PART II
GENERAL PHYSICAL OCEANOGRAPHY

Chapter 3. Tides and Tidal Streams

Early Knowledge

The rhythmic rise and fall of sea level, called the tide, has long been part of the seafarer's life. The early Greeks and Romans lived on the almost tideless Mediterranean Sea and learned of tides and their association with the sun and moon from voyages to the Atlantic Ocean. Pytheas of Marseille, who is reported to have circumnavigated Britain around 320 B.C., was one of the first to actually record the existence of tides and to note the close relationship between the time of high water and the transit of the moon. Julius Caesar also noted this relationship during a campaign to Britain, although his initial ignorance of tides caused his fleet to be wrecked on the beach during an invasion attempt just prior to a high tide. A somewhat similar fate befell Alexander the Great at the mouth of the Ganges River a few centuries earlier. Arab and Persian pilots who began to sail the coasts of India and China around the 9th century A.D knew of the twice daily rise and fall of sea level in certain river estuaries and the associated reversals in currents. Although the close association of the tides with the lunar phases is mentioned in Arab writings of the 9th century, one prominent scholar, Ibn al-Fakih (902 A.D.), linked tides in Canton Harbour to an angel that dips his finger in the China Sea so the sea rises and then ebbs on its removal, or to a whale that "inhales water causing ebb and exhales it causing flow."

In 1325 the Arab scholar, Al Dimiski of Damascus, published remarkably accurate tidal predictions for farmers irrigating their fields near the mouth of Shatt al Arab, the river formed by the confluence of the Tigris and Euphrates rivers, that flows into the northern Persian Gulf. He mentions for example that there were two high and two low tides each day, that for each there was slightly less than 1 h lag of both the ebb and the flood from the day before, and that the period of ebb in the river estuary persisted longer than the period of flood. Similar records from the 16th century indicate that certain individuals in England made their livelihood predicting tides in areas of the coast.

It was not until Sir Isaac Newton published his *Equilibrium Theory of the Tides* in 1687 that a scientific understanding was begun. Even today, the study of tides is far from complete despite a good understanding of their cause. Scientists are only beginning to sort out the effects of tides in the deep ocean and complicated coastal regions like those of British Columbia and Washington. There are even times when tidal theorists seem to revert to medieval thinking. A recent proposal for instance suggested that the tides exist because "the oceans initially came from the moon and are trying to return."

Datum

It has been universally agreed that the reference level from which to measure the height of the tides is the chart datum, the level to which low water can be expected to fall during normal tides (Fig. 3.1). Datum is used as the reference level for depths plotted on nautical charts. It has been chosen in such a way that it is low enough that few tides will fall below it, but not so low that a chart will show less depth than the mariner will usually find. Negative values in the tide tables, indicating unusually low tides, serve as a warning that the water level will fall below datum. One of the main purposes of tide gage installations is to obtain measurements long enough to accurately establish the datum level.

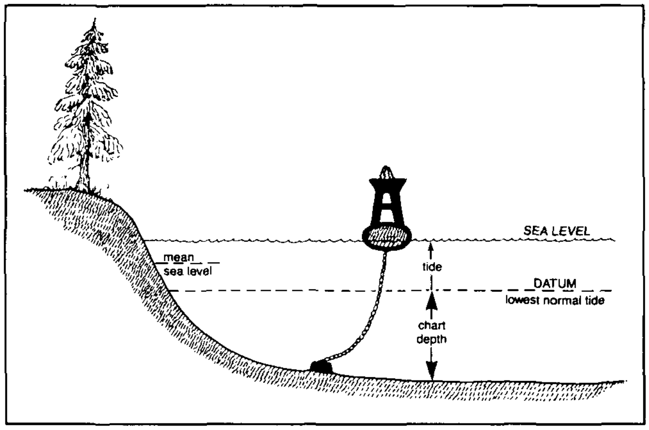

FIG. 3.1. Low water datum. Tides are measured upwards from this level.

A word of caution to those who use either American or Canadian charts — the two countries differ in their interpretation of "lower low water." Where Canadians define chart datum as the level of water at the "lowest normal tides," west coast Americans define it as "mean lower low water." Because averaged values are used, U.S. datums are slightly higher than Canadian datum. Soundings on American charts would, therefore, indicate a greater depth of water than a Canadian chart of the same area. In other words, Canadian hydrographers are more conservative when informing the mariner on the minimum depth of water he can normally expect at a particular location. No difficulty should arise if the appropriate set of tide tables is used; U.S. tables for U.S. charts and Canadian tables for Canadian charts. The main difference between the two is that U.S. tables tend to show negative tides more often than Canadian. The actual water depth in each case is found by adding the tide height for the particular area for the specified time to the chart datum; that is, actual depth = chart datum + tide height, as per the tide tables. It should be noted that datum is not the same as mean sea level. Datum is determined solely by the lowest tides whereas mean sea level is determined from the average of all tides, including high waters. Consequently, the datum level changes from place to place along the coast because of the changing nature of the tides. Datum elevations at Victoria and Vancouver, for example, are different because their low waters do not normally fall to the same levels.

Measurement of Tides

To determine the datum level and daily fluctuation in water depth at a particular location, measurements must be made. This is commonly done by tide gages — flotation devices and recorders placed within a boxlike container called a stilling well. The unit is affixed at a known height above the bottom to a permanent structure such as a wharf (Fig. 3.2a). A narrow opening near the submerged base of the well allows outside communication with the ocean and at the same time effectively damps-out rapid, unwanted fluctuations in sea level produced by surface waves. Gradual changes in water elevation due to the tide cause the float to move a pen across scaled chart paper that is slowly advanced by a clock mechanism. The resulting ink trace or "tidal" curve then gives the height of the water surface as a function of time (Fig. 3.2b). On the basis of such measurements the datum level is established and tide elevations subsequently determined.

Tidal records obtained from coastal locations deemed to be of major importance to navigation (reference ports) are analyzed by the Canadian Hydrographic Service and used to make the tidal predictions published in *Canadian Tide and Current Tables*. Predictions for these ports are generally based on a year or more of continuous observation, to ensure accurate times and heights of high and low water. Secondary ports are coastal locations of lesser navigational importance. Predicted tides for these ports are found by applying the appropriate time and height corrections given in the tide tables to the reference port predictions. These corrections are simply the measured differences in time and range for the tide between the two locations. In contrast to reference ports, tidal records at secondary ports typically only span periods of about 1 mo.

Though tide gages have been the traditional method of measurement around the world, they suffer from a variety of faults. They are susceptible to erroneous readings caused by swell waves and to partial plugging of the intake by debris and living organisms like starfish. Supporting structures are also expensive and often in danger of wave damage, despite the fact that gages are comparatively cheap. Lastly, installation of these gages is limited to shallow water adjacent to the coast. For these reasons, much more sophisticated instrumentation known as pressure gages has recently been developed. These are generally self-contained units that can be set on the seafloor at appreciable depths to record on tape the total bottom pressure of the combined weight of the air and water above. The principal component in many gages is a vibrating wire or crystal whose oscillation frequency varies with changes in the ambient pressure. Using an assumed density of sea water of 1.025 g/cm^3, this variation in frequency is then converted to a change in pressure and, hence, to a change in depth after allowance is made for any accompanying temperature alteration. The advantages of pressure gages are numerous: they can be placed at offshore locations, their records receive negligible contamination from wind waves, they are easy to handle, and they have comparatively good accuracy. Measurement resolution of these gages is typically around 1 part in 100,000, or about 10 times better than their accuracy, so a

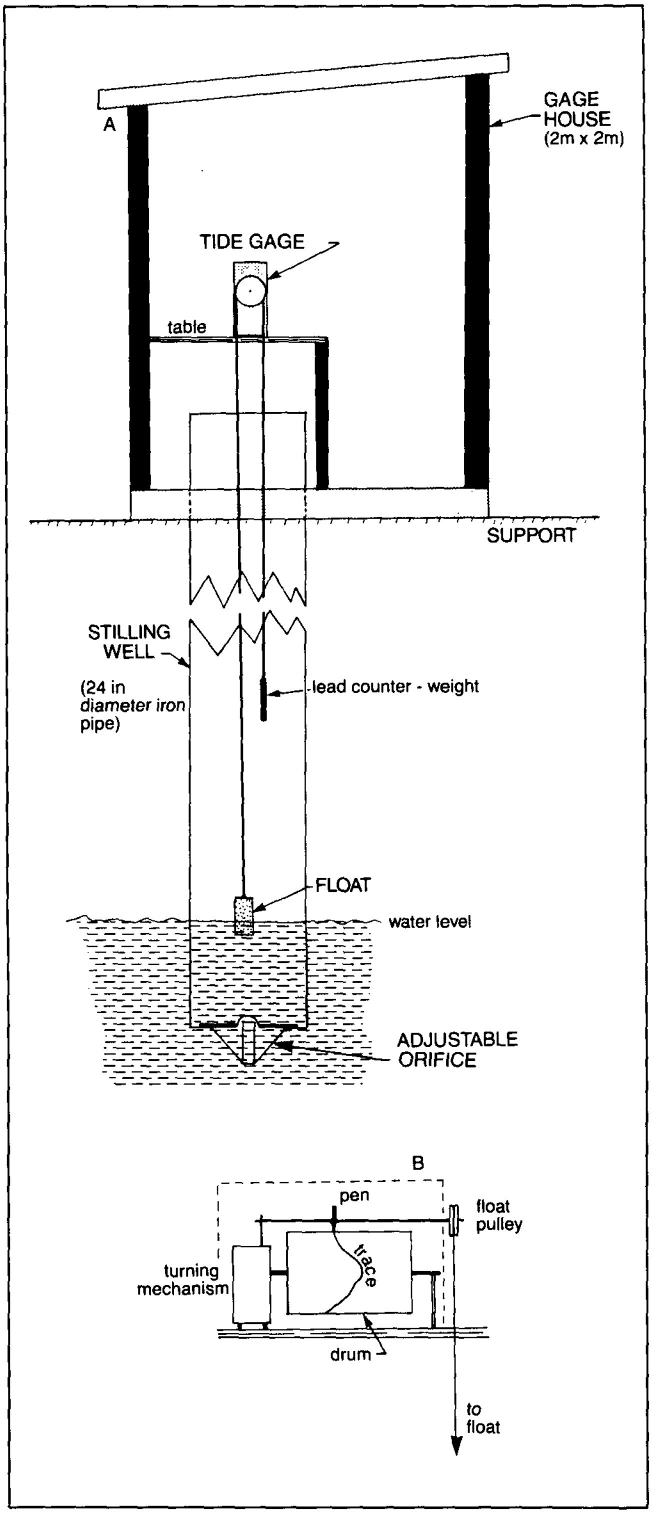

Fig. 3.2. (A) Stilling-well type tide gage. The oil-filled float adjusts with changing water level, mechanically moves pen (B) to trace a tidal curve on graph paper affixed to a drum rotating at a known rate. (Courtesy R. Brown)

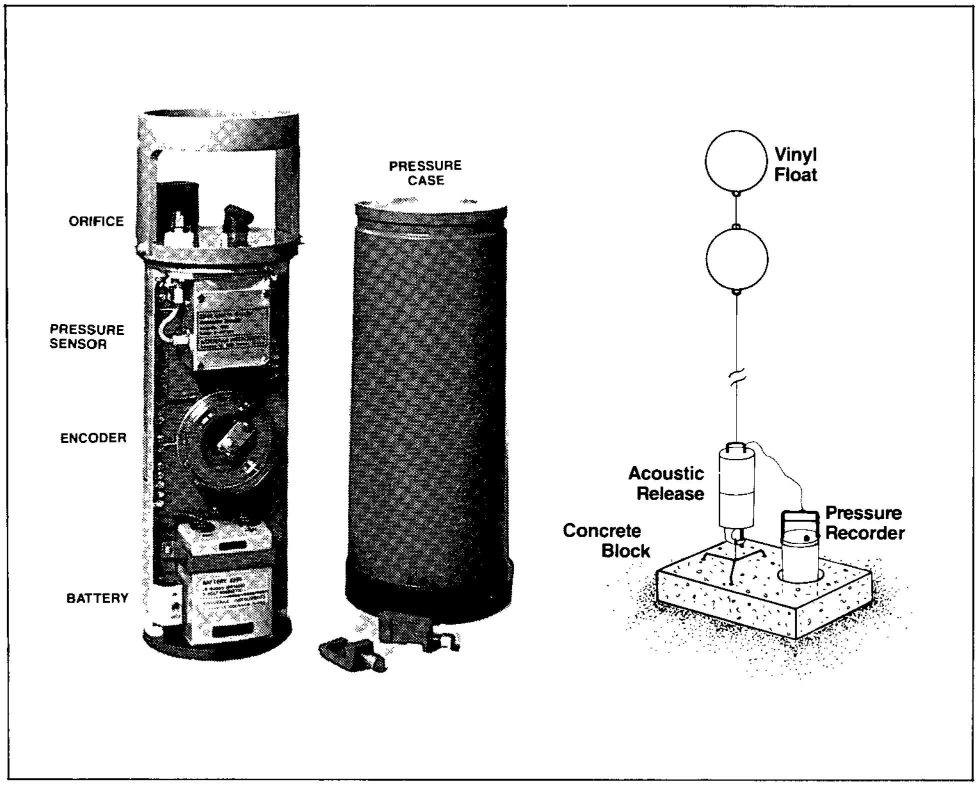

FIG. 3.3. Underwater pressure recorder and mooring arrangement. Water pressure at orifice is recorded internally and stored on magnetic tape. Recovery is effected by activating acoustic release via a surface transmitter; floats then lift recorder from concrete block to surface. (Courtesy Aanderaa Instruments Ltd., Victoria)

gage situated in 1000 m of water can detect changes in depth as small as 1 cm. The main disadvantages of these gages are their susceptibility to undesired dynamic pressure effects due to currents that flow past the orifice, pressure-to-depth conversion errors caused by departures of seawater density from the assumed value and the fact that data are not normally available in real time. As a consequence of the latter restriction, pressure gages are not employed in tsunami warning networks. Despite these limitations, pressure gages are essential to investigations of deep-sea tides and the offshore propagation of tsunamic waves. A modern pressure gage installation is shown in Fig. 3.3.

Nature of Tides

The tides nearly repeat themselves once every 24 h and 50 min. This is the lunar day and the time it takes a point on the earth to rotate back to the same position relative to the moon during each revolution. Therefore, the daily rhythm of the tidal cycle is governed by the lunar day and not by the solar day of 24 h, which paces the daily cycle of life on earth. Consequently, times of high and low water are roughly 50 min later from one solar day to the next. If the sun were the major cause of the tides they would repeat themselves nearly every 24 h, and high and low waters would occur at the same time each standard day. Henceforth, when tides are discussed "day" will usually refer to a lunar day.

The difference in depth between high and low water is called the range of the tide. Oceanic tidal ranges vary from typical low values of 10 cm or so in the Mediterranean Sea and Arctic Ocean to more than 15 m in the Bay of Fundy. Along the British Columbia–Washington coast, the range is commonly between 3 and 5 m, with greater ranges during June and December and smaller ranges during March and September.

To categorize the daily tide, the number of cycles (number of highs and lows) in a lunar day are counted. When the tide has 1 cycle (1 high and 1 low) per day it is said to be diurnal (Fig. 3.4a); when it has 2 cycles per day of nearly equal heights it is said to be semidiurnal (Fig. 3.4b). As the inserts in the two figures illustrate, tides on the west coast are purely diurnal or semidiurnal for only a few days each month. Most of the time they are a mixture of diurnal and semidiurnal tides and are said to be mixed (Fig. 3.4c). The tides at Victoria Harbour and Sooke, for example, are classified as mixed, predominantly diurnal, whereas tides at Seattle, Vancouver, Tofino, and Prince Rupert are mixed, predominantly semidiurnal. (For some examples of different types of mixed tides see Fig. 3.5.)

A property of all mixed tides on the west coast is the

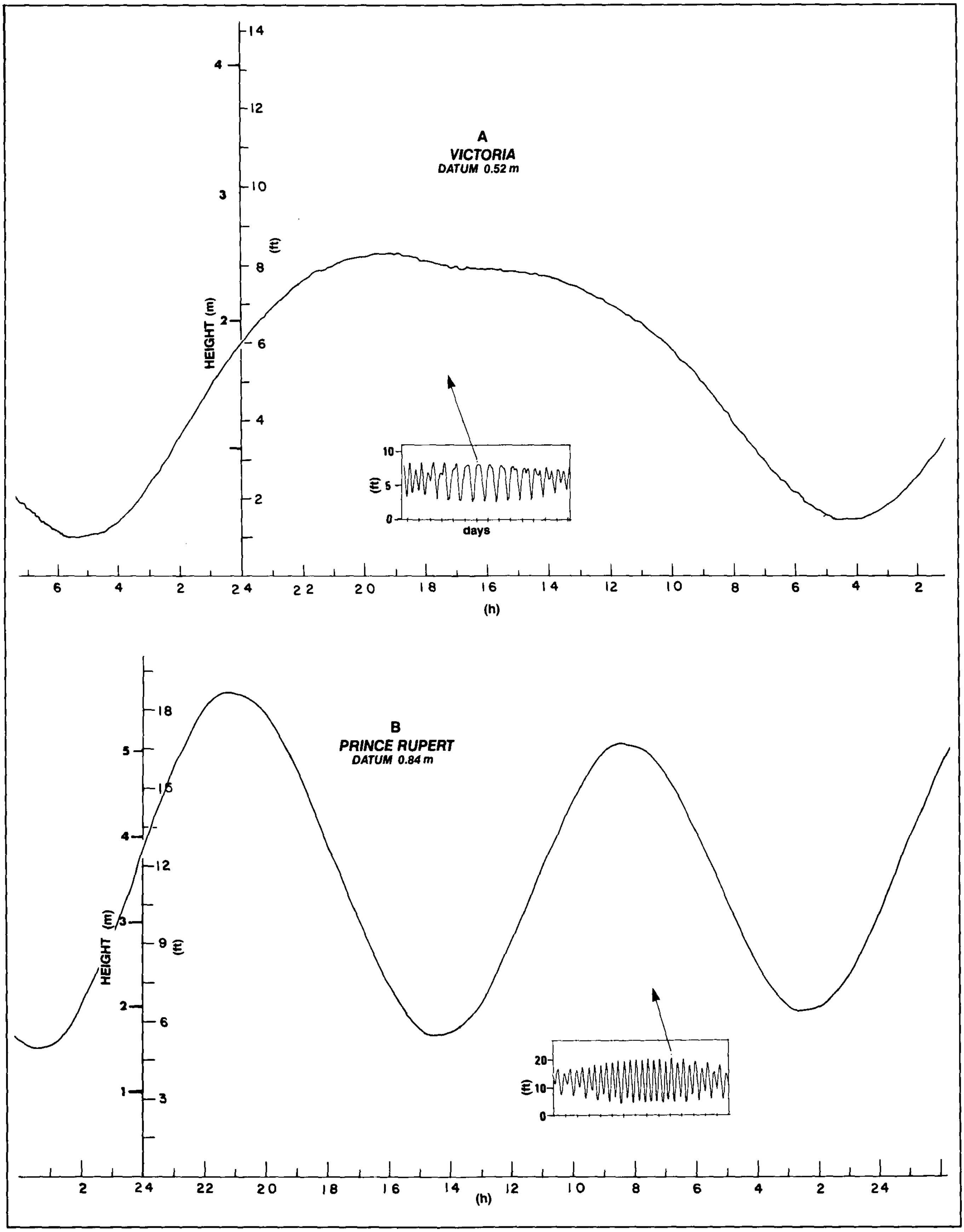

FIG. 3.4. Variations in water level over a 1-day period at three locations on British Columbia Coast. (A) diurnal tide, Victoria; (B) semidiurnal tide, Prince Rupert.

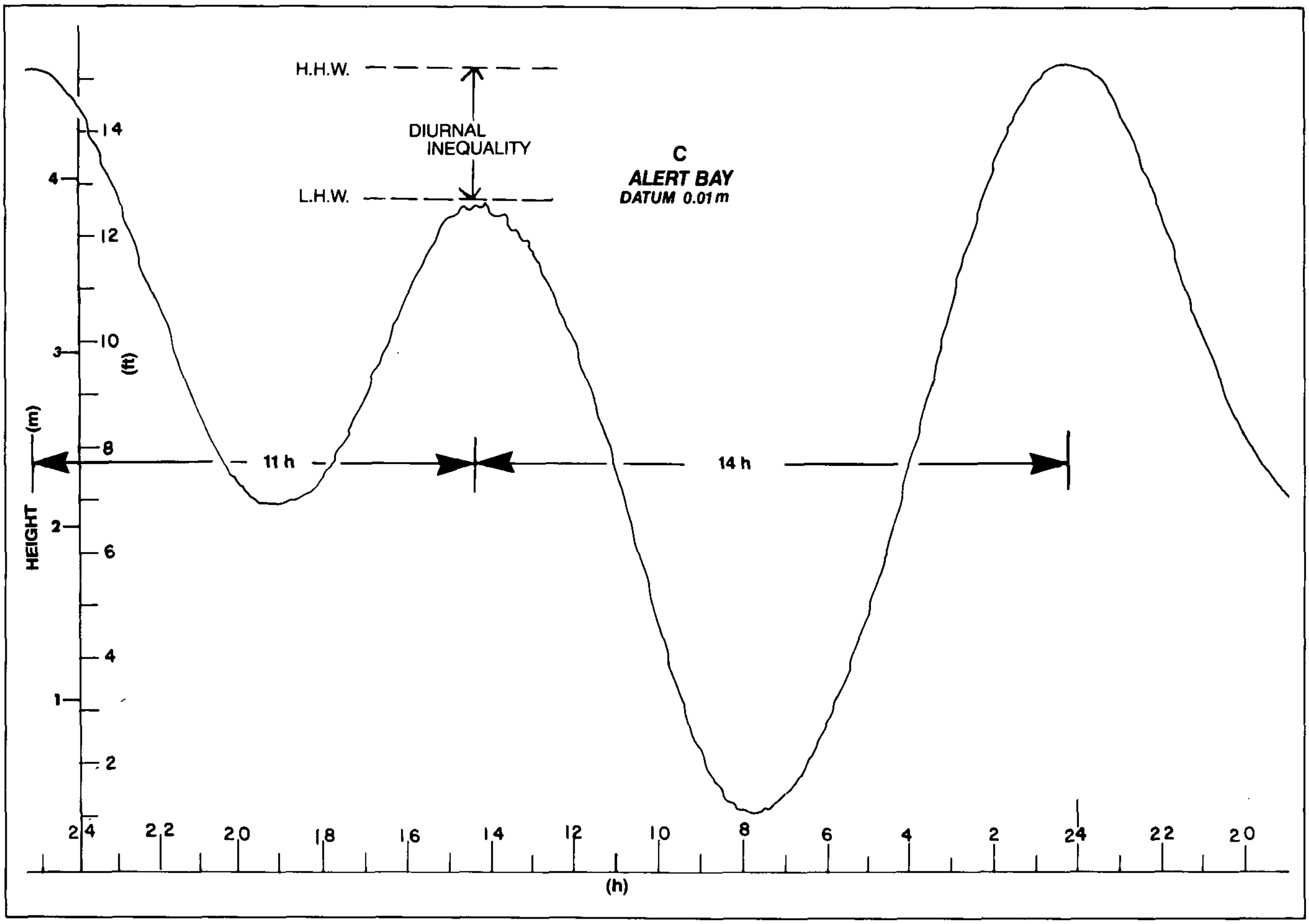

FIG. 3.4. (C) Mixed tide, Alert Bay. Time progresses from right to left in each diagram. Insets show evolution of tides at each location over 2 wk.

diurnal inequality, expressed in Fig. 3.4c as the difference in height between the higher high-water level and the lower high-water level. There is also a diurnal inequality between lower low water and higher low water. In fact, a characteristic feature of the inequality in tidal elevations within the basins that separate Vancouver Island from the mainland is that the difference in height between successive low waters is usually greater than the height difference between successive high waters. Also, whenever there is an inequality in the height between successive high tides there is a daily inequality in the time intervals between pairs of high tides, as shown by the intervals of 11 and 14 h (Fig. 3.4c). The same is true of successive low waters.

In addition to the differences in tidal range during any particular day, the range of the tide changes progressively from one day to the next with a cyclic period of around 2 wk (Fig. 3.5). Generally speaking, high tide becomes continually higher and low tide continually lower for about 7 days when the roles reverse, high tides become lower and low tides become higher for about the next 7 days. It is during such cyclic variations in range that comparatively high spring tides and comparatively low neap tides occur. These tides vary with the phases of the moon; spring tides occur near the times of a full or new moon and neap tides near the time of the moon's quarters.

The 2-wk cycle in the tidal range on the outer coast is closely linked to the moon's phases (Fig. 3.5). However, this obviously does not apply to tides within the protected waters of the southwest coast. (These differences will be discussed later.) Figure 3.5 further reveals an approximate 2-wk cycle in the magnitude of the diurnal inequality for both high and low tides. This inequality is always most pronounced when the moon is furthest north or south of the earth's equator (tropic tides), but is almost nonexistent when the moon is directly above the equator (equatorial tides). Within the Strait of Georgia, Puget Sound, and the eastern end of Juan de Fuca Strait, the diurnal inequality becomes so great that for a few days each month the tropic tides are essentially diurnal. At Sooke and Victoria, diurnal tides produced this way may persist for as long as 5 days. Along the outer coast, the diurnal inequality is smaller and the tides are never diurnal. Equatorial tides, on the other hand, are invariably semidiurnal throughout the entire world ocean. The latter is one of the few definitive statements that can be made regarding the tide without any exception or qualification.

The cyclic 14-day variation in the diurnal inequality is accompanied by a corresponding variation in the time intervals between successive high waters and between successive low waters. These time intervals alternate from equal, when the diurnal inequality is least (equatorial tides), to extreme values of 8½ and 16¼ h, when the

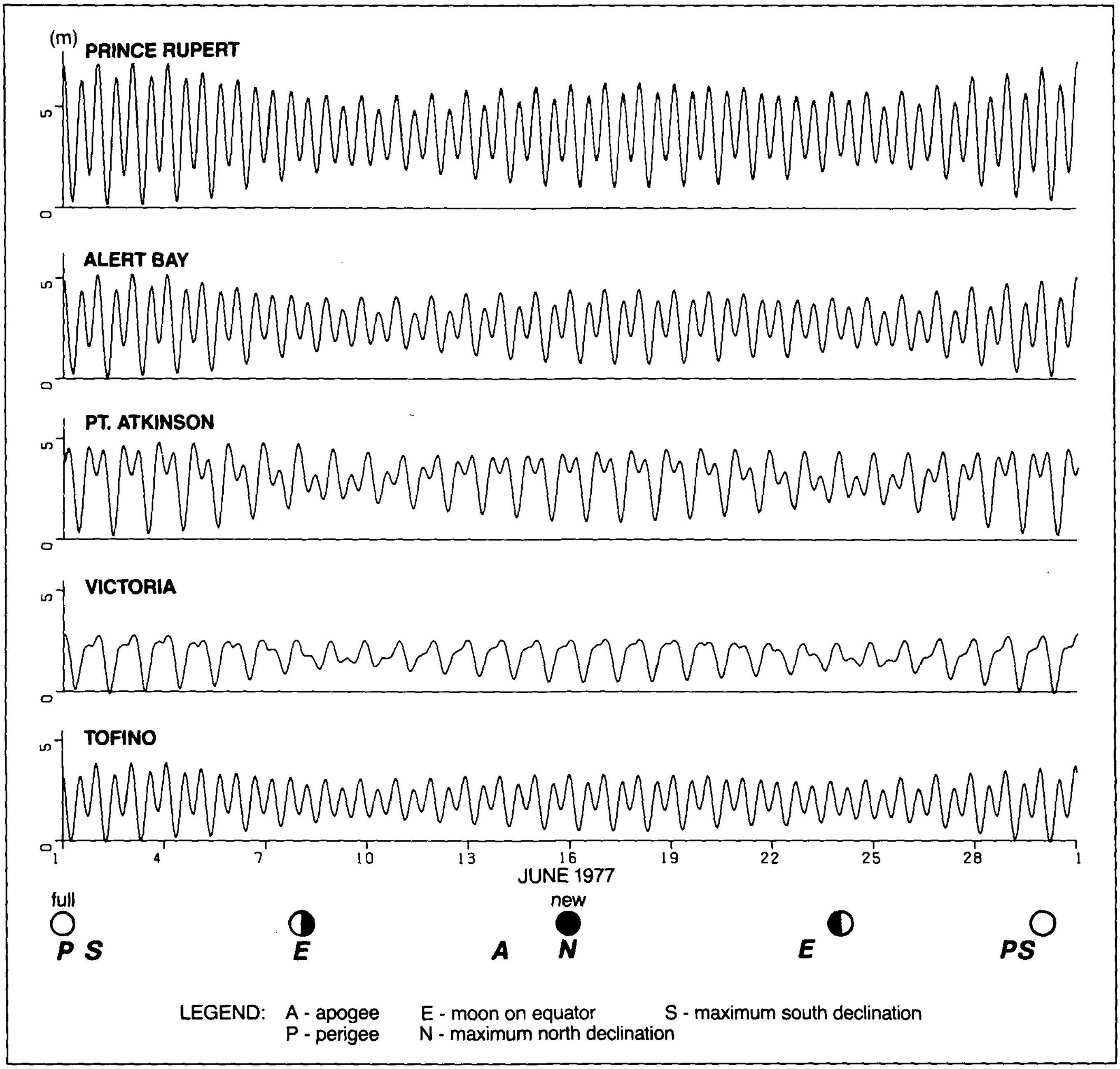

Fig. 3.5. Sea level variations over 1 mo at five locations on British Columbia coast (June 1977). Heights in metres. (Courtesy Canadian Hydrographic Service and A. Douglas)

diurnal inequality is greatest (tropic tides). In every configuration, of course, the combined time interval between the successive tides (the first interval plus the second) must always add up to the lunar day of just over 24¾ h. Thus, if the reader were to clock a 16½-h delay between two successive high tides, he would only need to wait another 8½ h until the next high tide for a total of 24¾ h.

When tidal curves like those in Fig. 3.5 are examined more closely, it can be seen that the tides are not repeated exactly every 14 days. Stages of the daily tide with comparatively large tidal ranges (or comparatively low tidal ranges) spaced 2 wk apart are not of identical magnitudes. This suggests that factors affect the tides whose cyclic variability is of longer duration than those that produce the approximate 14-day variation. If Fig. 3.5 were extended to cover an entire season, the tidal pattern would be more closely repeated every 29½ days than every 14 days. Therefore, some factor, or factors, influencing the rhythmic rise and fall of the sea must have a period of 29½ days. Even over a month, however, the tide is not repeated exactly, so the tidal record must be extended. If a full year's record were available and tides measured in January of one year were compared with those of the following January, would the tides match up exactly? Nearly, but not well enough for some purposes. Tide heights must be measured continuously for almost 19 yr before the tide would begin to repeat itself to an accuracy of a few centimetres. Although there are even longer period variations in the tidal range (see section Long-Period Tides), the whole exercise would, if extended, become somewhat academic as nontidal and nonperiodic fluctuations in sea level begin to mask the increasingly smaller differences in the measured tides. Nevertheless, the values listed in the tide tables take into consideration as many possible contributions as practical, in order to give accurate height predictions.

Equilibrium Theory of Tides

So far only the form that daily tidal variations can take has been examined; as was the situation before Newton's time. There is a set of observations but no theory to explain them. How are tides formed? Why are they almost always semidiurnal or mixed? And why does the range of the tide vary on such a regular basis about every 2 wk?

To obtain rudimentary answers to these basic questions it should be appreciated that the tide is really a combined or integrated response to a variety of natural phenomena. The single most important factor is the combined gravitational attraction of the moon and sun on the earth. Because of its greater proximity to the earth, the moon's gravitational attraction is twice as important as the sun's (Appendix D). Except for one important case, therefore, the influence of the sun can be disregarded when the main features of the tide are examined. As with Newton's Equilibrium Theory of the tides, moreover, matters are simplified by assuming that the earth is uniformly covered with water, with no continents or submarine mountains to interfere with the oceanic motions.

Because many simplifying assumptions have already been made, it is acceptable to go one step further and also ignore the presence of the moon. For the moment, imagine that the water-covered planet is drifting alone through a moonless, sunless universe! If the earth were not rotating on its own axis, its shape would then be a perfect sphere held together by its own gravity. Distances measured from its center to anywhere on the sea surface would be equal. In reality, of course, the earth spins on its axis once every 24 h and has subsequently been deformed into an ellipsoid with a diameter 42 km greater through the equator than the poles. There is a corresponding distortion of the ocean, so mean sea level decreases slightly from the equator toward either pole. An observer on earth would not notice any daily change in sea level because this distortion is uniform around a parallel of latitude. The forces produced by the earth's axial rotation, therefore, cannot generate tidelike variations in water level and, for the sake of simplicity, a nonrotating earth with a fixed orientation relative to the stars will be considered. Now the moon is returned to the picture.

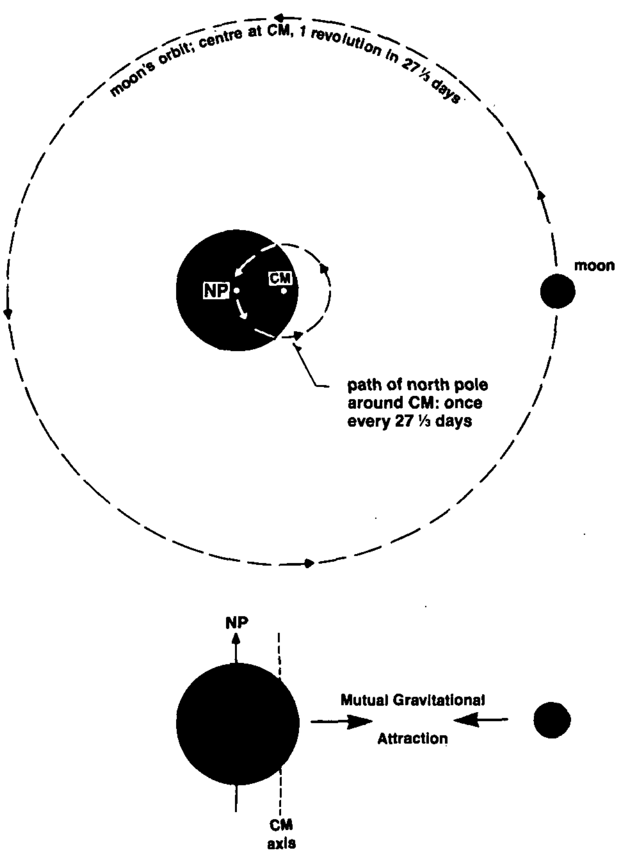

FIG. 3.6. Earth–moon system. *Top*: paths of moon and north pole (NP) around center of mass (CM) for combined system. *Bottom*: side view of center-of-mass axis and earth's axis of rotation when moon is at zero declination.

The earth and moon constitute a celestial unit held in a binding partnership by their mutual gravitational attraction. It is this gravitational force that disrupts the natural tendency for each of these bodies to move in a straight line through the heavens, and constrains their individual centers to travel in nearly circular orbits around a common center of mass once every 27⅓ days (Fig. 3.6). As the earth's mass is 82 times greater than the moon's, the center of mass, like the balance point of a teeter-totter, is shifted toward the earth and actually lies beneath the surface at a constant depth of roughly 1700 km. (The position of this imaginary point is not fixed at one particular spot for all times, but slowly circles inside the planet over a month, always keeping the same depth.) The rotation of the earth–moon system about its common center of mass causes each particle of the earth to trace out a circular path over a period of a month. Moreover, Fig. 3.7 shows the orbital radius for every particle is the same, regardless of distance from the center of mass. This has important consequences, for it means that the force associated with these circular orbits is of equal magnitude and direction for each bit of the earth. What is this force? A physicist would call it the centripetal force, the total force necessary to make each particle of the earth travel in its own particular circle in the presence of the moon's gravitational field. This centripetal force is uniform over the entire earth and is directed everywhere toward the moon at right angles to the axis of rotation drawn through the centre of mass. (To account for the tidal forces, the rotation of the earth about its own axis need not be considered). Unlike the centripetal force, however, the ever-present lunar gravitational attraction on a particle of the earth depends on that particle's distance and orientation measured relative to the center of the moon (Fig. 3.8). Only at the exact center of the earth are the moon's gravitational pull and the centripetal force identical in both strength and direction. At

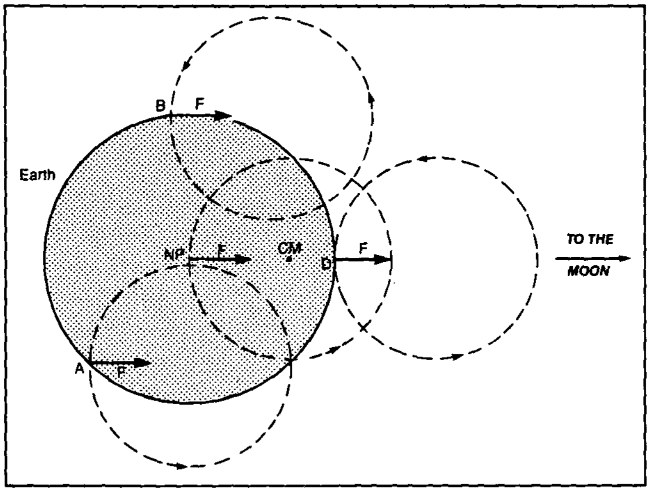

FIG. 3.7. Ignoring spin of earth about its axis of rotation, each segment of earth follows a circular path as it moves about center of mass (CM) for combined earth–moon system. Vectors (arrows) represent local centripetal force (F), which is the same over entire globe and equal to force of mutual attraction between centers of earth and moon. (see Fig. 3.6)

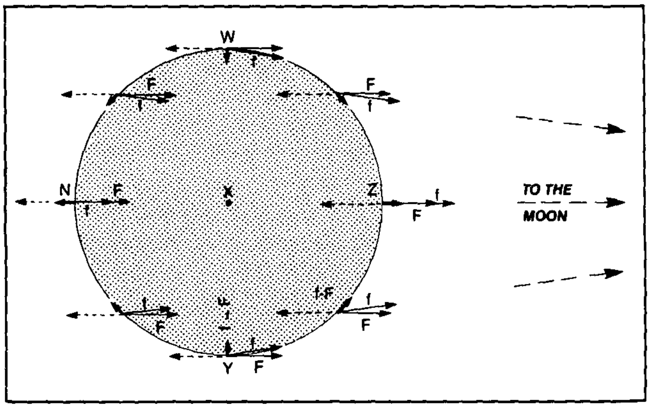

FIG. 3.8. Vector subtraction of centripetal force (F) from local attractive force (f) of moon on an element of water. Resultant vector (short stubby arrow) is tide-generating force, (f-F). Same result is obtained by vectorially adding centrifugal force (broken arrow) to f. Tide-generating force draws waters into bulges and hollows until balanced by pressure gradient associated with increasing sea surface slope. Low tides at W and Y; high tides at N (nadir) and Z (zenith). Compare to Fig. 3.9.

all other locations, there is a small difference between the two forces. It is this slight difference that causes the tide-generating force on the earth.

Before proceeding any further, a scientifically incorrect, though conceptually useful, approach will be taken to the ideas introduced so far. Instead of arguing on the basis of the inward centripetal force, each orbiting particle on the earth can be considered as experiencing an outward centrifugal force as the earth–moon system rotates around its center of mass. (In reality, there is no such force because to keep an object in a circular orbit requires an inward force not an outward one. What is commonly called the centrifugal force is nothing more than the "reluctance" of moving objects to travel in curved trajectories.) Turning a deaf ear to the critics, it is found that, because the centrifugal force originates from the same motions that produce the centripetal force, it will also be uniform in both magnitude and direction over the entire earth but will be directed away from the moon. The centrifugal force is then simply the reverse of the centripetal force (Fig. 3.8). Again the distance-dependent gravitational pull of the moon will only equal the centrifugal force at the exact center of the earth; away from the center the two opposing forces will be unbalanced and result in a net tide-generating force. The distribution of this force around a meridian of longitude is in Fig. 3.9a. It should be clear from this diagram that subtracting the moon's gravitational force from the centripetal force yields the same net force at each point as adding the moon's gravitational force to the outward centrifugal force. Similar, but weaker, tidal forces are also generated by the earth–sun pair.

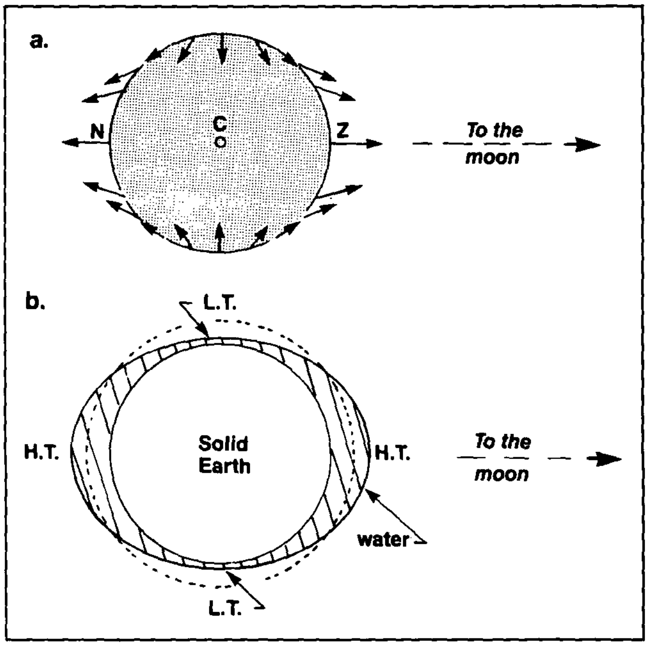

FIG. 3.9. *Top*: Simplified version of Fig. 3.8 shows distribution of tide-generating force along a meridian of longitude. ***Bottom***: Tidal bulges and hollows created by tide-generating force (highly exaggerated in vertical). Broken line corresponds to mean sea level or shape of ocean in absence of tidal forces. (H.T. = high tide; L.T. = low tide)

The lunar tide-generating force, the imbalance between the centrifugal (or centripetal) force and the moon's gravitational attraction, affects the formation of ocean tides in two basic ways: on the side of the earth that faces the moon, the moon's gravitational pull is slightly stronger than the opposing centrifugal push, so the water bulges toward the moon; on the side that faces away from the moon, the centrifugal force exceeds the moon's attraction so the water bulges away from the moon (Fig. 3.9b). To put it more concisely, the water is drawn away from the earth on one side and the earth away from the water on the other. The two bulges are associated with high tides. Midway between them, sea level is lowered below normal to create a low-tide hollow that girdles the earth. Maximum high tides, therefore, occur along a line joining the center of the earth to the center of the moon, whereas low tides are distributed along an earth-encircling swath at right angles to that line. To indicate the strength of the tide-generating force, note that at points N and Z in Fig. 3.9a, where the force is upward, a person weighing 90 kg (200 lbs) loses only 10 mg (0.00035 oz), or roughly the weight of a single tear.

Relative to the sun, the earth spins on its axis once every 24 h. Because the bulges and hollows maintain a fixed orientation relative to the moon, each will "almost" travel around the earth once each solar day. "Almost," because in 24 h the position of the moon will have shifted slightly in the sky relative to its position 24 h earlier. However, in 24 h 50 min the moon will be back to its original position as seen by an observer situated at a particular location on the earth (disregarding for the moment the daily change in the moon's elevation above the horizon). This explains why the tides occur 50 min later every solar day. The fact that there are two bulges and two hollows circling the earth explains the occurrence of semidiurnal tides with two high and two low tides per lunar day.

Other features of the tides can be readily accounted for by the equilibrium theory. But, before proceeding, it is important to appreciate the limitations of this concept.

The moving bulges and hollows can be considered a traveling or progressive wave, with a large crest-to-crest separation of approximately 22,000 km at the equator and a small amplitude of a metre or so. This is a true tidal wave and should not be confused with a tsunami, a Japanese word for the destructive waves commonly generated by earthquakes and often misnamed "tidal waves" (see Chapter 9). In reality, the tidal wave doesn't really keep pace with the moon as implied, but lags behind, sometimes by hours. Some reasons are: the water's inertia makes it impossible for the ocean to respond rapidly enough to the moon's movement; friction slows the wave as it rubs along the bottom; continents obstruct the wave's passage and force it to take many complicated pathways; and the ocean isn't deep enough. Unrestricted propagation of the tidal wave, in fact, would require the wave to move at speeds up to 1650 km/h (the rotational speed of a point on the earth's equator) and the ocean to have a depth of 22 km, whereas the actual depth rarely exceeds 5 km. Moreover, according to the equilibrium theory, maximum tidal heights should only be 0.8 m at equatorial latitudes and decrease toward the poles. Because actual tides are known to greatly exceed such values over most of the world's oceans, the theory is clearly much too simplistic to explain the detailed behavior of a complex phenomenon like the tide. A more sophisticated account of the tides was presented a century after Newton by the French mathematician Pierre Simon Marquis de Laplace. Called the *Dynamical Theory of the Tides,* it takes into consideration the effects of the limited depth and extent of ocean basins, as well as the influence of the earth's rotation and friction. Other scientists since then have further advanced understanding of the tides, though it is still far from complete.

Types of Tides

The notion of an equilibrium tide is a useful concept to account for the fundamental nature of tidal fluctuations. For example, the diurnal inequality and one reason for the biweekly tidal cycle are explained if the effect of the moon's declination is considered (its angle north or south of the earth's equatorial plane that originates from the average 23.5° tilt of the earth's axis with respect to the plane of the moon's orbit). Suppose the moon is at its maximum north or south declination (Fig. 3.10a, c). As usual, there will be two high tides a day, but because the bulges are not symmetric to the earth's axis of rotation, one bulge will create a higher tide than the other along a given parallel of latitude. (Only at the equator will the high tides be equal.) This difference is the diurnal inequality. As mentioned earlier, it is greatest at the time of tropic tides, when the moon has its largest declination. In Fig. 3.10b on the other hand, there will be no inequality in the two daily tides because the tidal bulges and hollows are now symmetric with respect to the earth's axis. These are the semidiurnal equatorial tides. Notice also that the range of the tide will be above normal during the tropic tides and below normal during equatorial tides.

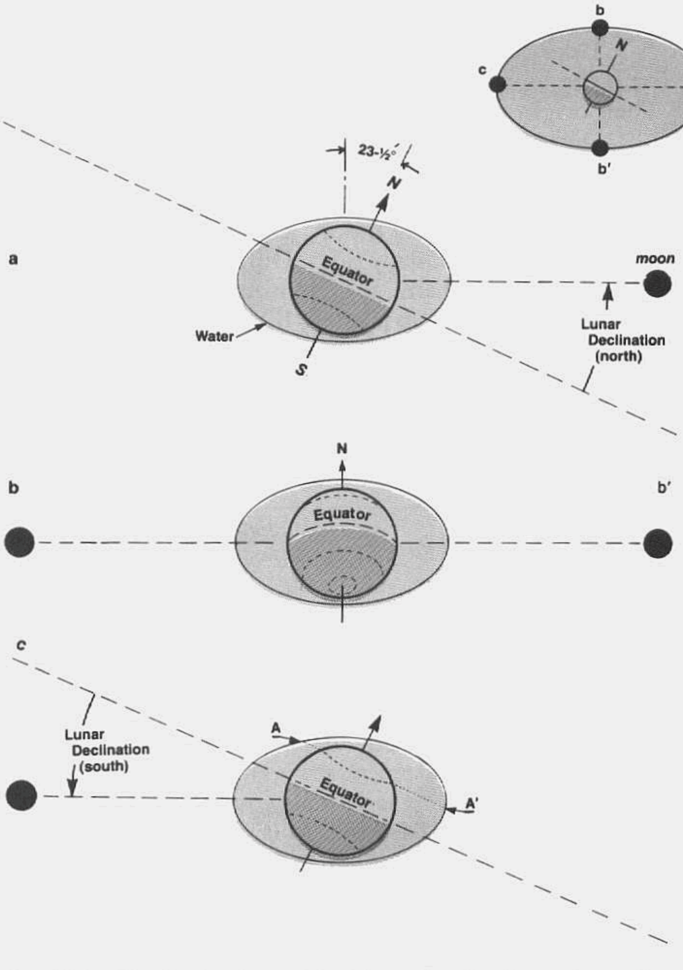

FIG. 3.10. Declinational-type tides. Except for a few days each month, tidal bulges are asymmetrical with respect to earth's NS-axis of rotation due to average 23½° tilt to moon's orbital plane (see inset, upper right). Because earth is revolving under fixed bulges, tides observed at latitudinal point, A', will be higher than at the same point, A, ½ (lunar) day earlier; difference in tidal heights is diurnal inequality. Only when moon is over equator (b and b') are tidal distortions symmetrical to earth's axis. Note: configurations a and c are viewed from point b' in inset; configurations b and b' are viewed from point c. There are roughly 15 days between a and c.

It takes the moon about 7 days to reach position b from position a in Fig. 3.10 (decreasing diurnal inequality) and another 7 days to reach position c (increasing diurnal inequality), causing a 2-wk cycle in the tidal range. Variations produced in this manner are called declinational-type tides and are an important aspect of coastal tidal patterns, particularly in Puget Sound and the eastern end of Juan de Fuca Strait. In the Strait of Georgia, declinational effects are strongest in the south and decrease northward.

There is also an important quasi-biweekly cycle in the tidal range produced by the sun's tidal pull which has been ignored until now. Its presence explains a feature of the tides that has been known for thousands of years: tides tend to have a greater range near a full or new moon than near the moon's quarters. Those with the large range are called spring tides, although they have nothing to do with the spring of the year; the term has a Saxon origin meaning "greater activity." The smaller range are neap tides, from a Saxon word meaning "inactive." Spring tides are about 20% greater than average tidal ranges, whereas neap tides are about 20% lower.

Tides of this nature, whose range varies in accordance with the lunar phases, are known as synodic-type tides. Their origin is easily explained.

During a full or new moon the sun is in line or in conjunction with the moon and earth (Fig. 3.11). The solar-generated tidal bulges and hollows then reinforce those produced by the moon, leading to a greater than normal range in the daily tides. During the moon's quarters, on the other hand, the sun is in opposition, and the tidal bulges it generates partially fill the hollows created by the moon. There is an accompanying decrease in the height of the lunar bulges by the solar-generated hollows and a subsequent reduction in the tidal range below normal. As there are approximately 15 days between a new moon and a full moon, there is a periodic 15-day variation in tidal range. Note that the period for synodic-type tides is roughly 1 day longer than that of the declinational type tides. Put another way, the synodic month is about 2 days longer than the usual sidereal month of 27⅓ days. This difference, though only one reason for the complex behavior of the tides, has an analogous origin to the difference of 50 min between the solar and lunar day. Suppose the moon and sun are in conjunction (Fig. 3.12). During the 27⅓ days it takes the moon to complete one orbit of the earth, the planet will have traveled 1/12 of its orbit around the sun. As a result, an extra 2 days are required for the sun and moon to again come into conjunction.

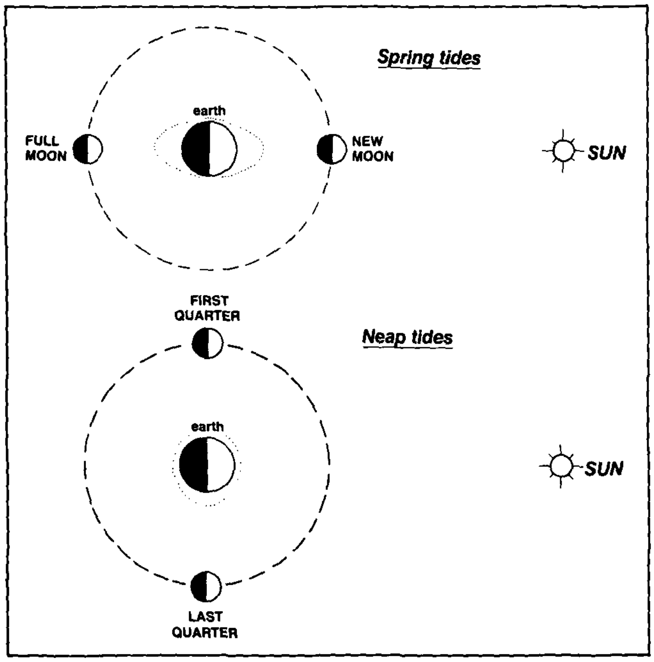

FIG. 3.11. Synodic-type tides. Plan view of relative alignment of sun, moon, and earth during spring and neap tides.

Comparatively short cyclic variations in the ranges of coastal tides also arise because of the eccentricity of the moon's orbit around the earth (Fig. 3.13). These, too, may be explained by equilibrium theory. In this case, the lunar tide-generating force causes tidal bulges and hollows to be more pronounced during the moon's closest approach to the earth at perigee and less pronounced when the moon is furthest away at apogee. Known as anomalistic-type tides, these variations in tidal range have a period of 27½ days. In protected inshore waters their importance almost equals that of the synodic tides, whereas on the outer coast they apparently only cause a 7-mo alteration in the range of the spring and neap tides.

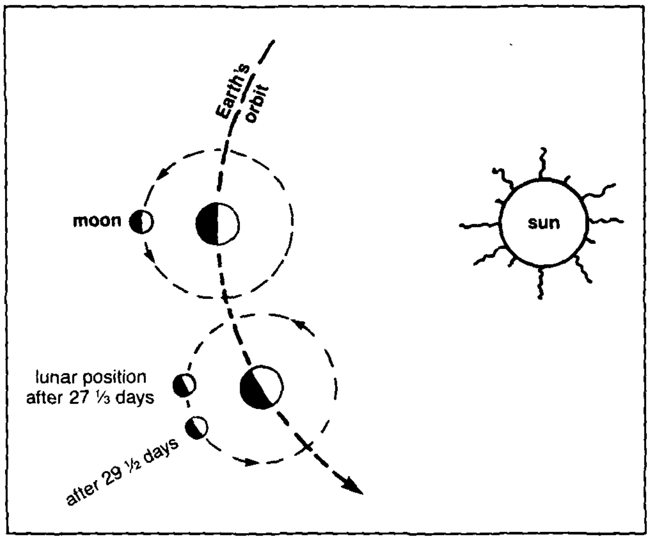

FIG. 3.12. Sidereal month (27⅓ days) versus synodic month (29½ days).

Figure 3.13 also shows that the eccentricity in the moon's orbit can lead to an inequality in the times between full to new moon and new to full moon, though the total time must equal one synodic month of 29½ days. This adds a further complication to an already complicated 2-wk tidal variation. Consider the situation when the sun is opposite point P so that the moon is new at that point. It will then be nearly full at point A; and vice versa. In either situation the length of time from new to full moon will equal that from full back to new moon, or exactly 14 days, 18 h, 22 min. But now suppose the sun is opposite point C. The moon will then be new at C and full at D, with quarters at P and A. Because the moon's orbital

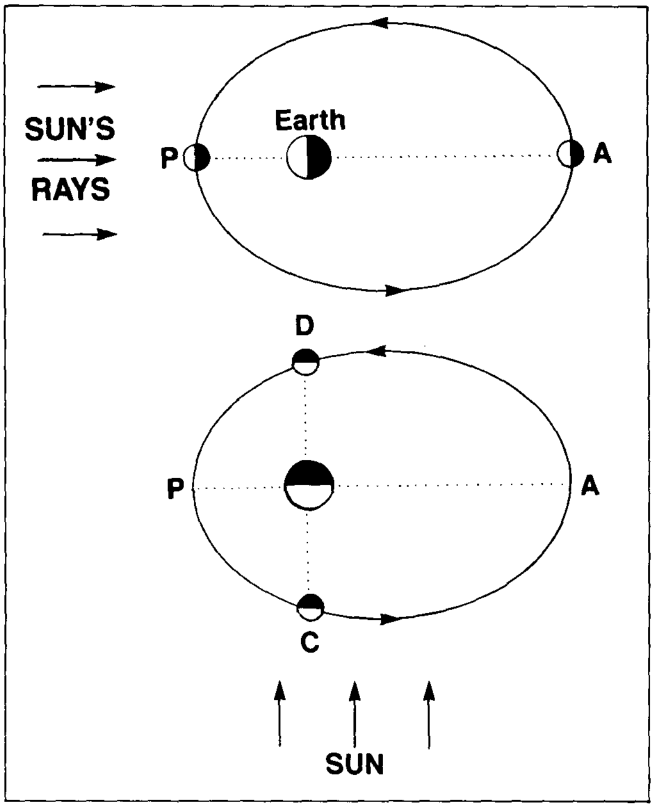

FIG. 3.13. Anomalistic-type tides. Moon's slightly elliptical orbit around earth, exaggerated for illustrative purposes, carries it closest at perigee (P) and farthest away at apogee (A).

speed is slowest near apogee (A) and greatest near perigee (P), and because the distance to A from C is greater than that from P to C, the time interval from new moon to full moon (C to D) will be longer than that from full moon back to new moon (D to C). To be precise, the intervals of time are 15 days, 14 h, 12 min; and 13 days, 22 h, 32 min respectively, the kind of extreme variability expected in the repeatability of the synodic-type tides associated with the phases of the moon.

Some concepts discussed can be applied to account for the tides in Fig. 3.5. For instance, maximum diurnal inequalities occur every time the moon has its maximum declination north (N) or south (S) of the equator while smallest inequalities occur when the moon is at the equator (E). These conditions are shown schematically in Fig. 3.10a–c, respectively. Tides along the outer coast are strongly synodic, attain greatest spring ranges immediately following the full and new moons, and smallest neap ranges follow the moon's quarters. The slight delay is due to the ever-present lag in the ocean's response to the tide-generating forces. In Fig. 3.5, one spring tide takes place when the moon is at perigee (P) and is subsequently enhanced; the other occurs near the moon's apogee and is diminished.

Within the protected coastal waters, the synodic tide is decidely less pronounced than along the outer coast, as indicated by the comparatively small range at the time of the new moon. The declinational tide, however, is more pronounced because small ranges occur just after the moon has crossed the equator (E) (Fig. 3.5). These are further augmented by the anomalistic tide. As a consequence, the greater range takes place near the moon's perigee and the lesser range near apogee. Examples from other months or locations would of course have somewhat different tidal modifications.

Long-Period Tides

In addition to the daily, biweekly, and monthly cycles in the tidal range, there are small cyclic modulations that take place over longer periods of time. Twice a year, for example, at the summer and winter solstices (June 21 and Dec. 22) the sun attains maximum declination from the equator (Fig. 3.14). When the sun and moon are in conjunction near this time, the spring tides along the British Columbia coast are at their greatest. During the fall and spring equinoxes, on the other hand, when the sun is on the equator (Sept. 22 and Mar. 21) and the moon at quarters, neap tides are at their lowest. This semiannual variation in the declination of the sun produces noticeable effects in waters adjacent to the Strait of Georgia where, as typified by Vancouver, lowest tides of the year occur near midnight just before Christmas and around noon near the end of June.

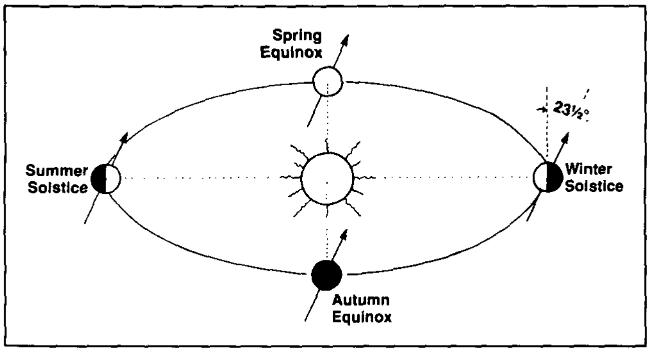

FIG. 3.14. Solar declination due to 23½°-tilt of earth's axis to orbital plane.

There is a small annual tidal range variation associated with the earth's slightly elliptical orbit around the sun and an average 14-mo cycle in the tidal range that amounts to about ½ cm (called the pole tide), caused by a wobble of the earth's axis of rotation known as the Chandler wobble. Winter–summer differences in mean atmospheric pressure and mean seawater density at the coast cause annual fluctuations in sea level of about 10–20 cm. (Effects of this type are called inverse barometer and steric effects, respectively.) A major long-period variation is caused by the 5° tilt of the moon's orbital plane with respect to the earth's (Fig. 3.15). The line of intersection formed where the plane of the moon's orbit crosses the plane of the earth's orbit slowly rotates, and requires 18.6 yr to complete one revolution. (Theoretically then, it should be possible to use 19-yr-old tide tables to accurately predict the tides today, though in practise things turn out to be more complex.) Once during each "19-yr cycle" the moon's orbital tilt adds to the 23.5° tilt of the earth's axis to permit maximum lunar declinations of 28.5° (= 23.5

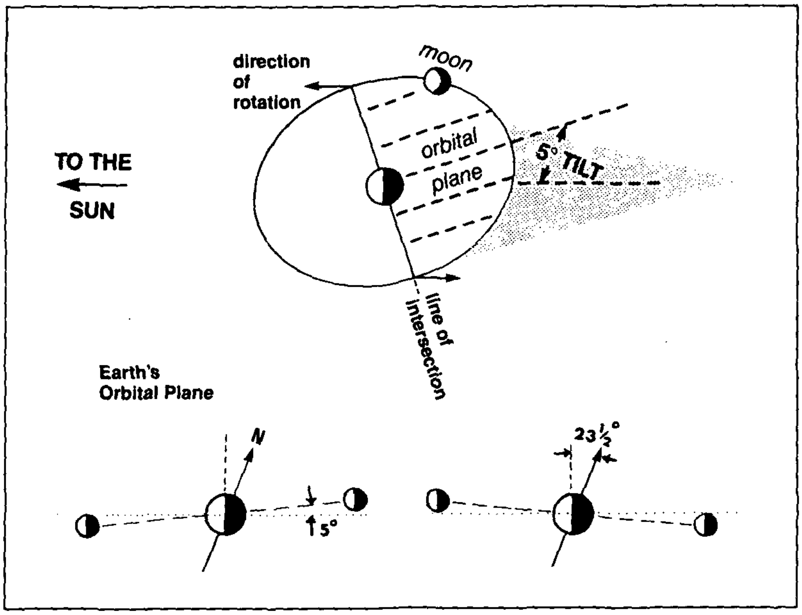

FIG. 3.15. The 5°-tilt of moon's orbital plane to plane of earth's orbit about sun. Line of intersection slowly rotates with center of earth as pivot point. Each complete wobble of lunar orbital plane takes 18.6 yr. Lower diagrams show how 5°-tilt increases maximum possible lunar declination to 28½° (right side) and 9.3 yr later decreases it to 18½° (left side).

+ 5°). At such times (the most recent 1969), the centerline of the moon's tide-generating force varies 57° north–south over the surface of the earth each month, so that diurnal inequalities attain their greatest values. After 9.3 yr, the lunar declination has decreased to only 18.5° (= 23.5–5°), monthly north–south variations are reduced to 37°, and diurnal inequalities are minimal.

Other long-period changes in the tidal range include an 8.8-yr cycle associated with alterations in the eccentricity of the moon's orbit about the earth and a 20,940-yr cycle due to a wobble in the earth's orbit about the sun. Over geological time scales, large, but irregular, variations in sea level are produced by the waxing and waning of the ice ages. In addition, prior to the fragmentation of Pangea into separate continents (Chapter 1), the supercontinent stood low with respect to sea level, and thereby allowed wide-spread inundation by the sea. Following breakup, the individual continents rose in relation to sea level and the degree of inundation decreased.

Tidal Constituents

The precise analysis of tides requires that observed records of sea level be systematically decomposed into their various constituents, each constituent or component with a specified amplitude and a specific cyclic period or frequency. Analysis of a tidal record is analogous to the decomposition of a sound wave or musical note into various harmonic frequencies. Along the Canadian west coast there are four main constituents which, when added together, account for almost all the variation in tidal range during a month. Foremost is the principal lunar semidiurnal constituent (or M_2 tide, where 2 is the number of cycles per day) associated with the moon's gravitational attraction. Except near Victoria and a few other special locations, this contribution generally accounts for around 50% of the tidal range. Next is the principal mixed diurnal constituent (or K_1 tide) with 1 cycle per lunar day which, like all diurnal effects, originates through the declination of the moon and, or, the sun. It accounts for roughly 25% of the tidal range and is largely responsible for the diurnal inequality (Fig. 3.16). Near Victoria, where it creates diurnal tides, this constituent becomes more important

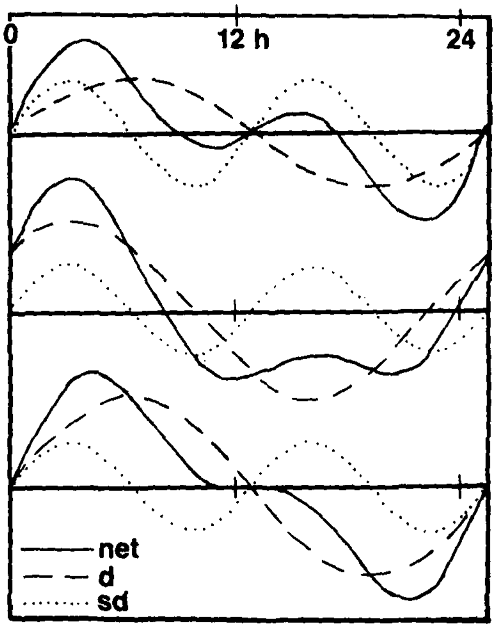

FIG. 3.16. Mixed-type tides (solid lines) result from addition of a diurnal tide (d) and a semidiurnal tide (sd). Height and diurnal inequality of net tide is different in each example due to differences in diurnal contribution relative to fixed semidiurnal contribution.

TABLE 3.1. Magnitudes (m) of the four main tide height constituents at various locations along the British Columbia–Washington coast. Each value is based upon 1 yr of tide height measurements except at Union Seamount where record was 110 days. Ratio determines type of tide. (MSD = mixed, predominantly semidiurnal; MD = mixed, predominantly diurnal). (Source: Canadian Hydrographic Survey)

Location	Constituent				Ratio	Type of Tide
	O_1	K_1	M_2	S_2	$(O_1 + K_1) \div (M_2 + S_2)$	
Union Seamount 49°35′ 132°47′	0.079	0.131	0.277	0.087	0.720	MSD
Tofino 48°09′ 125°55′	0.246	0.389	0.991	0.280	0.500	MSD
Port Renfrew 48°33′ 124°25′	0.287	0.458	0.712	0.210	0.809	MSD
Victoria 48°25′ 123°22′	0.370	0.627	0.373	0.102	2.100	MD
Port Townsend 48°08′ 122°46′	0.437	0.616	0.680	0.190	1.210	MSD
Seattle 47°36′ 122°20′	0.459	0.837	1.066	0.263	0.975	MSD
Sidney 48°39′ 123°24′	0.445	0.766	0.555	0.132	1.763	MD
Point Atkinson 49°20′ 123°15′	0.477	0.858	0.917	0.233	1.161	MSD
Comox 49°40′ 124°56′	0.489	0.885	1.002	0.253	1.095	MSD
Campbell River 50°01′ 125°14′	0.485	0.846	0.826	0.203	1.293	MSD
Alert Bay 50°35′ 126°56′	0.306	0.516	1.272	0.406	0.490	MSD
Queen Charlotte City 53°15′ 132°04′	0.315	0.511	1.975	0.651	0.315	MSD
Prince Rupert 54°19′ 130°20′	0.314	0.513	1.957	0.644	0.318	MSD

than the M_2 contribution. The next two major contributors to the tidal range are the principal solar semidiurnal constituent (S_2 tide), associated with the sun's gravitational attraction, and the diurnal (O_1) tide, linked to the moon's declination alone. These produce 15–20% of the tidal range, with about 5–10% caused by other more minor effects such as the P_1 tidal constituent due to the sun's declination alone, and the N_2 tidal constituent produced by the moon's varying distance from the earth during a month.

The ratio of magnitudes given by $(K_1 + O_1) \div (M_2 + S_2)$ determines the type of tide. When this ratio is less than 0.25 the tide is classified as semidiurnal; if it lies between 0.25 and 1.50 the tide is mixed, predominantly semidiurnal; between 1.50 and 3.0 the tide is mixed, predominantly diurnal; and finally, if the ratio exceeds 3.0 the tide is diurnal. Some values are presented in Table 3.1.

Coastal Tides

The previous explanations provide at best a bare framework for understanding the complexities of tides. Not only do continents block the free passage of the tidal wave, but variations in ocean depth alter direction and speed of propagation, and bottom friction hinders movement. In addition, the Coriolis force associated with the earth's rotation (see below) deflects tidal motions and forces them to move with the continents to the right of their direction of travel in the northern hemisphere. Consequently, the tidal wave always moves northward along the outer coasts of North America but southward along the coast of Asia. At the coasts, and within the protected coastal waters, the shape of the basin into which the tidal wave propagates plays a primary role in determining the response of sea level to the tidal wave. Each basin behaves differently. The classic example of how the shape of a basin affects the incoming ocean tide is the Bay of Fundy where a 3-m tidal range at the mouth of the bay can be amplified to a 18-m tidal range at the head of the bay.

Despite the complications, it is possible to present a general picture of tidal wave propagation into the coastal area of British Columbia and Washington. Before doing so, effects of the earth's rotation on the tide will be examined.

The Coriolis Force

Fluids such as air or water when moving freely over the solid crust of the earth's surface experience a slight deflection to the right of their direction of travel in the northern hemisphere and to the left in the southern hemisphere. Over short distances, these deflections are often too small to be noticeable, but over large enough distances they can accumulate to produce an appreciable "curving away" from the original path.

This effect is linked to the fact that motion is measured relative to a rotating frame of reference, the earth, which is rotating in space; if the earth were not spinning the effect would not exist.

If the rotating earth is compared to a merry-go-round and one person tries to throw a ball to another person on the opposite side of the carousel, Coriolis force is partially explained. When the ball is thrown, it naturally travels in a straight line (no knuckle balls allowed), yet it never reaches the receiver on the opposite side. Why not? Obviously, because the receiver has moved relative to where he was when the ball was thrown. But suppose neither player could see beyond the confines of the carousel and both were unaware they were traveling in circles, just as we are unaware that the earth is rotating. It would appear to them that the ball had taken a curved path, as if acted upon by some mysterious force! Of course, there isn't really a force, there only seems to be. Nevertheless, the effect is real enough to the players, as it is to observers of large-scale motions on the earth (Fig. 3.17). This apparent force that deflects freely moving objects away from what was thought to have been a straight line is the Coriolis force (named after the French mathematician, Gaspard Coriolis, who first described it in 1835). It has a strong influence in determining the prevailing wind patterns of the atmosphere and on the circulation of world oceans. Long-distance flying craft, such as jets and rockets, must continually adjust for it to reach their destinations. An automobile, however, does not respond to the Coriolis force because it is held at all times to the earth's surface by the friction between the tires and the road.

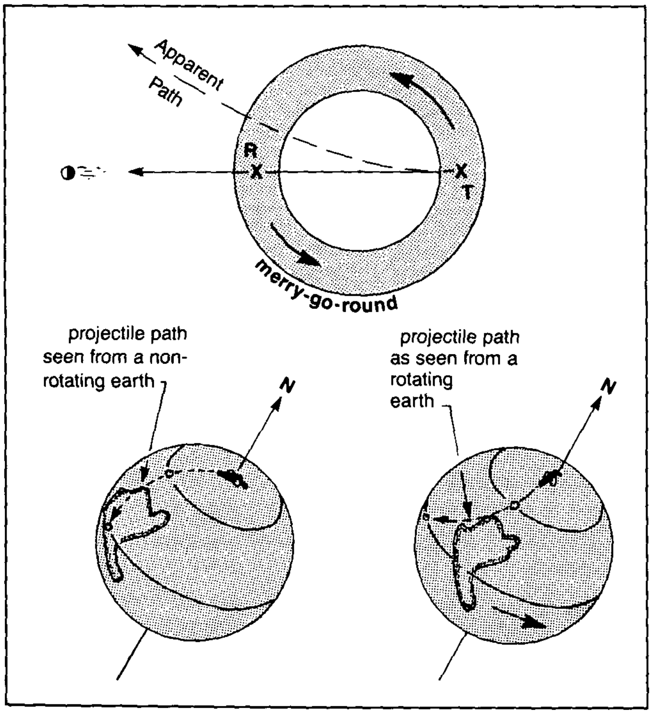

FIG. 3.17. The Coriolis force. *Top*: ball thrown in a straight line from T toward R on a merry-go-round appears to be deflected to right; i.e. to experience a force. *Bottom*: effect of earth's rotation on cannon ball fired from north pole toward equator (as seen from the ground).

Due to the deflection caused by the Coriolis force, the progressive tidal wave propagates northward along the west coast of North America "leaning" up against the coast (Fig. 3.18). This results in a seaward diminishing slope in the accompanying sea level and a corresponding decrease in the tidal range toward mid-ocean. Generally speaking, this slope amounts to an average decrease in the high- and low-tide levels by about 50 cm in 500 km, so that roughly 2500 km southwest of Vancouver Island the tidal range diminishes to a minimum. It then begins to increase again in the direction of Asia.

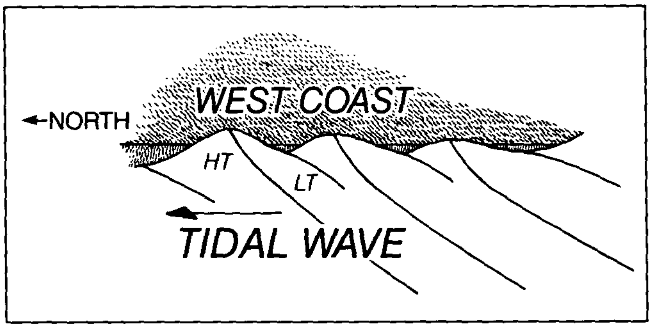

FIG. 3.18. Effect of Coriolis force on tide in Northeast Pacific Ocean. Tidal wave necessarily propagates northward, leaning against coast where the maximum tidal range takes place.

In the northeast corner of the Pacific Ocean, the speed of the tidal wave along the outer coast is typically around 740 km/h (400 kn), except over the continental shelf where it is somewhat slower due to the shallow depths. As a consequence of its great speed, there is little delay in a particular stage of the tide from one region to another and it is characterized by the following coastal features (Fig. 3.19) (modified after Dohler 1964).

From Barkley Sound to Cape Scott on the west coast of Vancouver Island, the tide occurs almost simultaneously and has an average range of about 3 m. As it moves into the numerous inlets in this region, there is a slight increase in range but no slowing down, except in the constricted passage at Quatsino Narrows where the tide is delayed by 45 min compared to the coast.

Along the west coast of the Queen Charlotte Islands and the mainland shores of Queen Charlotte Sound, the tide occurs simultaneously, but 30 min later than at Vancouver Island. The range of the tide on the mainland side of the Sound is increased even more as the tide moves into the various deep inlets, reaching 5 m at the heads. As the tide propagates northward from Queen Charlotte Sound into Hecate Strait, its range increases with the decreasing depth of the passage. This manifests itself as a south-to-north increase in the range and a time difference of the tide along the mainland side of Hecate Strait. A similar situation occurs as the tide enters Skidegate Channel that separates Graham and Moresby Island, and where a tidal range of 4.3 m at large tides at the coastal entrance becomes a range of 7.8 m at Queen Charlotte City two-thirds down the channel.

Around the northeastern end of Graham Island, the tide propagates more slowly and reaches Masset Inlet about 1 h later than Hecate Strait. At Prince Rupert, the tide arrives 1 h later than off Vancouver Island and has a mean range of 4.9 m.

It takes between 2 and 4 h for the tidal wave to sweep up Juan de Fuca Strait to the Haro Strait entrance. Along the way, the range decreases from the Cape Flattery entrance to Victoria, where the average range is around 2 m. As the tide rounds the corner between Victoria and the San Juan Islands it is partially deflected to the east, and produces slightly higher tides on the U.S. side than the Victoria side. There is then a delay of about 1 h, as the tide squeezes through the narrow channels of Haro and Rosario straits and attempts to negotiate the constricted

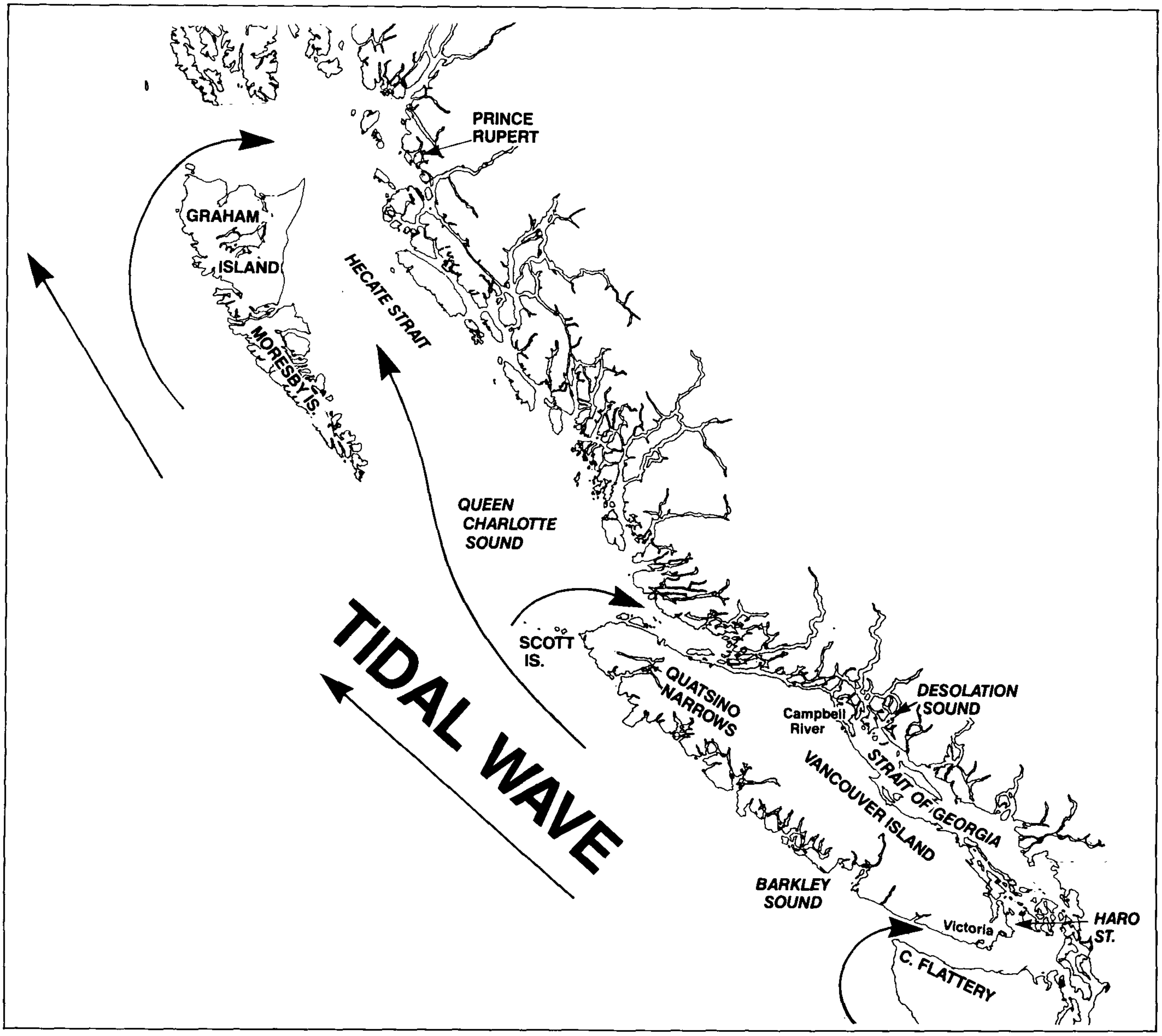

FIG. 3.19. Tide propagation along British Columbia–Washington coast.

passages between the Gulf Islands. The rapid tidal streams throughout this region are an obvious consequence of this delay. Inside the Strait of Georgia, however, water-level changes occur almost simultaneously with no more than a 30-min wait for two points to experience the same stage of the tide.

Between Cape Scott and Campbell River, the passage separating Vancouver Island from the mainland is narrow and strewn with islands. This, and the fact it takes the tide from the south almost 2 h longer to reach the narrowest parts of the channel than the tide from the north, explains the extremely rapid tidal streams associated with Seymour, Yuculta, Surge, Hole-in-the-Wall, Okosillo, and Arran rapids. This difference results in a 2-h lag in the times of the tide at Campbell River compared to Seymour Narrows.

Because propagation of the tidal wave from the south is essentially blocked by the tidal wave from the north at the northern end of the Strait of Georgia, mixing of water is minimized to the point where appreciable warming of the upper few metres can take place during the summer. In quiescent areas like Desolation Sound, surface temperatures may reach 25°C in late July.

Nonastronomical Tides

The kinds of tides described to now are known as astronomical tides as they are generated by gravitational forces. Although these are by far the most predominant type in the ocean, diurnal and semidiurnal changes in sea level can be produced by a variety of other mechanisms. Most are related to meteorological processes.

Due to the earth's daily rotation, the expansion and contraction of the ocean surface through daytime warming and nighttime cooling cause tidelike variations in sea level. (Similar fluctuations in the air are the main cause of large atmospheric tides. Though this heating–cooling cycle is primarily diurnal, the tides themselves are semidiurnal. This apparent paradox was resolved at the end of the 19th century when it was shown that the whole world atmosphere prefers to resonate with the weak semidiurnal constituent of the heating effect, but can never get into step with the much stronger diurnal component of this cyclic forcing.) Similar effects are produced by fluctuations in barometric pressure and by the daily rhythm of land and sea breezes, that move water away from and toward the coast. These so-called radiational tides are extremely difficult to separate from the astronomical tides and are usually lumped with the predictions published in the tide tables.

Storm surges are tidelike fluctuations in elevation initiated by strong onshore or offshore coastal winds during a severe storm (Fig. 3.20). Their effect is small at steep coasts but in low-lying countries, such as Holland, or Bangladesh on the delta of the Ganges River, they can cause extensive flooding, particularly if they occur during periods of high astronomical tides. A 1970 storm surge in Bangladesh is estimated to have killed half a million people. The highest recorded tide in British Columbia of 8.1 m at Queen Charlotte City on Jan. 12, 1967, was during a strong onshore blow.

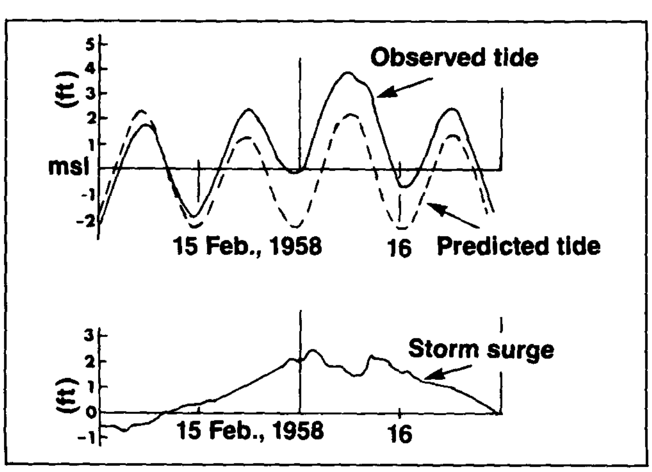

Fig. 3.20. Storm surge for Atlantic City, February 1958. Storm surge measures departure of actual tide from predicted astronomical tide (msl, mean sea level). (From Pore 1964)

Unusually low tides can also occur because of meteorological effects. In the shallow, constricted Torres Strait that separates the northeastern tip of Australia from Papua New Guinea, for example, storm surges have been known to leave large tankers stuck firmly in the mud.

In semienclosed basins such as bays and inlets, rapid but small sea-level changes called seiches are sometimes observed in the tide-gage records. Generated by tides or local winds, or by the propagation of certain types of oceanic waves across the mouth of a basin, the magnitude and periods of these oscillations are closely tied to the geometry of the particular embayment in which they arise. Within British Columbia coastal waters, seiches can have periods of minutes to hours and heights of a few centimetres; a typical period appears to be about 30 min. They are frequently observed in Whaler Bay, Campbell River, Pedder Bay, and Port San Juan as wiggles on the smoother tidal curves (Fig. 3.21; see also Fig. 3.4 for Alert Bay). As these fluctuations can cause the water level of a basin to move up and down a few centimetres over a few minutes, it is sometimes possible to see their presence directly. While sitting at the water's edge in Esquimalt Harbour on a flat-calm day, the author once watched such a gentle rise and fall along a wide stretch of the shoreline.

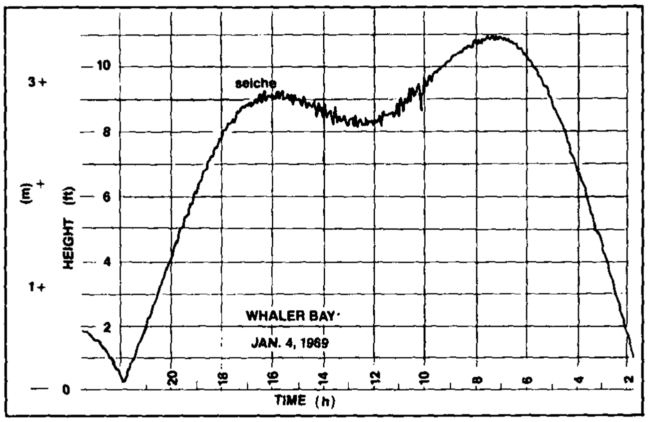

Fig. 3.21. Rapid, small-amplitude seiches superimposed on tide, Whaler Bay, southeast end of Galiano Island, Strait of Georgia. (From LeBlond 1972)

Tidal Friction

As part of a cosmological process, the dissipation of tidal energy at the open coasts of the world (and to a lesser extent the friction between tides and the ocean floor) is decelerating the earth's rate of rotation and causes a lengthening of the day by 0.001 s about every 100 yr. Studies of daily growth rings of ancient coral have shown that 370 million yr ago there were nearly 400 days a year rather than the present 365, so a day was 21.9 h long.

According to the natural laws governing the conservation of angular momentum, the reduction in the earth's spin rate has been accompanied necessarily by an increase in the orbital speed of the moon, causing it to swing out into an ever-widening orbit. This in turn has led to a decrease in the tidal range over the billions of years since the formation of the oceans. The present lunar recession of 3.3 cm/yr will continue until the earth ceases to spin on its polar axis some time in the remote future, just as earth-induced tides in the solid body of the moon have stopped the original spin and forced it to present the same side to the earth at all times. At that future date, the earth and moon will present constant faces to one another as they rotate about a common axis over a 47-day period. The solar tide will then take over, cause a steady reduction in the rotation rate of the earth–moon pair about their common center of mass, and reduce their distance of separation until, eventually, the moon will approach within the Roche Limit of 18,340 km, when its surface will be torn apart by the gravitational force of the earth. Shattered chunks of the moon will rapidly spiral to earth, and possibly halt further catastrophe by increasing the rotation rate of the earth–moon pair, allowing the moon to again recede from the earth.

Tidal Streams

Tidal streams are the horizontal currents associated with the vertical rise and fall of the tide. Unlike horizontal water motions induced by most other processes, these currents have horizontal speeds of almost uniform strength throughout the entire depth of water. In the deep ocean, speeds are only a fraction of a kilometre per hour and a parcel of water carried by the tidal flow has a typical excursion of about 1 km over each half of a tidal cycle. Near the coast, speeds exceed 1 km/h and the water moves over correspondingly greater distances. Where currents are observed to change with depth, a nontidal flow or residual current also will be present. Such flows may be associated with wind-driven ocean currents, surface-induced wind drift, internal oceanic motions generated by the interaction of the tidal streams with the bottom topography, or density currents.

Tidal streams are studied in connection with erosion, the movement of oil spills, the disposal of wastes, and navigation. The latter includes routing larger vessels that must consider time in addition to safety. The often violent tidal currents through Seymour Narrows between Vancouver Island and Quadra Island are an example of flow that must be respected by all ships. This was particularly true before the blasting of Ripple Rock on Apr. 5, 1958, by the biggest nonnuclear peace-time detonation on record at that time (Fig. 3.22). Prior to that date, the swift tidal streams and the twin peaks of the rock that protruded dangerously close to 3 m of the surface formed a deadly combination, wrecked 20 big ships, and took 114 lives.

The tidal stream associated with a rising tide is called the flood, an expression that probably derives from the flooding effect of the tide in low lying areas, whereas the current associated with a falling tide is called the ebb. Commonly used expressions like "flood tide" and "ebb tide" should be avoided, as they confuse the horizontal motions of tidal currents with the vertical displacements of the tide. (Records show that early seafarers also used the same words for tides and tidal currents.) Slack water occurs during the short time interval between the end of the flood and the beginning of the ebb, or vice versa, when the water has no horizontal motion. A further distinction is made between high-water slack at the end of the flood and low-water slack at the end of the ebb.

At any place and time, a tidal stream is specified by both direction and speed, which makes it a much more difficult and expensive quantity to measure than the tide. Measured speeds range from less than 0.5 m/s (1 kn) in the open ocean to a maximum of about 8 m/s (16 kn) in Nakwakto rapids that separate Seymour Inlet north of Vancouver Island from Queen Charlotte Sound. Speeds shown on nautical charts usually indicate the maximum values expected during spring and neap tides in the main channels only, because specification of the exact tidal current for all localities and times is an impossibility. Moreover, these values are often estimates and should be treated accordingly.

Although the connection between the tide and tidal streams is familiar (semidiurnal tides with two highs and two lows per day, for example, are associated with two floods, two ebbs, and four slack water periods a day) it

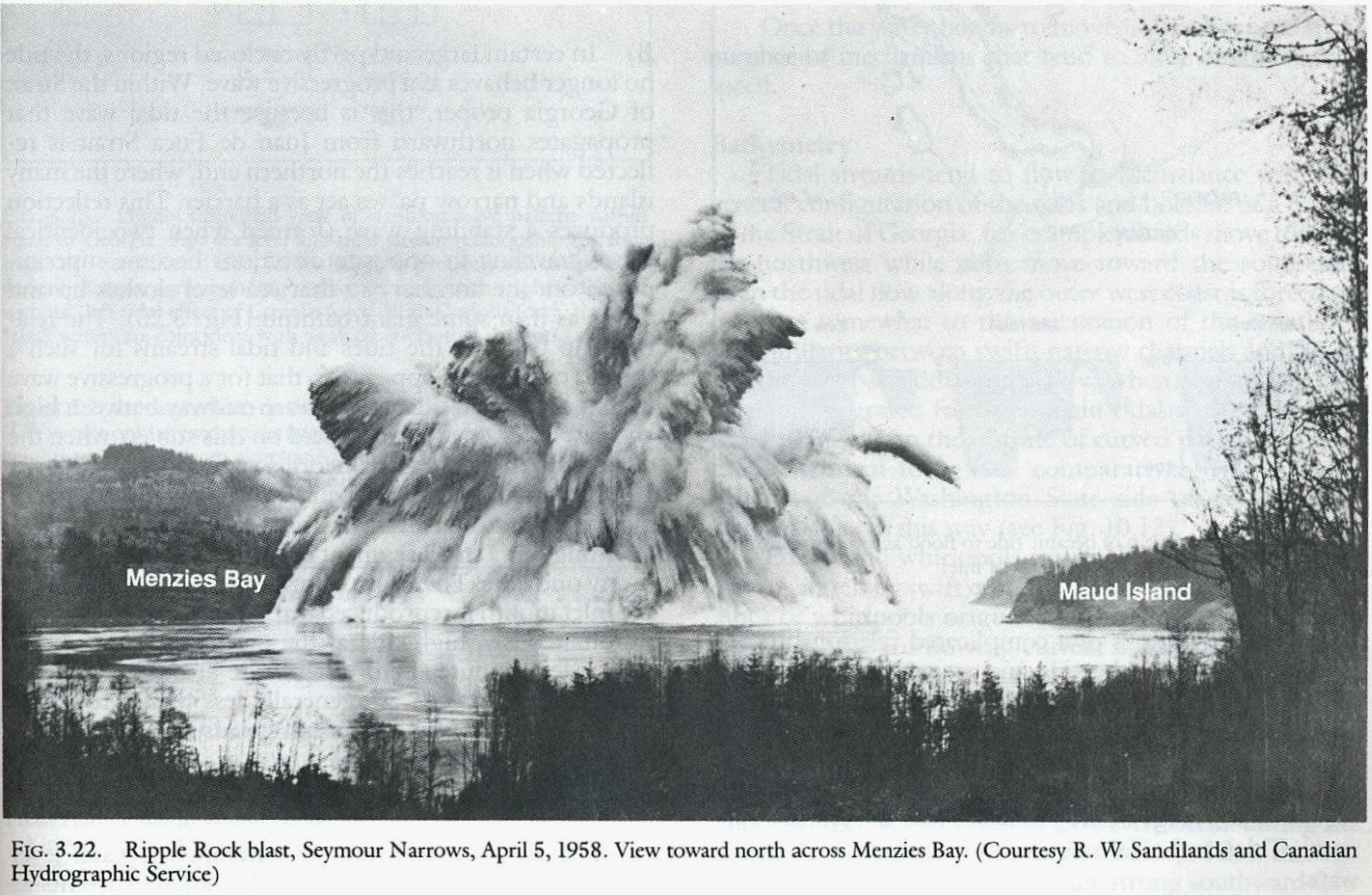

FIG. 3.22. Ripple Rock blast, Seymour Narrows, April 5, 1958. View toward north across Menzies Bay. (Courtesy R. W. Sandilands and Canadian Hydrographic Service)

must not be supposed that the strength of the current, nor the times of slack water, at any locality necessarily coincide with the corresponding vertical changes of the tide. Under certain conditions, in fact, slack water occurs midway between high and low tide, contrary to the commonly accepted notion that it occurs at high and low tide. A case in point is Nitinat Lake on the west coast of Vancouver Island (Fig. 3.23). Here, the channel that connects the lake to the ocean so constricts the passage of water that little enters or leaves the lake, the tidal range in the lake is about 0.3 m when the tidal range on the coast is 3.3 m. Therefore, because the elevation of the lake is nearly constant, slack water in the channel occurs when the coast tide has the same elevation as the lake, about midway between high and low water. Maximum currents, on the other hand, tend to develop when the difference in elevation between the lake and ocean is greatest, near low or high tide. Similar situations are common in B.C. waters, the fast currents through Nakwakto rapids and Skookumchuck rapids (Sechelt Inlet) are two of the more dramatic examples.

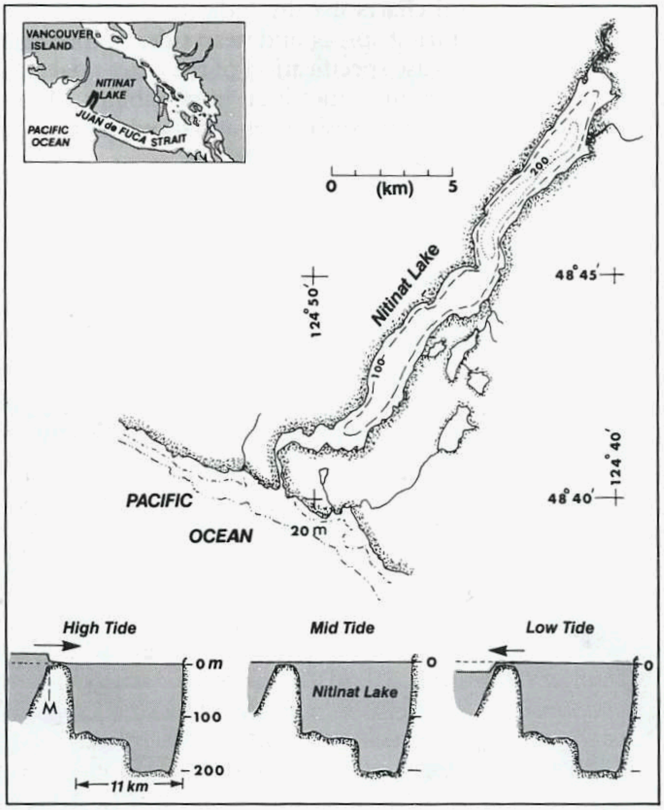

FIG. 3.23. Relationship of oceanic tide to flood and ebb streams into and out of Nitinat Lake. M = mouth of inlet.

Other indications of a complicated relationship between the stages of the tide and stages of the currents include the often nonexistence of "slack" water, and currents that continue to ebb when they should have turned to flood.

To understand the reasons for such complications requires knowledge of the processes that govern the currents. Tides and tidal streams are related in three basic ways.

A) As explained earlier, the oceanic tide propagates as a very long progressive wave, high tides correspond to the wave crests, and low tides to the wave troughs. Tidal streams in the ocean are the horizontal motion associated with the passage of this wave, floods correspond to the forward motions under the crests, and ebbs the backward motions under the troughs (Fig. 3.24). Given enough room to maneuver, this is how tides and tidal streams would be related. Once the motions become restricted by coastlines, however, the situation is usually altered, although tides and tidal currents in the main channels of larger regions like Queen Charlotte Sound, Hecate Strait, the Juan de Fuca Strait retain much of the behavior of a progressive tidal wave.

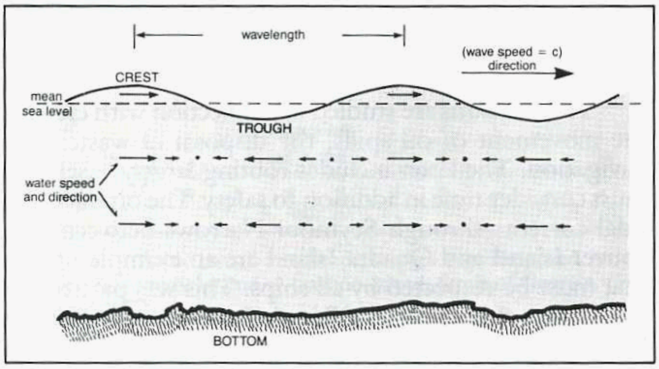

FIG. 3.24. Cross-section through a progressive wave with crests and troughs moving from left to right at speed, c. Dots indicate zero current speed at points midway between crests and troughs. Wave speed greatly exceeds current speeds associated with wave-induced motions.

B) In certain large and partly enclosed regions, the tide no longer behaves as a progressive wave. Within the Strait of Georgia proper, this is because the tidal wave that propagates northward from Juan de Fuca Strait is reflected when it reaches the northern end, where the many islands and narrow passes act as a barrier. This reflection produces a standing wave (formed when two identical waves traveling in opposite directions become superimposed on one another) so that sea level sloshes up and down as if in some giant bathtub (Fig. 3.25). The relationship between the tides and tidal streams for such a wave is completely opposite to that for a progressive wave in that maximum tidal streams are midway between high and low tides. More will be said on this subject when the circulation of the Strait of Georgia is discussed in Chapter 10. In deep inlets, the tide behaves somewhat like a standing wave, although all inlets are much too short to allow anything but a small segment of the wave to occupy them at any one time. Therefore, a variation in tidal elevation at the inlet mouth is reproduced throughout the inlet almost simultaneously, high tide at the mouth typically occurs only a few minutes before high tide at the head (Table 3.2). Current speeds are generally less than a few kilometres per hour, with maximum speeds midway between high and low tide, and slack waters at high and low tide (Fig. 3.26).

C) In passes and narrows, the difference in water level between the two ends of the channel induces currents as the water seeks a common level. This difference, called the hydraulic head, is caused by a difference in arrival time of the tide at either end or by a difference in the range of the tide at the two ends. The numerous passes within the San Juan and Gulf Islands have tidal streams produced in this way. In Active Pass, for example, the tidal streams are determined strictly by the hydraulic head, and have no resemblance to the ebbs and floods in the Strait of Georgia. Seymour Narrows is another region where tidal streams are produced by a delay in the tide between two ends of a channel. In this case, the time difference between the tide at the northern and southern ends of the Narrows amounts to as much as 2 h, with accompanying differences in elevation that can exceed a metre or more, and currents that can churn through the constricted passage at over 7 m/s (13 kn).

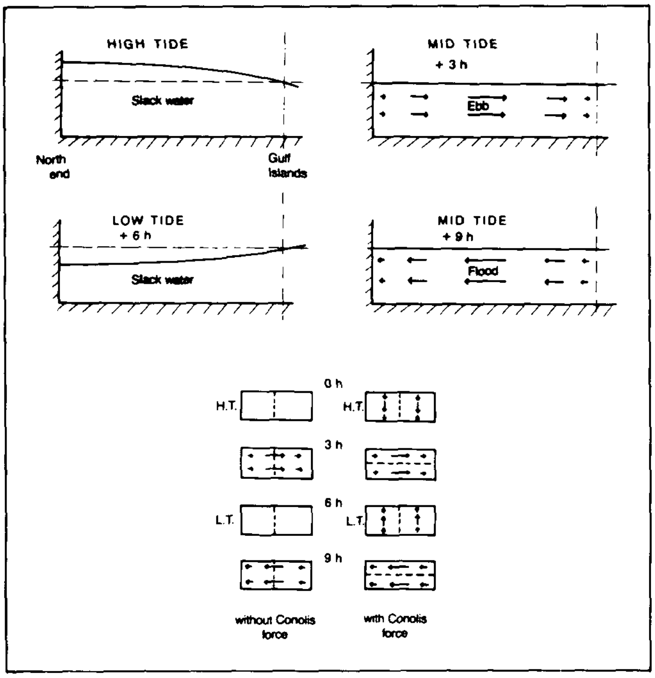

FIG. 3.25. Highly simplified view of standing wave pattern within Strait of Georgia. *Top*: sea level and tidal stream relationship for four stages of a semidiurnal tide; arrow length is proportional to speed of current. *Bottom*: corresponding surface currents with (right side) and without (left side) effect of Coriolis force. With Coriolis force, times of "slack water" are coincident with weak cross-channel currents.

TABLE 3.2. In deep inlets, times of high water (HW) and low water (LW) occur only a few minutes later at the head than at the mouth. Tidal heights were taken from simultaneous tide-gage records operated day and night; time kept accurately at the mouth and head by chronometers. (Adapted from Dawson 1920)

Long inlets (avg. tide range 4 m)	HW	LW
From Whaletown on Cortes Island to head of Bute Inlet, 84 km; from comparison of observations in two different seasons with same reference station	3 min later	9 min later
From Namu to Bella Coola by Burke Channel and Bentinck Arm, 111 km; from 144 simultaneous observations	2 min later	7 min later
From Hartley Bay in Wright Sound to Kitimat, by Douglas Channel, 79 km; from 222 simultaneous observations	4 min later	4 min later

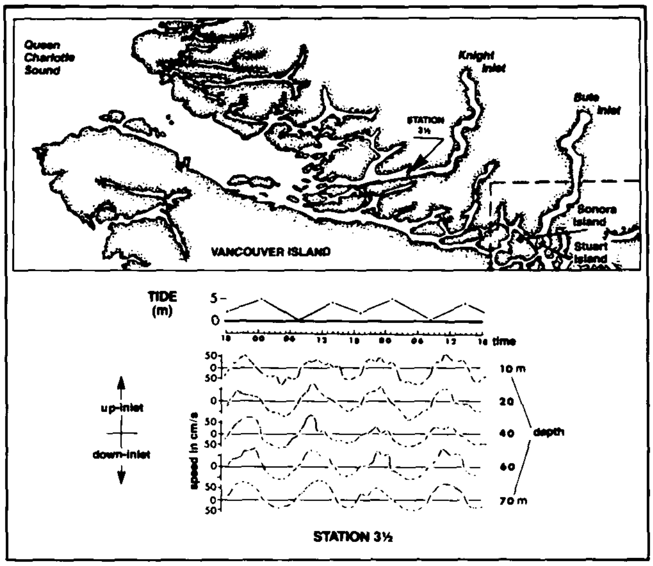

FIG. 3.26. Tides and observed along-channel currents at Station 3½ in Knight Inlet July 6–8, 1956. Speeds were measured almost simultaneously at the five different depths and plotted versus time in hours. Slack waters occur at low and high tides, maximum speeds midway between tides. (From Pickard and Rodgers 1959)

Factors that Influence Tidal Streams

Once the water begins to move, it is influenced by a number of mechanisms that tend to alter direction and speed.

Bathymetry

Tidal streams tend to flow in accordance with the general configuration of the sides and bottom of a basin. In the Strait of Georgia, for example, floods move toward the northwest while ebbs move toward the southeast. Even the tidal flow along the outer west coast is forced to conform somewhat to the orientation of the coastline. The similarity between swift, narrow channels and rivers accounts for backeddies in the flow, when nearshore water becomes separated from the main tidal stream, and for water piling up on the outside of curved passages due to the centrifugal force (the comparatively higher tidal heights on the Washington State side of the Strait of Georgia arise in this way (see Fig. 10.12).

Eddies and whirlpools are usually generated in narrow channels by swift currents. In Seymour Narrows, two lines of whirlpools originate during the flood where the rapid, southward flowing current comes in contact with the nearly motionless waters to the east and west (Fig. 3.27). In the Cordero Channel and Yuculta rapids area, between Sonora Island and Stuart Island at the entrance to Bute Inlet (Fig. 3.28), dangerous whirlpools are formed west of Little Dent Island and over certain shoals. The most dangerous whirlpools, however, occur during the flood where the rapid, 3 m/s (6 kn) eastward flow through Gillard Passage meets the equally strong southward flow

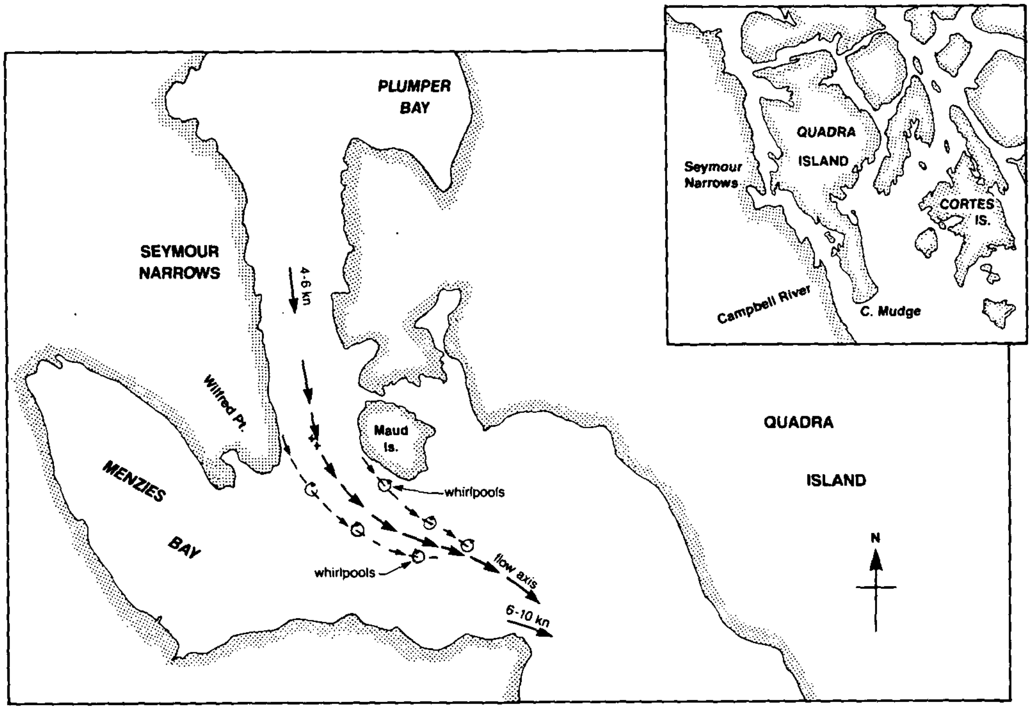

FIG. 3.27. Flood currents, Seymour Narrows. Whirlpools are formed where jetlike flow contacts slowly moving waters to right and left. Backeddies also formed in lee of shoreline promontories within the Narrows. Arrows denote typical maximum flood speeds. Plus (+) signs mark location of Ripple Rock peaks.

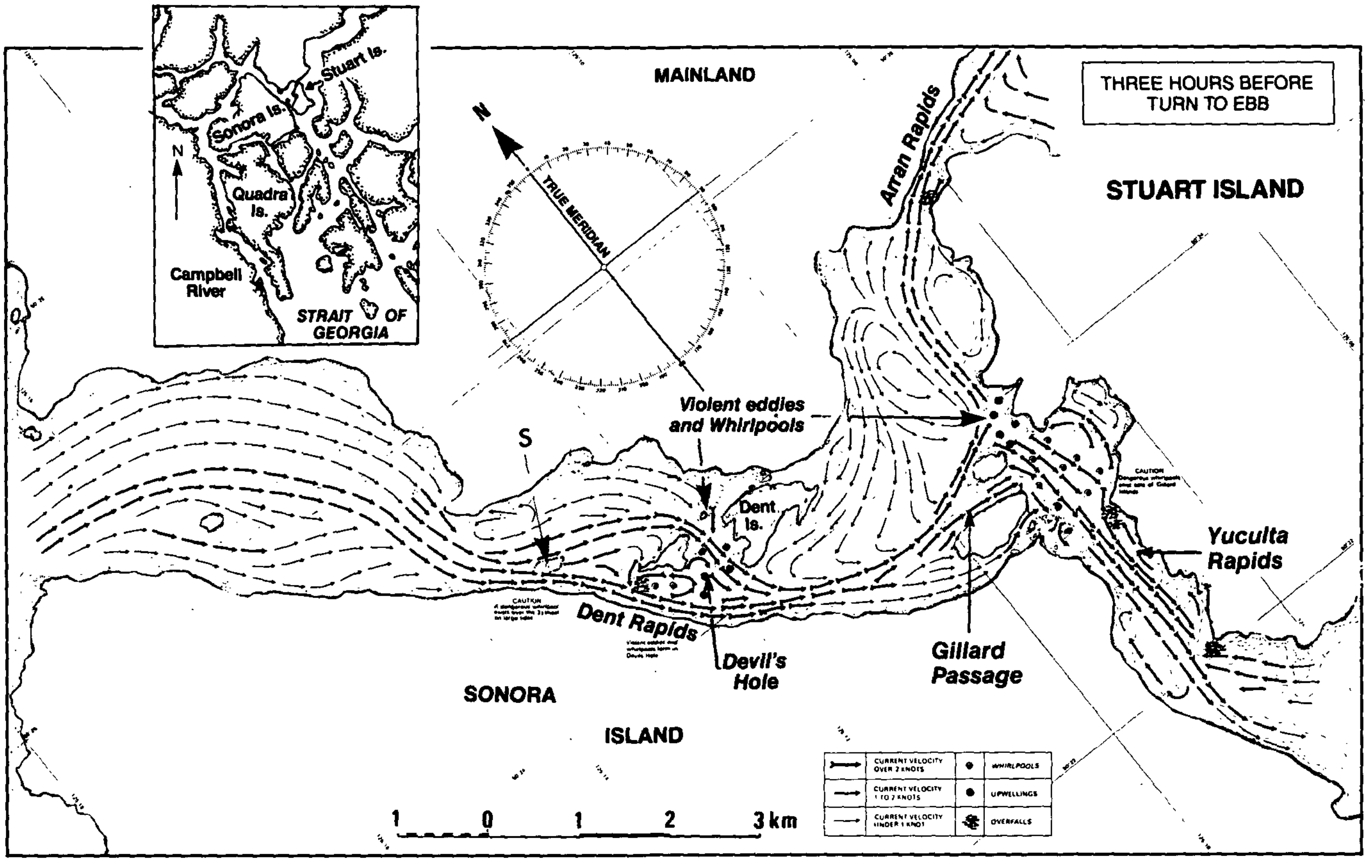

FIG. 3.28. Tidal streams 3 h before the turn to ebb in Cordero Channel region near entrance to Bute Inlet (see Inset: also Fig. 3.26). Dangerous whirlpool over 6.4-m (3½-fa) shoal (S) northwest of Dent Rapids on large tides. Violent eddies and whirlpools form in Devil's Hole and southeast of Gillard Passage. (Modified from Tidal Current Publication 23, Canadian Hydrographic Service 1970.)

toward Yuculta rapids. Hydrographer Stan Huggett recalls the unnerving feeling of being able to look down 4 m into the hole formed by one of these whirlpools as his survey launch struggled away from its edge!

Opposite to whirlpools are the smooth, dome-shaped surfaces called upwellings, produced when water is forced upward by features on the bottom of the tidal channel (see Fig. 5.2).

Finally, the bathymetric configuration of a channel may result in the flood moving in on one side while it is still ebbing on the other side. Although the ebb will eventually turn to flood, slack water within the channel will be virtually nonexistent.

Friction

A layer of water that is moving relative to another layer of water or over a solid surface is acted on by frictional drag, which opposes the motion. Consequently, the speed of a current decreases near the bottom and sides of a body of water, and eventually becomes zero at solid surfaces, a condition of the "no-slip" requirement for fluids in immediate contact with solids. Therefore, greatest speeds tend to be at the upper surface in midchannel where the effect of friction is least. A wide, deep channel will have more rapid tidal streams than a shallow, narrow one of the same hydraulic head, as exemplified by the adjacent, but different, Dodd Narrows and False Narrows between Gabriola and Vancouver islands.

If water is moving over or along a "smooth" surface (a sandy bottom, straight shoreline, or newly polished hull) the flow will also be smooth and friction minimal. On the other hand, if the surface is "rough" (a rocky bottom, an irregular shoreline, or a fouled hull) the flow will be irregular and friction large. Thus, rocky, island-strewn passages will slow tidal streams more than smooth obstacle-free ones. Irregular or turbulent motions such as whirlpools and upwellings also behave as effective frictional drags, and further reduce the water's speed. Without them, the average speed of the currents in places like Dodd Narrows or Active Pass would be greater. Obstacles such as log booms, bridge supports, and kelp beds also increase friction; the net drag on the Fraser River is significantly increased by the numerous man-placed obstacles within it.

Inertia and Momentum

Because it has mass, water possesses a certain inertia or indisposition to a change in motion; if stopped it tends to remain so, if moved it tends to continue moving, friction notwithstanding. Therefore, the speed of tidal currents associated with hydraulic heads will not adjust immediately to changes in water elevation along the channel, which partly accounts for the lag between tide and current in such areas (Fig. 3.29). Moreover, the speed of the water determines momentum in the same way that the speed of a boat determines momentum. The greater the momentum the greater the effort needed to bring the water to rest. Thus, water that runs out of a harbor during a falling tide will not immediately come to rest simply because the tide has begun to rise again; momentum will carry it against or around the advancing tide. Because the current speed is typically greatest in the deeper sections of the channel, the water there could still be ebbing after it had begun to flood elsewhere in the harbor. On the other hand, the momentum of any current is quickly lost once it moves into more open water from the confines of the channel, as Porlier Pass, where even strongest floods rapidly weaken a few kilometres into the Strait of Georgia.

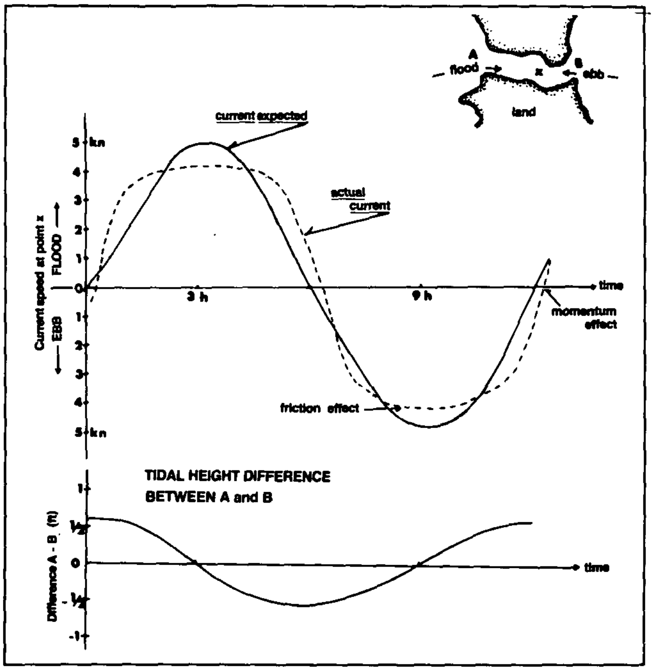

FIG. 3.29. A hypothetical example of expected and actual currents measured at a location X in a channel. Lower curve shows difference in sea level elevation (hydraulic head) between ends of channel, A to B, as it varies over a semidiurnal tidal cycle; upper curves give associated currents. Frictional effects reduce maximum possible speeds generated by hydraulic head, and momentum delays time of slack water from expected. Also, actual currents (broken line) reverse direction much more quickly than in absence of friction and momentum effects.

Coriolis Force

Provided there are not too many topographical restrictions on the flow, this force produces a continual deflection on the current to the right of its direction of motion, unlike friction drag and the hydraulic head that act along the direction of motion. In the open North Pacific, Coriolis force causes tidal streams to constantly change direction, whereby they turn in a complete revolution of 360° over each tidal cycle. More will be said on this topic shortly. In more confined regions, the Coriolis effect is often suppressed, and becomes negligible within the smaller passes of British Columbia. Nevertheless, its influence is marginally present in larger areas like the Strait of Georgia, Hecate Strait, and Juan de Fuca Strait where floods tend to be slightly stronger on the mainland sides (to the right of the direction of flood) and ebbs slightly stronger on the island sides (to the right of the direction of ebb).

River Runoff

Water brought into a region on the flood can be more dense (heavier) than water taken out during the ebb. If so, currents associated with the flood will be strongest closer to the bottom while currents associated with the ebb will be strongest nearer the surface. Vertical tidal mixing of Puget Sound and Strait of Georgia water with the more dense water of Juan de Fuca Strait at the turbulent passes connecting them produces such a situation. Thus, ebb currents are strongest in the upper half of Juan de Fuca Strait while flood currents are strongest in the lower half (see Fig. 11.12). This bias is, of course, nothing more than the estuarine circulation described in Chapter 2, which has been superimposed upon the "true" streams directly associated with the tidal forces of the moon and sun.

Winds

Wind-generated currents may upset the natural rhythm of tidal streams. Opposing winds can delay the turn of the tide near the surface, particularly in regions with a distinct brackish upper layer, and reduce the maximum speed of the tidal current. Winds that blow along the direction of flow on the other hand can augment the speed of the tidal current. This wind-effect may be appreciable; in some cases it accounts for the discrepancy between predicted and observed tidal currents.

Current Ellipses

Where the direction of tidal streams is constrained by the sides of a channel, the currents tend to be rectilinear. That is, ebbs and floods at all stages of the tide are directed parallel to the trend of the channel and there is negligible cross-channel set. Under such conditions, the stream simply decelerates without changing orientation, reaches slack water, and then accelerates in the opposite direction (Fig. 3.30a). Tidal streams of this kind are common to the narrow, regular basins within the inshore waters such as Johnstone Strait, Juan de Fuca Strait, and Puget Sound.

Where the tidal streams have room to maneuver, however, the influences of the earth's rotation, centrifugal forces, friction, and inertial effects cause the current's direction to turn with time as it changes between flood and ebb (Figs. 3.30b, c). For such rotary currents, the shape of the curve traced out by the tip of the current vector over a tidal cycle depends on the type of tide producing the flow. This effect can be shown by following the path of an imaginary ship riding at a deep-sea anchor, so that its distance from the pivot point is directly proportional to the speed of the tidal stream (Fig. 3.31); the stronger the current the farther the ship will be from the center, and vice versa. The direction of the current, of course, will be parallel to the anchor line. Suppose the tidal streams are purely semidiurnal and there is no change in their speed with time, only a change in direction. Then, over one complete cycle (12 h, 25 min), the ship would swing in a complete circle as it rotated with the top of the current vector. The swing would be counterclockwise if the tidal streams "backed" with time or clockwise if "veered." If tides are diurnal, the ship would complete the circle in just under 25 h.

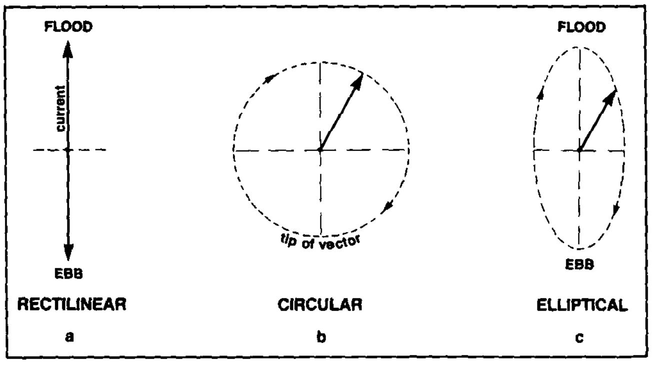

FIG. 3.30. Rectilinear versus rotary tidal streams, as measured at a fixed location. Vectors (arrows) give direction of flow at center point, and broken line gives location of tip over a complete tidal cycle. Length of arrow is proportional to speed of current.

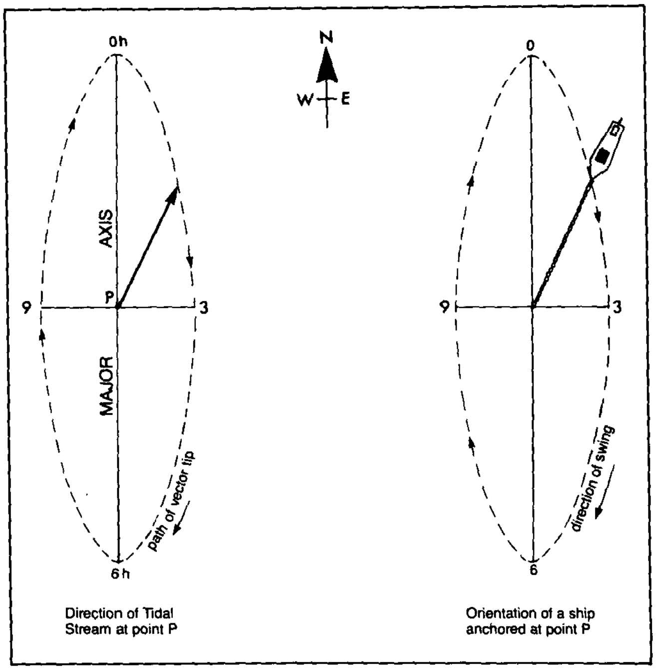

FIG. 3.31. Rotary tidal currents. Changes in speed and direction of tidal stream over a tidal cycle as measured at Point P are compared to length and orientation of an anchor line pivoted at Point P. Length and orientation of anchor line vary in accordance with strength and direction of flow.

The sense of rotation of the tidal stream in a given area can only be determined by direct observation. Moreover, the current usually has a preferred direction of flood and ebb, with appreciably lesser speeds at right angles to these directions. Instead of tracing out a circle, therefore, the tip of the current vector generally traces an ellipse, its major axis along the direction of maximum ebb and flood and its minor axis along the direction of weakest flow (Fig. 3.30). (A rectilinear tidal stream is one where the ellipse is so flattened that it has no minor axis.) As an example, consider a tidal stream with maximum flood to the north (0°True) and semidiurnal clockwise rotation, as in Fig. 3.31. As the flood swings to the east its speed gradually diminishes until, about 3.1 h later, it flows weakly to the east (90°T). The speed then begins to increase until the stream attains maximum ebb to the south (180°T), 3.1 h later. This in turn is followed by a reduction in speed to a minimum westerly flow (270°T) after an additional 3.1 h, and a return to maximum flood after a total time of about 12.4 h.

Because tides in the ocean are typically of the mixed variety, the picture presented here is an oversimplification. It is, nevertheless, valid under certain circumstances (a) on those days of the month when B.C. tides are almost purely diurnal or purely semidiurnal, (b) in inshore waters where the ellipses are constrained by channel topography, and (c) if attention is concentrated on each separate tidal constituent that makes up the mixed tide. In isolation then, the K_1 and M_2 constituents of the tidal streams behave as described. Therefore, tide analysts prefer to work with the "tidal current ellipses" for each constituent as a way to visualize the variation of currents with time in a particular location. However, the mariner should realize that the path traced out by the actual current vector associated with mixed tides is more complex than the simple ellipse traced out by each constituent, though from one day to the next the overall pattern is nearly repeated.

Tidelines

The assortment of driftwood, uprooted seaweed, foam, and other debris floating on the water is constantly shifted around by the different current regimes. Wherever these currents meet or "converge," accumulated material can form into discernible lines or rows called tidelines, although their formation may have nothing whatever to do with the tides.

Essentially four distinct mechanisms produce tidelines through a convergence of surface waters.

A) They may be formed where water in a particular current regime is sinking underneath or riding over top of the surface layer of another regime. In this case, the debris is skimmed off the advancing flow as it sinks below the surface of the opposing water mass, or alternatively the advancing current skims off debris as it moves over the adjoining water mass (Fig. 3.32). The former process is somewhat analogous to the subsidence of the oceanic crust beneath the continental margin discussed in Chapter 1. Tidelines of this type are commonly found in inlets with rivers at their heads, and at the boundary between clear Strait of Georgia water and the silty runoff from the Fraser River. In Pl. 2, tidelines of accumulated foam can be seen along the edges of the Squamish River runoff as it moves down Howe Sound. Tidelines at the entrances to tidal channels, or downstream of an island where different current regimes make contact, fall into this category, as do the tidelines south of Rosario and Haro straits and off Kelsey Bay in Johnstone Strait.

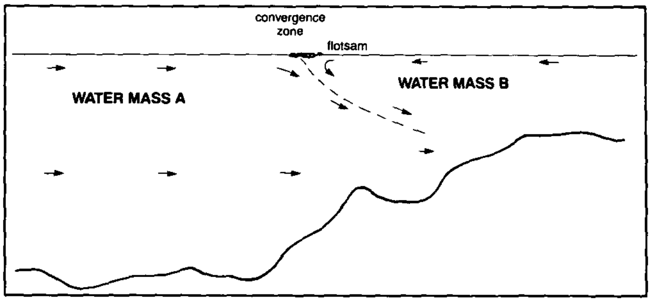

Fig. 3.32. Formation of tidelines at boundary between two opposing bodies of water. Surface debris gathers at convergence zone to form distinct lines that shift with changes in winds and tides.

Mariners should note that these tidelines often mark an abrupt change in the speed and direction of the near-surface currents, as illustrated by Fig. 3.33, based on measurements taken about 5 km west of Sturgeon Bank off the Fraser River delta. As the tidelines slowly passed

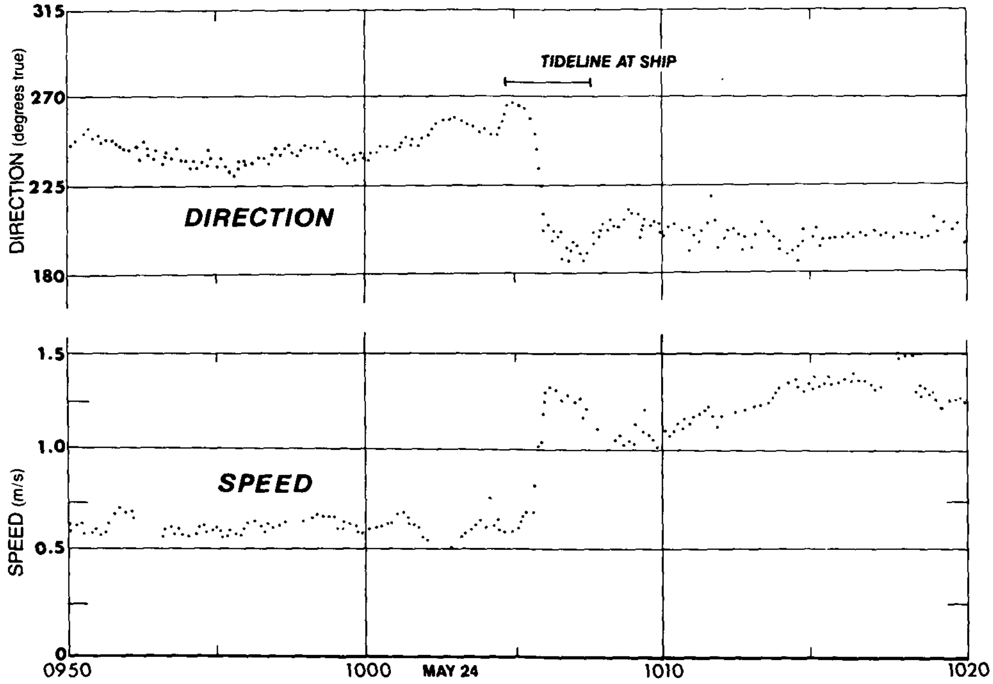

Fig. 3.33. Abrupt change in surface current speed and direction associated with passage of a tideline. Measurements taken from a research ship 5 km west of Sturgeon Bank, Fraser River, May 24, 1967. (Courtesy S. Tabata)

under the research ship, the current speed increased rapidly from 50 cm/s (1 kn) in the clear salty water to nearly 150 cm/s (3 kn) in the silty brackish water; the direction of the current swung almost 90° as it shifted from a westward flow to a southward flow.

B) Backeddies may also produce tidelines. In this case, the debris trapped within the eddy tends to become strung-out at the region of intersection between the slowly moving currents in the eddy and the more swiftly moving water in the main channel. Once established, this situation will persist until there is a reversal in flow or a disruption by the winds.

C) Surface convergence zones associated with internal gravity waves (Chapter 6) are another common source of tidelines. These subsurface waves are responsible for the well-defined series of bands or "slicks" that often characterize the surface of estuarine-type regions like the Strait of Georgia and mainland inlets during the summer. Even when there is little debris to accumulate, these bands may still be visible as relatively smooth stretches of water horizontally separated by many tens of metres of rippled water (and vice versa) that extend for many kilometres in twisting, disjointed ribbons. From the air they also appear as rows of light silty water separated by areas of dark blue oceanic water (Pl. 7). Sun glare associated with the relatively flat areas of the pattern may further help to delineate certain bands (Pl. 8). Logs caught in such convergence zones align themselves along the slicks while small ripples steepen rapidly and disappear as they approach the edge of the slick.

It should be noted that cat's paws, produced during periods of light winds, also will create rippled and nonrippled areas of water. These are usually patchy and without the lateral extent characteristic of internal wave slicks. Quite often the smooth regions of the slicks are due to a thin oily film of marine organisms, which damps the ripples and gives the sea a glossy look. Oil from motorboats has the same physical effect.

D) Lastly, winds can produce convergence zones by inducing cell-like circulation patterns in the surface water. Tidelines of this type are called windrows, and occur where the wind-blown foam and other material gathers at the junction of two adjacent cells set up at right angles to the wind direction (Chapter 4).

Red Tide

A visible feature of coastal areas in summer is the appearance of patches of reddish-brown water known as red tide (Pl. 9). Although it is essentially a biological phenomenon, the occurrence of red tide does bear some relation to the type of physical oceanic processes discussed. As in the formation of tidelines, red tide is usually associated with converging currents, except that the material brought together is not surface debris but a particular species of freely floating microscopic marine plants called phytoplankton (phyto meaning plant, and plankton meaning drifter). Found everywhere throughout the world's oceans, phytoplankton photosynthesize carbon dioxide and water into oxygen and carbohydrates in the presence of sunlight just as land plants do. In the spring, the increase in the amount of sunlight and accompanying increase in surface temperatures in coastal areas encourages multiplication of these plants, a process further hastened in regions of converging currents where nutrients are more abundant. Eventually, conditions become so favorable that their numbers exceed millions per litre of water and produce a display of color called a "bloom." In B.C. waters there are two such phytoplankton blooms, one associated with the rapid multiplication of diatoms (a form of phytoplankton with over 20,000 world species) and the second with the rapid multiplication of dinoflagellates (dinos, a whirling; and flagella, whips). The main diatom bloom takes place in spring and is responsible for the greenish tinge to the water in the Strait of Georgia at that time.

The dinoflagellate bloom takes place in early to late summer in coastal British Columbia and Washington, where it produces red tides. For the most part, this phenomena is beneficial to aquatic life because the microscopic plants provide both oxygen and food for other organisms. But there are times when its presence can lead to toxic effects. In the Gulf of Mexico, for example, a particular species of dinoflagellate releases a poison into the water which, during a red tide, becomes harmful to man and leads to massive fish kills. British Columbia species retain the toxin and there is no such problem. On the other hand, dinoflagellates are food for clams, oysters, mussels, and other shellfish that subsequently tend to concentrate the toxin within their tissue. Although immune to the poison, a healthy looking shellfish may in fact be a highly lethal morsel if red tide has been a persistent feature of the animal's habitat, and, if eaten, may lead to paralytic shellfish poisoning. The B.C. coast is potentially one of the worst areas in the world for such poisoning. Butterclams are one of the most serious offenders because they can retain the toxin for up to 2 yr. Mussels, on the contrary, quickly lose any toxin they take up and are usually safe to eat in late winter and early spring. Despite the possibilities, shellfish poisoning is rare. However, the prudent mariner wishing to live off the sea in the summer and autumn would be wise to heed all official shellfish poison warnings and to check with the Fisheries Inspection Branch, Fisheries and Marine Service, Fisheries and Oceans, 1090 West Pender St., Vancouver, B.C., or a local fisheries office before feasting on steamed clams and raw oysters. Symptoms of paralytic shellfish poisoning are generally noticed within 10 min of ingestion and include a tingling of the lips, tongue, and inside of the mouth. The tingling sensation gradually spreads to the fingers and toes, and leads to numbness in the legs and arms. Anyone experiencing these symptoms after eating shellfish should contact a doctor immediately. (On May 17, 1980, a resident of Health Bay, a small village on Gilford Island, died in Alert Bay hospital of paralytic shellfish poisoning after eating butterclams. Two others became seriously ill, but recovered within a few days.)

If the concentration of phytoplankton in a red tide becomes so great that the food supply becomes exhausted, the entire population will rapidly die out, sometimes in the short period of 1 day. As the decaying process uses up available oxygen, the water may become incapable of supporting life, which in turn can lead to mortality of marine life unfortunate enough to reside in the region, or to have had this poorly aerated water moved to their area by a change in the wind and current.

Bioluminescence

Anyone who has been boating or swimming in coastal waters at night will have noticed the brilliant "phosphorescence" that occurs whenever the water is disturbed. On moonless summer nights, a wake may become a faintly glowing path that extends far behind the ship, and each wave crest that laps against the hull produces a glittering display of brilliance against the darkness of the ocean.

Early mariners and scientists attributed these flashes of light to a wide variety of causes and it wasn't until the end of the 18th century that luminescence in the sea was found to be a biological process. It is now known that it originates with the same dinoflagellates that form red tides; each microscopic organism responds with a bright rapid flash lasting for ⅒ s whenever there is a disturbance in its aquatic environment. Contrary to popular belief, the processes are unlike phosphorescence (which depends on the prior absorption of light from an external source) as the organism generates the light internally through a chemical reaction. Moreover, it is "cold" light unaccompanied by the emission of heat that typifies most chemical reactions. Why these animals possess the capability for luminescence is unclear, but would appear to be some sort of alarm mechanism yet to be understood.