fluid—and the mixing of eddies at the margins of the jet entrain ambient water into the jet, causing it to widen.

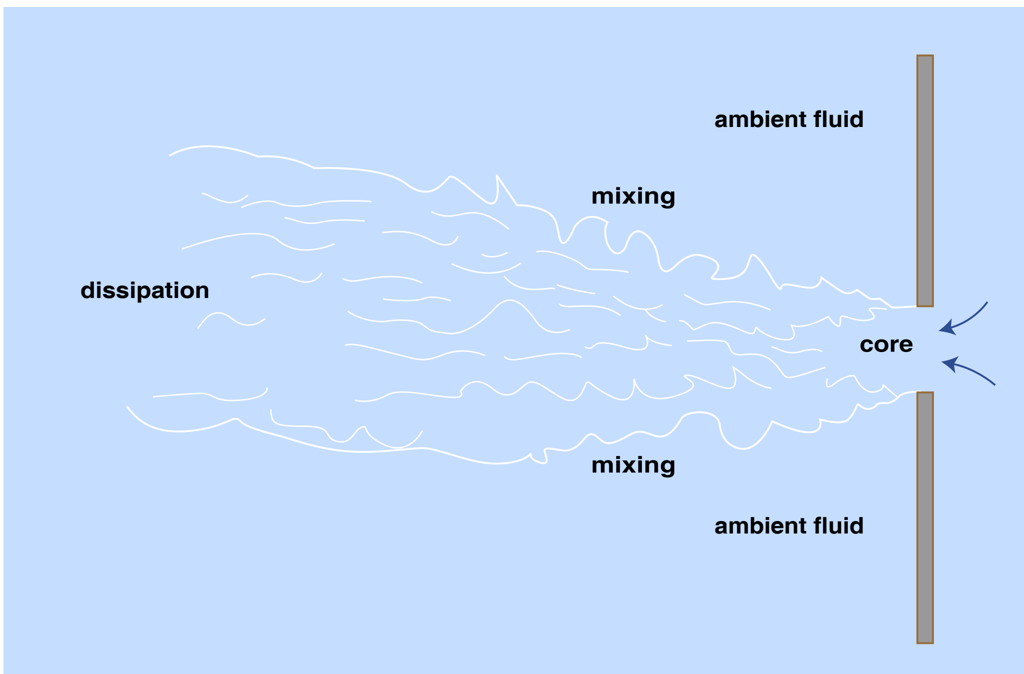

Figure by MIT OCW.

Figure 8-35. A jet entering a body of still fluid from a pipe. This is a cross-section view, cut through the axis of the pipe flow and the jet.

5.2.2 As you can imagine, the mixing at the margin of the jet tends to even out the properties of both the jet and the ambient fluid—which might be temperature, salinity, or concentration of suspended solids. Note in Figure 8-35 that there is a residual core in the jet, which has not yet been affected by the marginal mixing. This core shrinks to nothing in the downstream direction.

5.2.3 You're probably thinking: What does all of this hydrodynamics have to do with deltas? To make the connection more direct, think in terms of a jet with a somewhat different geometry: a free-surface channel flow entering a body of still water, as might a river flowing into a lake (Figure 8-36). The essential nature of the jet is no different; what is different is the geometry. The jet still mixes with the ambient fluid, but it can't mix upward, only sideways and downward. If the channel flow is muddy water, you can imagine how the jet would appear from the air: a spreading mass of muddy water, surrounded by clear water, with a core of muddiest water extending outward from the entrance and the muddiness gradually decreasing outward in all directions.

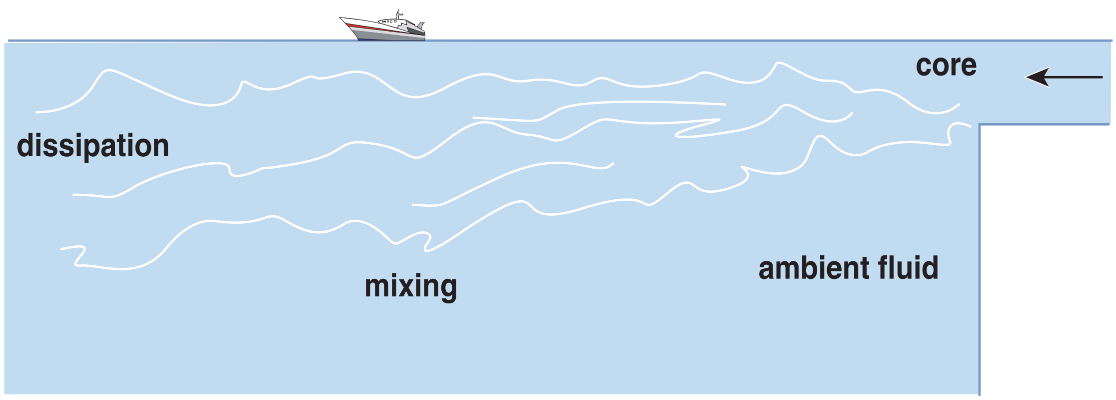

Figure by MIT OCW.
Figure 8-36. A jet formed where a channel flow enters a standing body of water. Vertical section view, through the axis of the jet. (A plan view would look much like what is shown in Figure 35, except that the shape of the jet is not axially symmetric.)

5.2.4 Now I'll throw in another complicating factor. The jet described in the preceding paragraph is a good model of a river entering a fresh-water lake or reservoir. But, as we all know, the ocean is salty, and, because it's salty its density is significantly greater, by an important few percent, than that of fresh water. That impedes vertical mixing of the jet with its surroundings, while not affecting its horizontal mixing greatly. The reason is that the density stratification that develops between the overlying fresh-water jet and the underlying salt-water medium is gravitationally stable, and it takes work to disrupt or break down that stratification by mixing. The jet may be able to do some vertical mixing, but the extent is much reduced. The jet ends up being largely in the form of a kind of horizontally oriented fan, spreading laterally but not downward (Figure 8-37).

5.3 Sediment Deposition in Deltas

5.3.1 Now that you have a good mental picture of the hydrodynamics of the delta environment, think about sediment deposition in the delta environment. Think in terms of the jet shown in Figure 8-36 or Figure 8-37, with the reservation that the depth of water in the water body is not infinitely deep—perhaps only several times the depth of flow in the approach channel (that is, the stream or river).

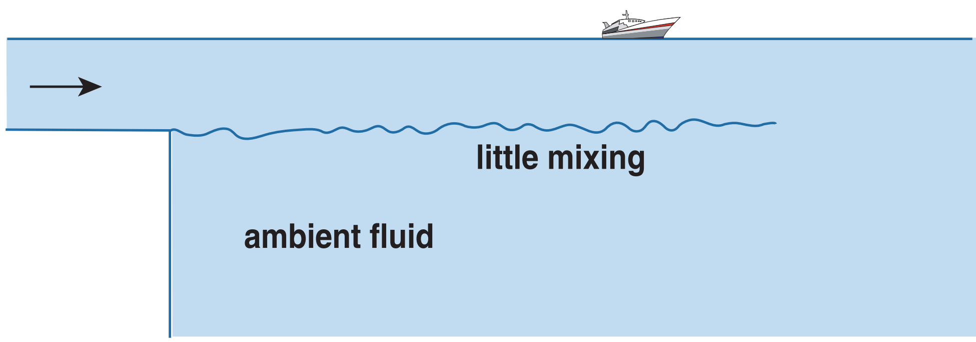

Figure by MIT OCW.

Figure 8-37. A jet formed where a fresh-water channel flow enters a standing body of saline water. Vertical section view, through the axis of the jet. (A plan view would look much like what is shown in Figure 35, except that the shape of the jet is not axially symmetric.)

5.3.2 Suppose that the stream is carrying both bed load and suspended load. You know what's going to happen: the bed load tends to be dropped out rapidly, at and near the entrance of the jet, because the sediment particles have relatively large settling velocities, but the finer part of the load is carried for greater distances away from the entrance, participating in the lateral mixing while it settles slowly to the bottom of the water body.

5.3.3 Think now specifically about the deposition of the bed load. At first, of course, it just makes a pile of sediment at base of the water body below the jet entrance. Eventually the pile builds up to reach the jet entrance. From then on, as the sediment comes to rest as the jet emerges into the water body, a wedge of coarse sediment builds forward into the water body, as sediment is deposited at the brink of a sediment body and slides down an angle-of-repose slope. Figure 8-38 shows three stages in the process, and Figure 8-39 shows a plan view.

5.3.4 Note, in particular, from Figure 8-38 that in the later stages the delta body consists of three rather distinct parts: ***topsets***, ***foresets***, and ***bottomsets***. The foresets we just dealt with. The bottomsets are readily understandable as well: as the foresets build forward, they bury earlier-deposited finer sediment, which fell from suspension at some distance from the jet entrance.

5.3.5 The topsets require a bit of further explanation. Keep in mind that the river flowing into the water body has some nonzero slope. (That's what keeps the water flowing, remember.) I purposely showed that slope in Figure 8-38. The outbuilding of the foresets has the effect of shifting the point of entrance of the jet in the direction out into the water body. That causes the downstream-most part of the river profile to rise at all points. That happens by sediment deposition. More specifically, a very small part of the passing bed load is extracted from the flow to come to permanent rest on the river bed, thereby building the profile of the river upward. These are the topsets of the delta.

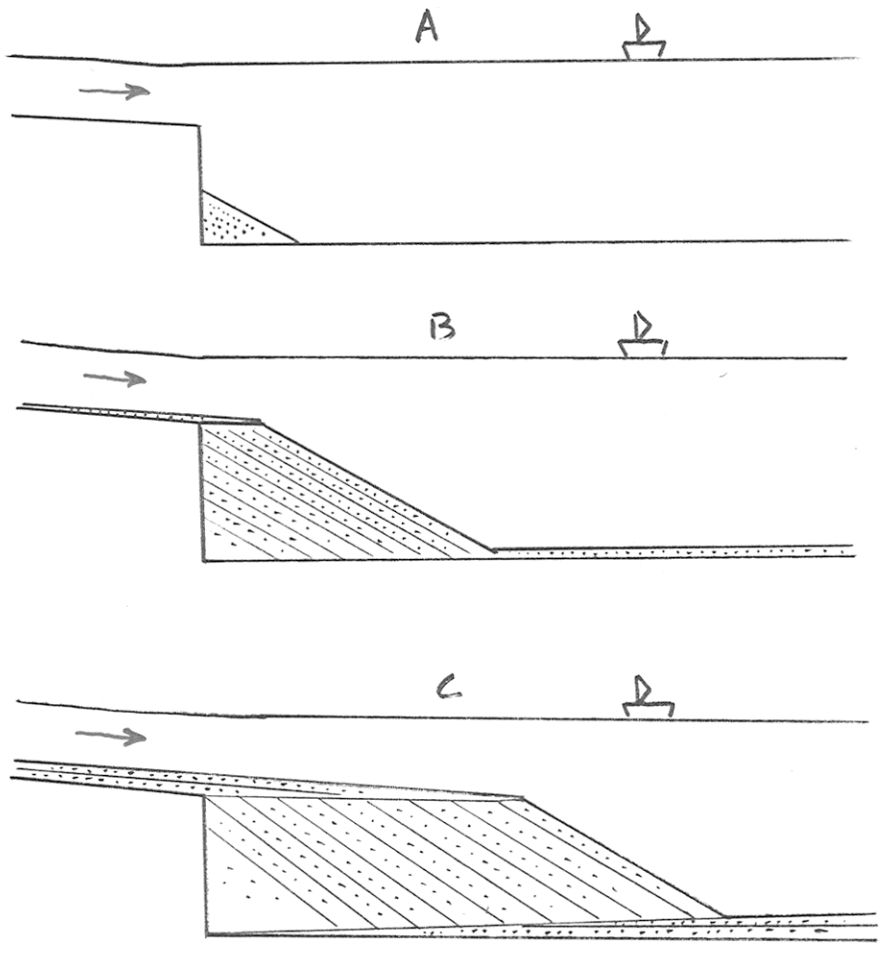

Figure 8-38. Building of a delta as a river enters a body of standing water. A) The earliest stage. B) A somewhat later stage. C) An even later stage, showing topsets, foresets, and bottomsets.

5.3.6 Finally, Figure 8-39 shows a plan view of the delta in Figure 8-38 after the delta has built out into the water body for an appreciable distance. The margin of the delta forms an arc, because sediment builds not just forward but also with a lateral component. Why? Basically because the tendency for deposition along the axis of the flow leads to a slight axial ridge, and then the flow tends to flow laterally off that ridge, down the slight sideways slopes of the delta body.

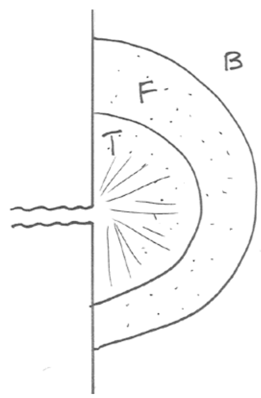

Figure 8-39. Plan view of the delta described in Figure 8-38. T, topsets; F, foresets; B, bottomsets.

5.4 Large Deltas in Nature

5.4.1 The delta described in the preceding section, and shown in Figures 8-38 and 8-39, is representative of the way deltas develop when a relatively small stream or river enters a water body in which currents and waves are minor. Such deltas are called ***Gilbert deltas***, in honor of an early geomorphologist, G.K. Gilbert, who made the first systematic studies of such small deltas in the western U.S. in the late 1800s.

5.4.2 At this point we need to address the various complicating factors that set in when the delta-building river is larger and the water body into which the delta is building is not as placid as was assumed in the preceding section. Figure 8-40 shows one of several classifications of deltas.

5.4.3 Let's deal first with the behavior of the river. The topset surface, which we'll call the ***delta plain***, is not a smooth surface. One reason for its non-smoothness, even if only sand and gravel bed load is being carried by the river, is that the delta plain is braided. (Go back to the chapter on rivers to review the nature of braiding in streams and rivers.)

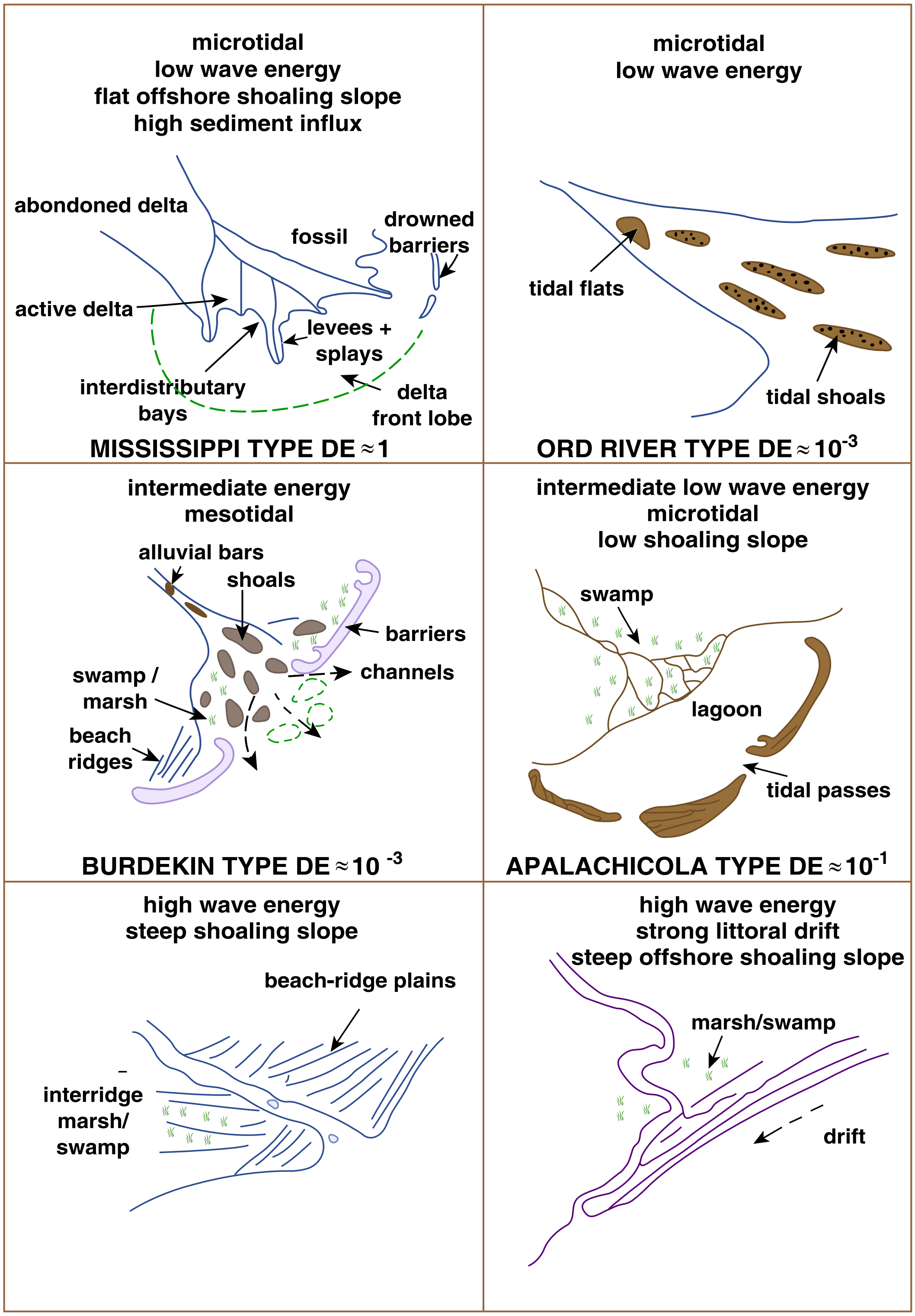

Figure by MIT OCW.

Figure 8-40. Classification of deltas. (From Carter, 1988.)

5.4.4 A more important reason, applicable to large mixed-load rivers, is that there is a strong tendency for upbuilding of the flow-carrying channel, deposition of natural levees, and occasional avulsion to relocate the channel to slightly lower areas on the aggrading delta plain. Typically there are two or more active channels, carrying water and sediment, operating on the delta plain at the same time. All the time, these shift around, avulse, and become abandoned, and new ones form. In that way, over time all of the delta plain is built up uniformly but with very complex internal structure in detail. The individual flow-carrying channels are called ***distributary channels***. The next time you look at a map of Louisiana, note how the most recent active area of the Mississippi delta is building out into the Gulf of Mexico in the form of a hand-like body with several active distributaries (locally called "passes").

5.4.5 Now we have to contend with the various effects that the water body itself has on the growing delta. We need to be concerned particularly with the action of waves, tides, and currents. Each of these factors is able to mold the morphology of the delta, in complex ways.

currents: The effects of throughgoing coast-parallel ocean currents are the easiest to understand. They tend to sweep the delivered sediment in one direction, causing slight to extreme asymmetry of the delta. Currents offshore of the mouth of the Amazon are so strong that even given the enormous sediment load of the river the delta does not protrude greatly into the ocean, the way the Mississippi delta does.

waves: ocean waves, approaching at a large angle to the coastline, tend to blunt the margin of the delta, giving it a fairly regular, arcuate shape. The Nile delta is the classic example.

tides: In regions with a large tidal range and strong tidal currents, deltas are shaped by strong back-and-forth flow in the various distributary channels, making for a strongly dissected delta body. The classic example is the Rhine delta, in the Netherlands.

The Mississippi delta is an excellent example of a delta that is dominated by the river itself rather than the water movements in the water body. That's why the delta is able to build outward in long salients. Such a delta is picturesquely called a ***bird's-foot delta***. (Look again at a map of the Mississippi delta to get the concept.)