DTU Space
National Space Institute

CP40

Algorithm Theoretical Baseline Document - Sea Floor Bathymetry

Ole Baltazar Andersen,

ATBD Version 0.2

August, 31th 2014

DTU Technical University of Denmark

Algorithm Theoretical Baseline Document - Sea Floor Bathymetry

Ole Baltazar Andersen

DTU Space,

Elektrovej bldg. 328.

DK-2800 Lyngby

Denmark

Email: oa@space.dtu.dk

Document history.

Version	Date	Comments
0.8	31th August	

Table of Contents

----

1. Background

This task deals with the use of Cryosat-2 SAR data to detect ocean bathymetry features, in particular the mapping of sea mounts, which should benefit from the finer along-track resolution of SAR altimeters and the closely spaced tracks of the Cryosat-2 mission.

As for the Coastal Ocean, the SAR re-tracking needed for mapping sea mounts will be the same as for Open Ocean. The addition to the processing is the need for further along-track processing and filtering to achieve the signal to noise needed to retrieve small gravity signals associated with underwater bathymetry.

The study has been performed in a region of the Pacific Ocean determined and installed during the CP40. As the across-track density of the sea surface height information is a crucial parameter for the accuracy of the derived bathymetry a full cycle of Cryosat-2 (369 days) of observations had to be awaited.

2. Region

The region used for this investigation was selected after a number of considerations and discussions in the project and it was decided to use a region close to Hawaii bounded by the geographical coordinates: 15-25N and 180-203E. The region is shown in Figure 1. The region was chosen due from the following considerations:

- Region should be a minimum of 200 by 200 km big
- Region should be covered with C2 SAR for AT LEAST 369 days
- No sea ice present in the area
- Region should not be too deep as upward continuation dampens short wavelength gravity.
- Region should contain strong geophysical signals.
- Region should contain a lot of marine gravity for validation.

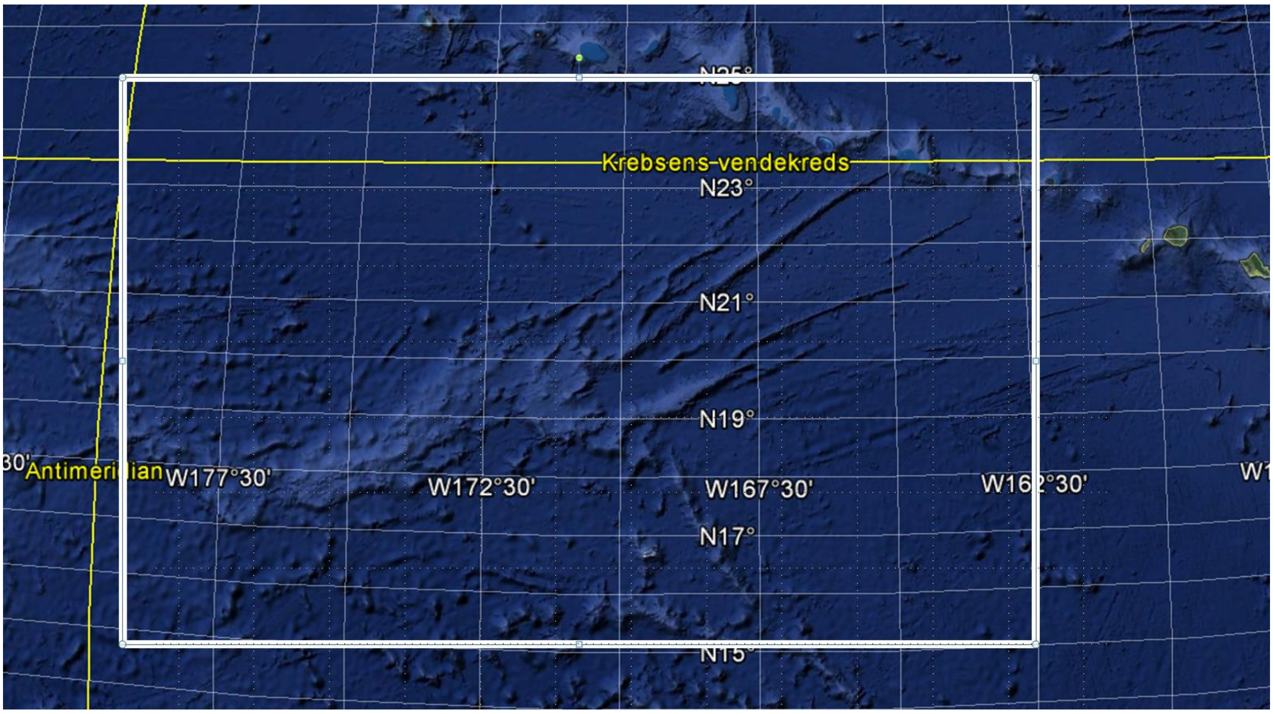

Figure 1: Region of investigation around the Hawaii'an island chain.

Figure 2 shows a zoom into the northern part of the region. This region is generally relative deep (4000 m) with the ocean bottom standing around 4-5 km below the surface. However, a number of prominent and named underwater sea mounts are showed.

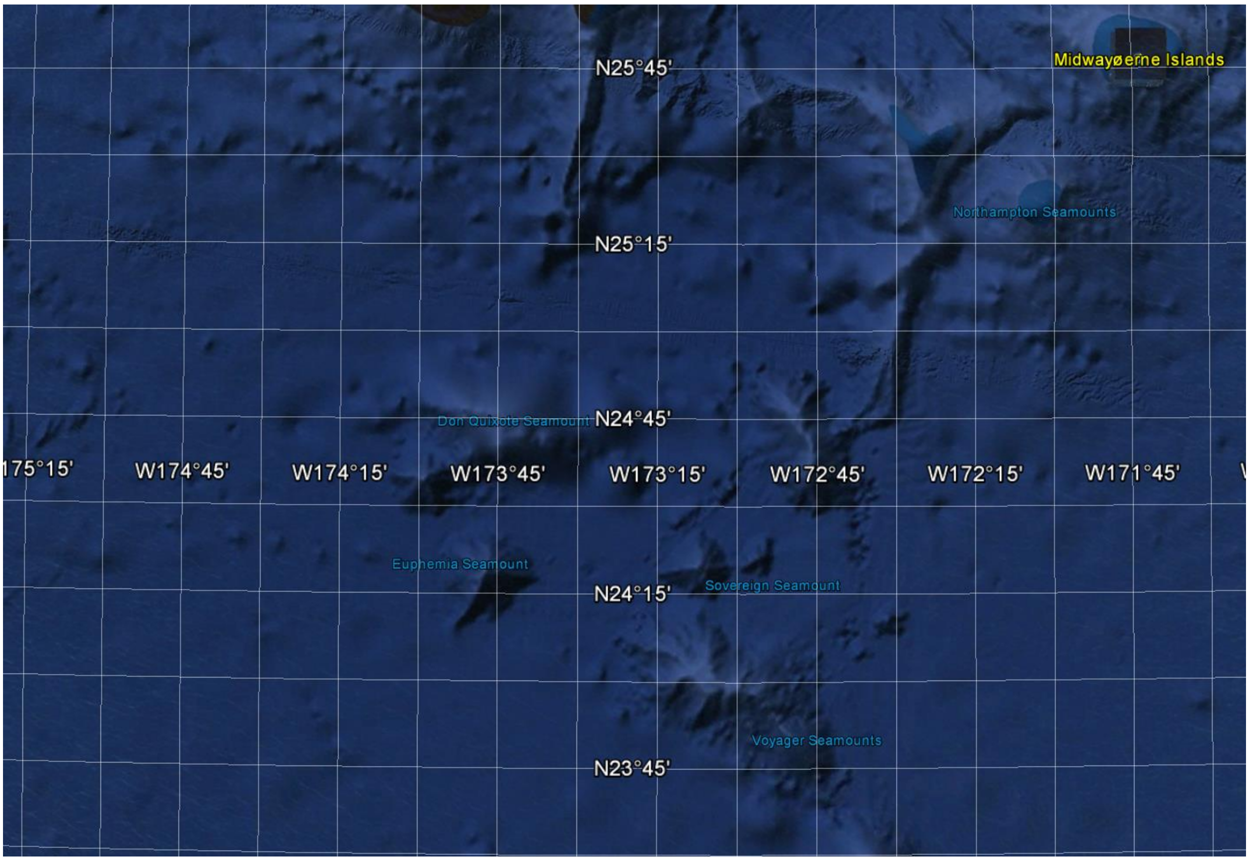

Figure 2: Northern part of the region of investigation around Hawaii Island chain. The Midway Island is seen in the upper right corner. A number of prominent and named underwater sea mounts are showed.

Two well-known limitations to bathymetry estimation from gravity should be mentioned here.

One is the phenomenon called isostacy which reveals the mechanics of the Earth tectonic plates. This reflect that the Earth's crust only has a finite strength to support overlaying material and that longer wavelength features will be isostatically compensated and in effect contain less mass that their surface topography. Various studies by i.e. Smith and Sandwell, 1994; 1997 concluded that for wavelength shorter than 100 km isostatic compensation is not an issue. In our processing of the data we consequently remove wavelength longer than 100 km and do not consider these in the processing.

While longer wavelength are isostatically compensated the short wavelength are affected by another phaenomenon. This is upward continuation which is a result of Newton's law that the gravity field falls off by the square of the distance between the source and the observer. This means that features on the sea bottom will be isostatically compensated as also demonstrated in the SAMOSA study for various sizes of ocean-bottom features (Andersen, 2011)

In essence anomalies with wavelength that are long compared to the mean water depth will suffer little attenuation while those that are shorter than pi times the water depth will suffer attenuation becoming stronger the shorter the wavelength are. Consequently very small sea mounts cannot be recovered from altimetric gravity if the water depth is too large.

Consequently the mean water depth is a limiting factor in the determination of bathymetric features and as mentioned then it is important to choose a region of moderate depth to study the impact of SAR altimetry on sea floor prediction.

Figure 3 illustrates the southern part of the region with only the Johnston Atoll named. A high amount of sea mounts are found and mapped throughout the region. Furthermore the northern part of this region only stands at a depth between 2 and 3 km making it a better test are for the use of SAR altimetry or investigation of mapping of smaller sea mounts.

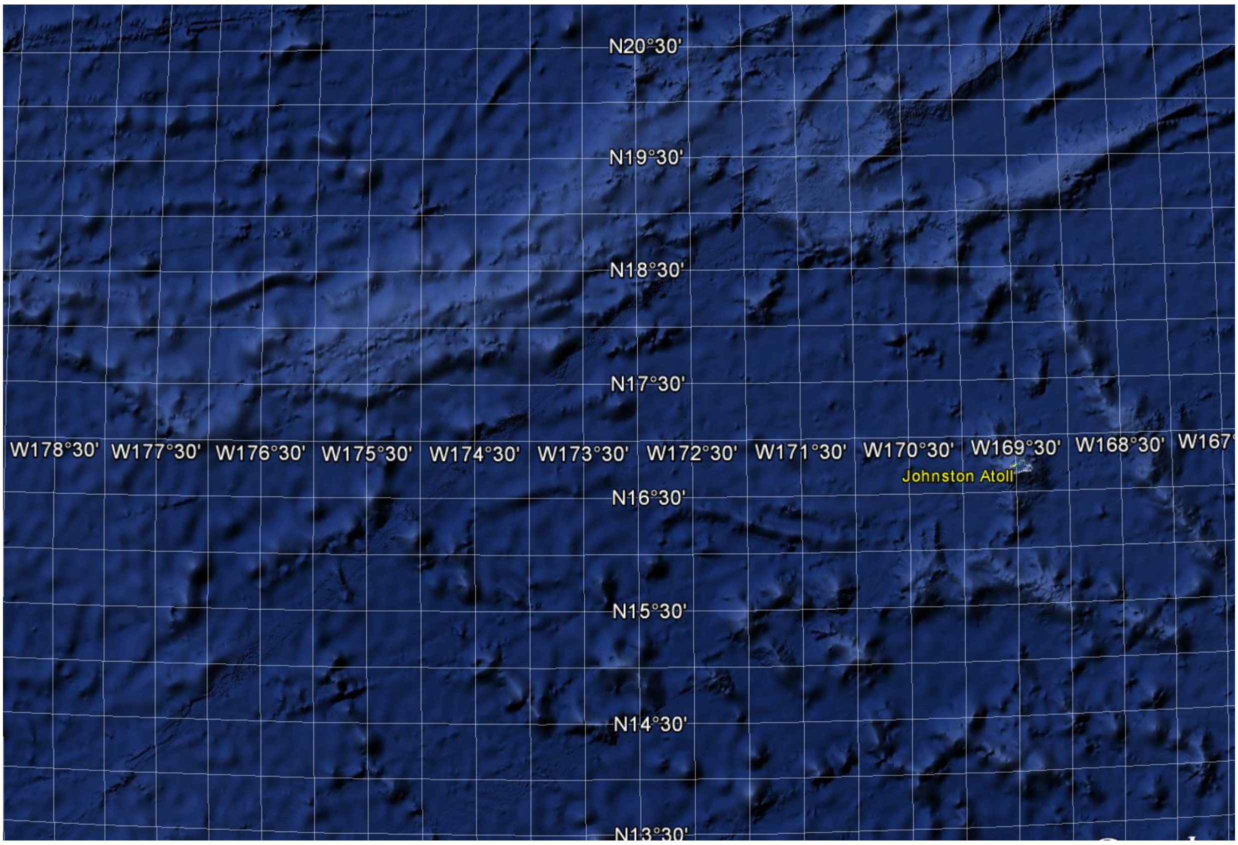

Figure 3: Southern part of the region with only the Johnston Atoll names. A high amount of sea mounts are seen throughout the region. Furthermore the northern part of this region only stands at depth between 2 and 3 km.

2.1 Mode Mask

A new Geographical Mode Mask (3.4) has been in place since Week 40 (01 October 2012) to accommodate this task. The mask is the basis of the CryoSat mission planning and defines the mode switching of the SIRAL instrument while the satellite revolves around the Earth. This new map takes into consideration the vast interest of the oceanographic, coastal altimetry and marine gravity community in the CryoSat mission. With respect to the previous version, the new SAR areas in the North Pacific Ocean were installed to map marine gravity field at high spatial resolution. The installed CP40 sea floor bathymetry mask is bounded by 15-25N and 180-203Eand is shown in the Figure 4 below.

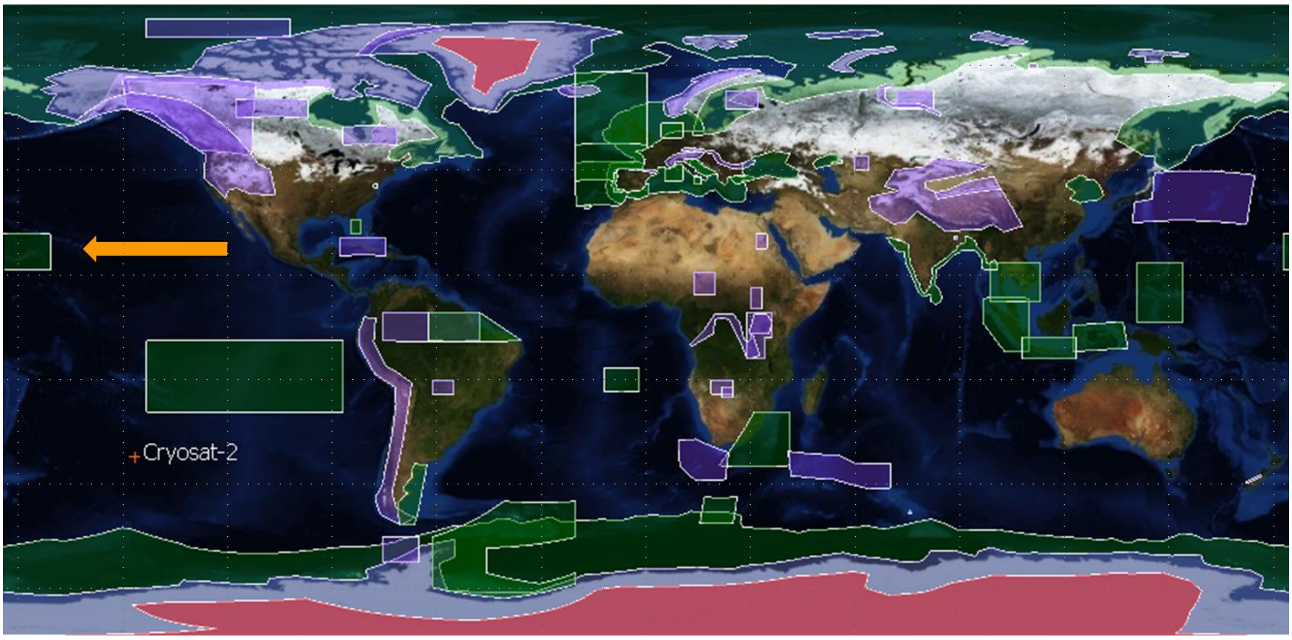

Figure 4. Cryosat-2 Mode mask V3.4 operational since week 40 of 2012 (http://earth.esa.int)

The region marked with an arrow was the region installed for the CP40 and the region used for the sea floor bathymetry prediction.

3 Algorithm Descriptions.

The method to determine bathymetry from altimetric observations is an important tool to assist the determination of bathymetry along ship tracks on board marine vessels. The marine observations will naturally not have the same sampling worldwide as satellite altimetry from geodetic mission, however these can be used to determine a smooth bathymetric field for the subsequent prediction. The fundamental difference is that from a ship the depth is directly inferred through the use of sounding methods. From altimetry the residual geoid height observations (or geoid slope information) will have to be used in an inversion process in which the height to the sea surface is estimated.

3.1 Algorithm Flow.

Initially the residual geoid heights observed from Cryosat-2 are converted to gravity following standard methods by DTU. Then long/short wavelength of the bathymetry and gravity anomalies are separated, local parameters for the inversion is established, and bathymetry is predicted. Finally the long and short wavelength and potentially more regions are combined to give the final bathymertry for the region.

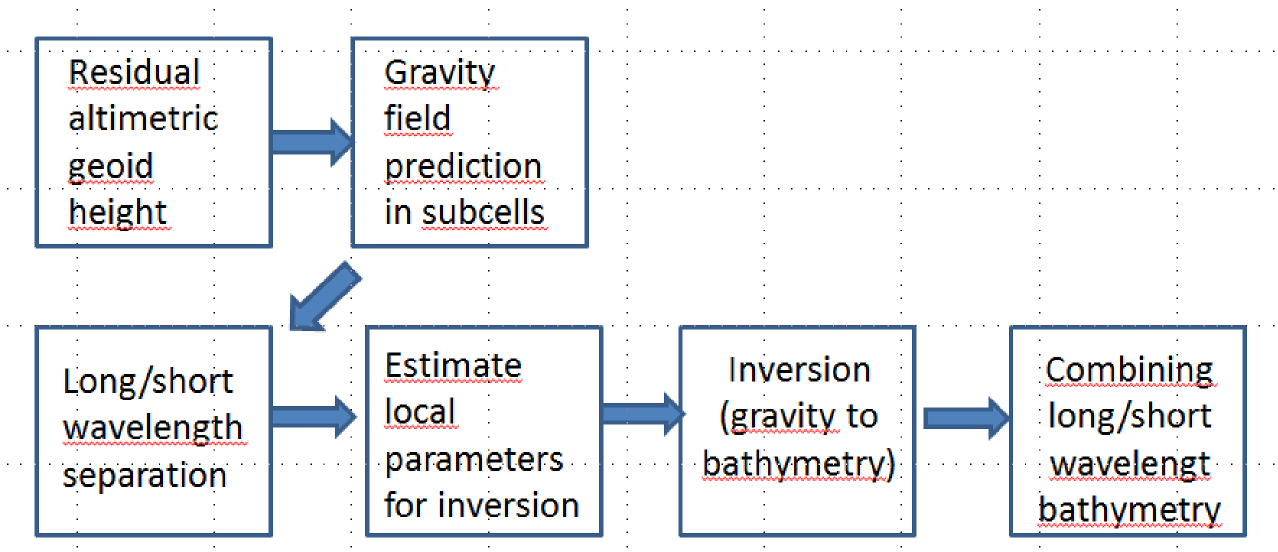

Figure 5: Algorithm flow.

3.2 Residual altimetric height.

The altimetric gravity was computed using the methods described in Andersen (2010); Andersen et al., 2009 and Andersen and Knudsen (1998) using altimetric sea surface height observations in overlapping tiles of 2° latitude by 6° longitude. The process of deriving gravity from the sea surface height applies a removerestore technique relative to EGM2008 and the dynamic topography DOT07A (Pavlis et al., 2012) to account for the long wavelengths.

Iterative local editing of the altimetric data is performed to ensure that there will be no outliers present. Subsequently a crossover adjustment is applied to remove ocean variability. This is followed by optimal interpolation onto a regular 1 minute grid using a covariance function with a correlation length of 6.5 km as described in Andersen,(2010).

3.3 Gravity field prediction:

Gravity is subsequently computed using Fast Fourier methods. As the conversion from geoid height to gravity enhances short wavelengths, a Wiener filter is applied to filter out wavelength shorter than 7 km. The setup is similar to that used for the derivation of the DTU10 gravity field. However, two important differences are implemented. The correlation length in the interpolation of geoid residuals was lowered from 9 to 6.5 km and the Wiener filter cut-off wavelength where the filter reaches 0.5 was lowered from 12 km to 7 km. These values are identical to the values used to derive the DTU13 global marine gravity field.

The lowering of the correlation length and the cut-off wavelength will allow significantly shorter wavelength gravity signal to be present in the new gravity field which again increases the fit with marine gravity.

In the following PVR document we have experimented with various ways of smoothing the short wavelength as this impacts which wavelength can be determined for the subsequent bathymetry prediction.

3.4 Bathymetric prediction.

The gravimetric response from an interface within the Earth can be expressed through a series expansion (Parkers FFT method or through numerical integration techniques where i.e. prism's are used.

Here we decided to apply the method by Parker (1976) in which gravity , Δg, located at a point on the surface of the earth (x,y,z = 0) is related to depth variation of an interface (h<0) relative to a depth (h0 <0) associated with a density anomaly ρ like:

$$\Delta g = -G \iiint_{h0}^{h} \frac{\rho z}{(x-x')^2 + (y-y')^2 + z^2)^{\ 3/2}} \, dz \, dx' \, dy'$$

Assuming constant density the formulation as par Parker can be submitted to a Fourier transform (Se Parker, 1976) and the formulation becomes

$$F(\Delta g) = 2\pi G \rho e^{2\pi q h0} \sum_{n=1}^{\infty} \frac{(2\pi q)^{n-1}}{n!} F((h-h0)^n)$$

The linear part of the expression is then:

$$\Delta g = 2\pi G \rho F^{-1}(e^{2\pi q h0} F(\Delta h)$$

Where Δh = h-h0. Hence in the linearized case the gravity anmlaies are obtained from the depth variations simply by scaling by 0.0419 mgal per density using in (g/ccm)) per meter. Actually this is equal to the Bouguer Correction for an infinite slab times the density contrast and an upward continuation. The beauty is that in the Fourier domain this is simply a multiplication.

The approach is simplified in two ways.

1) We have not taken into account horizontal variation in density. This is actually easily taken into account by keeping the density contrast as argument inside the Fourier transformation.
2) We will only work with one density contract (namely the sea bottom). The reason for this is twofold. First we have selected and process the altimetry data in relatively small areas regions (3 by 6 degrees) and secondly if we assure that the h0 in the formulation by Parker represents wavelength longer than 100 km through the use of low/high pass filtering and naturally filters the gravity field through the same filter prior to predicting bathymetry. Consequently we do not need to consider correlated multi-layers in the inversion and isostatic compensation (Smith and Sandwell, 1994).

3.5 Estimation of local parameters.

For the subsequent inversion the information about the covariance function is required. This was done using empirical covariance function fitting (Knudsen, 1993) in which homogenous covariance functions are used. This is made possible through the use of a linearized around the mean long wavelength depth initially determined. The estimated covariance functions were estimated from the high-pass filtered residual altimetric gravity field data and the corresponding filtered GEBCO 1 minute bathymetry (http://www.gebco.net/).

3.6 Inversion (sea floor depth prediction)

The estimation of the bathymetry parameters is normally ambiguous in the sense that different mass distributions are capable of creating the same gravity signal at the sea surface. Furthermore instabilities might occur when the correlations between the model parameters are not taken into account in the model. Here we used a 3-dimentional gravimetric inversion program to account for this. This program is called GRCOL and the program is part of the GRAVSOFT software library (Tscherning et al., 1994).

The inversion is carried out using the formulas of least squares collocation (Knudsen, 1993).

$$\Delta h_n = C_{\Delta h}^{T} \, (C + D)^{-1} \Delta g_n$$

Where the depths relative to the interface at h_0 are called Δh_n. The derived altimetric gravity anomalies are called Δg_n

For the test here the choice of a-priori spectrum of the covariance function were performed such that closed expressions could be used (Knudsen, 1993a). Using a-priori spectras similar to upward continuation this is the case. The upward continuation subsequently serves like a filtering *D (D > 0)*

The applied covariance function applied in the gravity to bathymetry inversion then becomes:

$$C_{\Delta h \Delta h(s)} = c \frac{2D}{(s^2 + (2D)^2)^{3/2}}$$

$$C_{\Delta g \Delta g(s)} = c(2\pi G\rho)^{\,2} \frac{2(D - h0)}{(s^2 + (2(D - h0)^2)^{3/2}}$$

$$C_{\Delta h \Delta h(s)} = c 2\pi G\rho \frac{2D - h0}{(s^2 + (2D - h0)^2)^{\frac{3}{2}}}$$

Where the noise factors c and D are determined empirically so that the covariance function fit the empirical covariance functions. The mean depth h0 is typically between 2 and 4 km depending on the region.

3.7 Combining short/long wavelength.

Once the predicted grid of depth anomalies are determined through the inversion this is combined with the long wavelength bathymetry field initially derived from GEBCO (http://www.gebco.net/) .

The way which altimetry is processed at the DTU in tiles of typically 3 by 6 degrees consequently means that the final bathymetry for the region of interest might have to be compiled by several smaller tiles.

4 Constrains and limitations

Several features like the enhanced surface following by the Cryosat-2 onboard tracker should enable more precise sea surface height estimated. However during the period of observations no extreme events of very high sea state was observed so this will likely have minor influence on the result.

The employment of Doppler shifts in the coherent reflections to slice the footprint into strips that are very narrow in the direction of flight (250m) is a feature that will be investigated through the use of higher frequency along-track observations than the traditional 1 Hz sea surface anomalies. One important limitation to recover sea floor bathymetry from satellite gravity is the precision of the derived sea surface height anomalies. In the DTU approach where sea surface height anomalies are directly used for the gravity field anomalies (different to the Sandwell and Smith approach where along track gradients are used), more focus on the range corrections must be made and the best state of the art corrections must be applied. For the Cryosat-2 SAR dataset is this not always the case and this might be a limiting factor in the bathymetry estimation.

The well-known limitations to bathymetry estimation from gravity should also be repeated here as they basically means that only a fraction of the remaining unknown sea mounts will be recoverable from satellite altimetry. Particulaly while longer wavelengths are isostatically compensated the short wavelength are affected by another phaenomenon. This is upward continuation which is a result of Newton's law that the gravity field falls off by the square of the distance between the source and the observer. This means that features on the sea bottom will be isostatically compensated as also demonstrated in the SAMOSA study for various sizes of ocean-bottom features (Andersen, 2011)

In essence anomalies with wavelength that are long compared to the mean water depth will suffer little attenuation while those that are shorter than pi times the water depth will suffer attenuation becoming stronger the shorter the wavelength are. Consequently very small sea mounts cannot be recovered from altimetric gravity if the water depth is too large.

Consequently the mean water depth is a limiting factor in the determination of bathymetric features and as mentioned then it is important to choose a region of moderate depth to study the impact of SAR altimetry on sea floor prediction

References

Andersen O. B (2011). SAMOSA report - Recovery of short wavelength geophysical signals and short spatial scale sea surface height/slope signals. Availale from http://www.satoc.eu/projects/samosa/

Andersen OB (2010) The DTU10 global gravity field and mean sea surface – improvements in the Arctic. Presented at the 2nd Int Gravity Field Service Symposium, Fairbanks, Alaska.

Andersen OB, Knudsen P, Berry PAM (2010) The DNSC08GRA global marine gravity field from double retracked satellite altimetry. J Geod 84: 191-199, doi: 10.1007/s00190-009-0355-9.

Andersen OB, Knudsen P (1998) Global Marine Gravity Field from the ERS-1 and GEOSAT Geodetic Mission Altimetry, J Geophys Res. 103:8129-8137

Andersen, OB, Scharroo R (2011) Range and geophysical corrections in coastal regions: And implications for mean sea surface determination. In: Vignudelli S et al. (eds) Coastal Altimetry (103-145), doi:10.1007/978-3-642-12796-0_5

Becker, J. J., D. T. Sandwell, W. H. F. Smith, J. Braud, B. Binder, J. Depner, D. Fabre, J. Factor, S. Ingalls, S-H. Kim, R. Ladner, K. Marks, S. Nelson, A. Pharaoh, R. Trimmer, J. Von Rosenberg, G. Wallace, P. Weatherall., (2009)Global Bathymetry and Elevation Data at 30 Arc Seconds Resolution: SRTM30_PLUS, Marine Geodesy, 32:4, 355-371.

Dinardo, S., Lucas, B., Gommenginger, C., Puig, C.M., Srokosz, M., Caparinni, M. (2013) Development of SAR Altimetry Mode Studies and Applications over Ocean, Coastal Zones and Inland Water (SAMOSA), Detailed Processing Model of the Sentinel-3 SRAL SAR altimeter ocean waveform retracker. Version 2.3.0.

Knudsen, P. (1993) Ingetrated inversion of gravity data. Report for Norsk Hydro, Kort- og Matrikelstyrelsen, Technical report no 7, 1993a.

Knudsen, P. Altimetry for Geodesy and Geophysics (1993b), in (ed) J. Kakkuri, lecture notes for NKG autumn school, Korpilampi, Finland, p. 87-126.

Pavlis NK, Holmes SA, Kenyon SC, Factor JK (2012) The development and evaluation of the Earth Gravitational Model 2008 (EGM2008). J Geophys Res, 117:B4

Sandwell DT, Smith WHF (2005) Retracking ERS-1 altimeter waveforms for optimal gravity field recovery, Geophys J Int, 163:79-89, doi 10.1111/j.1365-246X.2005.02724

Sandwell DT. Smith WHF (2009) Global marine gravity from retracked Geosat and ERS-1 altimetry: Ridge Segmentation versus spreading rate, J Geophys. Res, 114:B01411, doi: 10.1029/2008JB006008

Sandwell DT, Garcia E, Soofi K, Wessel P, Chandler M, Smith WHF (2013) Towards 1-mGal accuracy in global marine gravity from Cryosat-2, Envisat and Jason-1, The Leading Edge, 892-898

Smith, W. H. F., and F. Sandwell (1994) Bathymetric prediction from dense satellite altimetry and sparse shipboard bathymtry, J. Geophys. Res, 99 (B11), 21803-21824.

Smith, W. H. F and D. T. Sandwell, (1997) Global seafloor topography from satellite altimetry and ship depth soundings: evidence for stochastic reheating of the oceanic litosphere, Science, 277, 1956-1962.

Stenseng L, Andersen OB (2012) Preliminary gravity recovery from CryoSat-2 data in the Baffin Bay. Adv Space Res, 50(8): 1158-1163

Tscherning, C.C., P.Knudsen and R.Forsberg: Description of the GRAVSOFT package. Geophysical Institute, University of Copenhagen, Technical Report, 1991, 2. Ed. 1992, 3. Ed. 1993, 4. ed, 1994.

Wingham D et al. (2006) CryoSat-2: A mission to determine the fluctuations in Earths land and marine ice fields, Adv Space Res, 37(4): 841–871