TIDAL FLAT

See **SALT MARSHES AND MUD FLATS**

TIDES

D. T. Pugh, University of Southampton, Southampton, UK

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Introduction

Even the most casual coastal visitor is familiar with marine tides. Slightly more critical observers have noted from early history, relationships between the movements of the moon and sun, and with the phases of the moon. Several plausible and implausible explanations for the links were advanced by ancient civilizations. Apart from basic curiosity, interest in tides was also driven by the seafarer's need for safe and effective navigation, and by the practical interest of all those who worked along the shore.

Our understanding of the physical processes which relate the astronomy with the complicated patterns observed in the regular tidal water movements is now well advanced, and accurate tidal predictions are routine. Numerical models of the ocean responses to gravitational tidal forces allow computations of levels both on- and offshore, and satellite altimetry leads to detailed maps of ocean tides that confirm these. The budgets and flux of tidal energy from the earth–moon dynamics through to final dissipation in a wide range of detailed marine processes has been an active area of research in recent years. For the future, there are difficult challenges in understanding the importance of these processes for many complicated coastal and open ocean phenomena.

Tidal Patterns

Modern tidal theory began when Newton (1642–1727) applied his formulation of the Law of Gravitational Attraction: that two bodies attract each other with a force which is proportional to the product of their masses and inversely proportional to the square of the distance between them. He was able to show why there are two tides for each lunar transit. He also showed why the half-monthly spring-to-neap cycle occurred, why once-daily tides are a maximum when the Moon is furthest from the plane of the equator, and why equinoctial tides are larger than those at the solstices.

The two main tidal features of any sea-level record are the range (measured as the height between successive high and low levels) and the period (the time between one high (or low) level and the next high (or low) level). Spring tides are semidiurnal tides of increased range, which occur approximately twice a month near the time when the moon is either new or full. Neap tides are the semidiurnal tides of small range which occur between spring tides near the time of the first and last lunar quarter. The tidal responses of the ocean to the forcing of the moon and the sun are very complicated and tides vary greatly from one site to another.

Tidal currents, often called tidal streams, have similar variations from place to place. Semidiurnal, mixed, and diurnal currents occur; they usually have the same characteristics as the local tidal changes in sea level, but this is not always so. For example, the currents in the Singapore Strait are often diurnal in character, but the elevations are semidiurnal.

It is important to make a distinction between the popular use of the word 'tide' to signify any change of sea level, and the more specific use of the word to mean only regular, periodic variations. We define tides as periodic movements which are directly related in amplitude and phase to some periodic geophysical force. The dominant geophysical forcing function is the variation of the gravitational field on the surface of the earth, caused by the regular movements of the Moon–Earth and Earth–Sun systems. Movements due to these gravitational forces are termed gravitational tides. This is to distinguish them from the smaller movements due to regular meteorological forces which are called either meteorological or more usually radiational tides.

Gravitational Potential

The essential elements of a physical understanding of tide dynamics are contained in Newton's Laws of Motion and in the principle of Conservation of Mass. For tidal analysis the basics are Newton's

Laws of Motion and the Law of Gravitational Attraction.

The Law of Gravitational Attraction states that for two particles of masses m_1 and m_2, separated by a distance r the mutual attraction is:

$$F = G\frac{m_1 m_2}{r^2} \qquad [1]$$

G is the universal gravitational constant.

Use is made of the concept of the gravitational potential of a body; gravitational potential is the work which must be done against the force of attraction to remove a particle of unit mass to an infinite distance from the body. The potential at P on the Earth's surface (**Figure 1**) due to the moon is:

$$\Omega_p = -\frac{Gm}{MP} \qquad [2]$$

This definition of gravitational potential, involving a negative sign, is the one normally adopted in physics, but there is an alternative convention often used in geodesy, which treats the potential in the above equation as positive. The advantage of the geodetic convention is that an increase in potential on the surface of the earth will result in an increase of the level of the free water surface. Potential has units of L^2T^{-2}. The advantage of working with gravitational potential is that it is a scalar property, which allows simpler mathematical manipulation; in particular, the vector, gravitational force on a particle of unit mass is given by $-\mathrm{grad}(\Omega_p)$.

Applying the cosine law to $\triangle$OPM in **Figure 1**

$$\mathrm{MP}^2 = a^2 + r^2 - 2ar\cos\phi \qquad [3]$$

$$\therefore \mathrm{MP} = r\left[1 - 2\frac{a}{r}\cos\phi + \frac{a^2}{r^2}\right]^{1/2} \qquad [4]$$

Hence we have:

$$\Omega_p = -\frac{Gm}{r}\left[1 - 2\frac{a}{r}\cos\phi + \frac{a^2}{r^2}\right]^{-1/2} \qquad [5]$$

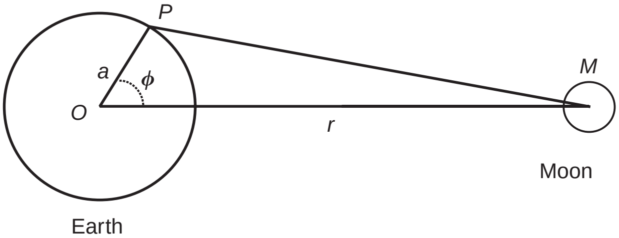

Figure 1 The general position of the point P on the Earth's surface, defined by the angle ϕ.

which may be expanded:

$$\Omega_p = -\frac{Gm}{r}\left[1 + \frac{a}{r}P_1(\cos\phi) + \frac{a^2}{r^2}P_2(\cos\phi) + \frac{a^3}{r^3}P_3(\cos\phi) + \cdots\right] \qquad [6]$$

The terms in $P_n(\cos\phi)$ are the Legendre Polynomials:

$$P_1 = \cos\phi \qquad [7]$$

$$P_2 = \tfrac{1}{2}(3\cos^2\phi - 1) \qquad [8]$$

$$P_3 = \tfrac{1}{2}(5\cos^3\phi - 3\cos\phi) \qquad [9]$$

The tidal forces represented by the terms in this potential are calculated from their spatial gradients $-\mathrm{grad}\ (P_n)$. The first term in the equation is constant (except for variations in r) and so produces no force. The second term produces a uniform force parallel to OM because differentiating with respect to $(a\cos\phi)$ yields a gradient of potential which provides the force necessary to produce the acceleration in the earth's orbit towards the center of mass of the Moon–Earth system. The third term is the major tide-producing term. For most purposes the fourth term may be neglected, as may all higher terms.

The effective tide-generating potential is therefore written as:

$$\Omega_p = -\tfrac{1}{2}Gm\frac{a^2}{r^3}(3\cos^2\phi - 1) \qquad [10]$$

The force on the unit mass at P may be resolved into two components as functions of ϕ:

vertically upwards: $-\dfrac{\partial\Omega_p}{\partial a} = 2g\Delta_l(\cos^2\phi - \tfrac{1}{3})$ [11]

horizontally in the direction of increasing ϕ:

$$-\frac{\partial\Omega_p}{a\delta\phi} = -g\Delta_l\sin 2\phi \qquad [12]$$

For the moon :

$$\Delta_l = \frac{3}{2}\frac{m_l}{m_e}\left(\frac{a}{R_l}\right)^3 \qquad [13]$$

m_1 is the lunar mass and m_e is the Earth mass. R_1, the lunar distance, replaces r. The resulting forces are shown in **Figure 2.**

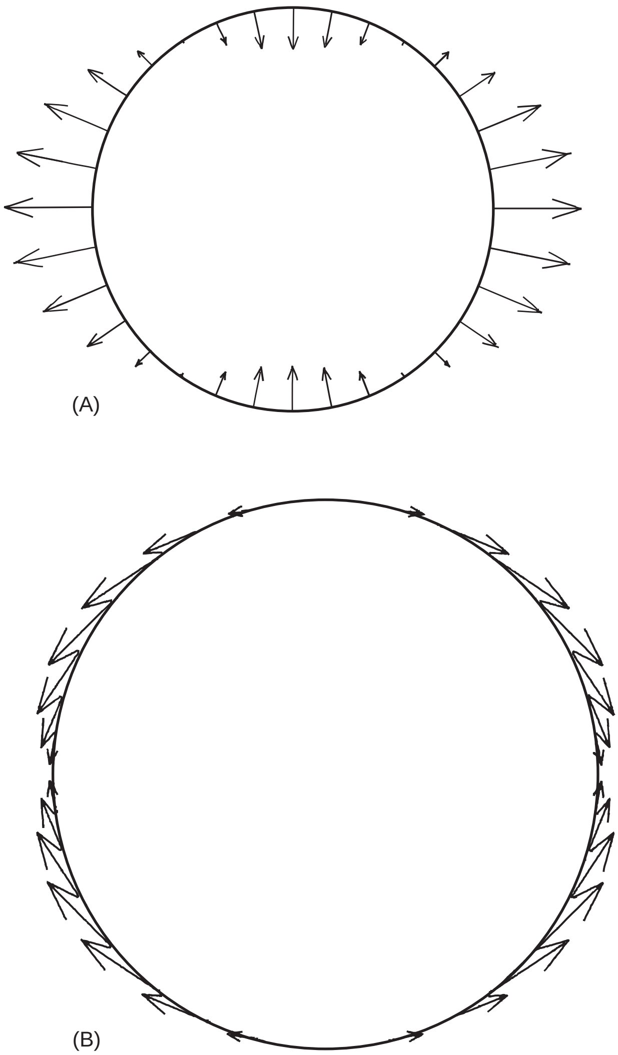

Figure 2 The tide-producing forces at the Earth's surface, due to the Moon. (A) The vertical forces, showing an outward pull at the equator and a smaller downward pull at the poles. (B) The horizontal forces, which are directed away from the poles towards the equator, with a maximum value at 45° latitude.

To generalize in three dimensions, the lunar angle ϕ must be expressed in suitable astronomical variables. These are chosen to be declination of the Moon north or south of the equator d_l, the north–south latitude of P, ϕ_p, and the hour angle of the moon, which is the difference in longitude between the meridian of P and the meridian of the sublunar point V on the Earth's surface.

The Equilibrium Tide

An equilibrium tide can be computed from eqn [10] by replacing $\cos^2\phi$ by the full astronomical expression in terms of d_1, ϕ_p and the hour angle C_1. The equilibrium tide is defined as the elevation of the sea surface that would be in equilibrium with the tidal forces if the Earth were covered with water and the response is instantaneous. It serves as an important reference system for tidal analysis.

It has three coefficients which characterize the three main species of tides: (1) the long period species; (2) the diurnal species at a frequency of one cycle per day (cos C); and (3) the semidiurnal species at two cycles per day (cos 2C).

The equilibrium tide due to the sun is expressed in a form analogous to the lunar tide, but with solar mass, distance, and declination substituted for lunar parameters.

The ratio of the two tidal amplitudes is:

$$\frac{m_s}{m_l}\left(\frac{R_l}{R_s}\right)^3$$

For the semidiurnal lunar tide at the equator when the lunar declination is zero, the equilibrium tidal amplitude is 0.27 m. For the sun it is 0.13 m. The solar amplitudes are smaller by a factor of 0.46 than those of the lunar tide, but the essential details are the same.

The maximum diurnal tidal ranges occur when the lunar declination is greatest. The ranges become very small when the declination is zero. This is because the effect of declination is to produce an asymmetry between the two high- and the two low-water levels observed as a point P rotates on the earth within the two tidal bulges.

The fortnightly spring/neap modulation of semidiurnal tidal amplitudes is due to the various combinations of the separate lunar and solar semidiurnal tides. At times of spring tides the lunar and solar forces combine together, but at neap tides the lunar and solar forces are out of phase and tend to cancel. In practice, the observed spring tides lag the maximum of the tidal forces, usually by one or two days due to the inertia of the oceans and energy losses. This delay is traditionally called the age of the tide.

The observed ocean tides are normally much larger than the equilibrium tide because of the dynamic response of the ocean to the tidal forces. But the observed tides do have their energy at the same frequencies (or periods) as the equilibrium tide. This forms the basis of tidal analysis.

Tidal Analysis

Tidal analysis of data collected by observations of sea levels and currents has two purposes. First, a good analysis provides the basis for predicting tides at future times, a valuable aid for shipping and other operations. Secondly, the results of analyses can be mapped and interpreted scientifically in terms of the hydrodynamics of the seas and their responses to tidal forcing. In tidal analysis the aim is to produce significant time-stable tidal parameters which describe the tidal regime at the place of observation. These parameters are often termed tidal constants on the assumption that the responses of the oceans and seas to tidal forces do not change with time. A good tidal analysis seeks to represent the data by a few significant stable numbers which mean something physically. In general, the longer the period of data included in the analysis, the greater the number of constants which can be independently determined. If possible, an analysis should give some idea of the confidence which should be attributed to each tidal constant determined.

The close relationship between the movement of the moon and sun, and the observed tides, make the lunar and solar coordinates a natural starting point for any analysis scheme. Three basic methods of tidal analysis have been developed. The first, which is now generally of only historical interest, the non-harmonic method, relates high and low water times and heights directly to the phases of the moon and other astronomical parameters. The second method which is generally used for predictions and for scientific work, harmonic analysis, treats the observed tides as the sum of a finite number of harmonic constituents with angular speeds determined from the astronomical arguments. The third method develops the concept, widely used in electronic engineering, of a frequency-dependent response of a system to a driving mechanism. For tides, the driving mechanism is the equilibrium potential. The latter two methods are special applications of the general formalisms of time series analysis.

Analyses of changing sea levels (scalar quantities) are obviously easier than those of currents (vectors), which can be analysed by resolving into two components.

Harmonic Analysis

The basis of harmonic analysis is the assumption that the tidal variations can be represented by a finite number N of harmonic terms of the form:

$$H_n \cos(\sigma_n t - g_n)$$

where H_n is the amplitude, g_n is the phase lag on the equilibrium tide at Greenwich and σ_n is the angular speed. The angular speeds σ_n are determined by an expansion of the equilibrium tide into harmonic terms. The speeds of these terms are found to have the general form:

$$\omega_n = i_a\omega_1 + i_b\omega_2 + i_c\omega_3 + (\omega_4, \omega_5, \omega_6 \text{ terms}) \quad [14]$$

Table 1 The basic astronomical periods which modulate the tidal forces

	Period	*Symbol*
Mean solar day (msd)	1.0000 msd	ω_0
Mean lunar day	1.0351 msd	ω_1
Sidereal month	27.3217 msd	ω_2
Tropical year	365.24222 msd	ω_3
Moon's perigee	8.85 years	ω_4
Regression of Moon's nodes	18.61 years	ω_5
Perihelion	20 942 years	ω_6

where the values of ω_1 to ω_6 are the angular speeds related to astronomical parameters and the coefficients, i_a to i_c are small integers (normally 0, 1 or 2) (**Table 1**).

The phase lags g_n are defined relative to the phase of the corresponding term in the harmonic expansion of the equilibrium tide.

Full harmonic analysis of the equilibrium tide shows the grouping of tidal terms into species (ω_1; diurnal, semidiurnal ...), groups (ω_2; monthly) and constituents (ω_3; annual).

Response Analysis

The basic ideas involved in response analysis are common to many activities. A system, sometimes called a 'black box', is subjected to an external stimulus or input. The output from a system depends on the input and the system response to that input. The response of the system may be evaluated by comparing the input and output functions at various forcing frequencies. These ideas are common in many different contexts, including mechanical engineering, financial modeling and electronics.

In tidal analysis the input is the equilibrium tidal potential. The tidal variations measured at a particular site may be considered as the output from the system. The system is the ocean, and we seek to describe its response to gravitational forces. This 'response' treatment has the conceptual advantage of clearly separating the astronomy (the input) from the oceanography (the black box). The basic response analysis assumes a linear system, but weak nonlinear interactions can be allowed for with extra terms.

Tidal Dynamics

The equilibrium tide consists of two symmetrical tidal bulges directly opposite the moon or sun. Semidiurnal tidal ranges would reach their maximum value of about 0.5 m at equatorial latitudes. The individual high water bulges would track around the earth, moving from east to west in steady progression. These theoretical characteristics are clearly *not* those of the observed tides.

The observed tides in the main oceans have much larger mean ranges, of about 1 m, but there are considerable variations. Times of tidal high water vary in a geographical pattern which bears no relationship to the simple ideas of a double bulge. The tides spread from the oceans onto the surrounding continental shelves, where even larger ranges are observed. In some shelf seas the spring tidal range may exceed 10 m: the Bay of Fundy, the Bristol Channel and the Argentine Shelf are well-known examples. Laplace (1749–1827) advanced the basic mathematical solutions for tidal waves on a rotating earth. More generally, the reasons for these complicated ocean responses to tidal forcing may be summarized as follows.

1. Movements of water on the surface of the earth must obey the physical laws represented by the hydrodynamic equations of continuity and momentum balance; this means that they must propagate as long waves. Any propagation of a wave, east to west around the earth, is impeded by the north–south continental boundaries.
2. Long waves travel at a speed that is related to the water depth; oceans are too shallow for this to match the tracking of the moon.
3. The various ocean basins have their individual natural modes of oscillation which influence their response to the tide-generating forces. There are many resonant frequencies. However, the whole global ocean system seems to be near to resonance at semidiurnal tidal frequencies, as the observed semidiurnal tides are generally much bigger than the diurnal tides.
4. Water movements are affected by the rotation of the earth. The tendency for water movement to maintain a uniform direction in absolute space means that it performs a curved path in the rotating frame of reference within which our observations are made.
5. The solid Earth responds elastically to the imposed gravitational tidal forces, and to the ocean tidal loading. The redistribution of water mass during the tidal cycle affects the gravitational field.

Long-Wave Characteristic, No Rotation

Provided that wave amplitudes are small compared with the depth, and that the depth is small compared with the wavelength, then the speed for the wave propagation is:

$$c = (gD)^{1/2} \qquad [15]$$

where g is gravitational acceleration, and D is the water depth. The currents u are related to the instantaneous level ζ by:

$$u = \zeta(g/D)^{1/2} \qquad [16]$$

Long waves have the special property that the speed c is independent of the frequency, and depends only on the value of g and the water depth; any disturbance which consists of a number of separate harmonic constituents will not change its shape as it propagates – this is nondispersive propagation. Waves at tidal periods are long waves, even in the deep ocean, and so their propagation is nondispersive. In the real ocean, tides cannot propagate endlessly as progressive waves. They undergo reflection at sudden changes of depth and at the coastal boundaries.

Standing Waves and Resonance

Two progressive waves traveling in opposite directions result in a wave motion, called a standing wave. This can happen where a wave is perfectly reflected at a barrier.

Systems which are forced by oscillations close to their natural period have large amplitude responses. The responses of oceans and many seas are close to semidiurnal resonance. In nature, the forced resonant oscillations cannot grow indefinitely because friction limits the response.

Because of energy losses, tidal waves are not perfectly reflected at the head of a basin, which means that the reflected wave is smaller than the ingoing wave. It is easy to show that this is equivalent to a progressive wave superimposed on a standing wave with the progressive wave carrying energy to the head of the basin. Standing waves cannot transmit energy because they consist of two progressive waves of equal amplitude traveling in opposite directions.

Long Waves on a Rotating Earth

A long progressive wave traveling in a channel on a rotating Earth behaves differently from a wave traveling along a nonrotating channel. The geostrophic forces that affect the motion in a rotating system, cause a deflection of the currents towards the right of the direction of motion in the Northern Hemisphere. The build-up of water on the right of the channel gives rise to a pressure gradient across the channel, which in turn develops until at equilibrium it balances the geostrophic force. The resulting Kelvin wave is described mathematically:

$$\zeta(y) = H_o \exp\left(\frac{-fy}{c}\right); \quad u(y) = \left(\frac{g}{D}\right)^{1/2}\zeta(y) \qquad [17]$$

where $\zeta(y)$ is the amplitude at a distance y from the right-hand boundary (in the Northern Hemisphere) and H_o is the amplitude of the wave at the boundary. The effect of the rotation appears only in the factor $\exp(-fy/c)$, which gives a decay of wave amplitude away from the boundary with a length scale of $c/f = [(gD)^{1/2}/f]$, which depends on the latitude and the water depth. This scale is called the Rossby radius of deformation. At a distance $y = c/f$ from the boundary the amplitude has fallen to $0.37\ H_o$. At 45°N in water of 4000 m depth the Rossby radius is 1900 km, but in water 50 m deep this is reduced to 215 km.

Kelvin waves are not the only solution to the hydrodynamic equations on a rotating Earth: a more general form, called Poincaré waves, gives amplitudes which vary sinusoidally rather than exponentially in the direction transverse to the direction of wave propagation.

The case of a standing-wave oscillation on a rotating Earth is of special interest in tidal studies. Away from the reflecting boundary, tidal waves can be represented by two Kelvin waves traveling in opposite directions. The wave rotates about a nodal point, which is called an amphidrome (**Figure 3**). The cotidal lines all radiate outwards from the amphidrome and the co-amplitude lines form a set of nearly concentric circles around the center at the amphidrome, at which the amplitude is zero. The amplitude is greatest around the boundaries of the basin.

Ocean Tides

Dynamically there are two essentially different types of tidal regime; in the wide and relatively deep ocean basins the observed tides are generated directly by the external gravitational forces; in the shelf seas the tides are driven by co-oscillation with the oceanic tides. The ocean response to the gravitational forcing may be described in terms of a forced linear oscillator, with weak energy dissipation.

A global chart of the principal lunar semidiurnal tidal constituent M_2 shows a complicated pattern of amphidromic systems. As a general rule these conform to the expected behavior for Kelvin wave propagation, with anticlockwise rotation in the Northern Hemisphere, and clockwise rotation in the Southern Hemisphere.

For example, in the Atlantic Ocean the most fully developed semidiurnal amphidrome is located near 50°N, 39°W. The tidal waves appear to travel around the position in a form which approximates to a Kelvin wave, from Portugal along the edge of the north-west European continental shelf towards Iceland, and thence west and south past Greenland to Newfoundland. There is a considerable leakage

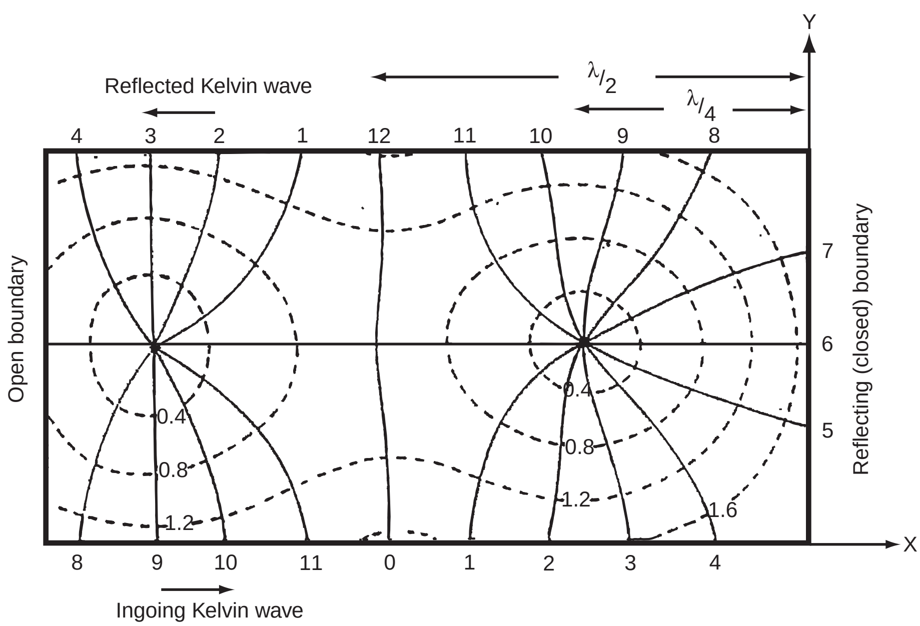

Figure 3 Cotidal and co-amplitude lines for a Kelvin wave reflected without energy loss in a rectangular channel. The incoming wave travels from left to right. Continuous lines are cotidal lines at intervals of 1/12 of a full cycle. Broken lines are lines of equal amplitude. Progression of the wave crests in the Northern Hemisphere is anticlockwise for both the amphidromic systems shown. In practice the reflected wave is weaker and so the amphidromes are moved towards the upper wall in the diagram.

of energy to the surrounding continental shelves and to the Arctic Ocean, so the wave reflected in a southerly direction, is weaker than the wave traveling northwards along the European coast.

The patterns of tidal waves on the continental shelf are scaled down as the wave speeds are reduced. In the very shallow water depths (typically less than 20 m) there are strong tidal currents and substantial energy losses due to bottom friction. Tidal waves are strongly influenced by linear Kelvin wave dynamics and by basin resonances. Energy is propagated to the shallow regions where it is dissipated.

Energy Fluxes and Budgets

The energy lost through tidal friction gradually slows down the rate of rotation of the earth, increasing the length of the day by one second in 41 000 years. Angular momentum of the earth–moon system is conserved by the moon moving away from the earth at 3.7 mm per year. The total rate of tidal energy dissipation due to the M_2 tide can be calculated rather exactly from the astronomic observations at 2.50 ± 0.05 TW, of which 0.1 TW is dissipated in the solid Earth. The total lunar dissipation is 3.0 TW, and the total due to both sun and moon is 4.0 TW. For comparison the geothermal heat loss is 30 TW, and the 1995 total installed global electric capacity was 2.9 TW. Solar radiation input is five orders of magnitude greater.

Most of the 2.4 TW of M_2 energy lost in the ocean is due to the work against bottom friction which opposes tidal currents. Because the friction increases approximately as the square of current speed, and the energy loses as the cube, tidal energy loses are concentrated in a few shelf areas of strong tidal currents. Notable among these are the northwest European Shelf, the Patagonian Shelf, the Yellow Sea, the Timor and Arafura Seas, Hudson Bay, Baffin Bay, and the Amazon Shelf. It now appears that up to 25% (1 TW) of the tidal energy may be dissipated by internal tidal waves in the deep ocean, where the dissipation processes contribute to vertical mixing and the breakdown of stratification. Again, energy losses may be concentrated in a few areas, for example where the rough topography of midocean ridges and island arcs create favorable conditions.

One of the main areas of tidal research is increasingly concentrated on gaining a better understanding of the many nonlinear ocean processes that are driven by this cascading tidal energy. Some examples, outlined below, are considered in more detail elsewhere in this Encyclopedia.

- Generation of tidal fronts in shelf seas, where the buoyancy forces due to tidal mixing compete with the buoyancy fluxes due to surface heating; the ratio of the water depth divided by the cube of the tidal current is a good indicator of the balance between the two factors, and fronts form along lines where this ratio reaches a critical value.
- River discharges to shallow seas near the mouth of rivers. The local input of freshwater buoyancy may be comparable to the buoyancy input from summer surface warming. These regions are called ROFIs (regions of freshwater influence). Spring–neap variations in the energy of tidal mixing strongly influence the circulation in these regions.
- The mixing and dispersion of pollutants often driven by the turbulence generated by tides. Maximum turbulent energy occurs some hours after the time of maximum tidal currents.
- Sediment processes of erosion and deposition are often controlled by varying tidal currents, particularly over a spring–neap cycle. The phase delay in suspended sediment concentration after maximum currents may be related to the phase lag in the turbulent energy, which has important consequences for sediment deposition and distribution.
- Residual circulation is partly driven by nonlinear responses to tidal currents in shallow water. Tidal flows also induce residual circulation around sandbanks because of the depth variations. In the Northern Hemisphere this circulation is observed to be in a clockwise sense. Near headlands and islands which impede tidal currents, residual eddies can cause marked asymmetry between the time and strength of the tidal ebb and flow currents.
- Tidal currents influence biological breeding patterns, migration, and recruitment. Some types of fish have adapted to changing tidal currents to assist in their migration: they lie dormant on the seabed when the currents are not favorable. Tidal mixing in shallow seas promotes productivity by returning nutrients to surface waters where light is available. Tidal fronts are known to be areas of high productivity. The most obvious example of tidal influence on biological processes is the zonation of species found at different levels along rocky shorelines.
- Evolution of sedimentary shores due to the dynamic equilibrium between waves, tides and other processes along sedimentary coasts which results in a wide range of features such as lagoons, sandbars, channels, and islands. These are very complicated processes which are still difficult to understand and model.
- Tidal amphidrome movements. The tidal amphidromes as shown in **Figure 4** only fall along the

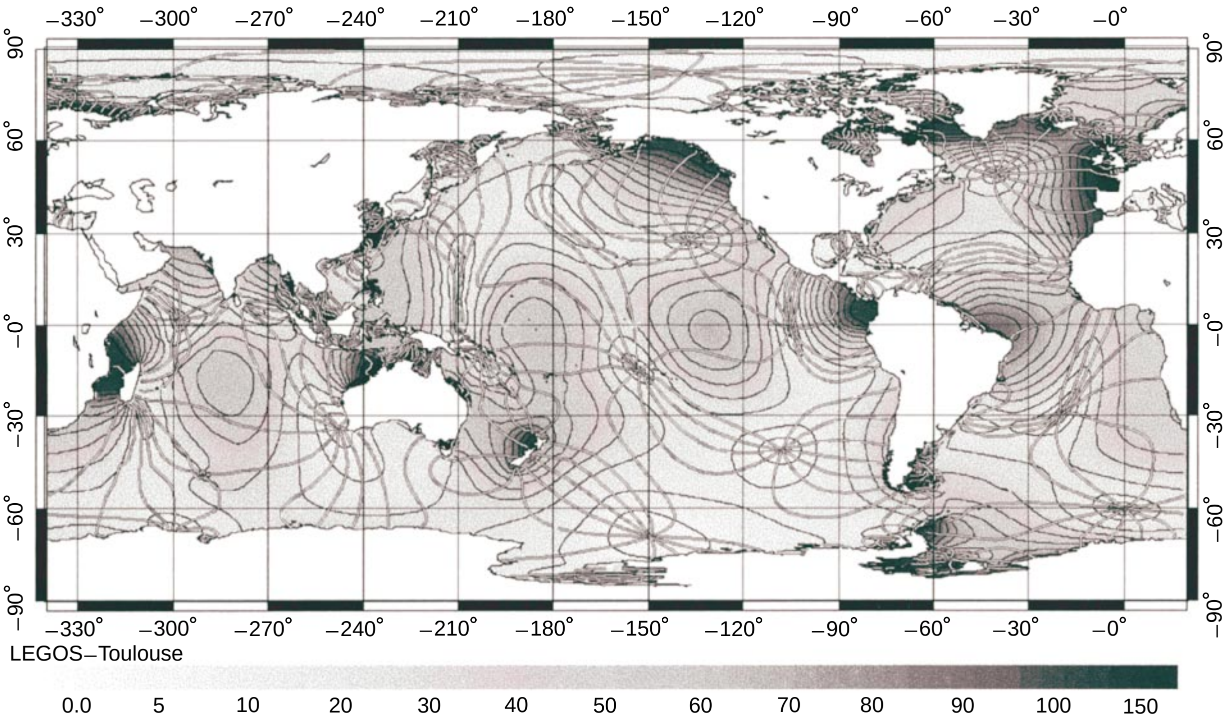

Figure 4 Map of the principal lunar semidiurnal tide produced by computer. Ocean tides observed by satellite and *in situ* now agree very closely with computer modeled tides. The dark areas show regions of high tidal amplitude. Note the convergence of cophase lines at amphidromes. In the Northern Hemisphere the tides normally progress in an anticlockwise sense around the amphidrome; in the Southern Hemisphere the progression is usually clockwise.

center line of the channel if the incoming tidal Kelvin wave is perfectly reflected. In reality, the reflected wave is weaker, and the amphidromes are displaced towards the side of the sea along which the outgoing tidal wave travels. Proportionately more energy is removed at spring tides than at neap tides, so the amphidromes can move by several tens of kilometers during the spring–neap cycle.

Nonlinear processes on the basic M_2 tide generate a series of higher harmonics, $M_4, M_6 \ldots$, with corresponding terms such as MS_4 for spring–neap interactions. A better understanding of the significance of the amplitudes and phases of these terms in the analyses of shallow-water tides and tidal phenomena will be an important tool in advancing our overall knowledge of the influence of tides on a wide range of ocean processes.

See also

Beaches, Physical Processes Affecting. Coastal Trapped Waves. Dispersion in Shallow Seas. Estuarine Circulation. Fish Migration, Horizontal. Geomorphology. Intertidal Fishes. Internal Tidal Mixing. Internal Waves. Lagoons. River Inputs. Salt Marshes and Mud Flats. Satellite Altimetry. Sea Level Change. Shelf-sea and Slope Fronts. Tidal Energy. Upper Ocean Vertical Structure. Waves on Beaches.

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