The Earth gravity field in the time of satellites

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1 Introduction

1.1 Satellites and satellite geodesy

On October 4, 1957, Sputnik 1 was launched from the Baikonur cosmodrome (in nowadays Kazakhstan): it was the first artificial satellite ever orbiting around the Earth. The history of satellite geodesy had started. In 1958, observations from Explorer 1 (the first satellite launched by the USA) and Sputnik 2 allowed an accurate determination of the Earth flattening (Buchar 1958; Merson and King-Hele 1958).

The first dedicated geodetic satellite was Anna 1B, launched from Cape Canaveral in 1962 on behalf of the US Navy for geodetic and navigation purposes: the payload, a SECOR (Sequential Collation of Range) instrument, allowed to estimate the coefficients of the main spherical harmonic components of a geopotential model (Nichols 1974).

The first computation of a model of the Earth gravity field exploiting satellite-tracking data was completed by the Smithsonian Astrophysical Observatory, which at the same time computed the position of the observation sta-tions on the Earth surface (Lundquist and Veis 1966). The approach based on which the estimation of the coefficients of the model was performed is still applied nowadays. Of course, new data have been collected and many more are being continuously acquired, while mathematical models used in the computation procedures have become more and more refined, e.g., to take into account also relativistic corrections and the effects of non-gravitational forces.

However, estimating the gravity field from the analysis of satellite orbits data can only provide models below a well-defined spatial resolution; the problem has always been well known from the beginning (Rapp 1997; Barlier and Lefebvre 2001). For this reason, while the

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development of measurement techniques made advances, specific missions for the recovery of the Earth gravity field were devised. In particular, satellite geodesy could greatly benefit from the improvements in accelerometers manufacturing: CHAMP, GRACE and GOCE are the most advanced projects in the field, supplying a fundamental contribution to the knowledge of the geoid and of the physical phenomena occurring over, on and below the surface of the Earth.

In this paper, the main steps in over 50 years of satellite geodesy will be remembered, together with the technological improvements that made it possible to obtain quite extraordinary results in the past few years. At the end, possible ideas for future developments and follow-ons of successful missions will also be described.

2 Studying the Earth gravity field by tracking probe satellites

The principle of physical geodesy based on tracking measurement of a satellite S from a ground station T can be simply described by the following fundamental equation:

$$\underline{x}_T + \underline{\rho} = \underline{x}_S, \tag{2.1}$$

where $\underline{x}_T$ represents the position vector of the ground station, $\underline{\rho}$ is the station-to-satellite range vector resulting from the tracking data and $\underline{x}_S$ is the position vector of the satellite.

Besides, Newton's equation of dynamics (in an inertial reference system) must also be taken into consideration:

$$m_S\ddot{\underline{x}}_S = \underline{g}_S + \underline{f}_S, \tag{2.2}$$

where $\underline{g}_S$ represents the sum of gravitational forces acting on the satellite, while $\underline{f}_S$ is the sum of non-gravitational forces such as atmospheric drag, solar radiation pressure, radiation reflected by the Earth's surface, etc.

The problem unknowns are: the initial conditions (position and velocity of the satellite) and the forces acting on the satellite (Earth's gravitational field, Earth and ocean tides, atmospheric drag, solar radiation pressure, etc.). If the functions connecting measured quantities and parameters are known, the equations can be linearized (provided that approximate values of the parameters are available) and a least squares adjustment can be applied in an iterative procedure, in order to estimate the parameters. In particular, the observations are represented by the satellite motion perturbations, while the Earth's gravity field parameters are the coefficients in a spherical harmonic expansion of the Earth's gravitational potential, representing the mathematical model of the geoid. As it is known (Heiskanen and Moritz 1967), this equation is obtained by solving Laplace's equation in spherical coordinates and can be written as:

$$V(r,\varphi,\lambda) = \frac{GM}{R}\sum_{\ell=0}^{\infty}\sum_{m=0}^{\ell}\left(\frac{R}{r}\right)^{\ell+1}\bar{P}_{\ell m}(\sin\varphi) \times [\bar{C}_{\ell m}\cos m\lambda + \bar{S}_{\ell m}\sin m\lambda] \tag{2.3}$$

where G is the gravitational constant; M and R are the total mass and the mean radius of the Earth; (r,φ,λ) are the satellite geo-centric spherical coordinates; $\bar{P}_{\ell m}(sin\varphi)$ are the fully normalized Legendre's functions of degree ℓ and order m; the $\bar{C}_{\ell m}$,$\bar{S}_{\ell m}$ coefficients represent the parameters to be estimated.

Of course, in order to solve the problem the spherical harmonic expansion has to be truncated at a maximum degree $L_{\max}$; this maximum degree basically depends on the quantity and distribution of the available data and on the type of analyzed observations.

3 Data from satellite tracking: optical, laser and Doppler data

The first applications of satellite data to physical geodesy were based on optical tracking techniques and Doppler techniques to measure range rates; later, it was also possible to exploit laser techniques, as such instruments were being developed to obtain range measurements.

In the case of optical tracking, cameras were used to photograph the position of particular satellites with fixed stars on the background (in order to provide a reference system). It was a purely geometric approach and it needed a large enough satellite to be targeted from the Earth surface. At the time (the years 1960s) telecommunications satellites were launched on orbits with mean height of about 1,600 km; the probes had the shape of balloons and a diameter of some tens of meters, with a highly reflective deployable surface made of Mylar, a PET film 12.7 µm thick. They were the ECHO satellites (see Table 1). Thanks to their particular dimensions and surface, the ECHO satellites could be seen from any position over most of the Earth surface.

Subsequently, a dedicated geodetic program was launched by NASA in 1966 (Teichman 1968): PAGEOS

Table 1 Characteristics of the ECHO and PAGEOS satellites

Satellite	Lifetime	Diameter (m)	Orbital height (km)
ECHO-1	1960–1968	30.50	1,600
ECHO-2	1964–1969	41.10	1,600
PAGEOS	1966–1974	30.48	4,000

(PAssive Geodetic Earth Orbiting Satellite) had similar characteristics to those of the ECHO satellites (see again Table 1 and Fig. 1). However, its higher orbit made it possible to observe the probe from a network of more than 40 stations around the world; in the course of the experiment, also the stations positions were determined with an accuracy of 5–15 m (20 times better than what could be achieved at the time, with terrestrial measurements). This is considered the first global and homogeneous reference system established on the Earth surface.

For a long time, optical observations were the only data from which a satellite position could be determined with a high enough level of accuracy.

At a later stage, the laser technology began to be exploited in satellite geodesy: in this case, the ground stations are equipped with a laser system and the probe is represented by a spherical satellite having the surface covered by retro-reflecting prisms (corner cube reflectors). The observation is primarily constituted by the time needed for the signal to go from ground to the satellite and back; from it, the station-to-satellite distance can be derived. In the first years, the accuracy was about 1.0–1.5 m; nowadays it is as good as a few millimeters, making laser ranging the most accurate satellite tracking technique. By it, not only the satellite position along the orbit can be determined, but also the position of the geo-center, the Earth orientation parameters and the displacement of the ground stations.

Satellite missions based on laser ranging were and are (e.g.):

- STARLETTE (CNES, 1975);
- LAGEOS 1 (NASA, 1976), the first dedicated laser ranging satellite;
- AJISAI (Japan, 1986);
- ETALON 1 e 2 (USSR, 1989);
- LAGEOS 2 (NASA-ASI, 1992), twin satellite of LAGEOS 1;
- STELLA (CNES, 1993);
- LARES (ASI, 2012)

The orbits of such satellites range from about 800–19,000 km. In particular, the LAGEOS 1 and LAGEOS 2 satellites (see Fig. 1) are covered with 426 corner cube reflectors of silica glass; the orbit heights are 5,860 and 5,620 km, respectively; inclinations are 109.84° and 52.64°, respectively.

At the moment, the program of the International Laser Ranging Service (established in 1998) includes observations of more than 40 satellites (Pearlman et al. 2002).

To conclude this short summary of satellite tracking techniques, it must be recalled that during the 1960s the research was developed on radio-transmitting systems where the transmitter is on board the satellite while the receiving station is on the Earth surface; the Doppler effect of the signal is measured. The first system of this kind was TRANSIT, developed for the US Navy: it was the first satellite navigation system, the forefather of GPS (Hofmann-Wellenhof et al. 2001).

4 Estimating Earth gravity field models by satellite tracking and ground gravity data

Based on Eqs. (2.1) and (2.2) and exploiting satellite tracking observations such as those described in the previous chapter, it is possible to obtain "satellite-only"

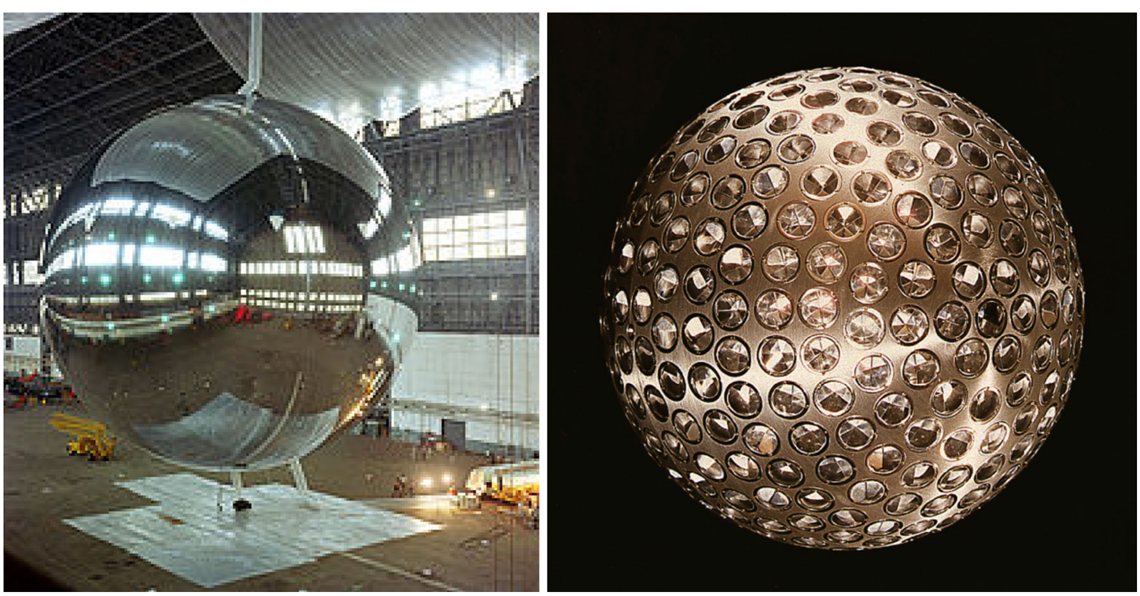


Fig. 1 Satellites for optical and laser observations: *left panel* PAGEOS, *right panel* LAGEOS 1 (both images: NASA)

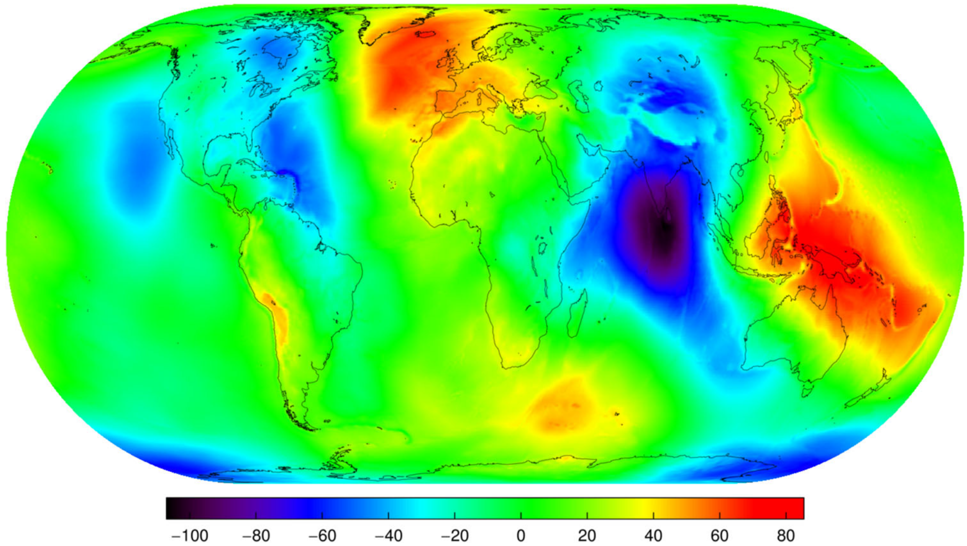


Fig. 2 The EGM96 gravity model represented in terms of geoid heights (m)

gravity models. The first one was computed by the Smithsonian Astrophysical Observatory in 1966 (Lundquist and Veis 1966): it was the SE 1 model (Standard Earth) with harmonic coefficients estimated up to a maximum degree $\ell = 8$. Later, many different gravity models were computed. Among them, the GEM (Goddard Earth Models) series by GSFC (Goddard Space Flight Center) and the GRIM (GRGS and German geodetic research Institute Munich) models estimated by CNES/GRGS (Groupe de Recherche de Géodésie Spatiale) and GFZ (GeoForschungZentrum Potsdam). In more recent years, satellite-only models were highly improved by integrating ground tracking data with satellite-to-satellite tracking data in the so-called high-low mode: the satellite on a low orbit carries a GPS antenna and receiver and is tracked by the GPS constellation satellites. Also data from satellite altimetry could be incorporated in the estimated models. One of the models obtained by combining different types of satellite data is represented by EGM96S (the Satellite-Only Earth Geopotential Model 1996), with maximum degree $\ell = 70$.

As it is evident, models estimated by exploiting satellite data could not achieve higher spherical harmonic degrees, and were just able to catch the low-frequency representation of the Earth gravity field. That was until dedicated gravity missions were conceived in more recent years.

Before then, ground gravity data were necessary to estimate the high frequency part of the gravity signal. However, these observations display some drawbacks: they do not cover homogeneously the Earth surface (just like satellite data) and have a different level of precision with respect to satellite-tracking data. This implies the necessity to define proper weights for the observations during the adjustment procedure, as it was done, e.g., for the EGM96 model, having a maximum degree of $\ell = 360$ (see Fig. 2).

Finally, in order to improve the low-frequency component of the estimated gravity model, the only possible strategy was to devise dedicated satellite gravity missions, capable of supplying observations with homogeneous spatial distribution and precision.

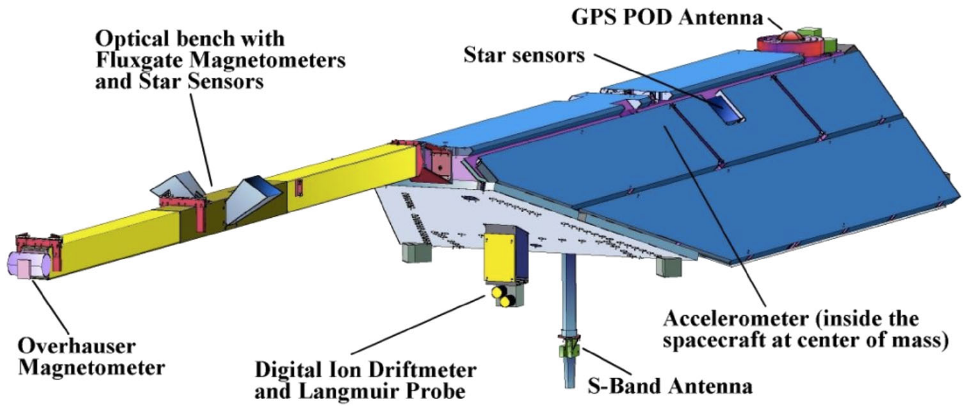
5 Dedicated gravity missions

5.1 CHAMP, GRACE, GOCE

The beginning of the twenty-first century marked a turn in satellite techniques for the observation of the gravity field. Starting form the year 2000, the CHAMP (CHAllenging Minisatellite Payload), GRACE (Gravity Recovery And Climate Experiment) and GOCE (Gravity field and steady-state Ocean Circulation Explorer) LEO (Low Earth Orbit) satellites were launched in rapid succession.

A low orbit means that the signal of the Earth gravity field is still quite strong, but also that strong frictions act on the satellite surface due to the Earth atmosphere and to other forces (albedo, solar radiation pressure, etc.). So it was necessary to have one or more accelerometers available on board to measure non-gravitational forces acting on the satellite: this was achieved with sufficient accuracy only in very recent years. So the history of satellite geodesy is deeply related to that of accelerometry, from the times of the CACTUS (Capteur Accélérométrique Capacitif à Trois axes UltraSensitifs) accelerometer developed in France by Onera (Office National d'Etudes et de Recherches Aérospatiales) for the Castor satellite mission (1975). CACTUS could measure accelerations with an accuracy of 10^{-9} m/s^2 (Beaussier et al. 1977). Nowadays, the accelerometers that constituted part of the payload of the GOCE mission measured accelerations of the order of 10^{-12} m/s^2.

Fig. 3 The CHAMP mission satellite and payload (image: GFZ)



5.2 CHAMP

CHAMP was the first mission ever dedicated to the observation of the Earth gravity field. It was conceived by GFZ (GeoForschungsZentrum) in Potsdam (Germany) and launched on July 15, 2000 from the Russian cosmodrome of Plesetsk on a quasi-circular, quasi-polar orbit ($i = 87°$) at an altitude of 454 km. The mission ended on September 19, 2010, after more than 10 years and 60,000 orbital revolutions around the Earth, while its lifetime had been expected not to exceed 5 years.

The payload was constituted by: a GPS antenna and receiver, an accelerometer, a star camera, a corner cube prism for satellite tracking from ground and a magnetometer for the estimate of the magnetic field of the Earth (see Fig. 3).

CHAMP was based on the principle of SST-hl, that is Satellite-to-Satellite Tracking in the high-low mode. The position of the satellite was tracked not only exploiting laser measurements from ground, but also GPS measurements, thanks to the on-board antenna and receiver. Non-gravitational forces were determined by means of the accelerometer.

The observations acquired allowed to estimate several models of the geoid with a commission error of the order of 5 cm at harmonic degree $\ell = 50$ (spatial resolution of about 400 km). One of such models is EIGEN-CHAMP03S (see Fig. 4), computed by GFZ and CNES/GRGS, incorporating 33 months of data and reaching a full power resolution equal to $\ell = 60$ (Reigber et al. 2002, 2004). A subsequent release of this solution was EIGEN-CHAMP05S incorporating 72 months of data and extending up to degree $\ell = 150$ (Flechtner et al. 2010); other CHAMP-only solutions were computed, e.g., by Bern University (Prange et al. 2009). These solutions were obtained by applying a regularization procedure to the normal system of equations.

These models are of lower resolution than the ones estimated from subsequent gravity missions (GRACE and GOCE), but it can be asserted that the mission was valuable in successfully testing the functioning of electrostatic accelerometers when flying on a low orbit.

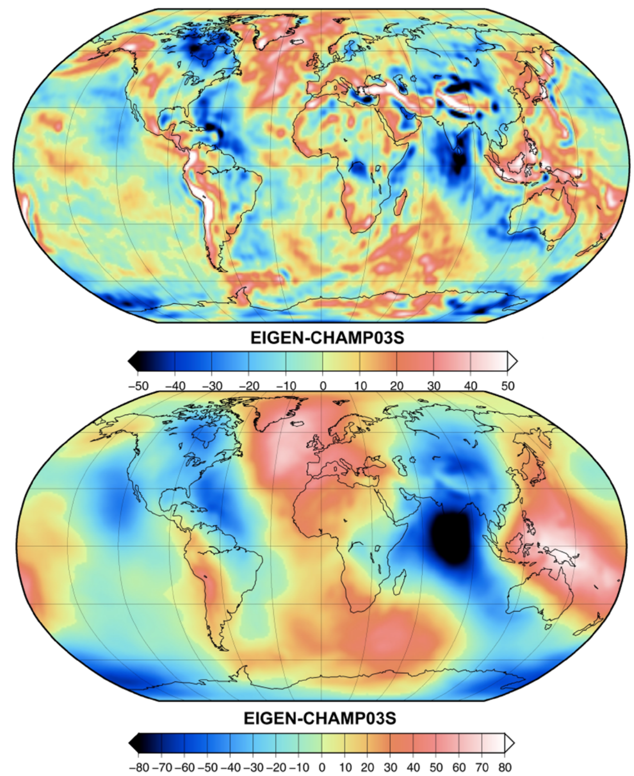


Fig. 4 The EIGEN-CHAMP03 model represented in terms of gravity anomalies (mgal) (*upper panel*) and geoid undulations (m) (*lower panel*)

5.3 GRACE

GRACE (Gravity Recovery And Climate Experiment) is an American–German mission planned and designed by NASA (National Aeronautics and Space Administration) and DLR (Deutsche Forschungsanstalt fur Luft und Raumfahrt). It consists of twin satellites flying on a quasi-circular, quasi-polar orbit ($i = 89°$) at an initial altitude of about 494 km and at a reciprocal distance of about 220 km. Both satellites were launched on March 17, 2002, from Plesetsk; the mission lifetime, expected to be about 5 years, is now foreseen to end in 2015–2016.

Each GRACE satellite is equipped with GPS antenna and receiver, a triaxial electrostatic accelerometer, a star camera and a distance meter in K-band. The last instrument is the one that allows to perform SST measurements in the low–low mode (SST–ll): this means that gradiometric measurements are realized by the pair of spacecrafts continuously tracking each other (see Fig. 5).

The mission data allowed so far to estimate global geoid models with a commission error equal to or below 1 cm at harmonic degree $\ell = 120$ (spatial resolution of about 170 km). In particular, it has been possible to estimate also time variations of the gravity field. Among GRACE mod-els, there are the EIGEN series computed at GFZ Potsdam and the GGM series computed at the University of Texas. The GRACE Gravity Model 03 (GGM03S) is represented in Fig. 6. Published in 2008, it is based on 4 years of GRACE observations (from January 2003 to December 2006) and has a harmonic resolution equal to $\ell = 180$. The GGM03C model (2009) also includes radar altimetry data and ground gravity data and has a harmonic resolution equal to $\ell = 360$.

Moreover, the ITG-GRACE03 model computed at the University of Bonn (Mayer-Gürr 2006) was selected to be combined with ground gravity data in the estimate of the most largely used gravity model nowadays, namely EGM2008 (Pavlis et al. 2008).[1]

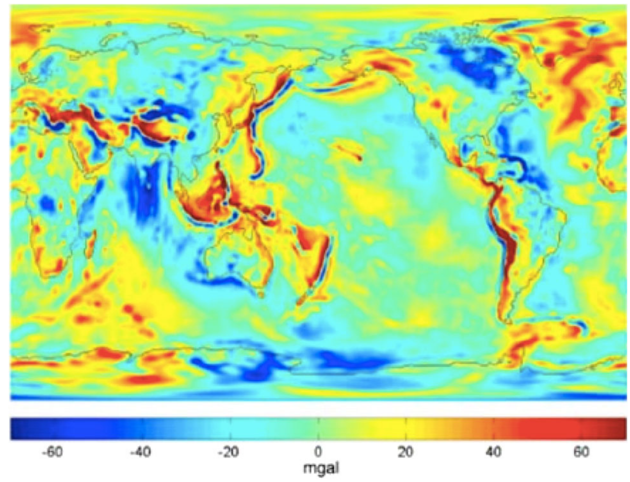


Fig. 6 The GRACE GGM03S model represented in terms of gravity anomalies (mgal)

Among the applications that benefit from GRACE results, hydro-geological studies at a regional scale should be cited, such as the ones on monthly water accumulation in the Amazon River and Orinoco basins which are based on the monthly variations of the values of gravity anoma-lies, such as delivered by GRACE (see Fig. 7) (Tapley et al. 2004).

Other remarkable results from GRACE observations were represented by the evident signature in gravity anomalies variations after megathrust earthquakes, such as the Sumatra earthquake of December 26, 2004 (see Fig. 8) (Han et al. 2006), and the estimates of the mass balance in Antarctica and Greenland ice sheets (Ramillien et al. 2006).

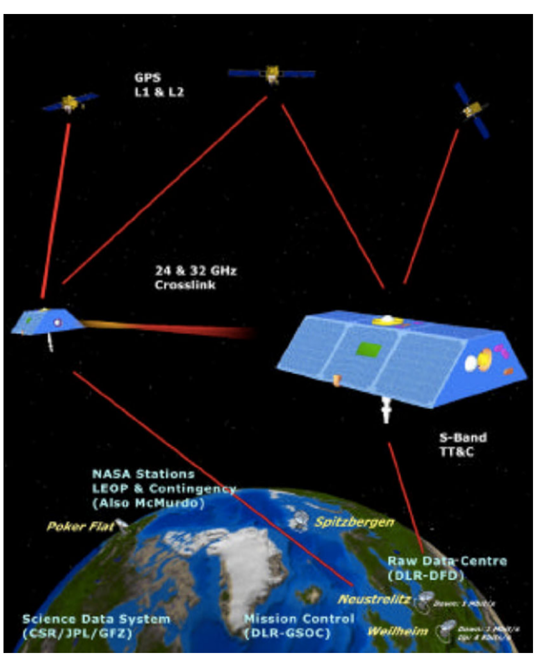


Fig. 5 The GRACE mission concept (image: NASA)

5.4 GOCE

The GOCE mission of the European Space Agency (Floberghagen et al. 2011) represents the most expensive and sophisticated project ever devised to study the Earth gravity field. The satellite was launched from Plesetsk on March 17, 2009, on a quasi-circular, quasi-polar orbit

[1] The spherical harmonic model of the Earth's gravitational potential EGM2008 was estimated by least squares combination of the ITG-GRACE03S gravitational model and gravitational information obtained from a global set of mean free-air gravity anomalies given on a 5 arc-minute equiangular grid. This grid was estimated by merging terrestrial, altimetry-derived, and airborne gravity data. EGM2008 is complete to degree and order 2159, plus additional coefficients up to degree 2190 and order 2159. Over areas having good gravity data coverage, the discrepancies between EGM2008 geoid undulations and GPS/Levelling derived undulations are on the order of ±5 to ±10 cm.

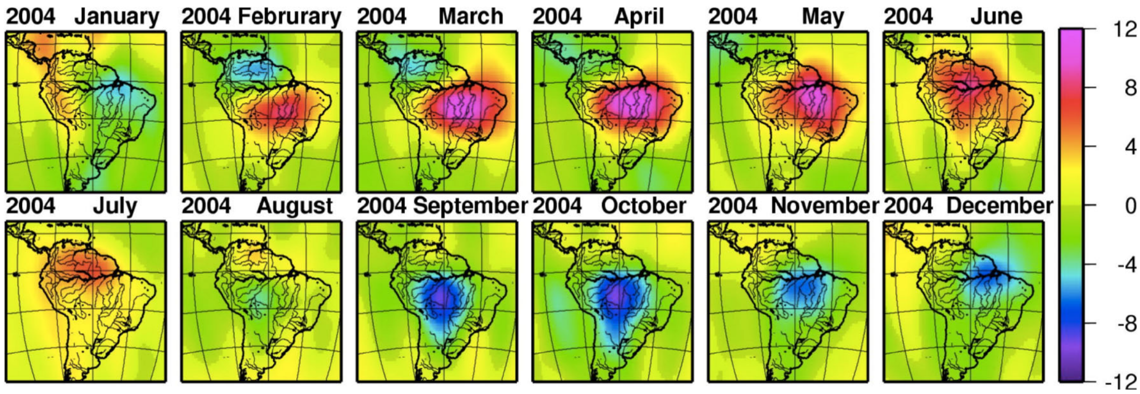


Fig. 7 Monthly variations of the gravity anomaly (mgal) in the Amazon River basin from GRACE data in 2004

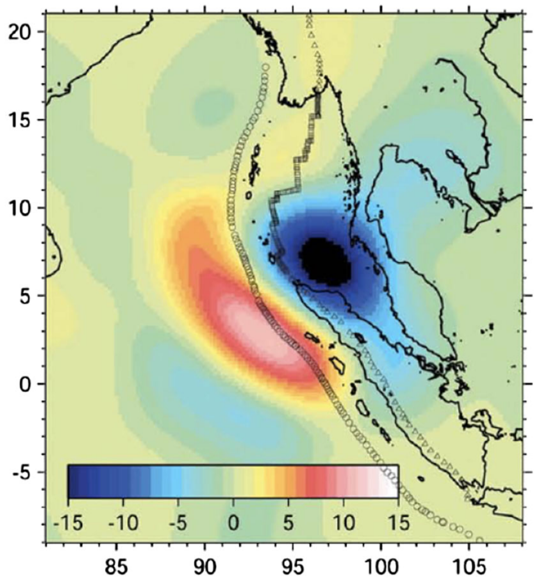


Fig. 8 Variation of the gravity anomaly (mgal) due to the Sumatra earthquake (Mw = 9.3)

Fig. 9 The distinctive shape of the GOCE satellite (image: ESA)

($i = 96.7°$) at an altitude of about 250 km. The mission lifetime ended after more than four and a half years on November 11, 2013 (it had been expected to be no longer than 2 years).

It was the first (and so far the only one) satellite mission based on the principle of gradiometric measurements. In fact, among other instruments such as a GPS antenna and receiver, star camera and corner cube prism, the GOCE spacecraft (see Fig. 9) carried an array of three pairs of tri-axial electrostatic accelerometers arranged along three orthogonal axes, constituting a gradiometer allowing to determine the tensor of the second derivatives of the gravitational potential. It must also be mentioned that GOCE was equipped with a Xenon ion propulsion engine, to counter act the effect of atmospheric drag in almost real time.

The observations from the GOCE mission have been treated according to different data analysis strategies, namely the so-called direct approach, time-wise approach and space-wise approach (Migliaccio et al. 2004; Pail et al. 2011). The first GOCE-only geoid model, computed in July 2010, is shown in Fig. 10. The latest models extend up to harmonic degree $\ell = 280$, and have geoid undulation accuracy of the order of 1–2 cm and gravity anomaly accuracy of the order of 1 mgal up to harmonic degree $\ell = 200$ (spatial resolution of about 100 km). The pro-gressive improvement from the first to the last release of the time-wise model is displayed in Fig. 11, mainly depending on the increasing number of processed data and, in the case of the last release, also on the lower orbits of the GOCE satellite.

GOCE models, attaining a quite higher resolution than the ones from CHAMP and GRACE, allow for many interesting applications, such as the possibility to define a unique height datum at a global level and to improve geoid modeling at a local scale. As an example, the case study of the Italy–Switzerland border is reported where local geoids can differ by as much as 40 cm (see Fig. 12), mainly due to height datum inconsistencies. Using GOCE models a

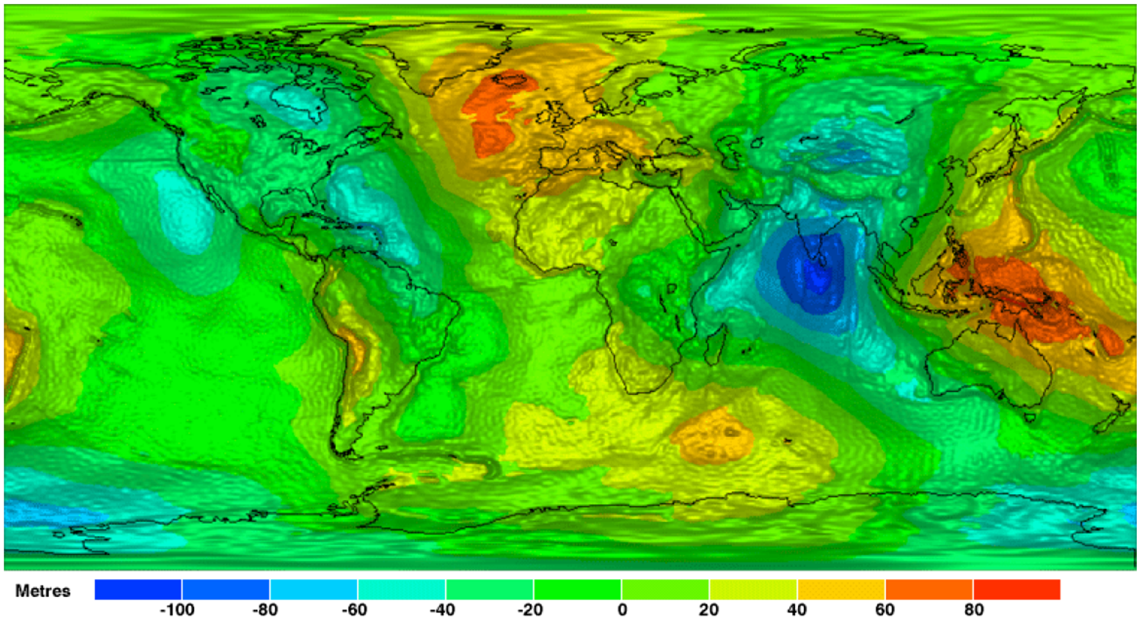


Fig. 10 The first GOCE gravity model computed in July 2010, geoid heights in (m)

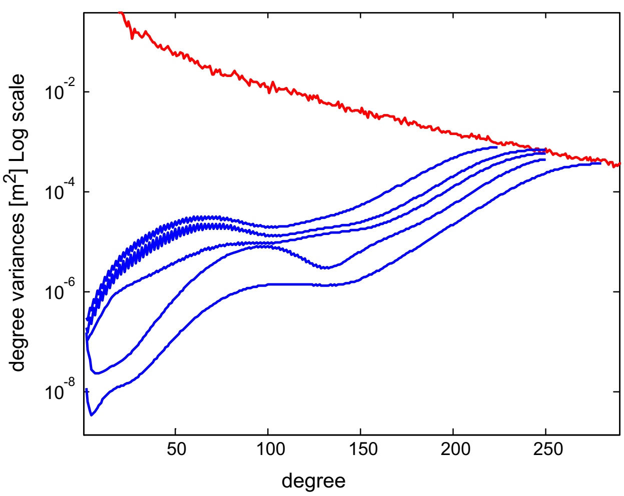


Fig. 11 Improvements of error degree variances (m^2) of the GOCE gravity models from the first to the last time-wise release (from the higher to the lower *blue curve*). EGM2008 empirical signal degree variances in *red* (color figure online)

procedure can be set up to estimate and remove this discrepancy (Gilardoni et al. 2013).

6 Data and applications from satellite altimetry

Going back to the 1980s, it must be recalled that other satellite missions were launched based on different measuring techniques, designed for oceanographic and geophysical purposes, to derive models of the ocean circulation and tectonics. However, also geodesy could take a great advantage from the observations acquired over the oceans by radar-altimetric satellites such as Topex/Poseidon and Jason1/Jason2 (to mention the most famous ones). The first radar-altimeters, flying on an orbit at an altitude of about 800 km, allowed to measure the distance of the satellite from the sea surface with a relative precision of about 5 cm. The satellites were equipped with a radar-altimeter, two radars for infrared and microwaves wavelengths, a synthetic aperture radar (SAR) and an infrared image sensor. The data for the determination of the geoid and of the ocean bathymetry were measured by that radar-altimeter. In Fig. 13, the principle of radar altimetry is illustrated.

The observation equation of radar altimetry can be written as:

$$\mathrm{SSH} = N + \mathrm{DOT} = r_S - \rho - r_e \tag{6.1}$$

where SSH is the Sea Surface Height with respect to the ellipsoid, N is the geoid undulation, DOT is the Dynamic Ocean Topography, r_S is the radial distance of the satellite (computed from the ephemerides of the satellite), r_e is the radial distance of the sub-satellite point on the ellipsoid (known from the ephemerides of the satellite) and finally ρ is the satellite–sea surface range (measured) (Rummel 1993).

Thus, altimetry cannot supply the estimates of N and DOT independently, but just their sum. So, for oceanographic purposes an accurate model of geoid is needed to derive the DOT, while for geodetic purposes a good model of the DOT must be available in order to apply inversion techniques to estimate values of Δg over the oceans. In fact, estimating gravity anomalies over the sea surface amounts to solving Stokes' inverse problem to obtain

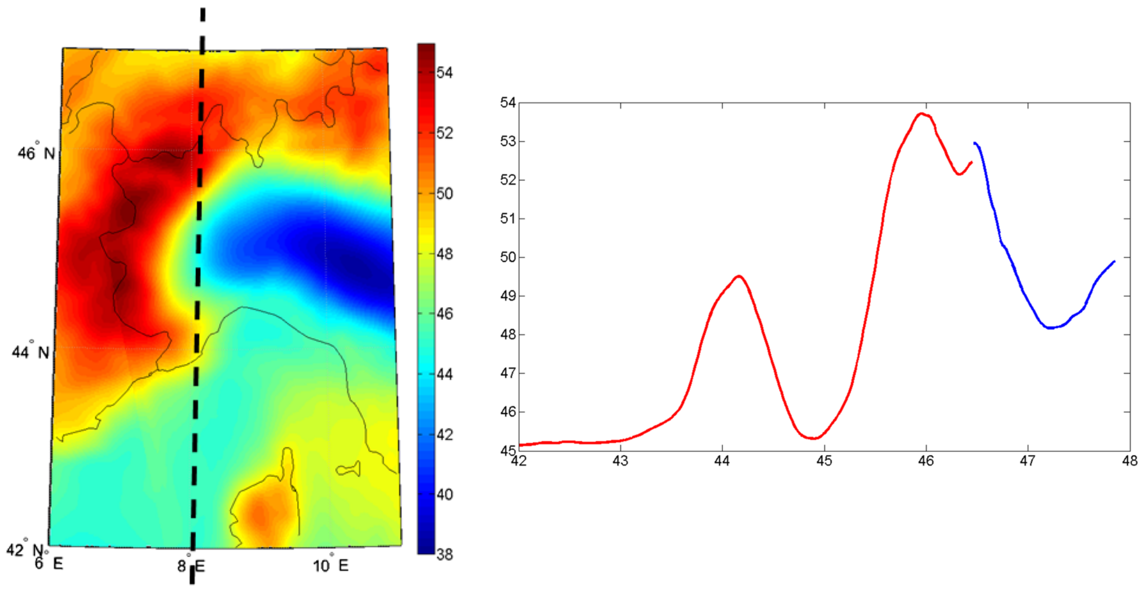


Fig. 12 Height datum problem when comparing Italian and Swiss local geoids (m): the profile along the meridian indicated with a *dash line* on the *left panel* is shown on the right with a *red line* for the Italian geoid and a *blue line* for the Swiss one (color figure online)

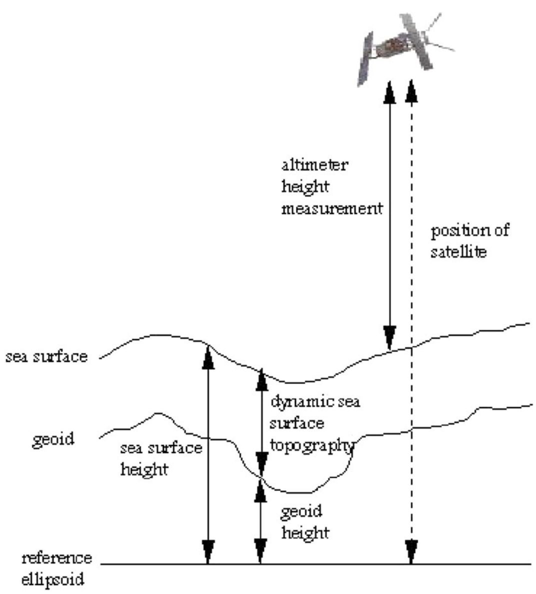


Fig. 13 The radar altimetry observation concept

models such as the DTU10 model (Andersen 2010) computed by the National Space Institute of Denmark (see Fig. 14).

The same Institute also derives estimates of the dynamic ocean topography at a global scale, by subtracting the GOCE geoid model from altimeter observations. Furthermore, the components of the geostrophic velocity (v_E, v_N) can be computed by applying the following equations:

$$v_N = -\frac{g}{f}\frac{1}{R}\frac{\partial E_{t(\mathrm{DOT})}}{\partial \vartheta} \qquad v_E = \frac{g}{f}\frac{1}{R\sin\vartheta}\frac{\partial E_{t(\mathrm{DOT})}}{\partial \lambda} \tag{6.2}$$

where R = mean Earth radius; $f = 2\ \omega\ \cos\vartheta$ = Coriolis term; ω = Earth rotation ratio; ϑ = co-latitude; $E_{t(\mathrm{DOT})}$ = time average of the DOT.

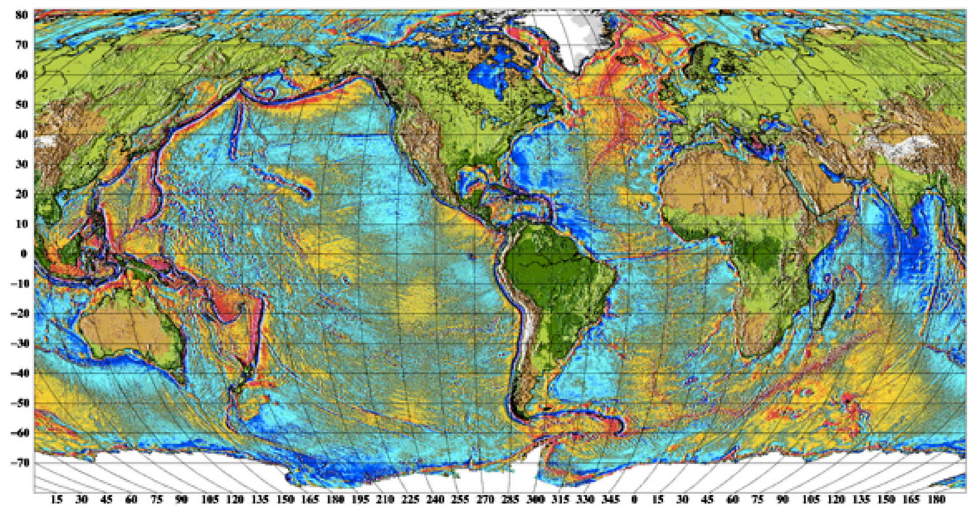
In Fig. 15 an example of a map showing geostrophic current velocities is reported, for the area of the Drake Passage. The results obtained from drifter data are compared to those obtained from radar altimetry: the rms of the differences is of the order of 0.06 m/s.

7 Future perspectives and concluding remarks

As for future gravity missions, both the GRACE and the GOCE measurement principle are under revision in order to improve the presently achieved results.

In the former case, the main step forward is the use of the laser interferometry instead of microwaves to measure the inter-satellite distance. Furthermore, satellite constellations different from the standard couple of chasing satellites are under investigation. These constellations should allow to significantly improve resolution and accuracy of the estimated gravity field (both for the static and the time variable part). However, the practical implementations of these ideas are quite expensive and hard to realize, e.g., for the need of continuously changing the orientation of the laser beam depending on the reciprocal position of the satellites constituting the flight constellations. A much easier solution from the technical point of view is to launch two couples of GRACE-like satellites flying on two circular orbits with different inclinations (Bender et al. 2008). This solution will anyway guarantee an improvement of the

Fig. 14 Free air gravity anomalies from satellite altimetry: the DTU10 model, spatial resolution $1' \times 1'$



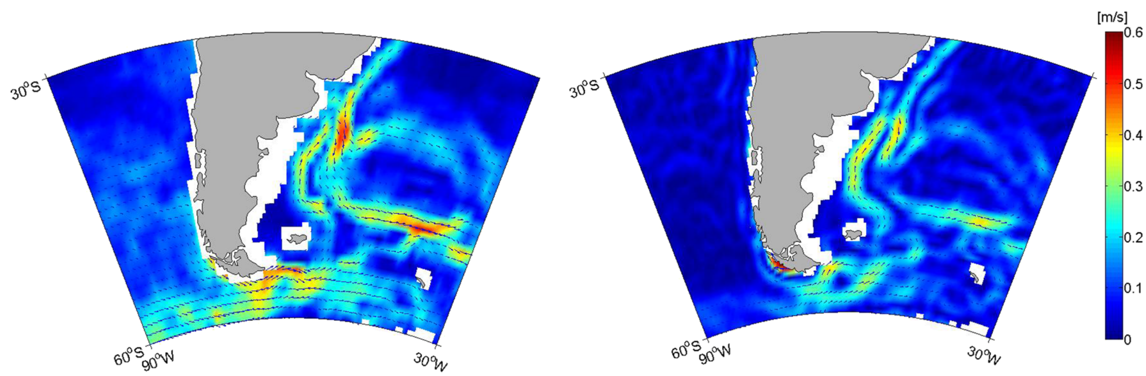


Fig. 15 Geostrophic current velocities in the area of the Drake Passage; *left panel* estimated from drifter data; *right panel* estimated from radar altimetry

present knowledge of the gravity field and it is actually the most likely candidate for a GRACE follow-on.

In the case of satellite gradiometry, researches are mainly oriented towards the atom interferometry (Peters et al. 1999). By exploiting this technology it could be possible to build gradiometers with about 1 $\mu E/\sqrt{Hz}$ accuracy, namely three orders of magnitude better than the electrostatic gradiometer on board GOCE. However, atom interferometry at the current state-of-art would require further developments before being launched in space, e.g., in the miniaturization and in the integration of optical and laser components.

These dedicated gravity missions will lead to high precision estimates of the gravity field. As an example, the GRACE follow-on mission by NASA-GFZ, to be launched in August 2017, will provide enhanced gravity field estimates both in the static field and in the time varying component (Bryant et al. 2012). The outcomes of this and other future missions will have strong impacts on many ongoing researches such as those related to the height system unification. Furthermore, the procedures for computing regional and local geoids will benefit of satellite-based global estimates that will provide more reliable low–medium frequency reference models to be used in the remove-restore procedure.

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