

crust may introduce some variability, older passive margins typically have heat flows similar to old oceanic crust, whereas young margins, such as in the Red Sea, have high heat flow.

Gas Hydrates

Gas hydrates, solid composites of biogenically derived gasses (mainly methane) combined with water ice, may be present in marine sediments, especially at continental margins, under the correct pressure and temperature conditions. Gas hydrates are detected by direct sampling, and inferred from seismic reflection data when a strong bottom-simulating reflector (BSR) is produced by the seismic velocity contrast between the gas hydrate and the sediment below. Given a depth (and thus pressure) of a bottom-simulating reflector, its temperature can be predicted from known relationships and a heat flow calculated. Alternatively, heat flow data can be used to estimate the bottom-simulating reflector temperature. The volume of hydrocarbons contained in marine gas hydrates is large. Changes in eustatic sea level (hence pressure) or bottom seawater temperatures could result in their release and thus increase greenhouse gasses and affect global climate.

Further Reading

- Humphris S, Mullineaux L, Zierenberg R and Thomson R (eds) (1995) Seafloor hydrothermal systems, physical, chemical, biological, and geological interactions. *Geophysical Monographs* 91: 425–445.
- Hyndman RD, Langseth MG and Von Herzen RP (1987) Deep Sea Drilling Project geothermal measurements: a review. *Review of Geophysics* 25: 1563–1582.
- Langseth MG, Jr and Von Herzen RP (1971) Heat flow through the floor of the world oceans. In: Maxwell AE (ed.) *The Sea*, vol. IV, part 1, pp. 299–352. New York: Wiley-Interscience.
- Lowell RP, Rona PA and Von Herzen RP (1995) Seafloor hydrothermal systems. *Journal of Geophysical Research* 100: 327–352.
- Parsons B and Sclater JG (1977) An analysis of the variation of ocean floor bathymetry and heat flow with age. *Journal of Geophysical Research* 82: 803–827.
- Pollack HN, Hurter SJ *et al.* (1993) Heat flow from the earth's interior: analysis of the global data set. *Review of Geophysics* 31: 267–280.
- Stein CA and Stein S (1992) A model for the global variation in oceanic depth and heat flow with lithospheric age. *Nature* 359: 123–129.
- Von Herzen RP (1987) Measurement of oceanic heat flow. In: Sammis C and Henyey T (eds) *Methods of Experimental Physics – Geophysics*, vol. 24, Part B, pp. 227–263. London: Academic Press.
- Wright JA and Loudon KE (eds) (1989) *CRC Handbook of Seafloor Heat Flow*. Boca Raton: CRC Press.

GLACIAL CRUSTAL REBOUND, SEA LEVELS AND SHORELINES

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Introduction

Geological, geomorphological, and instrumental records point to a complex and changing relation between land and sea surfaces. Elevated coral reefs or wave-cut rock platforms indicate that in some localities sea levels have been higher in the past, while observations elsewhere of submerged forests or flooded sites of human occupation attest to levels having been lower. Such observations are indicators of the relative change in the land and sea levels: raised shorelines are indicative of land having been uplifted or of the ocean volume having decreased, while submerged shorelines are a consequence of

land subsidence or of an increase in ocean volume. A major scientific goal of sea-level studies is to separate out these two effects.

A number of factors contribute to the instability of the land surfaces, including the tectonic upheavals of the crust emanating from the Earth's interior and the planet's inability to support large surface loads of ice or sediments without undergoing deformation. Factors contributing to the ocean volume changes include the removal or addition of water to the oceans as ice sheets wax and wane, as well as addition of water into the oceans from the Earth's interior through volcanic activity. These various processes operate over a range of timescales and sea level fluctuations can be expected to fluctuate over geological time and are recorded as doing so.

The study of such fluctuations is more than a scientific curiosity because its outcome impacts on a number of areas of research. Modern sea level

change, for example, must be seen against this background of geologically–climatologically driven change before contributions arising from the actions of man can be securely evaluated. In geophysics, one outcome of the sea level analyses is an estimate of the viscosity of the Earth, a physical property that is essential in any quantification of internal convection and thermal processes. Glaciological modeling of the behavior of large ice sheets during the last cold period is critically dependent on independent constraints on ice volume, and this can be extracted from the sea level information. Finally, as sea level rises and falls, so the shorelines advance and retreat. As major sea level changes have occurred during critical periods of human development, reconstructions of coastal regions are an important part in assessing the impact of changing sea levels on human movements and settlement.

Tectonics and Sea Level Change

Major causes of land movements are the tectonic processes that have shaped the planet's surface over geological time. Convection within the high-temperature viscous interior of the Earth results in stresses being generated in the upper, cold, and relatively rigid zone known as the lithosphere, a layer some 50–100 km thick that includes the crust. This convection drives plate tectonics — the movement of large parts of the lithosphere over the Earth's surface — mountain building, volcanism, and earthquakes, all with concomitant vertical displacements of the crust and hence relative sea level changes. The geological record indicates that these processes have been occurring throughout much of the planet's history. In the Andes, for example, Charles Darwin identified fossil seashells and petrified pine trees trapped in marine sediments at 4000 m elevation. In Papua New Guinea, 120 000-year-old coral reefs occur at elevations of up to 400 m above present sea level (Figure 1).

One of the consequences of the global tectonic events is that the ocean basins are being continually reshaped as mid-ocean ridges form or as ocean floor collides with continents. The associated sea level changes are global but their timescale is long, of the order 10^7 to 10^8 years, and the rates are small, less than 0.01 mm per year. Figure 2 illustrates the global sea level curve inferred for the past 600 million years from sediment records on continental margins. The long-term trends of rising and falling sea levels on timescales of 50–100 million years are attributed to these major changes in the ocean basin configurations. Superimposed on this are smaller-amplitude and shorter-period oscillations that reflect more re-



Figure 1 Raised coral reefs from the Huon Peninsula, Papua New Guinea. In this section the highest reef indicated (point 1) is about 340 m above sea level and is dated at about 125 000 years old. Elsewhere this reef attains more than 400 m elevation. The top of the present sea cliffs (point 2) is about 7000 years old and lies at about 20 m above sea level. The intermediate reef tops formed at times when the rate of tectonic uplift was about equal to the rate of sea level rise, so that prolonged periods of reef growth were possible. Photograph by Y. Ota.

gional processes such as large-scale volcanism in an ocean environment or the collision of continents. More locally, land is pushed up or down episodically in response to the deeper processes. The associated vertical crustal displacements are rapid, resulting in sea level rises or falls that may attain a few meters in amplitude, but which are followed by much longer periods of inactivity or even a relaxation of the original displacements. The raised reefs illustrated in Figure 1, for example, are the result of a large number of episodic uplift events each of typically a meter amplitude. Such displacements are mostly local phenomena, the Papua New Guinea example extending only for some 100–150 km of coastline.

The episodic but local tectonic causes of the changing position between land and sea can usually be identified as such because of the associated seismic activity and other tell-tale geological signatures. The development of geophysical models to describe these local vertical movements is still in a state of infancy and, in any discussion of global changes in sea level, information from such localities is best set aside in favor of observations from more stable environments.

Glacial Cycles and Sea Level Change

More important for understanding sea level change on human timescales than the tectonic contributions — important in terms of rates of change and in terms of their globality — is the change in ocean volume driven by cyclic global changes in climate from glacial to interglacial conditions. In Quaternary time, about the last two million years, glacial and interglacial conditions have followed each other on

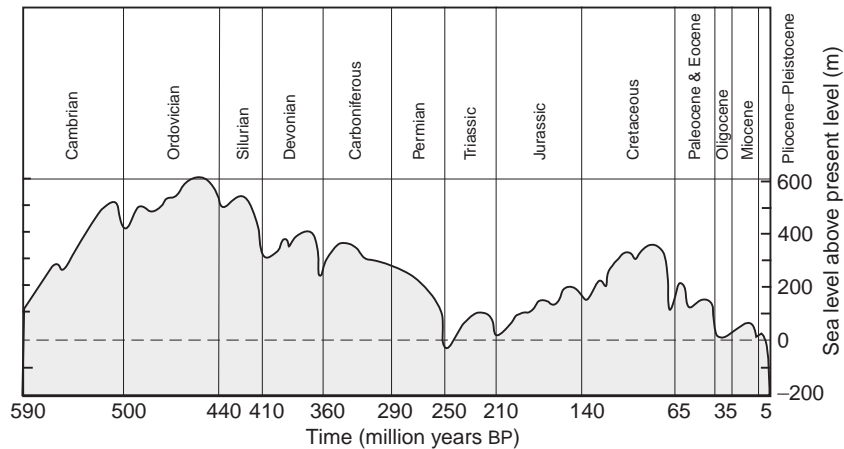


Figure 2 Global sea level variations through the last 600 million years estimated from seismic stratigraphic studies of sediments deposited at continental margins. Redrawn with permission from Hallam A (1984) Pre-Quaternary sea-level changes. *Annual Review of Earth and Planetary Science* 12: 205–243.

timescales of the order of 10^4 – 10^5 years. During interglacials, climate conditions were similar to those of today and sea levels were within a few meters of their present day position. During the major glacials, such as 20 000 years ago, large ice sheets formed in the northern hemisphere and the Antarctic ice sheet expanded, taking enough water out of the oceans to lower sea levels by between 100 and 150 m. **Figure 3** illustrates the changes in global sea level over the last 130 000 years, from the last interglacial, the last time that conditions were similar to those of today, through the Last Glacial Maximum and to the present. At the end of the last interglacial, at 120 000 years ago, climate began to fluctuate; increasingly colder conditions were

reached, ice sheets over North America and Europe became more or less permanent features of the landscape, sea levels reached progressively lower values, and large parts of today’s coastal shelves were exposed. Soon after the culmination of maximum glaciation, the ice sheets again disappeared, within a period of about 10 000 years, and climate returned to interglacial conditions.

The global changes illustrated in **Figure 3** are only one part of the sea level signal because the actual ice–water mass exchange does not give rise to a spatially uniform response. Under a growing ice sheet, the Earth is stressed; the load stresses are transmitted through the lithosphere to the viscous underlying mantle, which begins to flow away from the stressed area and the crust subsides beneath the ice load. When the ice melts, the crust rebounds. Also, the meltwater added to the oceans loads the seafloor and the additional stresses are transmitted to the mantle, where they tend to dissipate by driving flow to unstressed regions below the continents. Hence the seafloor subsides while the interiors of the continents rise, causing a tilting of the continental margins. The combined adjustments to the changing ice and water loads are called the glacio- and hydro-isostatic effects and together they result in a complex pattern of spatial sea level change each time ice sheets wax and wane.

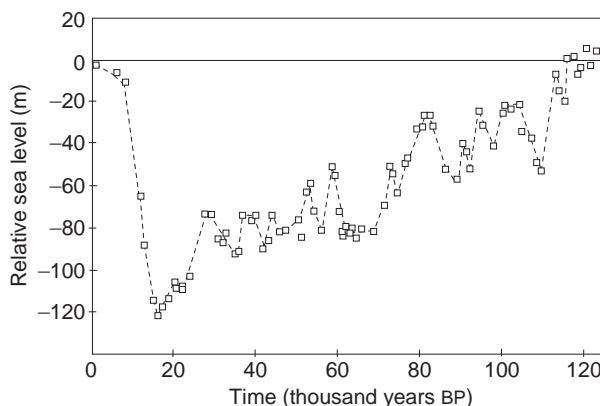


Figure 3 Global sea level variations (relative to present) since the time of the last interglacial 120 000–130 000 years ago when climate and environmental conditions were last similar to those of the last few thousand years. Redrawn with permission from Chapell J *et al.* (1996) Reconciliation of Late Quaternary sea level changes derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records. *Earth and Planetary Science Letters* 141b: 227–236.

Observations of Sea Level Change Since the Last Glacial Maximum

Evidence for the positions of past shorelines occurs in many forms. Submerged freshwater peats and tree stumps, tidal-dwelling mollusks, and archaeological sites would all point to a rise in relative sea level

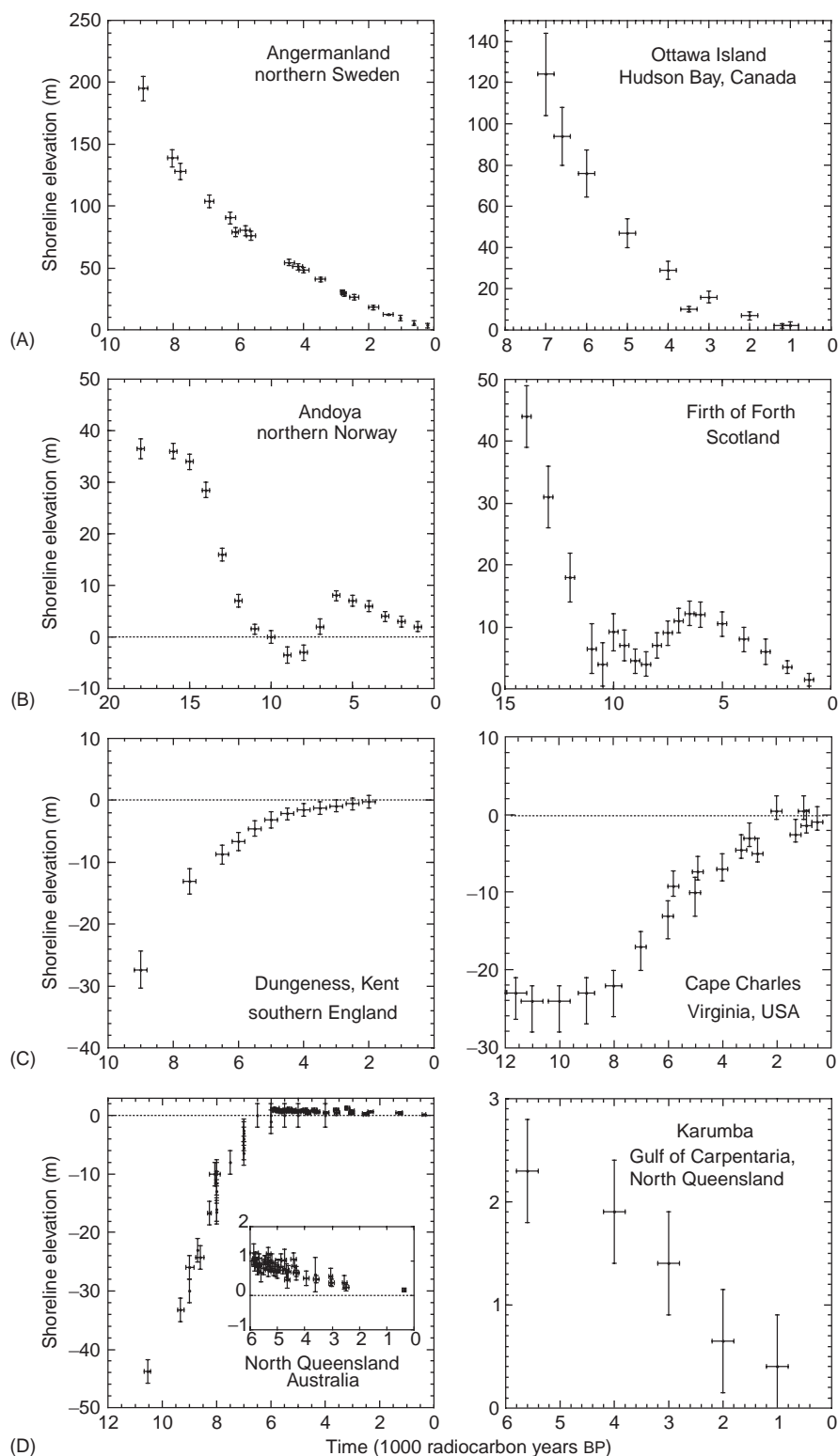


Figure 4 Characteristic sea level curves observed in different localities (note the different scales used). (A) From Ångermanland, Gulf of Bothnia, Sweden, and from Hudson Bay, Canada. Both sites lie close to centers of former ice sheet, over northern Europe and North America respectively. (B) From the Atlantic coast of Norway and the west coast of Scotland. These sites are close to former ice-sheet margins at the time of the last glaciation. (C) From southern England and Virginia, USA. These sites lie outside the areas of former glaciation and at a distance from the margins where the crust is subsiding in response to the melting of the nearby icesheet. (D) Two sites in northern Australia, one for the Coral Sea coast of Northern Queensland and the second from the Gulf of Carpentaria some 200 km away but on the other side of the Cape York Peninsula. At both localities sea level rose until about 6000 years ago before peaking above present level.

since the time of growth, deposition, or construction. Raised coral reefs, such as in **Figure 1**, whale bones cemented in beach deposits, wave-cut rock platforms and notches, or peats formed from saline-loving plants would all be indicative of a falling sea level since the time of formation. To obtain useful sea level measurements from these data requires several steps: an understanding of the relationship between the feature's elevation and mean sea level at the time of growth or deposition, a measurement of the height or depth with respect to present sea level, and a measurement of the age. All aspects of the observation present their own peculiar problems but over recent years a substantial body of observational evidence has been built up for sea level change since the last glaciation that ended at about 20 000 years ago. Some of this evidence is illustrated in **Figure 4**, which also indicates the very significant spatial variability that may occur even when, as is the case here, the evidence is from sites that are believed to be tectonically stable.

In areas of former major glaciation, raised shorelines occur with ages that are progressively greater with increasing elevation and with the oldest shorelines corresponding to the time when the region first became ice-free. Two examples, from northern Sweden and Hudson Bay in Canada, respectively, are illustrated in **Figure 4A**. In both cases sea level has been falling from the time the area became ice-free and open to the sea. This occurred at about 9000 years ago in the former case and about 7000 years later in the second case. Sea level curves from localities just within the margins of the former ice sheet are illustrated in **Figure 4B**, from western Norway and Scotland, respectively. Here, immediately after the area became ice-free, the sea level fell but a time was reached when it again rose. A local maximum was reached at about 6000 years ago, after which the fall continued up until the present. Farther away from the former ice margins the observed sea level pattern changes dramatically, as is illustrated in **Figure 4C**. Here the sea level initially rose rapidly but the rate decreased for the last 7000 years up to the present. The two examples illustrated, from southern England and the Atlantic coast of the United States, are representative of tectonically stable localities that lie within a few thousand kilometers from the former centers of glaciation. Much farther away again from the former ice sheets the sea level signal undergoes a further small but significant change in that the present level was first reached at about 6000 years ago and then exceeded by a small amount before returning to its present value. The two examples illustrated in **Figure 4D** are from nearby localities in northern Aus-

tralia. At both sites present sea level was reached about 6000 years ago after a prolonged period of rapid rise with resulting highstands that are small in amplitude but geographically variable.

The examples illustrated in **Figure 4** indicate the rich spectrum of variation in sea level that occurred when the last large ice sheets melted. Earlier glacial cycles resulted in a similar spatial variability, but much of the record before the Last Glacial Maximum has been overwritten by the effects of the last deglaciation.

Glacio-Hydro-Isostatic Models

If, during the decay of the ice sheets, the meltwater volume was distributed uniformly over the oceans, then the sea level change at time t would be

$$\Delta\zeta_c(t) = (\text{change in ice volume/ocean surface area}) \times \rho_i/\rho_w \quad [1a]$$

where ρ_i and ρ_w are the densities of ice and water, respectively. (See end of article – symbols used.) This term is the ice-volume equivalent sea level and it provides a measure of the change in ice volume through time. Because the ocean area changes with time a more precise definition is

$$\Delta\zeta_c(t) = \frac{dV_i}{dt} \frac{\rho_i}{\rho_0} \int_t \frac{\rho_i}{A_0(t)} \quad [1b]$$

where A_0 is the area of the ocean surface, excluding areas covered by grounded ice. The sea level curve illustrated in **Figure 3** is essentially this function. However, it represents only a zero-order approximation of the actual sea level change because of the changing gravitational field and deformation of the Earth.

In the absence of winds or ocean currents, the ocean surface is of constant gravitational potential. A planet of a defined mass distribution has a family of such surfaces outside it, one of which – the geoid – corresponds to mean sea level. If the mass distribution on the surface (the ice and water) or in the interior (the load-forced mass redistribution in the mantle) changes, so will the gravity field, the geoid, and the sea level change. The ice sheet, for example, represents a large mass that exerts a gravitational pull on the ocean and, in the absence of other factors, sea level will rise in the vicinity of the ice sheet and fall farther away. At the same time, the Earth deforms under the changing load, with two consequences: the land surface with respect to

which the level of the sea is measured is time-dependent, as is the gravitational attraction of the solid Earth and hence the geoid.

The calculation of the change of sea level resulting from the growth or decay of ice sheets therefore involves three steps: the calculation of the amount of water entering into the ocean and the distribution of this meltwater over the globe; the calculation of the deformation of the Earth's surface; and the calculation of the change in the shape of the gravitational equipotential surfaces. In the absence of vertical tectonic motions of the Earth's surface, the relative sea level change $\Delta\zeta(\varphi, t)$ at a site φ and time t can be written schematically as

$$\Delta\zeta(\varphi, t) = \Delta\zeta_e(t) + \Delta\zeta_1(\varphi, t) \quad [2]$$

where $\Delta\zeta_1(\varphi, t)$ is the combined perturbation from the uniform sea level rise term [1]. This is referred to as the isostatic contribution to relative sea level change.

In a first approximation, the Earth's response to a global force is that of an elastic outer spherical layer (the lithosphere) overlying a viscous or viscoelastic mantle that itself contains a fluid core. When subjected to an external force (e.g., gravity) or a surface load (e.g., an ice cap), the planet experiences an instantaneous elastic deformation followed by a time-dependent or viscous response with a characteristic timescale(s) that is a function of the viscosity. Such behavior of the Earth is well documented by other geophysical observations: the gravitational attraction of the Sun and Moon raises tides in the solid Earth; ocean tides load the seafloor with a time-dependent water load to which the Earth's surface responds by further deformation; atmospheric pressure fluctuations over the continents induce deformations in the solid Earth. The displacements, measured with precision scientific instruments, have both an elastic and a viscous component, with the latter becoming increasingly important as the duration of the load or force increases. These loads are much smaller than the ice and water loads associated with the major deglaciation, the half-daily ocean tide amplitudes being only 1% of the glacial sea level change, and they indicate that the Earth will respond to even small changes in the ice sheets and to small additions of meltwater into the oceans.

The theory underpinning the formulation of planetary deformation by external forces or surface loads is well developed and has been tested against a range of different geophysical and geological observations. Essentially, the theory is one of formulating the response of the planet to a point load and

then integrating this point-load solution over the load through time, calculating at each epoch the surface deformation and the shape of the equipotential surfaces, making sure that the meltwater is appropriately distributed into the oceans and that the total ice-water mass is preserved. Physical inputs into the formulation are the time-space history of the ice sheets, a description of the ocean basins from which the water is extracted or into which it is added, and a description of the rheology, or response parameters, of the Earth. For the last requirement, the elastic properties of the Earth, as well as the density distribution with depth, have been determined from seismological and geodetic studies. Less well determined are the viscous properties of the mantle, and the usual procedure is to adopt a simple parametrization of the viscosity structure and to estimate the relevant parameters from analyses of the sea level change itself.

The formulation is conveniently separated into two parts for schematic reasons: the glacio-isostatic effect representing the crustal and geoid displacements due to the ice load, and the hydro-isostatic effect due to the water load. Thus

$$\Delta\zeta(\varphi, t) = \Delta\zeta_e(t) + \Delta\zeta_i(\varphi, t) + \Delta\zeta_w(\varphi, t) \quad [3]$$

where $\Delta\zeta_i(\varphi, t)$ is the glacio-isostatic and $\Delta\zeta_w(\varphi, t)$ the hydro-isostatic contribution to sea level change. (In reality the two are coupled and this is included in the formulation.) If $\Delta\zeta_i(\varphi, t)$ is evaluated everywhere, the past water depths and land elevations $H(\varphi, t)$ measured with respect to coeval sea level are given by

$$H(\varphi, t) = H_0(\varphi) - \Delta\zeta(\varphi, t) \quad [4]$$

where $H_0(\varphi)$ is the present water depth or land elevation at location φ .

A frequently encountered concept is eustatic sea level, which is the globally averaged sea level at any time t . Because of the deformation of the seafloor during and after the deglaciation, the isostatic term $\Delta\zeta(\varphi, t)$ is not zero when averaged over the ocean at any time t , so that the eustatic sea level change is

$$\Delta\zeta_{\text{eus}}(t) = \Delta\zeta_e(t) + \langle \Delta\zeta_1\{\varphi, t\} \rangle_0$$

where the second term on the right-hand side denotes the spatially averaged isostatic term. Note that $\Delta\zeta_e(t)$ relates directly to the ice volume, and not $\Delta\zeta_{\text{eus}}(t)$.

The Anatomy of the Sea Level Function

The relative importance of the two isostatic terms in eqn. [3] determines the spatial variability in the sea level signal. Consider an ice sheet of radius that is much larger than the lithospheric thickness: the limiting crustal deflection beneath the center of the load is $I\rho_i/\rho_m$ where I is the maximum ice thickness and ρ_m is the upper mantle density. This is the local isostatic approximation and it provides a reasonable approximation of the crustal deflection if the loading time is long compared with the relaxation time of the mantle. Thus for a 3 km thick ice sheet the maximum deflection of the crust can reach 1 km, compared with a typical ice volume for a large ice sheet that raises sea level globally by 50–100 m. Near the centers of the formerly glaciated regions it is the crustal rebound that dominates and sea level falls with respect to the land.

This is indeed observed, as illustrated in **Figure 4A**, and the sea level curve here consists of essentially the sum of two terms, the major glacio-isostatic term and the minor ice-volume equivalent sea level term $\Delta\zeta_e(t)$ (**Figure 5A**). Of note is that the rebound continues long after all ice has disappeared, and this is evidence that the mantle response includes a memory of the earlier load. It is the decay time of this part of the curve that determines the mantle viscosity. As the ice margin is approached, the local ice thickness becomes less and the crustal rebound is reduced and at some stage is equal, but of opposite sign, to $\Delta\zeta_e(t)$. Hence sea level is constant for a period (**Figure 5B**) before rising again when the rebound becomes the minor term. After global melting has ceased, the dominant signal is the late stage of the crustal rebound and levels fall up to the present. Thus the oscillating sea level curves observed in areas such as Norway and Scotland (**Figure 4B**) are essentially the sum of two effects of similar magnitude but opposite sign. The early part of the observation contains information on earth rheology as well as on the local ice thickness and the globally averaged rate of addition of meltwater into the oceans. Furthermore, the secondary maximum is indicative of the timing of the end of global glaciation and the latter part of the record is indicative mainly of the mantle response.

At the sites beyond the ice margins it is the meltwater term $\Delta\zeta_e(t)$ that is important, but it is not the sole factor. When the ice sheet builds up, mantle material flows away from the stressed mantle region and, because flow is confined, the crust around the periphery of the ice sheet is uplifted. When the ice sheet melts, subsidence of the crust occurs in

a broad zone peripheral to the original ice sheet and at these locations the isostatic effect is one of an apparent subsidence of the crust. This is illustrated in **Figure 5C**. Thus, when the ocean volumes have stabilized, sea level continues to rise, further indicating that the planet responds viscously to the changing surface loads. The early part of the observational record (e.g., **Figure 4C**) is mostly indicative of the rate at which meltwater is added into the ocean, whereas the latter part is more indicative of mantle viscosity.

In all of the examples considered so far it is the glacial-rebound that dominates the total isostatic adjustment, the hydro-isostatic term being present but comparatively small. Consider an addition of water that raises sea level by an amount D . The local isostatic response to this load is $D\rho_w/\rho_m$, where ρ_w is the density of ocean water. This gives an upper limit to the amount of subsidence of the sea floor of about 30 m for a 100 m sea level rise. This is for the middle of large ocean basins and at the margins the response is about half as great. Thus the hydro-isostatic effect is significant and is the dominant perturbing term at margins far from the former ice sheets. This occurs at the Australian margin, for example, where the sea level signal is essentially determined by $\Delta\zeta_e(t)$ and the water-load response (**Figure 4D**). Up to the end of melting, sea level is dominated by $\Delta\zeta_e(t)$ but thereafter it is determined largely by the water-load term $\Delta\zeta_w(\varphi, t)$, such that small highstands develop at 6000 years. The amplitudes of these highstands turn out to be strongly dependent on the geometry of the water-load distribution around the site: for narrow gulfs, for example, their amplitude increases with distance from the coast and from the water load at rates that are particularly sensitive to the mantle viscosity.

While the examples in **Figure 5** explain the general characteristics of the global spatial variability of the sea level signal, they also indicate how observations of such variability are used to estimate the physical quantities that enter into the schematic model (3). Thus, observations near the center of the ice sheet partially constrain the mantle viscosity and central ice thickness. Observations from the ice sheet margin partially constrain both the viscosity and local ice thickness and establish the time of termination of global melting. Observations far from the ice margins determine the total volumes of ice that melted into the oceans as well as providing further constraints on the mantle response. By selecting data from different localities and time intervals, it is possible to estimate the various parameters that underpin the sea level eqn. [3] and to use these models to predict sea level and shoreline change for

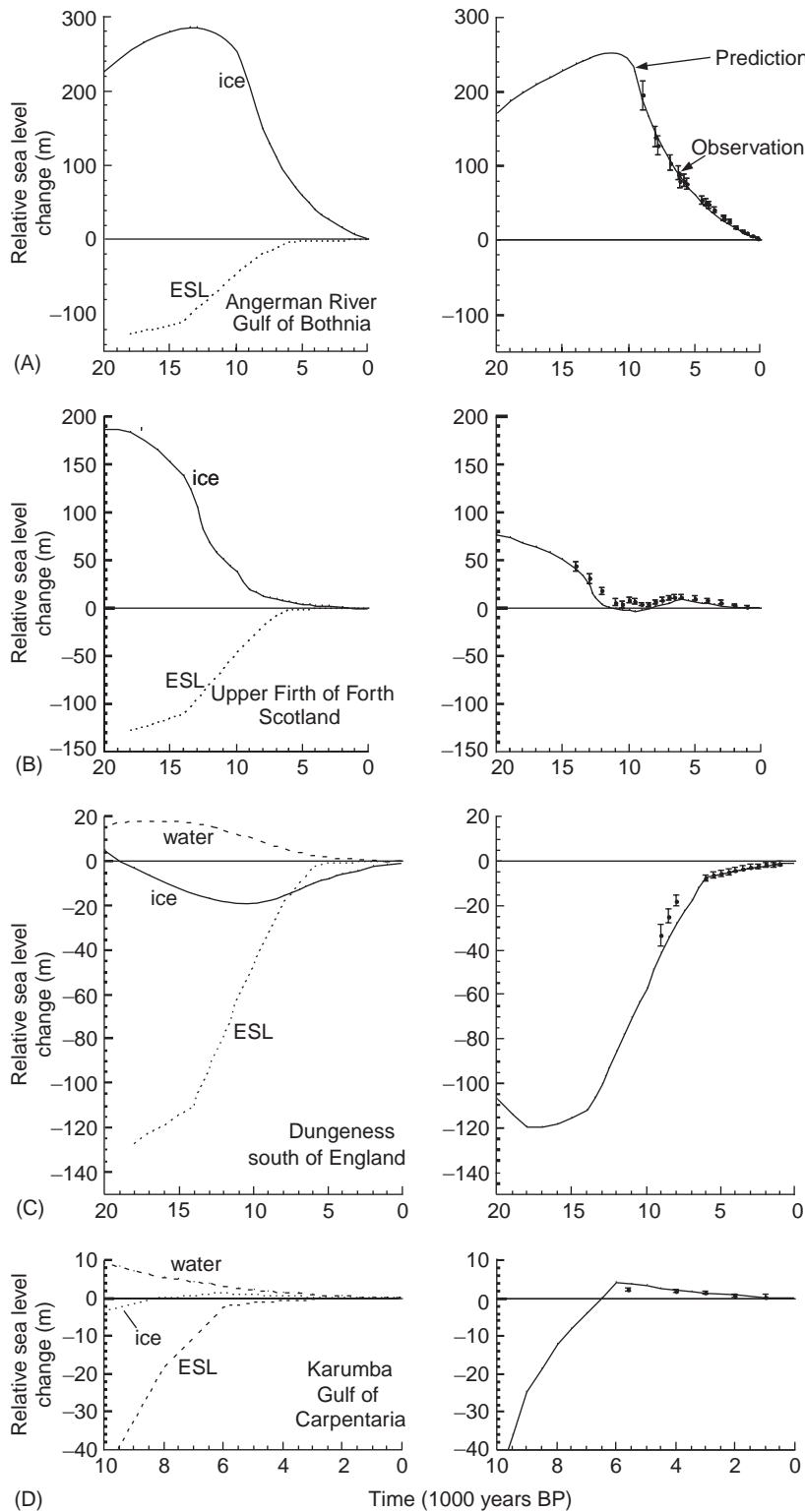


Figure 5 Schematic representation of the sea level curves in terms of the ice-volume equivalent function $\Delta\zeta_e(t)$ (denoted by ESL) and the glacio- and hydro-isostatic contributions $\Delta\zeta_i(\varphi, t)$, $\Delta\zeta_w(\varphi, t)$ (denoted by ice and water, respectively). The panels on the left indicate the predicted individual components and the panels on the right indicate the total predicted change compared with the observed values (data points). (A) For the sites at the ice center where $|\Delta\zeta_i(\varphi, t)| \gg |\Delta\zeta_e(t)| \gg |\Delta\zeta_w(\varphi, t)|$. (B) For sites near but within the ice margin where $|\Delta\zeta_i(\varphi, t)| \sim |\Delta\zeta_e(t)| \gg |\Delta\zeta_w(\varphi, t)|$, but the first two terms are of opposite sign. (C) For sites beyond the ice margin where $|\Delta\zeta_e(t)| > |\Delta\zeta_i(\varphi, t)| \gg |\Delta\zeta_w(\varphi, t)|$. (D) For sites at continental margins far from the former ice sheets where $|\Delta\zeta_w(\varphi, t)| > |\Delta\zeta_i(t)|$. Adapted with permission from Lambeck K and Johnston P (1988). The viscosity of the mantle: evidence from analyses of glacial rebound phenomena. In: Jackson ISN (ed.) *The Earth's Mantle — Composition, Structure and Evolution*, pp. 461–502. Cambridge: Cambridge University Press.

unobserved areas. Analyses of sea-level change from different regions of the world lead to estimates for the lithospheric thickness of between 60–100 km, average upper mantle viscosity (from the base of the lithosphere to a depth of 670 km) of about $(1-5)10^{20}$ Pa s and an average lower mantle viscosity of about $(1-5)10^{22}$ Pa s. Some evidence exists that these parameters vary spatially; lithosphere thickness and upper mantle viscosity being lower beneath oceans than beneath continents.

Sea Level Change and Shoreline Migration: Some Examples

Figure 6 illustrates the ice-volume equivalent sea level function $\Delta\zeta_e(t)$ since the time of the last maximum glaciation. This curve is based on sea level indicators from a number of localities, all far from the former ice sheets, with corrections applied for the isostatic effects. The right-hand axis indicates the corresponding change in ice volume (from the relation [1b]). Much of this ice came from the ice sheets over North America and northern Europe

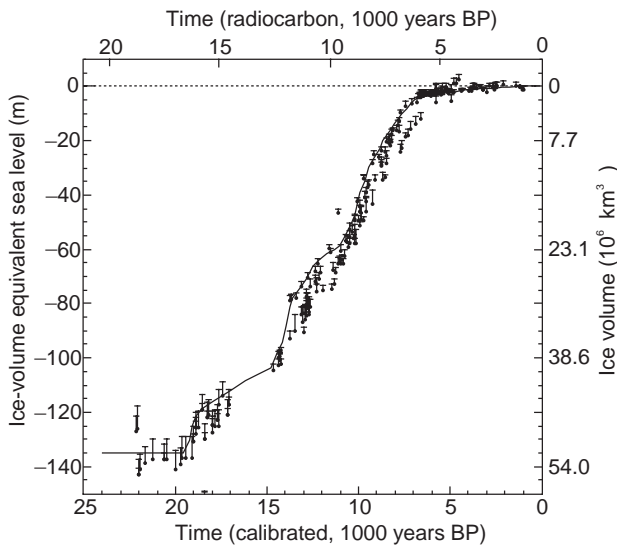


Figure 6 The ice-volume equivalent function $\Delta\zeta_e(t)$ and ice volumes since the time of the last glacial maximum inferred from corals from Barbados, from sediment facies from north-western Australia, and from other sources for the last 7000 years. The actual sea level function lies at the upper limit defined by these observations (continuous line). The upper time scale corresponds to the radiocarbon timescale and the lower one is calibrated to calendar years. Adapted with permission from Fleming K *et al.* (1998) Refining the eustatic sea-level curve since the LGM using the far- and intermediate-field sites. *Earth and Planetary Science Letters* 163: 327–342; and Yokoyama Y *et al.* (2000) Timing of the Last Glacial Maximum from observed sea-level minimum. *Nature* 406: 713–716.

but a not insubstantial part also originated from Antarctica. Much of the melting of these ice sheets occurred within 10 000 years, at times the rise in sea level exceeding 30 mm per year, and by 6000 years ago most of the deglaciation was completed. With this sea level function, individual ice sheet models, the formulation for the isostatic factors, and a knowledge of the topography and bathymetry of the world, it becomes possible to reconstruct the paleo shorelines using the relation [4].

Scandinavia is a well-studied area for glacial rebound and sea level change since the time the ice retreat began about 18 000 years ago. The observational evidence is quite plentiful and a good record of ice margin retreat exists. Figure 7 illustrates examples for two epochs. The first (Figure 7A), at 16 000 years ago, corresponds to a time after onset of deglaciation. A large ice sheet existed over Scandinavia with a smaller one over the British Isles. Globally sea level was about 110 m lower than now and large parts of the present shallow seas were exposed, for example, the North Sea and the English Channel but also the coastal shelf farther south such as the northern Adriatic Sea. The red and orange contours indicate the sea level change between this period and the present. Beneath the ice these rebound contours are positive, indicating that if shorelines could form here they would be above sea level today. Immediately beyond the ice margin, a broad but shallow bulge develops in the topography, which will subside as the ice sheet retreats. At the second epoch selected (Figure 7B) 10 500 years ago, the ice has retreated and reached a temporary halt as the climate briefly returned to colder conditions; the Younger Dryas time of Europe. Much of the Baltic was then ice-free and a freshwater lake developed at some 25–30 m above coeval sea level. The flooding of the North Sea had begun in earnest. By 9000 years ago, most of the ice was gone and shorelines began to approach their present configuration.

The sea level change around the Australian margin has also been examined in some detail and it has been possible to make detailed reconstructions of the shoreline evolution there. Figure 8 illustrates the reconstructions for northern Australia, the Indonesian islands, and the Malay–IndoChina peninsula. At the time of the Last Glacial Maximum much of the shallow shelves were exposed and deeper depressions within them, such as in the Gulf of Carpentaria or the Gulf of Thailand, would have been isolated from the open sea. Sediments in these depressions will sometimes retain signatures of these pre-marine conditions and such data provide important constraints on the models of sea level change.

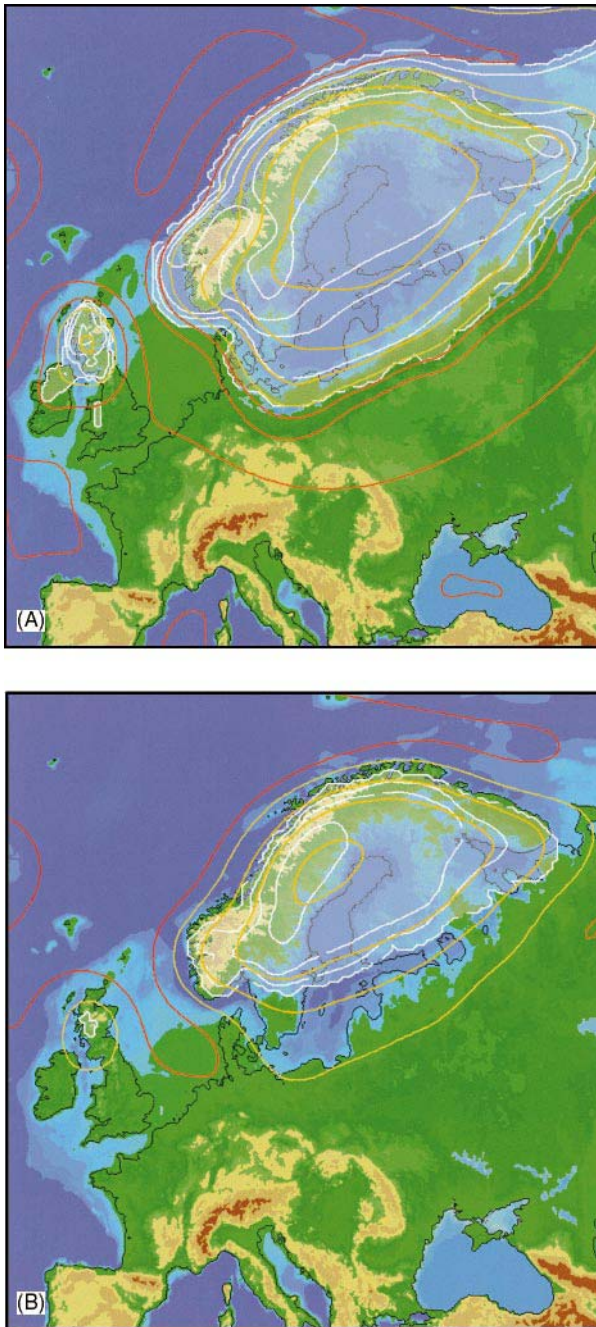


Figure 7 Shoreline reconstructions for Europe (A) at 16000 and (B) at 10500 years ago. The contours are at 400m intervals for the ice thickness (white) and 25m for the water depths less than 100m. The orange, yellow, and red contours are the predicted lines of equal sea level change from the specified epoch to the present and indicate where shorelines of these epochs could be expected if conditions permitted their formation and preservation. The zero contour is in yellow; orange contours, at intervals of 100m, are above present, and red contours, at 50m intervals, are below present. At 10500 years ago the Baltic is isolated from the Atlantic and its level lies about 25m above that of the latter.

Part of the information illustrated in **Figure 6**, for example, comes from the shallow depression on the Northwest Shelf of Australia. By 12000 years ago

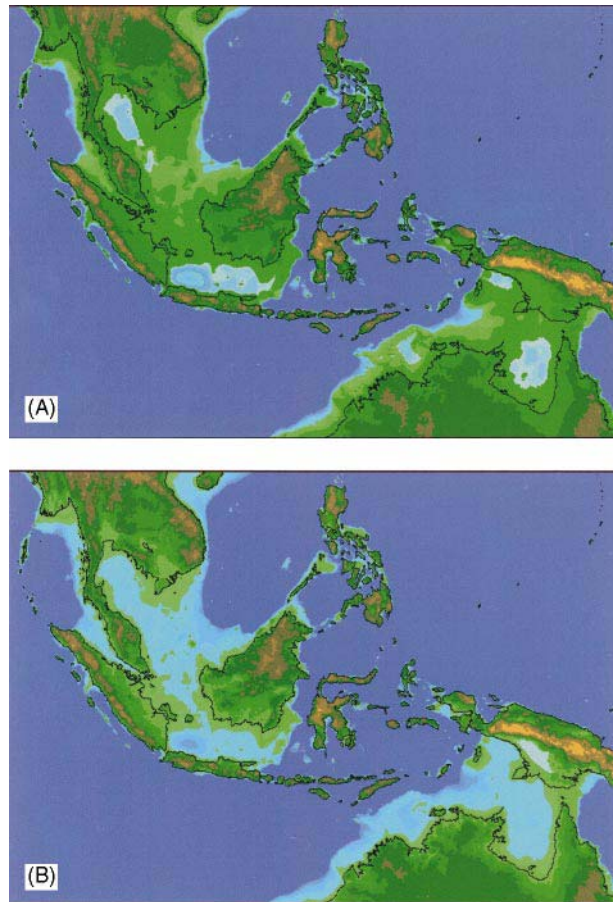


Figure 8 Shoreline reconstructions for South East Asia and northern Australia at the time of the last glacial maximum, (A) at about 18000 years ago and (B) at 12000 years ago. The water depth contours are the same as in **Figure 7**. The water depths in the inland lakes at 18000 years ago are contoured at 25m intervals relative to their maximum levels that can be attained without overflowing.

the sea has begun its encroachment of the shelves and the inland lakes were replaced by shallow seas.

Symbols used

- $\Delta\zeta_e(t)$ Ice-volume equivalent sea level. Uniform change in sea level produced by an ice volume that is distributed uniformly over the ocean surface.
- $\Delta\zeta_i(t)$ Perturbation in sea level due to glacio-isostatic $\Delta\zeta_i(t)$ and hydro-isostatic $\Delta\zeta_w(t)$ effects.
- $\Delta\zeta_{eus}(t)$ Eustatic sea level change. The globally averaged sea level at time t .

See also

Beaches, Physical Processes affecting. Coral Reefs. Fiordic Ecosystems. Geomorphology.

Lagoons. Mangroves. Past Climate From Corals. Rocky Shores. Salt Marshes and Mud Flats. Salt Marsh Vegetation. Sandy Beaches, Biology of. Sea Level Change.

Further Reading

- Lambeck K (1988) *Geophysical Geodesy: The Slow Deformations of the Earth*. Oxford: Oxford University Press.
- Lambeck K and Johnston P (1999) The viscosity of the mantle: evidence from analysis of glacial-rebound phenomena. In: Jackson ISN (ed.) *The Earth's Mantle - Composition, Structure and Evolution*, pp. 461–502. Cambridge: Cambridge University Press.
- Lambeck K, Smither C and Johnston P (1998) Sea-level change, glacial rebound and mantle viscosity for northern Europe. *Geophysical Journal International* 134: 102–144.
- Peltier WR (1998) Postglacial variations in the level of the sea: implications for climate dynamics and solid-earth geophysics. *Reviews in Geophysics* 36: 603–689.
- Pirazzoli PA (1991) *World Atlas of Holocene Sea-Level Changes*. Amsterdam: Elsevier.
- Plassche O van de (ed.) (1986) *Sea-Level Research: A Manual for the Collection and Evaluation of Data*. Norwich: Geo Books.
- Sabadini R, Lambeck K and Boschi E (1991) *Glacial Isostasy, Sea Level and Mantle Rheology*. Dordrecht: Kluwer.

GLOBAL MARINE POLLUTION

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Selected topics on pollution of the oceans are covered in detail by articles listed in the cross-references at the end. This brief overview is an introduction and guide to the appended list of Further Reading, which is designed to provide an entry to the historical and global context of these issues.

Until recently the size and mobility of the oceans encouraged the view that they could not be significantly affected by human activities. Fresh water lakes and rivers had been degraded for centuries by effluents, particularly sewage, but, although from the 1920s coastal oil pollution from shipping discharges was widespread (Pritchard, 1987), it was felt that in general the sea could safely dilute and disperse anything added to it. Erosion of this view began in the 1950s, when fallout from the testing of nuclear weapons in the atmosphere resulted in enhanced levels of artificial radionuclides throughout the world's oceans (Park *et al.*, 1983). At about the same time, the effluent from a factory at Minamata in Japan caused illness and deaths from consumption of mercury-contaminated fish (Kutsuna, 1968), focusing global attention on the potential dangers of toxic metals. In the early 1960s, build-up in the marine environment of residues from synthetic organic pesticides poisoned top predators such as fish-eating birds (Tolba *et al.*, 1992), and in 1967 the first wreck of a supertanker, the *Torrey Canyon*, highlighted the threat of oil from shipping accidents,

as distinct from operational discharges (Eighth Report of the Royal Commission on Environmental Pollution, 1981).

It might therefore be said that the decades of the 1950s and 1960s saw the beginnings of marine pollution as a serious concern, and one that demanded widespread control. It attracted the efforts of national and international agencies, not least those of the United Nations. The fear of effects of radioactivity focused early attention, and initiated the establishment of the International Commission on Radiological Protection (ICRP) (Brackley, 1990), which produced a set of radiation protection standards, applicable not just to fallout from weapons testing but also to the increasingly more relevant issues of operational discharges from nuclear reactors and reprocessing plants, from disposal of low-level radioactive material from a variety of sources including research and medicine, and from accidents in industrial installations and nuclear-powered ships. Following Minamata, other metals, in particular cadmium and lead, joined mercury on the list of concerns. Since this, like radioactive wastes, was seen as a public health problem, immediate action was taken. Metals in seafoods were monitored and import regulations were put in place. As a result, metal toxicity in seafoods is no longer a major issue, and since most marine organisms are resilient to metals, this form of pollution affects ecosystems only when metals are in very high concentrations, such as where mine tailings reach the sea.

Synthetic organics, either as pesticides and antifoulants (notably tributyl tin (TBT); de Mora, 1996) or as industrial chemicals, are present in sea water, biota, and sediments, and affect the whole spectrum of marine life, from primary producers to mammals