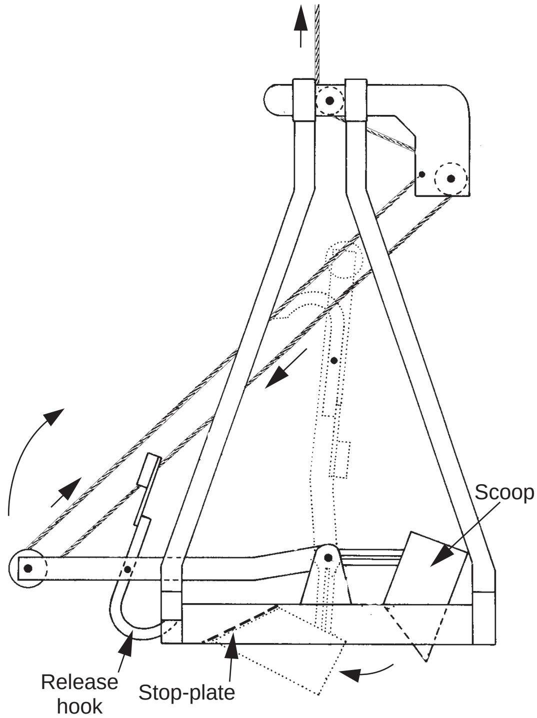


**Figure 18** Diagram of a Hamon scoop.

## Sampling Difficult Sediments

Most of the samplers so far discussed operate reasonably well in mud or sand substrata. Few operate satisfactorily in gravel or stony mixed ground either because the bottom is too hard for the sampler to penetrate the substratum or because of the increased likelihood of a stone holding the jaws open when they are drawn together. To get around this problem various types of scoops have been devised. The Holme grab has a double scoop action with two buckets rotating in opposite directions to minimize any lateral movement during digging. The scoops are closed by means of a cable and pulley arrangement (**Figure 17**) and simultaneously take two samples of 0.05 $m^2$ surface area.

The Hamon grab, which has proved to be very effective in coarse, loose sediments, takes a single rectangular scoop of the substratum covering a surface area of about 0.29 $m^2$. The scoop is forced into the sediment by a long lever driven by pulleys that are powered by the pull of the warp (**Figure 18**). Although the samples may not always be as consistent as those from a more conventional grab sampler, the Hamon grab has found widespread use where regular sampling on rough ground is impossible by any other means.

## See also

**Benthic Organisms Overview. Benthic Boundary Layer Effects.**

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# GRAVITY

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## Introduction

The gravity field varies over the oceans on account of lateral variations in density beneath the ocean surface. The most prominent anomalies arise from undulations on density interfaces, such as occur at the water–rock interface at the seafloor or at the crust–mantle interface, also known as the Moho discontinuity. Because marine gravity is relatively easy to measure, it serves as a remote sensing tool for exploring the earth beneath the oceans. The interpretation of marine gravity anomalies in terms of the Earth's structure is highly nonunique, however, and thus requires simultaneous consideration of other geophysically observed quantities. The most useful auxiliary measurements include depth of the ocean from echo sounders, the shape of buried reflectors from marine seismic reflection data, and/or the density of ocean rocks as determined from dredge samples or inferred from seismic velocities.

Depending on the spatial wavelength of the observed variation in the gravity field, marine gravity observations are applied to the solution of a number of important problems in earth structure and dynamics. At the very longest wavelengths of 1000 to 10 000 km, the marine gravity field is usually combined with anomalies over land to infer the dynamics of the entire planet. At medium wavelengths of several tens to hundreds of kilometers, the gravity field contains important information on the thermal and mechanical properties of the lithospheric plates and on the thickness of their sedimentary cover. At even shorter wavelengths, the field reflects local irregularities in density, such as produced by seafloor bathymetric features, magma chambers, and buried ore bodies. On account of the large number of potential contributions to the marine gravity field, modern methods of analysis include spectral filtering to remove signals outside of the waveband of interest and interpretation within the context of models that obey the laws of phyics.

## Units

Gravity is an acceleration. The acceleration of gravity on the earth's surface is about $9.81\,\mathrm{m\,s^{-2}}$. Gravity anomalies (observed gravitational acceleration minus an expected value) are typically much smaller, about 0.5% of the total field. The SI-compatible unit for gravity anomaly is the gravity unit (gu): $1\,\mathrm{gu} = 10^{-6}\,\mathrm{m\,s^{-2}}$. However, the older c.g.s unit for gravity anomaly, the milligal (mGal), is still in very wide use: $1\,\mathrm{mGal} = 10\,\mathrm{gu}$. Typical small-scale variations in gravity over the ocean range from a few tens to a few hundreds of gravity units. Lateral variations in gravitational acceleration (gravity gradients) are measured in Eötvös units (E): $1\,\mathrm{E} = 10^{-9}\,\mathrm{s^{-2}}$. Another quantity useful in gravity interpretation is the density of earth materials, measured in $\mathrm{kg\,m^{-3}}$. In the marine realm, relevant densities range from about $1000\,\mathrm{kg\,m^{-3}}$ for water to more than $3300\,\mathrm{kg\,m^{-3}}$ for mantle rocks.

A close relative of the marine gravity field is the marine geoid. The geoid, measured in units of height, is the elevation of the sea level equipotential surface. Geoid anomalies are measured in meters and are the departure of the true equipotential surface from that predicted for an idealized spheroidal Earth whose density structure varies only with radius. Geoid anomalies range from 0 to more than $\pm 100\,\mathrm{m}$. The direction of the force of gravity is everywhere perpendicular to the geoid surface, and the magnitude of the gravitational attraction is the vertical derivative of the geopotential $U$ (eqn [1]).

$$g = -\frac{\partial U}{\partial z} \qquad [1]$$

Geoid height $N$ is related to the same equipotential $U$ via Brun's formula (eqn [2]).

$$N = -\frac{U}{g_0} \qquad [2]$$

in which $g_0$ is the acceleration of gravity on the spheroid.

For a ship sailing on the sea surface (the equipotential), it is easier to measure gravity. From a satellite in free-fall orbit high over the Earth's surface, radar altimeters can measure with centimeter precision variations in the elevation of sea level, an excellent approximation to the true geoid that would follow the surface of a motionless ocean. Regardless of whether geoid or gravity is the quantity measured directly, simple formulas in the wavenumber domain allow gravity to be computed from geoid and vice versa. Given the same equipotential, the gravity representation emphasizes the power in the high-frequency (short-wavelength) part of the spectrum, whereas the geoid representation emphasizes the longer wavelengths. Therefore, for investigations of high-frequency phenomena, gravity is generally the quantity interpreted even if geoid is what was measured. The opposite is true for long-wavelength phenomena.

## Measurement of Marine Gravity

Marine gravity measurements can be and have been acquired with several different sorts of sensors and from a variety of platforms, including ships, submarines, airplanes, and satellites. The ideal combination of sensor system and platform depends upon the needed accuracy, spatial coverage, and available time and funds.

### Gravimeters

The design for most marine gravimeters is borrowed from their terrestrial counterparts and are either absolute or relative in their measurements. Absolute gravimeters measure the full acceleration of gravity $g$ at the survey site along the direction of the local vertical. Modern marine absolute gravimeters measure precisely the vertical position $z$ of a falling mass (e.g., a corner cube reflector) as a function of time $t$ in a vacuum cylinder using laser interferometry and an atomic clock. The acceleration is then calculated as the second derivative of the position of the falling mass as a function of time (eqn [3]).

$$g = \frac{d^2 z(t)}{dt^2} \quad [3]$$

Absolute gravimeters tend to be larger, more difficult to deploy, costlier to build, and more expensive to run than relative gravimeters, and thus are only used when relative gravimeters are inadequate for the problem being addressed.

Most gravity measurements at sea are relative measurements, $\Delta g$: the instrument measures the difference between gravity at the study site and at another site where absolute gravity is known (e.g., the dock where the expedition originates). Modern relative gravimeters are based on Hooke's law for the force $F$ required to extend a spring a distance $x$ (eqn [4]), where $k_s$ is the spring constant, calculated by extending the spring under a known force.

$$F = -k_s x = mg \quad [4]$$

If a mass $m$ is suspended from this spring at a site where gravity $g$ is known (e.g., by deploying an absolute gravimeter at that base station), then gravity at other locations can be calculated by observing how much more or less that same spring is stretched at other locations by the same mass. Although such systems are relatively inexpensive to build and easy to deploy, they suffer from drift: in effect, the spring constant changes with time because no physical spring is perfectly elastic. To first order, the drift can be corrected by returning to the same or another base station with the same instrument, and assuming that the drift was linear with time in between. The accuracy of this linear drift assumption improves with more frequent visits to the base station, but this is usually impractical for marine surveys. Through clever design, the latest generation of marine gravimeters has greatly reduced the drift problem as compared with earlier instruments.

The measurement of the gravity gradient tensor was widely used early in the twentieth century for oil exploration, but fell into disfavor in the 1930s as scalar gravimeters became more reliable and easy to use. Gravity gradiometry at sea is currently making a comeback as the result of declassification of military gradiometer technology developed for use in submarines during the Cold War. Gravity gradiometers measure the three-dimensional gradient in the gravity vector using six pairs of aligned gravimeters, with accuracies reaching better than 1 Eötvös. In comparison with measurements of gravity, the gravity gradient has more sensitivity to variations at short wavelengths ($\sim$5 km or less), making it useful for delimiting shallow structures buried beneath the seafloor.

Geoid anomalies can be directly measured from orbiting satellites carrying radar altimeters. The altimeters measure the travel time of a radar pulse from the satellite to the ocean surface, from which it is reflected and bounced back to the satellite. Tracking stations on Earth solve for the position of the satellite with respect to the center of the Earth. These two types of information are then combined to calculate the height of the sea level equipotential surface above the center of the Earth. Because the solid land surface does not follow an equipotential, altimeters cannot be used to constrain the terrestrial geoid. Furthermore, it is difficult to extract geoid from ocean areas covered by sea ice. However, in the near future, laser altimeters deployed from satellites hold the promise of extracting geoid information even over ocean surfaces marred with sea ice, on account of their enhanced resolution.

## Platforms

Marine gravity data can be acquired either from moving or from stationary platforms. Because the gravity field from variations in the depth of the sea floor is such a large component of the observed signal, most marine gravity surveys have relied on ships or submarines that enable the simultaneous acquisition of depth observations. However, airborne gravity measurements have been acquired successfully over ice-covered areas of the polar oceans, and orbiting satellites have measured the marine geoid from space.

A major challenge in acquiring gravity data from a floating platform at the sea surface is in separating the acceleration of the platform in the dynamic ocean from the acceleration of gravity. This problem is overcome by mounting marine gravimeters deployed from ships on inertially stabilized tables. These tables employ gyroscopes to maintain a constant attitude despite the pitching and rolling of the ship beneath the table. The nongravitational acceleration is somewhat mitigated by mounting marine gravimeters deep in the hold and as close to the ship's center of motion as possible. Special damping mechanisms also prevent the spring in the gravimeter from responding to extremely high-frequency changes in the force on the suspended mass.

Instruments deployed in submersibles resting on the bottom of the ocean or in instrument packages lowered to the bottom of the ocean do not suffer from the dynamic accelerations of the moving ocean surface, but bottom currents can also be an important source of noise in submarine gravimetry. Installing instruments in boreholes is the most effective way to counter this problem, but it is also an expensive solution.

## Reduction of Marine Data

A number of standard corrections must be applied to the raw gravity data (either $g$ or $\Delta g$) prior to interpretation. In addition to any drift correction, as mentioned above for relative gravity measurements, a latitude correction is immediately applied to account for the large change in gravity between the poles and the Equator caused by Earth's rotation. Near the Equator, the centrifugal acceleration from the Earth's spin is large, and gravity is about 50 000 gu less, on average, than at the poles. Because this effect is 5000 times larger than typical regional gravity signals of interest, it must be removed from the data using a standard formula for the variation of gravity $g_0$ on a spheroid of revolution best fitting the shape of the Earth (eqn [5]; $\Theta$ = latitude).

$$g_0(\Theta) = 9.780\,318\,5(1 + 5.278\,895 \times 10^{-3} \sin^2 \Theta + 2.3462 \times 10^{-5} \sin^4 \Theta) \quad \mathrm{m\,s^{-2}} \qquad [5]$$

A second correction that must be made if the gravity is measured from a moving vehicle, such as a ship or airplane, accounts for the effect on gravity of the motion of the vehicle with respect to the Earth's spin. A ship steaming to the east is, in effect, rotating faster than the Earth. The centrifugal effect of this increased rate of rotation causes gravity to be less than it would be if the ship were stationary. The opposite effect occurs for a ship steaming to the west. This term, called the Eötvös correction, is largest near the Equator and involves only the east–west component $v_{\mathrm{EW}}$ of the ship's velocity vector (eqn [6]), in which $\omega$ is the angular velocity of the earth's rotation.

$$g_{\mathrm{EOT}} = 2\omega v_{\mathrm{EW}} \cos \Theta \qquad [6]$$

The free air gravity correction, which accounts for the elevation of the measurement above the Earth's sea level equipotential surface, is obviously not needed if the measurement is made on the sea surface. The free air correction $g_{\mathrm{FA}}$ is required if the measurement is made from a submersible or an airplane: eqn [7], where $h$ is elevation above sea level in meters.

$$g_{\mathrm{FA}} = 3.1\,h \quad \mathrm{gu} \qquad [7]$$

This correction is added to the observation if the sensor is deployed above the Earth's surface, and subtracted for stations below sea level.

For land surveys, the Bouguer correction accounts for the extra mass of the topography between the observation and sea level. For its marine equivalent, it adds in the extra gravitational attraction that would be present if rock rather than water existed between sea level and the bottom of the ocean. Except in areas of rugged bathymetry, the Bouguer correction $g_{\mathrm{B}}$ is calculated using the slab formula (eqn [8]).

$$g_{\mathrm{B}} = -2\pi\,\Delta\rho\,Gz \qquad [8]$$

Here $\Delta\rho$ is the density difference between oceanic crust and sea water, $G$ is Newton's constant, and $z$ is the depth of the sea floor. This correction is seldom used because it produces very large positive gravity anomalies. Furthermore, there are more accurate corrections for the effect of bathymetry that do not make the unrealistic assumption that the expected state for the oceans should be that the entire depth is filled with crustal rocks displacing the water. The Bouguer correction is necessary, however, when gravity measurements are made from a submarine, in order to combine those data with more conventional observations from the sea surface. In this case, the Bouguer correction is applied twice: once to remove the upward attraction of the layer of water above the submarine, and once more to add in that layer's gravitation field below the sensor.

Satellite measurements of sea surface height go through a different processing sequence to recover marine geoid anomalies. The most important step is in calculating precise orbits. Information from tracking stations is supplemented with a 'crossover analysis' that removes long-wavelength bias in orbit elevation by forcing the height valves to agree wherever orbits cross. Corrections are then made for known physical oceanographic effects such as tides, and wave action is averaged out. The height of the sea level geoid above the Earth's center, assuming the standard spheroid, is subtracted from the data to create geoid anomalies.

## History

A principal impediment to the acquisition of useful gravity observations at sea was the difficulty in separating the desired acceleration of gravity from the acceleration of the platform floating on the surface of the moving ocean. For this reason, the first successful gravity measurements to be acquired at sea were taken from a submarine by the Dutch pioneer, Vening Meinesz, in 1923. He used a pendulum gravimeter, which was the state of the art for measuring absolute gravity at that time. By accurately timing the period, $T$, of the swinging pendulum, the acceleration of gravity, $g$, can be recovered according to eqn [9], in which $l$ is the length of the pendulum arm.

$$T = 2\pi\sqrt{\frac{l}{g}} \qquad [9]$$

By 1959, five thousand gravity measurements had been acquired from submarines globally. These measurements were instrumental in revealing the large gravity anomalies associated with the great trenches along the western margin of the Pacific. However, these gravity observations were very time-consuming to acquire because of the long integration times needed to achieve a high-precision estimate of the pendulum's period, and could not be adapted for use on a surface ship.

Gravity measurements at sea became routine and reliable in the late 1950s with the development of gyroscopically stabilized platforms and heavily damped mass-and-spring systems constrained to move only vertically. The new platforms compensated for the pitch and roll of the ship such that simple mass-and-spring gravimeters could collect time series of variations in gravity over the oceans from vessels under way. Without any need to stop the ship on station, a time series of gravity measurements could be obtained at only small incremental cost to ship operations. With the advent of the new instrumentation, the catalogue of marine gravity values has grown in the past 40 years to more than 2.5 million measurements.

A new era of precision in marine gravity began with the advent of the Global Positioning System (GPS) in the late 1980s. Prior to this time, the largest source of uncertainty in marine gravity lay in the Eötvös correction. Older navigation systems (dead reckoning, celestial, and even the TRANSIT satellite system) were too imprecise in the absolute position of the ship and too infrequently available to allow accurate velocity estimation from minute to minute, especially if the ship was maneuvering. Typically, gravity data had to be discarded for an hour or so near the time of any change in course. The high positioning accuracy and frequency of GPS fixes now allows such precise calculation of the Eötvös correction that it is no longer the limiting factor in the accuracy of marine gravity data.

A breakthough in determining the global marine gravity field was achieved with the launching of the GEOS-3 (1975–1977) and Seasat (1978) satellites, which carried radar altimeters. Altimeters were deployed for the purpose of measuring dynamic sea surface elevation associated with physical oceanographic effects. The Seasat satellite carried a new, high-precision altimeter that characterized the variations in sea surface elevation with unprecedented detail. The satellite failed prematurely, but not before it returned a wealth of data on the marine gravity field from its observations of the marine geoid. The geoid variations at mid- and short-wavelength were so large that the dynamic oceanographic effects motivating the mission could be considered a much smaller noise term. The success of the Seasat mission led to the launch of Geosat, which measured the geoid at even higher precision and resolution. Unfortunately, most of that data remained classified by the US military until the results from a similar European mission were about to be released into the public domain. The declassification of the Geosat data in 1995 fueled a major revolution in our understanding of the deep seafloor (**Figure 1**).

The latest developments in marine gravity stem from the desire to detect the shortest spatial wavelengths of gravity variations by taking gravimeters to the bottom of the ocean. Gravity is one example of a potential field, and as such the amplitude, $A$, of the signal of interest decays with distance, $z$, between source and detector as in eqn [10], where $k$ is the modulus of the spatial wavenumber, the reciprocal of the spatial wavelength.

$$A \sim e^{-2\pi kz} \qquad [10]$$

For sensors located on a ship at the sea surface in average ocean depths of 4.5 km, it is extremely difficult to detect short-wavelength variations in gravity of a few kilometers or less. Even lowering the gravimeter to the cruising depth of most submarines (a few hundred meters) does little to overcome the upward attenuation of the signal from localized sources on and beneath the seafloor. The solution to this problem recently has been to take gravimeters to the bottom of the ocean, either in a deep-diving submersible such as *Alvin*, or as an instrument package lowered on a cable. Most gravity measurements at sea are relative measurements. However, recent advances in instrumentation now allow absolute gravimeters to be deployed on the bottom of the

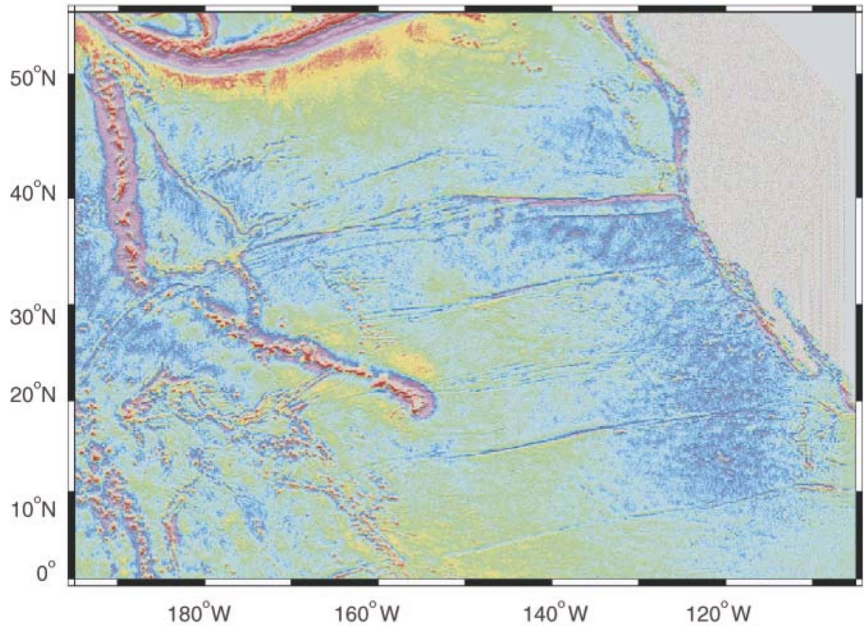


**Figure 1** Gravity field over the North Pacific. This view is based on satellite altimetry data from the Geosat and other missions. (Data from Sandwell and Smith (1997).)

ocean, avoiding the problem of instrument drift that adds error to relative gravity measurements. However, noise associated with the short baselines required for operation in the deep sea remains problematic.

## Interpretation of Marine Gravity

### Short-Wavelength Anomalies

The shortest-wavelength gravity anomalies over the oceans (less than a few tens of kilometers) are the least ambiguous to interpret since they invariably are of shallow origin. The upward continuation factor guarantees that any spatially localized anomalies with deep sources will be undetectable at the ocean surface. Near-bottom gravity measurements are able to improve somewhat the detection of concentrated density anomalies buried at deeper levels, but most are assumed to lie within the oceanic crust.

One of the most useful applications of short-wavelength gravity anomalies has been to predict ocean bathymetry (**Figure 2**). Radar altimeters deployed on the Seasat and Geosat missions measured with centimeter accuracy the height of the underlying sea surface, an excellent approximation to the marine geoid, over all ice-free marine regions. The accuracy and spatial coverage was far better than had been provided from more than a century of marine surveys from ships. At short wavelengths, undulations of the rock–water interface are the largest contribution to the short-wavelength portion of the geoid spectrum, which opened up the possibility of predicting ocean depth from the excellent geoid data. For example, an undersea volcano, or seamount, represents a mass excess over the water it displaces. The extra mass locally raises the equipotential surface, such that positive geoid anomalies are seen over volcanoes and ridges while geoid lows are seen over narrow deeps and trenches. The prediction of bathymetry from marine geoid or gravity data is tricky: the highest frequencies in the bathymetry cannot be estimated because of the upward attenuation problem, and the longer wavelengths are canceled out in the geoid by their isostatic compensation (see following section). These longer wavelengths in the bathymetry must be introduced into the solution using traditional echo soundings from sparse ship tracks. Nevertheless, the best map we currently have of the depth of the global ocean is courtesy of satellite altimetry.

### Mid-Wavelength Anomalies

The mid-wavelength part of the gravity spectrum (tens to hundreds of kilometers) is dominated by the effects of isostatic compensation. Isostasy is the process by which the Earth supports variations in topography or bathymetry in order to bring about a condition of hydrostatic equilibrium at depth. The definition of isostasy can be extended to include both static and dynamic compensation mechanisms,

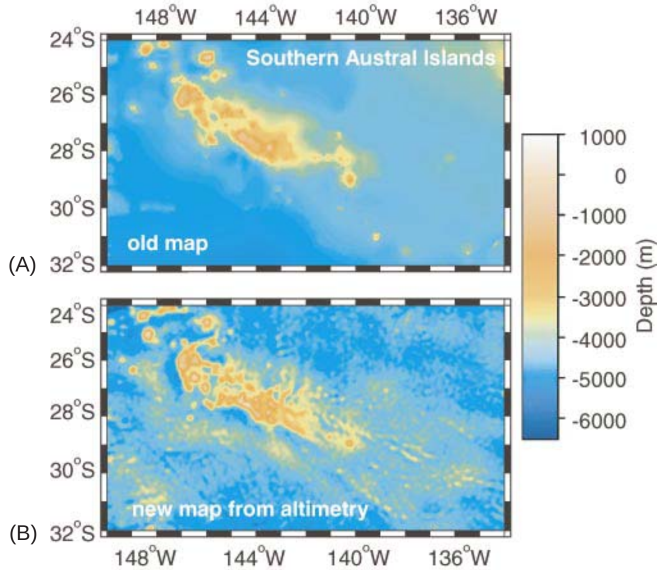


**Figure 2** Example of bathymetric prediction from gravity anomalies in a largely unexplored region of the South Pacific. (A) The best available bathymetry from sparse echo soundings available in the early 1990s. (B) A diagram shows a dramatic improvement in definition of the bathymetry when satellite gravity observations are used to constrain the short-wavelength component of the bathymetry. (Adapted from McNutt and Bonneville (1996).)

but at these wavelengths the static mechanisms are most important. There are a number of different types of isostatic compensation at work in the oceans, and the details of the gravity field can be used to distinguish them and to estimate the thermo-mechanical behavior of oceanic plates.

One of the simplest mechanisms for isostatic compensation is Airy isostasy: the oceanic crust is thickened beneath areas of shallow bathymetry. The thick crustal roots displace denser mantle material, such that the elevated features float on the mantle much like icebergs float in the ocean. Of the various methods of isostatic compensation, this mechanism predicts the smallest gravity anomalies over a given feature. From analysis of marine gravity, we now know that this sort of compensation mechanism is only found where the oceanic crust is extremely weak, such as on very young lithosphere near a midocean ridge. For example, large plateaus formed when hotspots intersect midocean ridges are largely supported by Airy-type isostasy. Elsewhere the oceanic lithosphere is strong enough to exhibit some lateral strength in supporting superimposed volcanoes and other surface loads.

An extremely common form for support of bathymetric features in the oceans is elastic flexure. Oceanic lithosphere has sufficient strength to bend elastically, thus distributing the weight of a topographic feature over an area broader than that of the feature itself (**Figure 3**). Analysis of marine gravity has been instrumental in establishing that the elastic strength of the oceanic lithosphere increases with increasing age. Young lithosphere near the midocean ridge is quite weak, in some cases hardly distinguished from Airy-type isostasy. The oldest oceanic lithosphere displays an effective thickness equivalent to that of a perfectly elastic plate 40 km thick. The fact that this thickness is less than that of the commonly accepted value for the thickness of the mechanical plate that drifts over the

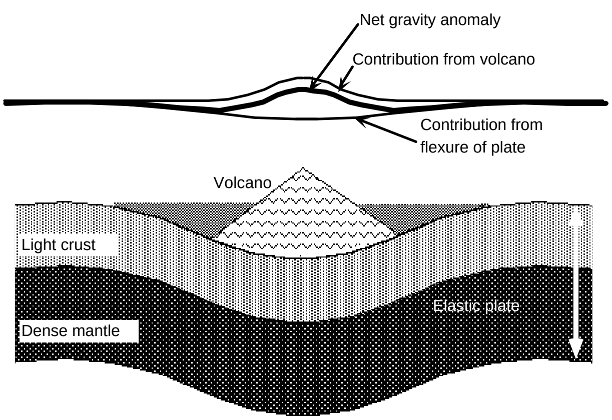


**Figure 3** Cartoon showing how the seafloor is warped as an elastic plate under the weight of a small volcano. The gravity anomaly that would be detected by a ship sailing along the sea surface over this feature is the net difference between the positive gravity perturbation from the extra mass of the volcano and the negative gravity perturbation produced when the elastically flexed light crust replaces denser mantle. (Adapted from McNutt and Bonneville (1996).)

asthenosphere indicates that the base of the oceanic lithosphere is not capable of sustaining large deviatoric stresses (of the order of 100 MPa or more) over million-year timescales.

Another very important method of isostatic compensation in the ocean is Pratt isostasy. This method of support supposes that the height of a vertical column of bathymetry is inversely proportional to its density. Low-density columns can be higher because they are lighter, whereas heavy columns must be short in order to produce the same integrated mass at some assumed depth of compensation. In the oceans, variations in the temperature of the lithosphere produce elevation changes in the manner of Pratt isostasy. For example, ridges stand 4 km above the deep ocean basins because the underlying lithosphere is hotter when the plate is young. The bathymetric swells around young hotspot volcanoes may also be supported by Pratt-type isostasy, although some combination of crustal thickening and dynamic isostasy may be operating as well. Again, gravity and geoid anomalies have been principal constraints in arguing for the mechanism of support for bathymetric swells.

### Long-Wavelength Anomalies

At wavelengths from 1000 to several thousand kilometers, gravity anomalies are usually derived from satellite observations and interpreted using equations appropriate for a spherical earth. Geoid is interpreted more commonly than gravity directly, as it emphasizes the longer wavelengths in the geopotential field. Isostatic compensation for smaller-scale bathymetric features, such as seamounts, can usually be ignored in that the gravity anomaly from bathymetry is canceled out by that from its compensation when spatially averaged over longer wavelengths.

The principal signal at these wavelengths arises from the subduction of lithospheric slabs and other sorts of convective overturn within the mantle. Three sorts of gravitational contributions must be considered: (1) the direct effect of mass anomalies within the mantle, either buoyant risers or dense sinkers which drive convection; (2) the warping of the surface caused by viscous coupling of the risers or sinkers to the earth's surface; and (3) the warping of any deeper density discontinuities (such as the core–mantle boundary) also caused by viscous coupling.

In the 1980s, estimates of the locations and densities of mass anomalies in the mantle responsible for the first contribution above began to become available courtesy of seismic tomography. Travel times of earthquake waves constrained the locations of seismically fast and slow regions in the mantle. By assuming that the seismic velocity variations were caused by temperature differences between hot, rising material and cold, sinking material, it was possible to convert velocity to density using standard relations. Knowing the locations of the mass anomalies driving convection inside the Earth led to a breakthrough in understanding the long-wavelength gravity and geoid fields.

The amount of deformation on density interfaces above and below the mass anomalies inferred from tomography (contributions (2) and (3) above) depends upon the viscosity structure of Earth's mantle. Coupling is more efficient with a more viscous mantle, whereas a weaker mantle is able to soften the transmission of the viscous stresses from the risers and sinkers. Therefore, one of the principal uses of marine gravity anomalies at long wavelengths has been to calibrate the viscosity structure of the oceanic upper mantle. This interpretation must be constrained by estimates of the dynamic surface topography over the oceans, which is actually easier to estimate than over the continents because of the relatively uniform thickness of oceanic crust.

A fairly common result from this sort of analysis is that the oceanic upper mantle must be relatively inviscid. The geoid shows that there are large mass anomalies within the mantle driving convection that are poorly coupled to variations in the depth of the seafloor. If the upper mantle were more viscous, there should be a stronger positive correlation between marine geoid and depth of the seafloor at long wavelengths.

## See also

**Manned Submersibles, Deep Water. Satellite Altimetry. Satellite Oceanography, History and Introductory Concepts.**

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# GRAVITY CURRENTS

*See* **NON-ROTATING GRAVITY CURRENTS; ROTATING GRAVITY CURRENTS**

# GRAVITY WAVES

*See* **SURFACE, GRAVITY AND CAPILLARY WAVES**

# GULF STREAM

*See* **FLORIDA CURRENT, GULF STREAM AND LABRADOR CURRENT**

# GULLS

*See* **LARIDAE, STERNIDAE AND RYNCHOPIDAE**

# GYRE

*See* **OCEAN GYRE ECOSYSTEMS**