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High-resolution residual geoid and gravity anomaly data of the northern Indian Ocean – An input to geological understanding

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ABSTRACT

Geoid data are more sensitive to density distributions deep within the Earth, thus the data are useful for studying the internal processes of the Earth leading to formation of geological structures. In this paper, we present much improved version of high resolution ($1' \times 1'$) geoid anomaly map of the northern Indian Ocean generated from the altimeter data obtained from Geodetic Missions of GEOSAT and ERS-1 along with ERS-2, TOPEX/POSIDEON and JASON satellites. The geoid map of the Indian Ocean is dominated by a significant low of -106 m south of Sri Lanka, named as the Indian Ocean Geoid Low (IOGL), whose origin is not clearly known yet. The residual geoid data are retrieved from the geoid data by removing the long-wavelength core–mantle density effects using recent spherical harmonic coefficients of Earth Gravity Model 2008 (EGM2008) up to degree and order 50 from the observed geoid data. The coefficients are smoothly rolled off between degrees 30–70 in order to avoid artifacts related to the sharp truncation at degree 50. With this process we observed significant improvement in the residual geoid data when compared to the previous low-spatial resolution maps. The previous version was superposed by systematic broad regional highs and lows (like checker board) with amplitude up to ±12 m, though the trends of geoid in general match in both versions. These methodical artifacts in the previous version may have arisen due to the use of old Rapp's geo-potential model coefficients, as well as sharp truncation of reference model at degree and order 50. Geoid anomalies are converted to free-air gravity anomalies and validated with cross-over corrected ship-borne gravity data of the Arabian Sea and Bay of Bengal. The present satellite derived gravity data matches well with the ship-borne data with Root Mean Square Error (RMSE) of 5.1–7.8 mGal, and this is found to be within the error limits when compared with other globally available satellite data. Spectral analysis of ship-borne and satellite data suggested that the satellite gravity data have a resolution down to 16–18 km. Further, the geoid, residual geoid and gravity anomalies are integrated with seismic data along two profiles in the Bay of Bengal and Arabian Sea, and inferences have been made in terms of density distributions at different depths. The new residual geoid anomaly map shows excellent correlation with regional tectonic features such as Sunda subduction zone, volcanic traces (Chagos–Laccadive, Ninetyeast and 85°E ridges) and mid-ocean ridge systems (Central Indian and Carlsberg ridges).

1. Introduction

Geoid is the equi-potential surface of gravity which best approximate the mean sea level in the absence of external gravitational field, and undulations of this surface from the reference ellipsoid are called geoid anomalies. These anomalies can be computed by removing the mean dynamic ocean topography from the time averaged ocean surface height measurements of satellite altimeter data sets. High spatial and temporal resolution of the geoid data are achieved by combining the altimeter data obtained from Geodetic and Exact Repeat missions. While, the geoid models computed directly using GRACE and CHAMP satellites are accurate only wavelengths >160 km (Tapley and Kim, 2001), the precise measurements (accuracy of $\sim$1 cm at 800 km) of short wavelength variations in the Sea Surface Height (SSH) by the satellite altimeters allow us to compute the high-resolution geoid.

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The geoid data are, in general, dominated by very long-wavelength (>3000 km) anomaly components contributed by the deep seated sources. In order to study the lithospheric-scale structures, these long-wavelength core–mantle density effects have to be removed from the geoid data. For this purpose, the spherical harmonic coefficients of reference geoid models are considered. The resulting geoid is then generally referred to as the residual geoid, which are sensitive to the anomalous density distributions lying within the lithosphere and generally correspond to mantle convection processes (McKenzie et al., 1980). However, care must be taken while removing the long-wavelength reference geoid as it significantly affects the amplitude and wavelength characteristics of the residual geoid anomalies. Sandwell and Renkin (1988) cautioned that improper removal of long-wavelength reference geoid can generate regular pattern of highs and lows in the residual geoid data and highlighted an example of such data artifact in the Pacific Ocean which was attributed to the mantle convection processes by McKenzie et al. (1980).

In this paper, we made an attempt to improve the existing geoid and residual geoid anomaly maps of the northern Indian Ocean (Majumdar and Bhattacharyya, 2004; Majumdar et al., 2006 hereafter called as previous maps) by incorporating a refined data reduction procedure that dwells on the enhanced resolution of recent Earth Gravity Model-2008 (EGM2008) and adept truncation of model coefficients for the removal of long-wavelength reference geoid. Thus, the major objectives of the present study are to (1) to present updated high resolution ($1' \times 1'$) geoid and residual geoid anomaly maps of the northern Indian Ocean, (2) demonstrate the efficacy of the present maps with the 'previous maps', (3) use the geoid data to generate high-resolution gravity anomalies, and validate them with ship-borne gravity anomalies and (4) finally, to carry out a preliminary analysis of these maps together with a comparative study of ship–track as well as satellite derived free-air gravity anomalies and the residual geoid anomalies along two regional multi-channel seismic reflection profiles across the Bay of Bengal and Arabian Sea.

2. Data and methodology

Geodetic Missions of GEOSAT and ERS-1 along with ERS-2, TOPEX/POSIDEON and JASON satellites (Andersen and Knudsen, 2009) are utilized for recovery of geoid and gravity data over the Indian Ocean. The procedure for retrieval of gravity from the altimeter data is briefly explained in Fig. 1. Two types of corrections (i) instrumental and (ii) environmental, are necessary for the altimeter range measurements. The instrument corrections account for the Doppler-shift errors, oscillator-shift errors, acceleration errors, etc. The sources of instrument errors and their effect on the SSH measurements for different satellites are described in their respective Geophysical Data Records (GDRs). The environmental corrections account for the path-delay in travel time due to water vapor (wet tropospheric correction), dry gases (dry tropospheric corrections) and electrons in ionosphere (ionospheric correction). Apart from these atmospheric effects, the returned altimeter is affected by instantaneous effects of the sea surface due to tidal forces and pressure of specular reflectors, winds and waves. The errors in SSH due to ocean tides are corrected by interpolating surface tide height obtained from various tidal models along satellite track (Andersen and Knudsen, 2009). The atmospheric pressure loading (inverse barometric effect) is corrected using a global mean pressure of 1013 mbar (Andersen and Knudsen, 2009). Re-tracking of the raw waveforms of GM altimeter data is generally considered to improve the coverage and hence resolution of the geoid or gravity (Maus et al., 1998; Bansal et al., 2005; Sandwell and Smith, 2009). The mean sea surface was derived using re-tracked products

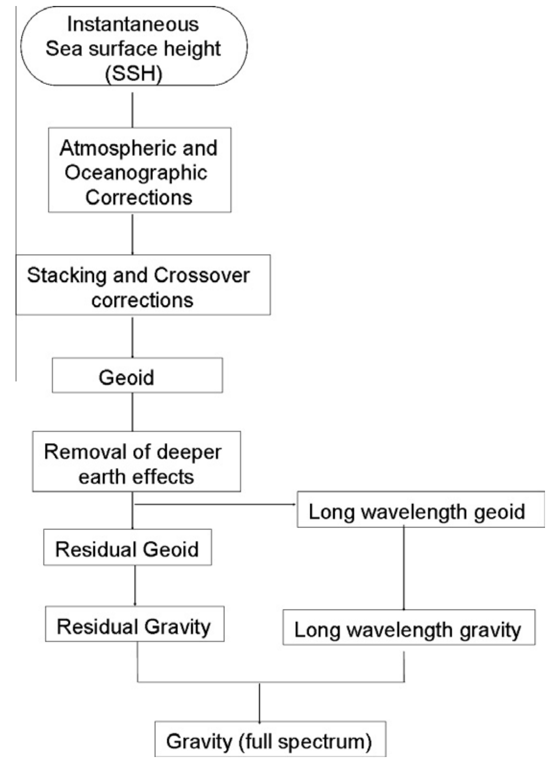

Fig. 1. Processing steps involved in Marine geoid and gravity retrieval from the satellite altimeter data.

of GEOSAT and ERS-1 data sets. The ERS-1 GM data were double retracked using the rule-based expert re-tracking system which has significantly enhanced the number of data, particularly in coastal regions (Andersen et al., 2010) and probably remove the spurious jumps in ERS-1 profiles as reported earlier (Bansal et al., 2005). The re-tracked GM and mean EM have been corrected for seasonal, orbital and other long-wavelength effects by cross-over adjustments with minimum variance criteria (Andersen and Knudsen, 2009). Finally, the geoid is determined by removing mean dynamic topography from the time averaged mean sea-surface data. The effect of environmental errors in SSH measurements and details of the models used for the corrections are different for different satellite systems. These are discussed in detail in Andersen and Knudsen (2009) and hence not reproduced here.

In the present study, the EGM2008 coefficients (Pavlis et al., 2008) up to degree and order 50 are utilized to remove the long-wavelength components of the geoid to obtain the residual geoid. As the sharp truncation of spherical harmonic coefficients at a particular degree would cause side lobes in the spatial structures of the truncated fields (create artifacts), the EGM2008 coefficients are smoothly rolled off between degrees 30–70. The gentle cutting of spherical harmonic coefficients is achieved by smooth rolling of function $f(n)$ from 1 to 0 between the lower to higher orders. This can be represented as $f(n) = \left(\frac{n-n_a}{n_b-n_a}\right)^4 - 2\left(\frac{n-n_a}{n_b-n_a}\right)^2 + 1$.

Where n_a and n_b are lower and higher orders (ICGEM, http://icgem.gfz-potsdam.de/ICGEM/potato/gentlecut_engl.pdf).

The conversion of geoid grid to gravity anomaly is carried out using 2D Fourier transform method (Haxby et al., 1983). EGM2008 geoid model were removed from the altimeter derived geoid data before transforming to gravity data. In order to avoid the edge effects, the Fourier transform is applied for extended area of 1° on either side. The full spectrum gravity field is recovered by restoring the gravity effect of the EGM2008 geoid field.