12.710 Marine Geology and Geophysics
Global Tectonics and the Earth System: Measuring the Sea Surface

Turcotte and Schubert, Chapter 5

Introduction

Much effort in Earth sciences right now is in understanding interaction between various major components of the Earth system: the solid Earth, ocean, atmosphere, cryosphere, hydrosphere, etc. Very often the questions of interest involve how one part of the system changes in response to processes that are occurring within the others. Although changes on many timescales occur and are important, there is currently a particular emphasis on changes that are observable over decadal timescales. Such changes may be important in themselves or provide the basis for discerning longer-term variations.

One of the measurable quantities that has significant importance for a wide variety of questions in marine science is the shape of the sea surface. The height of the sea surface at a range of spatial scales provides information on:

- ocean circulation and the nature of currents
- ocean-atmosphere interactions (*e.g.* waves)
- the structure of the seafloor
- climate (sea level change).

Because of its utility for investigating a range of topics useful for ocean science we will discuss how the elevation of the sea surface is measured. For the particular application the required accuracy of the sea surface measurement in a spatial and vertical sense varies. For example, regional marine geophysical studies require very high spatial (short wavelength) resolution to understand detailed structure and local processes, while measurements of sea level change require precise characterization of the sea surface at the longest wavelengths. As we shall see, short wavelength measurements are best made from the surface while determination of shape with long wavelength accuracy requires uniform measurements obtained from orbit.

The key in using sea height measurements as a data set for evaluating Earth structure and change is having a proper reference from which to measure variations. Gravity happens to provide a natural reference frame for monitoring many aspects of global change. We therefore begin by reviewing some basics about gravity to provide background for understanding how the sea surface is measured.

<u>A Primer on Gravity and the Geoid</u>

Gravity

The force that is exerted on any small mass in the Earth has two components, due to:

- gravitational attraction of the mass
- acceleration due to rotation.

Note that gravity is an acceleration, and without the proper information is not immediately distinguished from other accelerations in a system.

Consider the gravitational force on a mass m' located at a point P outside the Earth by a small mass element inside the Earth, dm. The gravitational attraction, df, is goverened by **Newton's law of Gravitation**

$$df = \frac{Gm'dm}{b^2} \qquad (1)$$

where G is the **Universal constant of gravitation** (=6.67x10^{-11} m^3 kg^{-1} s^{-2}) and b is the distance between dm and P. The small gravitational acceleration at P due to the attraction of dm is the force per unit mass exerted on m' in the direction of P

$$dg = \frac{df}{m'}. \qquad (2)$$

Combining (1) and (2) we see that

$$dg = \frac{Gdm}{b^2}. \qquad (3)$$

We can obtain a straightforward expression for the **acceleration of gravity** *(g)* by assuming that the entire mass of the Earth, M, is concentrated at its center. We obtain

$$g = \frac{GM}{r^2} \qquad (4)$$

where GM=3.986005x10^{14} m^3 s^{-2}, and r is the distance from the center of the Earth at which the gravitational attraction is being measured. Note g is defined as positive if directed radially inward. The "central point" is the Earth's **center of mass**, and this is the **coordinate system origin** for all studies related to the global gravity field and sea surface elevation.

The value of g varies with position on Earth's surface. The **rotation** of the Earth produces a **flattening** of the Earth which is manifest as an **equatorial bulge**, and this leads to a **latitudinal dependence of gravity**. (In fact, the gravity at the equator is about 1/200's less than that at the poles because the equator with its greater radius is farther from the center of the Earth.) The Earth's shape is said to be an **oblate spheroid** and the gravity field of this spheroid (sometimes called a **reference ellipsoid)** is taken to be the reference gravitational field of the Earth. Mass variations due to **surface topography** and **internal mass variations** (due to temperature variations, compositional variations, phase changes, etc.) result in regional variations in the gravity field that are called **gravity anomalies.**

Much of our understanding of the structure of the oceanic (and continental) crust and upper mantle comes from measurement and analysis of gravity anomalies. In practice, gravity is interpreted in combination with surface topography, the latter of which allows removal of gravitational attraction of surface mass to leave the signal corresponding to subsurface mass variations.

The SI unit of gravity is the **gal** where 1 gal= 1 cm s^{-2}. Gravity anomalies are often expressed as **milligals** where 100 mgal=1 gal.

The Geoid

When gravity is used to measure height in the oceans it is inevitably expressed in terms of the **gravitational potential** or **geoid**. The reason is quite simple**: sea level is approximately a surface of constant geopotential**. This is true except for tides, waves, and currents, which generally produce deviations of several meters. A practical visualization of the geoid over land is where a pattern of canals is cut through the continents. The geoid is the level to which water would rise in the canals. The shape of the geoid reflects hydrodynamic equilibrium of the accelerations due to rotation and the mass distribution. The geoid has units of **meters**.

To first order the geoid *(V)* can be obtained by integrating the gravitational acceleration

$$V = GM\int \frac{dr}{r^2} = -\frac{GM}{r} \tag{5}$$

(Actually, in geophysics we usually start with the potential and from that derive gravity anomalies.) Note that the geoid has a negative sign so a mass excess produces an **outwarping** of the geopotential.

We may write down an expression for the long wavelength shape of the geopotential *(U)* that takes into account contributions from the Earth's rotation and long wavelength internal mass distribution. Essentially what we are doing is adding terms to the original "point mass" term to take into account the oblate shape of the Earth.

$$U = -\frac{GM}{r} + \frac{GMa^2 J_2}{2r^3}(3\sin^2\phi - 1) - \frac{1}{2}\omega^2 r^2 \cos^2\phi \tag{6}$$

where a is the equatorial radius, ϕ is latitude, ω is the rotational velocity, and

$$J_2 = \frac{C - A}{Ma^2} \tag{7}$$

is the **oblateness parameter** where C and A are the maximum and minimum moments of inertia. The moments are integrals of the Earth's internal mass distribution. The moments can be determined by summing over a mass distribution M the infinitesimal mass components dm times the square of the perpendicular distance from the mass point to the axes of inertia such as

$$C = \int \left(x'^2 + y'^2\right) dm$$

where

$$x' = r' \sin\theta' \cos\psi'$$
$$y' = r' \sin\theta' \sin\psi'.$$

The maximum and minimum moments can be re-written

$$\begin{aligned} C &= \int r_1^2 \sin^2\theta \, dm \\ A &= \int r_1^2 \left(\sin^2\theta \sin^2\psi + \cos^2\theta\right) dm \end{aligned} \tag{8}$$

where *(r, θ, ψ)* is the spherical coordinate system used in the integration of the mass elements *dm*. In practice J_2 is measured from perturbations of orbiting spacecraft and equals $J_2 = 1.08270 \times 10^{-3}$. After the "point mass" term in the potential the oblateness is the biggest term. A plot of the geoid that includes J_2 will show a horizontally "striped" pattern because this term swamps all of the others. For this reason maps of the geoid (and gravity field) always have J_2 subtracted out.

Keep in mind that equation (6) only retains terms that describe the Earth's geopotential shape at wavelengths up to those that describe its oblateness or flattening. These terms are of order the length scale of the circumference. To describe shorter wavelength features it would be necessary to solve **Laplace's Equation** for a potential

$$\nabla^2 V = 0 \qquad (9)$$

where

$$\nabla^2 V = \frac{1}{r^2}\frac{\partial}{\partial r}\left(r^2 \frac{\partial V}{\partial r}\right) + \frac{1}{r^2 \sin\theta}\frac{\partial}{\partial \theta}\left(\sin\theta \frac{\partial V}{\partial \theta}\right) + \frac{1}{r^2 \sin^2\theta}\frac{\partial^2 V}{\partial \lambda^2} \qquad (10)$$

in spherical geometry. The solution for equation (9) in spherical geometry consists of **spherical harmonics**, which are series functions. When taken to high degree and order they can represent the solution to (9) for as short a wavelength as desired (or the data are capable of resolving). The gravitational potential, *V,* at the altitude of an orbiting spacecraft that is measuring gravitational perturbations may be represented in spherical harmonic form as

$$V(r,\phi,\lambda) = \frac{GM}{r} + \frac{GM}{r}\sum_{l=2}^{N}\sum_{m=0}^{l}\left(\frac{a}{r}\right)^{l}\overline{P}_{lm}(\sin\varphi)(\overline{C}_{lm}\cos m\lambda + \overline{S}_{lm}\sin m\lambda) \qquad (11)$$

where r is the radial distance from the center of mass of Earth to the spacecraft, ϕ and λ are the latitude and longitude of the spacecraft, $\overline{P}_{lm}$ are the normalized associated Legendre functions of degree l and order m, $\overline{C}_{lm}$ and $\overline{S}_{lm}$ are Stokes coefficients which are estimated from spacecraft tracking observations (the overbars indicate normalized coefficients), and N is the maximum degree representing the size or resolution of the field. High resolution corresponds to high degree and order.

Measuring Sea Surface Topography

Sea Surface -- Definitions

Earlier we said the sea surface was a good approximation of Earth's geoid or gravitational equipotential. This is true but the difference between the sea surface and geoid is an important distinction. Here are some definitions to clarify what is meant by various measurements of the elevation of the ocean surface:

- sea surface height - the elevation of the surface of the ocean with respect to the center of mass or a reference ellipsoid.
- mean sea surface - the average height of the measured sea surface; smoothes out time variations.
- geoid - repeating from above: an equipotential surface that corresponds to the state of hydrostatic equilibrium of the accelerations due to rotation and the mass distribution.

Measured with respect to an ellipsoid. RMS amplitudes are of order 30 m but the largest deviation is ~100 m.
- dynamic topography - the difference between the sea surface height and the marine geoid. The differences are due to tides, currents and waves. Amplitudes are of order 1-2 m.

--> The marine geoid is determined from the global gravity field.
--> Sea surface height comes from satellite altimetry (described below).

Physical Oceanography Aside:
- The **surface geostrophic velocity** is proportional to the **slope of dynamic topography**. This can be related to important quantities in ocean dynamics such as the near-surface advective heat transport. Thus measurements of the ocean surface provide an indication of the energetics of the oceans.
- But note that the **geotrophic estimation is valid only where the error in the marine geoid is insignificant compared to the undulation amplitudes of the mean dynamic topography**. Currently, errors in the marine geoid prevent determination of absolute dynamic topography for λ<2500 km. The accuracy of the marine geoid is a limiting factor in advancing certain important dynamic oceanographic studies.

Making the Measurement

We shall focus on satellite observations because these provide near-global maps of the sea surface. (In contrast airborne or shipboard observations are important and provide high spatial resolution but can obtain only regional coverage.) Indeed**, a comparison of satellite and ship-board gravity show long wavelength biases in the latter**, that are a consequence of the inability to properly provide an absolute reference the latter measurements.

Globally distributed measurements of sea surface height are made with **radar altimeters**. **Radar** is an **active remote sensing technique** in which pulse trains of electromagnetic radiation are radiated toward the surface of the ocean by an RF (radio frequency) generator and some part of the reflected signal is detected at an onboard receiver. In principle, the radar pulses can be analyzed in two different general applications:

- **ranging**- measure the time delay between transmission and reception of signal to yield the range to the surface.
- **scatterometry** - measure the strength of the returning signal to characterize the roughness/reflectivity of the surface at the spatial scale of the radar wavelength.

The geometric configuration of the radar system must be optimized to perform ranging and/or scatterometry. When the system is approximately **nadir-viewing** the **radar** backscatter cross-section is most **sensitive to surface slopes and therefore topography**. This is the configuration used in ranging. When the **incidence angle** is pointed more than about 10° **off-nadir** the return is best used to analyze backscatter magnitude, which provides information on the **roughness** of the surface at the radar wavelength (λ~10 cm) as well as the surface **reflectivity** (--> dielectric constant). For example, radar scatterometry provides measurements of **significant wave height** from which oceanic wind speeds can be inferred. For determination of mean sea surface the ranging function is used.

Satellite Altimetry

The technique of **satellite altimetry** represents the precise measurement of range from radar altimeters to the ocean surface. Satellite altimeters **measure the two-way travel**

time of outgoing radar pulses from a satellite to the Earth's surface and back. The spacecraft on which the radar transmitter is mounted naturally moves with respect to the Earth's center of mass. By tracking the spacecraft it is possible to determine the spacecraft orbit. By subtracting the range of the spacecraft above the ocean from the distance of the spacecraft from the center of mass, a measurement of the elevation of the sea surface is obtained.

Radar altimetry has greatly improved knowledge about the short wavelength marine gravity. After correction for pressure and tidal variations and measurement biases, altimetric data, in combination with a very precise orbit, measure the ocean geoid, except for meter-scale dynamic topography.

Determination of satellite gravity from satellite altimetry:
- Differentiate the ascending and descending altimeter profiles in the along-track directions to provide a measure of the geoid slope or "deflections from the vertical".
- From the geoid slopes gravity anomalies can be obtained by standard methods.

The **major error sources** associated with this technique are due to **radial orbit error**, **atmospheric** (tropospheric and ionospheric) **effects**, **sea state bias and inverted barometer** (*i.e.* the pressure of the atmosphere acting on the ocean). The range from the spacecraft to the ocean surface has now been achieved to levels of 3-4 cm, and the major error source is the radial orbit uncertainty. The combined measure of the geoid and dynamic topography is 4-5 cm. Note that range error is not the limiting error source. The limiting factor is the geoid.

Orbits for Altimetric Mapping

- **Altimetric orbit -** Ideally, one would like to make interpretations about the sea surface that are not based on azimuthal variations in errors. Therefore, if a satellite is used for altimetry it is desirable that the ascending and descending sub-satellite tracks cross at 90^o so that orthogonal components of surface slope can be determined with equal accuracy.

- **Exact-repeat orbit** - An exact repeat orbit retraces the spacecraft groundtracks over the Earth's surface after a certain period. In this case spatial resolution is sacrificed for the purpose of temporal resolution. Thus this is the kind of orbit used to measure change of the ocean surface over a period of time. Exact repeat orbits have also been used for military spacecraft that release data to the science community to limit across-track resolution for reasons of "national security".

Results

Early mean sea level calculations were determined from the **GEOS-3** and **SEASAT** satellites. GEOS-3 had an along-track (half wavelength) resolution of about 40 km but operated 3.5 years and its across-track resolution varied from 10-200 km. SEASAT had an along-track resolution of 25 km but operated for only three months so there was poor across-track resolution (80-120 km). The combined data sets performed to an accuracy of 15 cm and a horizontal resolution of 25 km. In these early studies sea surface height measurements were corrected for altimeter measurement bias, ocean tide effects, and solid Earth tide effects.

The Navy launched the **Geosat** satellite in 1985. Very high resolution data, with 5-10 km along-track resolution, have been obtained from the **Geosat Geodetic Mission (GM)**, which was initially classified. The GM has an across-track resolution of as little as 2 km. At the end of the geodetic mission Geosat was placed into an orbit with a 17-day repeat in the spacecraft groundtrack, referred to as the **Geosat Exact Repeat Mission (ERM)**, which overlaid the SEASAT 17-day ground tracks. The across-track resolution was 160 km.

The **ERS-1** spacecraft was launched by the European Space Agency (ESA) and surveyed the oceans in 1994-1995. The spacing of spacecraft ground tracks was 8 km, very high resolution. However, the data is not very useful for marine gravity studies because of errors and ocean dynamical effects. The data obtained from this mission are useful for reconnaissance of marine seafloor features but for detailed modeling the accuracy of the data is not state of the art compared to what is now being achieved with satellite or airborne sensors.

More recently, observations from the **TOPEX/POSEIDON** satellite were used to estimate the mean sea surface to an accuracy of 3-4 cm with a maximum spherical harmonic degree of 70 (half wavelength resolution of 290 km). Here the mean surface was measured every 10 days and then averaged over a 2-year period. This data set is the most accurate one in existence for vertical control of the sea surface, however, the across-track resolution is only 315 km.

The best current marine geoid comes from **Earth Gravity Model '96 (EGM96)**, a joint global gravitational field model solution produced by the Goddard Space Flight Center and the Defense Mapping Agency. EGM96 is a 360x360 spherical harmonic model that contains almost all known (useful) observations. It includes satellite-to-satellite tracking data, surface gravity observations and satellite altimetry. A 70x70 subset solution that combines satellite tracking, surface gravimetry and altimetry is characterized by a 18-cm global geoid error, making it highly desirable for oceanic applications that require a high accuracy reference surface.

The Future

In the spring of 1997 the NASA Earth System Science Pathfinder program approved the **Gravity Recovery And Climate Explorer (GRACE)**. GRACE will improve the long wavelength geoid and analyze time variable gravity, including subtle but important vertical tectonic signals such as **post-glacial rebound**, which until now have confounded attempts to measure sea level change. GRACE will also provide a measure of changes in continental reservoirs, and thus monitor the input into the ocean, which is an essential component of sea level determination.

Sea Level

Climate models used to study the effects of atmospheric greenhouse gases predict an increase in the overall global temperature of 1°-3.5° in the next century. An increase of this magnitude would have major effects, including melting of the ice caps and thermal expansion of ocean water and associated sea level rise. **Tide gauge** measurements show that sea level appears to have been rising 1.0-2.5 mm yr^{-1} averaged over the past century. The sources of the rise are not well-understood, but likely include contributions from the transfer of (H_2O) mass from the continents to the oceans as well as thermal expansion of seawater.

Table 1. Estimated ranges of likely contributions to sea level rise averaged over the past century (from Warrick et al., 1996)

Contribution	Rate of Increase or Decrease (mm/yr)

Thermal expansion of ocean	0.2 - 0.7
Antarctica	-1.4 - 1.4
Greenland	-0.4 - 0.4
Glaciers and small ice caps	0.2 - 0.5
Surface and ground water	-0.5 - 0.7

Sea level rise = mass influx from ice caps, glaciers, groundwater + thermal expansion of seawater

Note: The 1st term on RHS corresponds to a mass exchange and produces a change in the gravity field. The second term on the RHS involves no net transfer of mass and so has no associated signature in the geoid or gravity field.

--> By comparing satellite gravity observations with tide gauge or altimeter results it should be possible to discriminate between the two RHS contributions.

*** To detect sea level rise will require vertical resolution of a global-scale signal of few tenths of mm/yr.**

Tectonic Signals: Post-glacial Rebound

Obviously, to detect the signature of sea level rise will be a challenge. A critical issue is isolating possible tectonic signatures in the vertical dimension that could conceivably be falsely interpreted as arising due to sea level change. Tectonic signatures due to processes such as orogeny are of order 1 cm yr^{-1} and are now being detected by techniques such as GPS. An obvious potential challenge comes from unloading of the Earth due to **post-glacial rebound**, which has a magnitude in the vertical direction of order mm/yr, comparable to that of proposed sea level change.

During the last ice age, the extensive ice loads over Canada and Scandinavia caused deep depressions in the Earth's surface. Although this ice had mostly vanished by about 9,000 years ago, the depressions have not yet completely disappeared, but are filled by Hudson Bay and the Baltic Sea. The **Earth's surface continues to rise as a result of horizontal inflow of mass from the surrounding mantle**, because over long periods of time, the Earth behaves more as a viscous fluid than as an elastic solid. Post-glacial rebound is the manifestation of this process of readjustment.

The ability of the Earth to flow is what permits the existence of convection in the Earth's mantle. The **primary goal of post-glacial rebound analyses for many years has been to provide constraints on the viscosity profile of the mantle** (though more recently the results have been recognized to have great relevance to sea level change). Workers who model post-glacial rebound, however, have yet to reach a consensus on the results as applied to mantle structure, particularly on the how much the viscosity increases in the lower mantle (*i.e.,* at depths below 670 km). Progress has been hampered by the difficulty of finding observational constraints that are sensitive to lower mantle viscosity but unaffected by other geophysical processes.

For example, the 40-50 mGal depression in the Earth's gravity field over Northern Canada has sometimes been interpreted as the result of post-glacial rebound, but has also been argued to reflect mass anomalies related to mantle convection. Similarly, a secular drift in the position of the Earth's rotation axis, a non-tidal acceleration of the Earth's rotation rate, and a time variation of the Earth's geopotential flattening have all been observed, and could all be due in part to post-glacial rebound. But there could also be important effects from other geophysical processes, including changes in polar ice mass, and from the slow, steady mass-redistribution caused by mantle convective flow.

Crustal uplift rates, as determined mainly from **geological dating of raised beaches**, are probably the **least ambiguous data type** in terms of their relative insensitivity to other processes, and these form the basic observational foundation of post-glacial rebound studies. Still, it is difficult to unravel the competing effects of lower mantle viscosity, upper mantle viscosity, lithospheric thickness, and uncertainties in the time history and spatial distribution of the ice sheet, using this data set alone.

The **elevations of ancient shorelines** around lakes produced by glaciation provided the first information on rebound associated with the great ice sheets. What is desired is an **understanding of the Earth's response to unloading of long wavelength surface loads corresponding to the ice sheets**. In the earliest studies that attempted to measure the time history of uplift to estimate **mantle viscosity**, the models assumed that the Earth's lithosphere behaves as a linear elastic plate with longer term viscous response. The governing rheology is thus that of a **viscoelastic or Maxwell material**. Later, more refined studies attempted to derive a distribution of mantle viscosity over depth that was consistent with the observed uplift history.

Note that for either approach the problem requires knowledge of the spatial extent, thickness distribution and temporal change of the Pleistocene ice sheet, which is considered to be a known input parameter. It should be apparent **that errors in the ice model could map into errors in the inferred lower mantle viscosity**. It is not clear how large the ice load errors of current models might be.

A popular current **deglaciation model** has been the **ICE-3G** model of **Tushingham and Peltier** (1991). This model is constrained by the locations of glacial features (tills, striations, moraines) as well as by observations of regional sea level for various sites. **Fortunately only the large-scale components of the ice model are important for lower mantle viscosity studies**. Thus, the problem of separating ice sheet and viscosity parameters is less of an issue in the interpretation than it is for geological crustal uplift data, where short wavelength terms in the load are more important.

Mantle viscosity estimates have also been made from crustal rebound associated with **Pleistocene lakes**. While ice sheet studies are associated with larger loads and large lateral extent, thus sampling greater lower mantle) depths, lakes have the advantage of having a much better known load. The reason is simple: the top of the lake forms an (approximate) equipotential surface and lake loads can be determined to within a few percent, much better than ice sheet models whose errors are mostly indeterminate. The results from mantle viscosity determined from well-constrained lake models can be taken into account in post-glacial models.

Some early attempts focused on **Lake Bonneville, Utah**, which had a well-preserved stratigraphic record of ancient shorelines. Such models produced solutions that called for a lithosphere thickness of 25-30 km, with an upper mantle viscosity of 10^{18}-10^{19} Pa s. The theory is similar to used in loading studies of oceanic lithosphere by seamounts or other surface loads. Solutions are sought for the flexural equation for an elastic plate

$$D\nabla^4 w + kw = q \quad (12)$$

where ∇^4 is the biharmonic operator, *w* is the vertical displacement, *q* is the applied load, *k* is a foundation modulus *(=gρ)* and

$$D = \frac{Eh^3}{12\left(1-\upsilon^2\right)} \tag{13}$$

is the flexural rigidity which is related to the **effective elastic lithosphere thickness** *(h)*, where ν is Young's modulus. In practice, a viscous term will need to be added to the elastic flexure equation to take into account the influence of the viscous mantle.

The post-glacial rebound of **Hudson's Bay** has been extensively studied and models predict approximately 11 mm yr^{-1} vertical uplift per year. Note that the **rebound signal has to be isolated from other geophysical processes that could cause vertical motion**, such as inflation or deflation due to **ground water**. Such processes are being taken into account in studies of the area. The amplitude of the geoid perturbation from the hydrology in north-central Canada is about 0.2 mm yr^{-1}, small in comparison to the post-glacial rebound signature. In addition, because the hydrological and post-glacial rebound contributions have different spatial patterns, the hydrological effects on the inferred rebound signal would be further reduced.

Some models have started to take into account **horizontal motions** predicted by ice sheet unloading. These are smaller than the vertical component of motion, but still at the level **of several mm/yr**. However, since the signal is small compared to other tectonic motions (recall that plate motions due to seafloor spreading and subduction are cm-10 cm yr^{-1}) and since sea level rise requires knowledge of the vertical component of motion, the question arises as to why the horizontal component of motion is important. The answer is that models of post-glacial rebound seem to have a sensitivity to elastic lithosphere thickness and upper mantle viscosity that assists in the discrimination of Earth structure models. Understanding the horizontal component of motion will become more important as real-time position data are collected in areas once covered by ice sheets in stable platforms that do not undergo other significant horizontal tectonic motions. Predicted horizontal motions for the former Hudson Bay ice sheet are predicted be of order 4 mm yr^{-1} in the direction of the load center with the maximum amplitude 2400 km from the load center.

GPS monitoring of vertical motions along the sea coast in combination with tide gauge observations represent a powerful combination to sort out the sea level problem. Excellent reference frame information from satellite information combined with the ground observations are what is required to sort out the extent of sea level change.