Consistency of geoid models, radar altimetry, and hydrodynamic modelling in the North Sea

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Abstract. Radar altimetry, when corrected for tides, atmospheric forcing of the sea surface, and the effects of density variations and mean and time-variable currents, provides an along-track realization of the marine geoid. In this study we investigate whether and how such an 'altimetric-hydrodynamic' geoid over the North Sea can serve for validating satellite-gravimetric geoids.

Our results indicate that, using ERS-2 and ENVISAT along-track altimetry and water levels from the high-resolution operational circulation model BSHcmod, we do find distinct differences in RMS fits for various state-of-the art satellite-only models (beyond degree 145 for GRACE-only, and beyond degree 185 for GOCE-models) and for combined geoid models, very similar as seen in GPS-levelling validations over land areas.

We find that, at spectral resolution of up to about 200, an RMS fit as low as about 7 cm can be obtained for the most recent GOCE-derived models such as GOCO05S. This is slightly above what we expect from budgeting individual errors. Key to the validation is a proper treatment of the spectral mismatch between satellite-gravimetric and altimetric-hydrodynamic geoids. Comparing data fits and error budget suggests that geoid truncation errors residual to EGM2008 (i.e. EGM2008 commission and omission error) may amount up to few cm.

1. Introduction

Radar altimeters measure the instantaneous distance from a satellite-borne nadir-pointing antenna to the sea surface. Given a precise orbit and state-of-the-art media, instrument and geophysical corrections, radar altimetry provides an along-track realization of the instantaneous sea surface height (SSH) with respect to an ellipsoid.

When correcting such SSHs for water level variations associated with tides, atmospheric forcing of the sea surface, salinity and density contrasts, and mean and time-variable currents, one should arrive at a realization of the marine geoid. Traditionally, however, corrected altimetric SSHs have been used more often to derive marine gravity anomalies, and thus contribute to geoid determination indirectly at medium to short wavelengths.

Vice versa, satellite-gravimetric geoid modelling has been viewed as central in enabling the full potential of radar altimetry for improving our knowledge of ocean processes. Major recent improvements in spatial resolution and accuracy of the gravimetric geoid, in particular following the data analysis of the GOCE mission, have lead to a wealth of publication in this field (e.g. recent Special Issue in Newton's Bulletin, 2015). These approaches ultimately seek to improve ocean modelling through combining radar altimetry and satellite gravimetry.

However, we feel it is reasonable to ask whether, for a well-monitored region like the North Sea, altimetry combined with high-resolution operational hydrodynamic modelling can be used to validate the satellite geoids, without relying on separate and possibly inconsistent corrections for tides, inverse barometric effect, and dynamic topography.

Our approach is straightforward: We apply all standard corrections to along-track altimetry, except tide modelling and the correction for the inverse barometric effect. Subsequently, altimetric SSHs are further reduced to geoid heights by applying output from the regional high-resolution 3D hydrodynamic model BSHcmod (Dick et al., 2001) forced by atmospheric pressure and wind, astronomical constituents, and open-Atlantic boundary conditions. These geoid heights are then temporally averaged and compared to satellite-only and combined geoid models. Yet, an assumption we make is that the effect of omission errors of the underlying geoid models can be mitigated in our assessment, in that 'filling-up' with EGM2008 (i.e. adding signal from high-degree coefficients) allows unbiased model comparisons at the same degree, and to some extent also across different resolution.

Finally, we also investigate the ability of the hydrodynamic model to reproduce water levels at a set of tide gauges. This, together with uncertainty estimates for altimetry and geoid, is then combined into an error budget and compared to the fits obtained from the data sets.

2. Data

2.1. Geoid Models

With the advent of the GRACE and GOCE satellite missions, a variety of geoid models have been determined and the need for validating (or at least comparing) them with independent observations has become obvious.

In this study, we select a group of satellite-only and combined models as to cover the progress in geoid modelling over the last years. A summary is provided in table 1. All models are expressed in the zero tide reference system, and geoid heights are represented, using Bruns' first approximation, above the GRS80 ellipsoid. We are aware that altimetry data has been ingested in all combined models and that our consistency study should not be misunderstood as a fully independent validation. Tab. 1

Recent (i.e. release 5) GOCE-derived geoid models are thought to be accurate to about 1-3 cm at resolutions of degree 200 to 220 (from propagation of the GOCE variance covariance matrix, Gruber et al., 2015), and GNSS-levelling comparisons (Voigt and Denker, 2015) appear to confirm these errors for some well-observed regions of the world.

2.2. Radar Altimetry Data

1 Hz radar altimetry data for the years 2000.0–2012.0 has been obtained for the ERS-2 and ENVISAT satellites from the RADS data base (Naeije et al., 2008). ERS-2 and ENVISAT have been flying in identical 35 day repeat orbits with an average track spacing of less than 50 km in the North Sea.

We used cycles 0-169 of ERS-2 and 6-94 of ENVISAT data. Due to well-known problems like scattering of the radar pulse close to the coastline, on tidal flats and in shallow waters, and degraded quality of wet troposphere corrections, only data up to about 20 km off the coastline is considered. All standard corrections from RADS have been applied (ECMWF dry troposphere, MWR wet troposhere, solid Earth tide, GOT4.8 load tide, pole tide, reference frame offset, ERS-2: DGM-E04 orbits, JPL GIM ionosphere, BM3 SSB, ENVISAT: ESOC EIGEN-6C orbits, smoothed dual-frequency ionosphere, CLS SSB), but excluding tides and inverse barometric (IB) effect.

Finally, sea surface heights have been transformed from the Topex/Poseidon ellipsoid to the GRS80 ellipsoid. It is generally assumed that such (individual) sea surface heights from ERS-2 and ENVISAT are precise at the 2-3 cm level excluding the coastal region. Thus averaging of the order of $M = 100$ individual measurements (i.e. hydrodynamically corrected altimeter measurements per reference track location) would bring errors down to mm level, provided sea surface variability is taken care of.

2.3. Water levels

The mean dynamic topography in the North Sea reaches from about -40 cm off the UK coast to 10 cm in the Skagerrak; the main reason for this being the large difference in salinity between the North Sea and the Baltic, related to river inflow in the Baltic. Changes in atmospheric conditions affect sea levels at the level of 0.3 m (through pressure systems) and more (through wind), with surges reaching frequently up to 1 m and 2-3 m in extreme cases. Large deviations from the theoretical inverse barometric effects exist (e.g. Huess, 2001).

Tides in the North Sea represent a co-oscillating response to tides in the North Atlantic (Banner et al., 1979). Tidal waves enter the North Sea through its Northern boundary and through the strait of Dover. Subsequently, coastal geometry, bottom friction and resonances play a role in generating North Sea tides. Seasonal variations of major constituents such as M_2 are known to occur, and shallow-water constituents such as M_4 are visible in altimetry (Andersen 1999). Moreover, a gradual increase in tidal range has been observed at Dutch and German tide gauges in the Southern North Sea, starting around 1955 (Jensen and Mudersbach, 2006). The skills of conventional (deep-ocean) empirical tide models used in altimetry are therefore limited in this region (Madsen et al., 2007).

Hydrodynamic models have found widespread application in the North Sea; for operational forecast but also for assessing effects of coastal engineering or sea level rise on future tides. Along this line, BSHcmod (Dick et al., 2001) represents an operational baroclinic circulation and tide model for the North Sea, developed and run at the German Federal Maritime and Hydrographic Service (BSH). BSHcmod is nested into a North Atlantic model and driven by weather forecasts provided by the German Weather Service DWD. The temporal resolution of the model is 15 min. For version 3, the spatial resolution has been $10' \times 6'$ over the North Sea and $1.6' \times 1'$ over the German Bight, whereas for version 4 the resolution has been $5' \times 3'$ and $0.8' \times 0.5'$ accordingly. The model reproduces 14 partial tides. No radar altimetry data is assimilated in the model.

We used BSHcmod output for the same time frame as with altimetry, 2000.0–2012.0. However, BSHcmod outputs originate from forecast model runs, and model updates take place immediately with usually no overlapping time span that would allow data cross-checking. In the investigated time frame, such a model update occured on January 1, 2008, when BSHcmod switched from version 3 to version 4 at higher resolution. Unless it is explicitly stated, all our comparisons refer to the composite model timeseries.

2.4. Tide Gauge Data

In order to allow for an independent assessment of the errors of modelled water levels, we compare them to tide gauge observations (TGs). To this end, 15 TGs from the monitoring network of the German Federal Waterways and Shipping Administration (WSV), all equipped with GNSS receivers, had been selected (we disregarded some gauges in the Ems estuary after observing anomalous behaviour). The locations of the gauges are shown in figures 1-4 as red circles. All gauges except FINO1 (15.4.2008 – 21.6.2011) cover the whole time span of our investigation. TG values are recorded at 1min interval and downsampled to 15min, to be comparable with BSHcmod model output.

3. Methods

Our method is as follows: We compute altimetric-hydrodynamic geoid heights above the GRS80 ellipsoid along the satellite passes following Eq. (1),

$$N^{(a)} = h_{obs}(t) - \delta h(t) \ . \tag{1}$$

Here, $h_{obs}(t)$ is the altimetric SSH, and $\delta h(t)$ is the time-variable water elevation as computed by BSHcmod and spatially and temporally interpolated to the footprint location, thus including circulation, tides, wind stress and surges.

The $N^{(a)}$ relate to the altimeter footprints and thus, after this step, cover the North Sea with an along-track resolution of approximately 7 km, cross-track spacing of less than 50 km, up to a distance of about 20 km to the coast.

In fact, each overpass of the altimeter provides an estimate $N_i^{(a)}$ for a given location. Thus, in order to suppress altimetric noise and unmodelled sea surface variability, we average these estimates into

$$\overline{N}^{(a)} = \frac{1}{M} \sum_i N_i^{(a)} \tag{2}$$

and derive the standard deviation $\sigma_{N_i^{(a)}}$ of the individual estimates:

$$\sigma^2_{N_i^{(a)}} = \frac{1}{M} \sum_i (N_i^{(a)} - \overline{N}^{(a)})^2 \ . \tag{3}$$

As the ERS-2 and ENVISAT orbits vary by few km, evaluation of Eq. (2) requires that all altimetric heights within a certain region are binned onto a reference track position. To this end, all measurements within a cap of 5km size are assigned to a reference footprint and averaged in Eq. (2). Measurements exceeding five sigma are, finally, deemed as outliers and rejected in Eq. (2).

We then compute the corresponding gravimetric geoid heights above the GRS80 ellipsoid, at the reference footprint locations, following Eq. (4),

$$N^{(g)} = N^{(g)}_{j;n} + \delta N^{(g)} \ . \tag{4}$$

In the above, $N^{(g)}_{j,n}$ is the geoid height obtained from the satellite-gravimetric model $\texttt{MODEL}_j$, complete up to spherical harmonic degree n (which may be lower than the model's maximum degree).

In order to mitigate the spectral inconsistency between $N^{(g)}_{j,n}$and $N^{(a)}$, we add a high-resolution ('fill-up') geoid contribution $\delta N^{(g)}$ from EGM2008 (complete from degree n up to full model resolution, i.e. 2190). This is of course the same as if we would smooth the $N^{(a)}$ by removing the high-frequency geoid contribution. An alternative way would be applying spectral or spatial filtering to the $N^{(a)}$ directly; however, this poses problems in coastal regions and due to the limited size of the North Sea we refrain from this option.

However, our approach for mitigating spectral inconsistency is consistent with what is applied currently in land-based GNSS-levelling validations of the recent GOCE models (e.g. Voigt and Denker, 2015, for Germany, and Šprlak et al., 2015, for Norway).

Our metric of comparison is the spatial RMS of differences, from P altimeter footprints

$$RMS^2 = \frac{1}{P}\sum_{p=1}^{P}\left(\overline{N}^{(a)} - N^{(g)} - c\right)^2 \tag{5}$$

with c being the (weighted) spatial average of the $\overline{N}^{(a)} - N^{(g)}$,

$$c = \sum_{p=1}^{P}\omega_p\left(\overline{N}^{(a)} - N^{(g)}\right) \tag{6}$$

and ω_p following from the sigmas per location

$$\omega_p = \frac{\frac{1}{\sigma^2_{N_p^{(a)}}}}{\sum_{i=1}^{P}\frac{1}{\sigma^2_{N_i^{(a)}}}}\ . \tag{7}$$

The RMS, Eq. (5), could be compared to error propagation applied to Eqs. (2) and (4). Assuming we know the a priori errors of altimetric measurement $\sigma_{h_{obs}}$, water level modelling $\sigma_{\delta h}$, commission error $\sigma_{N^{(g)}}$ and omission error $\sigma_{\delta N^{(g)}}$ of the geoid model (or, in case of 'fill-up', of EGM2008), the predicted fit will be of the order of

$$\sigma^2_{\overline{N}^{(a)}-N^{(g)}} = \frac{1}{M}\sigma^2_{h_{obs}} + \frac{1}{M}\sigma^2_{\delta h} + \sigma^2_{N^{(g)}} + \sigma^2_{\delta N^{(g)}}\ . \tag{8}$$

With what has been said before ($\sigma_{h_{obs}}$ 2-3 cm, $\sigma_{N^{(g)}}$ 2-3 cm), assuming $\sigma_{\delta h}$ to 20-30 cm, see section 4.2, and averaging of the order of $M = 100$ individual measurements, we would expect to see a noise floor of 4-5 cm in the RMS difference between altimetric-hydrodynamic and gravimetric geoid heights. $\sigma_{\delta N^{(g)}}$ is particularly difficult to estimate, and we postpone a discussion to a later section. Higher RMS at a given location independent of the used geoid model may point at hydrodynamic modelling problems, while we would expect to see geoid model performance in varying RMS as per location, for different truncation and fill-up degrees.

We notice that in the above error budget, as well as for all metrics provided in the subsequent results section, errors are assumed as uncorrelated. In reality, altimetry measurements are known to be correlated along-track due to orbit errors, and geographically due to errors in media and sea state corrections. Hydrodynamically modelled water levels tend to be affected by systematic phase errors, which map into temporal correlation and large RMS values in error time series. Geoid model errors, finally, are inevitably spatially correlated due to the way they are computed. With these caveats in mind, our errors and RMS fits should be seen as worst case assumptions; yet since RMS fits provide the standard methodology in model evaluations we think for the sake of repeatability and comparability our approach is the most reasonable at the time of being.

4. Results

4.1. Altimetric-Hydrodynamic Geoid

Figs. 1 and 2 display maps of the temporal mean $N^{(a)}$ from BSHcmod version 3 (2000.0-2008.0) and version 4 (2008.0-2012.0).

Fig. 1
Fig. 2

It is obvious that a dm-size offset in mean sea surface between the two model versions exist over the North Sea, while this appears not the case for the Skagerrak and large parts of the Baltic Sea (not shown in the figure). This was noticed already in Weiss (2013) for a limited number of T/P and Jason crossover points, and we discuss it below. Apart from this, the mean modelled water level appears similar to earlier altimetric studies (Madsen et al., 2007). The mean water level or dynamic topography in BSHcmod is characterized by a general East-West sloping with gradients not exceeding 17 " (except coastal areas).

For reference, the full RMS water level variability of BSHcmod version 3 is shown in Fig. 3, with an average RMS of 44 cm (for version 4, this is very similar with a slightly higher average RMS of 47 cm, not shown here). Fig. 3

Next, in Fig. 4 a spatial representation of the empirical $\sigma_{N_i^{(a)}}$ is provided; this map tells where hydrodynamically corrected water levels from individual altimetry tracks fit less well (off the UK coast and in the strait of Dover, and thus where weights ω_p will be low), and where they fit very well (Norwegian coast, Skagerrak). Larger $\sigma_{N_i^{(a)}}$ are clearly associated with regions of higher variability of water levels due to tides and surges; our comparison may thus aid, outside of the scope of the present study, in guiding efforts directed at model improvements.

One may ask, whether and to what extent the mean water level offset between model version time series represents an artefact – and may adversely affect our comparisons – and to what extent it reflects a change in real conditions. In fact we have reasons to assume that changes in the model resolution, model forcing and data assimilation system, Fig. 4

and possibly in vertical referencing, are responsible for this effect. But since neither an extended overlap period for both model versions is provided, nor does a consistent reanalysis exist, we can only speculate about this. A comparison of Fig. 3 and 4 reveals that differences are generally below 8-10 cm in the North Sea interior, and larger along the UK shelf and the Channel where water level amplitudes are largest. Yet, we find that the spatial average of the $\sigma_{N_i^{(a)}}$ amounts to 15.0 cm; whereas when we evaluate only over the BSHcmod version 3 time span we arrive at little improvement with 14.8 cm, and 13.3 cm for model version 4. Future work would therefore likely benefit from concentrating on the version 4 time series alone, but at the time being, with only few years, we rather use the full time series for the generation of the $\overline{N}^{(a)}$.

4.2. Comparison at tide gauges

Modelled water levels are projected to TG locations and compared to 15min observations, after removing a local temporal mean. Again, we evaluate separately the two BSHcmod model time series: version 3 (2000.0-2008.0) and version 4 (2008.0-2010.0).

Table 2 shows the RMS difference between modelled and observed water levels, and Fig. 5 provides exemplary time series (Borkum-Fischerbalje, TGBF). We note that the model captures tidal and wind-driven water level ranges typically at the 5-20 cm level, but that it sometimes (here days 23-27) lags behind and this may dominate the tabulated RMS. Tab. 2

While individual (relative) TG readings are generally assumed to be accurate at the few cm level, we find RMS differences at the dm level, with the largest differences in the Ems and Weser estuaries. This suggests that the model has its largest difficulties in shallow regions, as may be expected. Furthermore, we find that application of model version

4 generally reduced the RMS, with greatest improvements for those stations that have exceptional high differences in version 3 (e.g. Bremerhaven TGBH, Knock TGKN, and Wilhelmshaven TGWH). With what has been said before, switching from model version 3 to 4 appears to have bigger effects on TG RMS reduction compared to altimetry. Fig. 5

This comparison suggests that modelled water levels, off the estuaries, can be assumed as having errors at the 20-30 cm level. This, however, applies to individual readings. Assuming errors as uncorrelated, and averaging of the order of 100 water levels, should reduce the error level down to 2-3 cm (except estuaries).

4.3. Satellite-Gravimetric Geoids

Fig. 6 displays the RMS fit of the global geoid models (see table 1), truncated at various degrees, when compared track-wise with the altimetric-hydrodynamic geoid heights. As discussed before, the gravimetric geoid error consists of the model or commission error and the omission error, due to missing real short-wavelength geoid signal. The figure can be interpreted as follows: As long as the error of an added coefficient of subsequent degree does not exceed the signal, increasing the degree of truncation will reduce the total error; thus resulting in a decreasing trend in the RMS error curve. When the error curve remains at the same level or rises, the error is likely at least as large as the signal, which means that the specific model has reached its maximum resolution. Fig. 6

As we will use the `EGM2008` model, subsequently, to mitigate the spectral mismatch between altimetric-hydrodynamic geoid heights and those from gravimetric models, it is appropriate to discuss the fit of the full `EGM2008` model to altimetry first, for $N^{(g)} = N_{EGM2008;2190}$. We find the RMS steeply decreasing from $>$ 40cm at degree 100, entirely dominated by truncation errors, levelling off to about 15cm at degrees 240 to 260 (for comparison, with the global ocean tide model GOT4.8, (Ray, 1999, updated), standard IB corrections, but no MDT model removed, at degrees 240 to 260 this would be about 22cm). At full model resolution (degree 2190), the RMS of EGM2008 amounts to 6.4cm.

We find that the GRACE-only models ITG-Grace2010S and ITSG-Grace2014S can resolve gravity signal up to d/o 160 and 170, respectively. One should note that hardly any difference between the two models would be visible in a global degree variances plot. Improvements through adding more data become obvious when looking at the GOCE models. We find that early GOCE models such as ITG-Goce02 contain signal up to d/o 205, while the latest models, which are based on the entire mission data, contribute to the North Sea geoid even up to d/o 245. Fig. 7

In Fig. 7, the same global geoid models are shown in comparison to the altimetric-hydrodynamic heights, but this time after reducing the omission error through fill-in with EGM2008. At lower degrees, the commission error, which is expected to grow with the degree of truncation, appears small and the misfit is likely dominated by errors in altimetry and/or hydrodynamic modeling. The various data sets agree at a level of 7cm (for GOT4.8 as above, this would be about 17cm), provided that the geoid is represented at moderate spatial resolution. Divergence of the error curves indicate at which degree a specific geoid model deteriorates in accuracy.

As expected, the errors (or rather misfits) of the latest GRACE-only solution ITSG-Grace2014S diverge at higher degree (d/o 150) compared to the earlier ITG-Grace2010S (d/o 145). For the GOCE models of the second and last generation, errors grow exponentially from 185 and 225, respectively. The error curves of the latest GOCE and GRACE/GOCE combination models are very similar, which indicates that the contribution of the combined data is mainly in the lower degrees.

The new GFZ solution EIGEN-6C4 has been chosen as an additional combination model besides EGM2008 as it includes GOCE data. The model performs quite well and it achieves RMS values nearly as low as the EGM2008 'noise floor' 6.4cm of while its curve never falls below that of EGM2008. Tab. 3

As expected, the regional models that are based on recent satellite-gravimetric models perform very well, for they have been adapted to the study area. For the gridded models the comparison was applied at full resolution only and the resulting RMS, min and max values are summarized in table 3. GCG2011 was warped to coincide with the geoid heights calculated at GPS/levelling points, but over the North Sea no corrector surface has been applied and the agreement is excellent. However, the model is defined in the area of the exclusive economic zone of Germany, which only covers a small part of the North Sea. It is thus questionable whether this result is comparable to the other models. EGG08 also achieves a good result, which is nearly as low as that of EGM2008. The RMS fit of NLGEO2013 is slightly higher, likely since the density of surface data used for this model is higher over the dutch mainland and waters compared to e.g. German and UK coast and waters (Slobbe, personal communication).

At the first glance EGM2008 seems to be superior to all the other models. However, one has to remember that EGM2008 already includes altimetry data and a good consistency is thus not surprising. Incidentally, the same is true for all combined models. Moreover, degree-wise 'augmentation' of harmonic coefficients is not optimal in a statistical sense. The coefficients, in particular those of the GOCE models, are highly correlated and provide a good solution only when used in linear combination. When this basic principle is ignored, the result can hardly be predicted.

Finally, we notice that estimating c in Eq. (5) led to an offset of about 17.3 cm, very consistent across all tested geoid models and nearly independent of the model expansion degree. A possible explanation for this could be that application of the hydrodynamic model realizes the reduction of water heights to an equipotential surface different from the geoid. BSHcmod has not been constrained by altimetry but likely by tide gauges in procedures not well known to us; so we hypothesize that unspecified vertical model referencing in combination with other possible reference system issues is responsible for the offset. Due to the difference between hydrodynamic model versions we used, our estimate for c depends on the time frame (19 cm with version 3 and 13 cm with version 4).

5. Conclusions

From budgeting the various error sources, we expect to see a noise floor of about 4-5 cm in the RMS difference between altimetric-hydrodynamic and (state-of-the art GOCE-derived) gravimetric geoid heights.

We find distinct differences in RMS fits for satellite-only models, combined geoid models, and regional geoid models. As maybe expected, GRACE-based models resolve up to d/o 170, early GOCE models up to about d/o 205, while latest models contribute to the North

Sea geoid up to d/o 245. When correcting for omission errors, the models agree with hydrodynamically corrected altimetry at the 7 cm level, provided the geoid is represented at moderate spectral resolution (up to d/o 230 for most recent GOCE models). This is slightly above what we expect from budgeting individual errors, and the reason may be related to a missing estimate of the geoid omission error (with respect to EGM2008). For higher truncation degrees, distinct differences between earlier and more mature models evolve. As expected, GRACE-only derived models perform less well beyond about degree 140 compared to recent GOCE models. We find best RMS fits for recent combined regional models, even down to 3-4 cm (but in a limited area); yet such models are not completely independent of altimetry. A caveat is that EGM2008 has been used as a fill-in in all our computations, in line with what is applied in terrestial GNSS/levelling validation studies.

Such land-based tests (Voigt and Denker, Gruber 2015) provide fits down to few cm and may be likely more accurate compared to our validation, but strong differences between regions exist, depending on quality of GNSS data and levelling, but in particular also on the areal extension.

Finally, we have found an offset of about 17cm between gravimetric and altimetric-hydrodynamic geoids, consistent across all tested models. We hypothesize that the reason for this may be related to the vertical referencing of the hydrodynamic model.

In summary, we believe that radar altimetry combined with high-resolution water level modelling provides a viable alternative for validation of current geoid models, at least for certain regions. Whether this holds for the future, with even lower geoid model errors, depends on whether hydrodynamic modelling will be able to keep improving. Moreover, with new altimetry technology we believe the observational gap between sea and land may be closed: E.g. Fenoglio-Marc et al. (2015) find Cryosat-2 SSH fits to tide gauges in the German Bight down to minimum RMS of 6cm (PRLM mode) and 7cm (SAR mode). This would, together with satellite-only geoid models and airborne gravimetry provide a basis for the validation of hydrodynamic modelling in the important coastal zone. Further improvement in the coastal zone may be associated with altimetry retracking (e.g. Passaro et al., 2015): on the one hand retracking may enable connecting altimetry and tide gauges, on the other hand retracked footprints would contribute new and independent information for validating geoid models.

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Table 1. Geoid models used in this study (S = satellite-only, C = combined, RC = regional combined, SLR = satellite laser ranging data, SURF = surface data, ALTI = altimetry data, KIN = kinematic GNSS orbit data

Type	Model	Reference	Data	Max. Res.
S	ITG-GRACE2010S	Mayer-Gürr et al. (2010)	7 yr GRACE	$n = 180$
	ITSG-GRACE2014S	Mayer-Gürr et al. (2014)	11 yr GRACE	$n = 200$
	ITG-GOCE02	Schall et al. (2014)	8 m GOCE	$n = 240$
	DIR_R5	Bruinsma et al. (2013)	4 yr GOCE + GRACE + LAGEOS	$n = 300$
	SPW_R4	Gatti et al. (2014)	3 yr GOCE	$n = 280$
	TIM_R5	Brockmann et al. (2014)	4 yr GOCE	$n = 280$
	DGM_1S	Hashemi Farahani et al. (2013)	7 yr GRACE + 14 m GOCE	$n = 250$
	GOCO05S	Mayer-Gürr et al. (2015)	11 yr GRACE+ 4 yr GOCE + SLR + KIN	$n = 280$
C	EGM2008	Pavlis et al. (2012)	GRACE + SURF + ALTI	$n = 2190$
	EIGEN-6C4	Förste et al. (2014)	GRACE + GOCE + LAGEOS + SURF + ALTI	$n = 2190$
RC	EGG2008	Denker et al. (2009, updated) Denker (2013)	CHAMP + GRACE + SURF + ALTI	1’×1’
	NLGEO2013	Slobbe et al. (2014)	DGM1-S + SURF + ALTI	1 km × 1 km
	GCG2011	BKG (2011)	EIGEN-5C + EGM2008 + SURF + ALTI	1.5’×1’

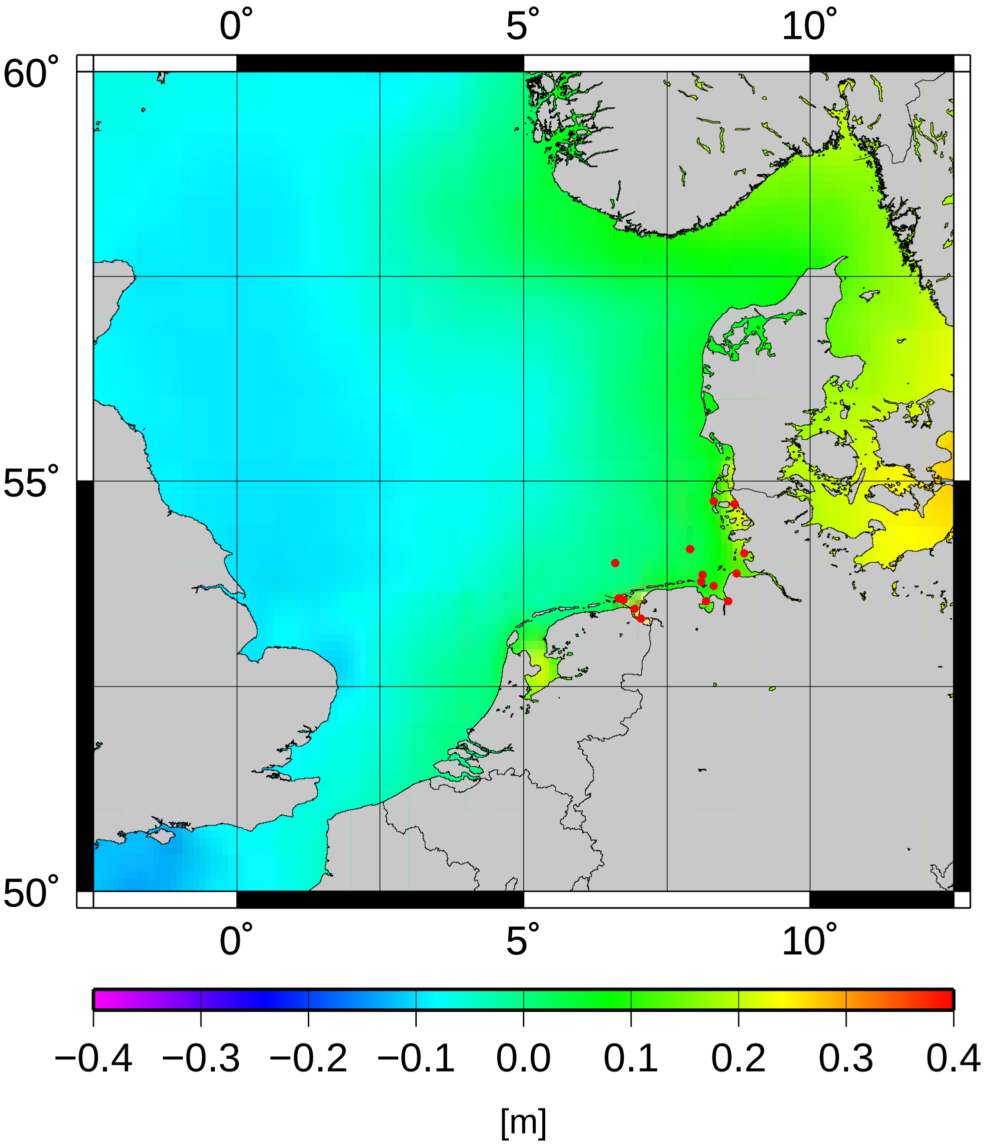

Figure 1. MDT from BSHcmod version 3, 2000.0-2008.0

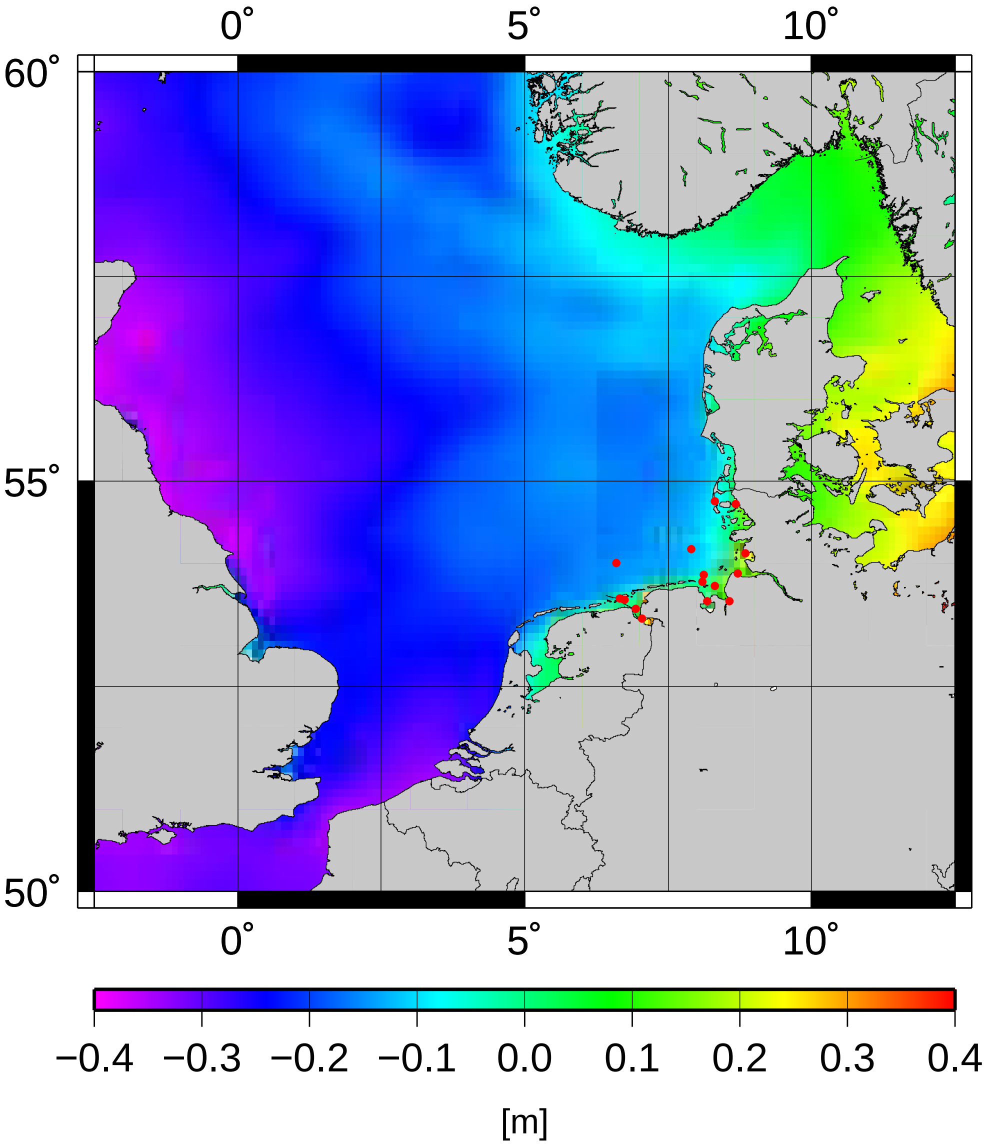

Figure 2. MDT from BSHcmod version 4, 2008.0-2012.0

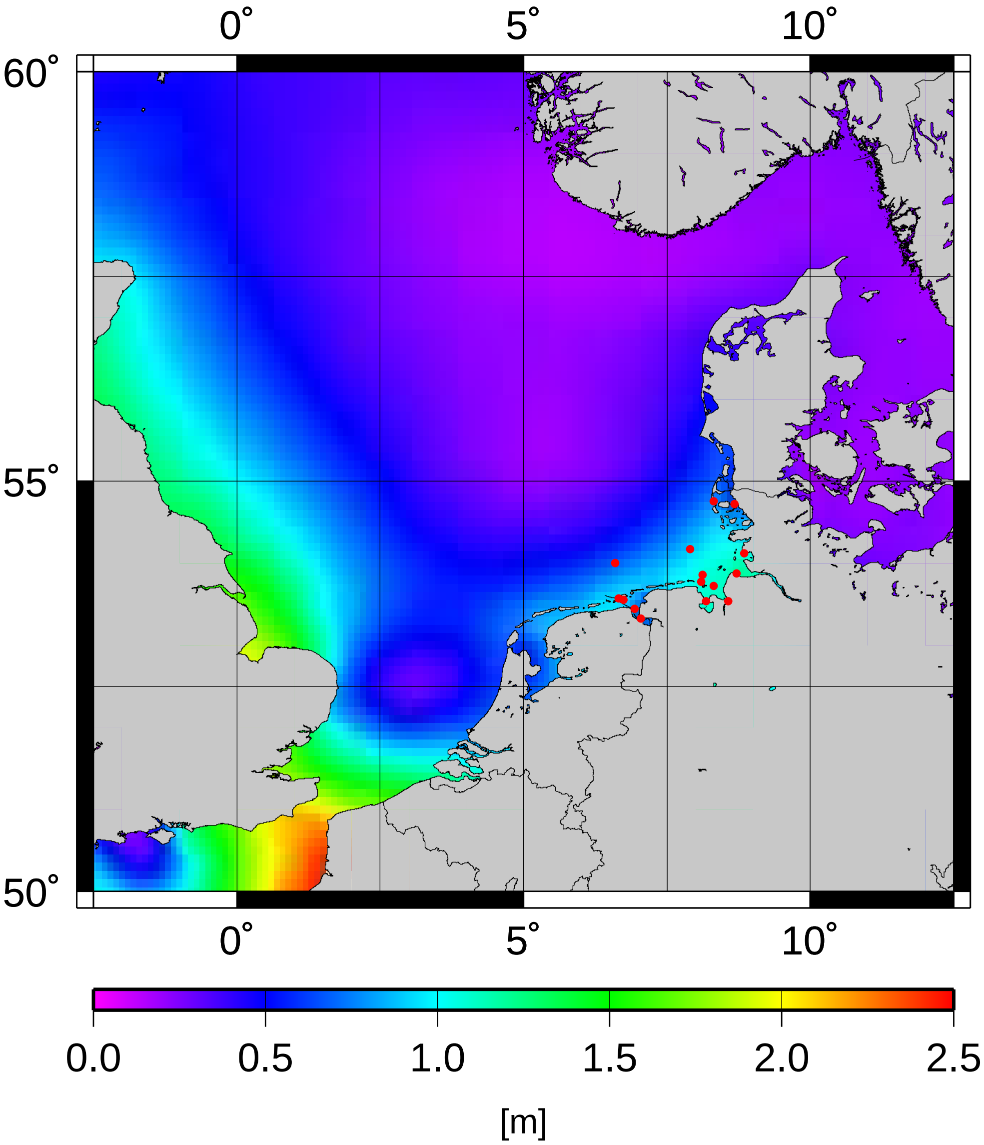

Figure 3. RMS water level variability from BSHcmod version 3, 2008.0-2012.0

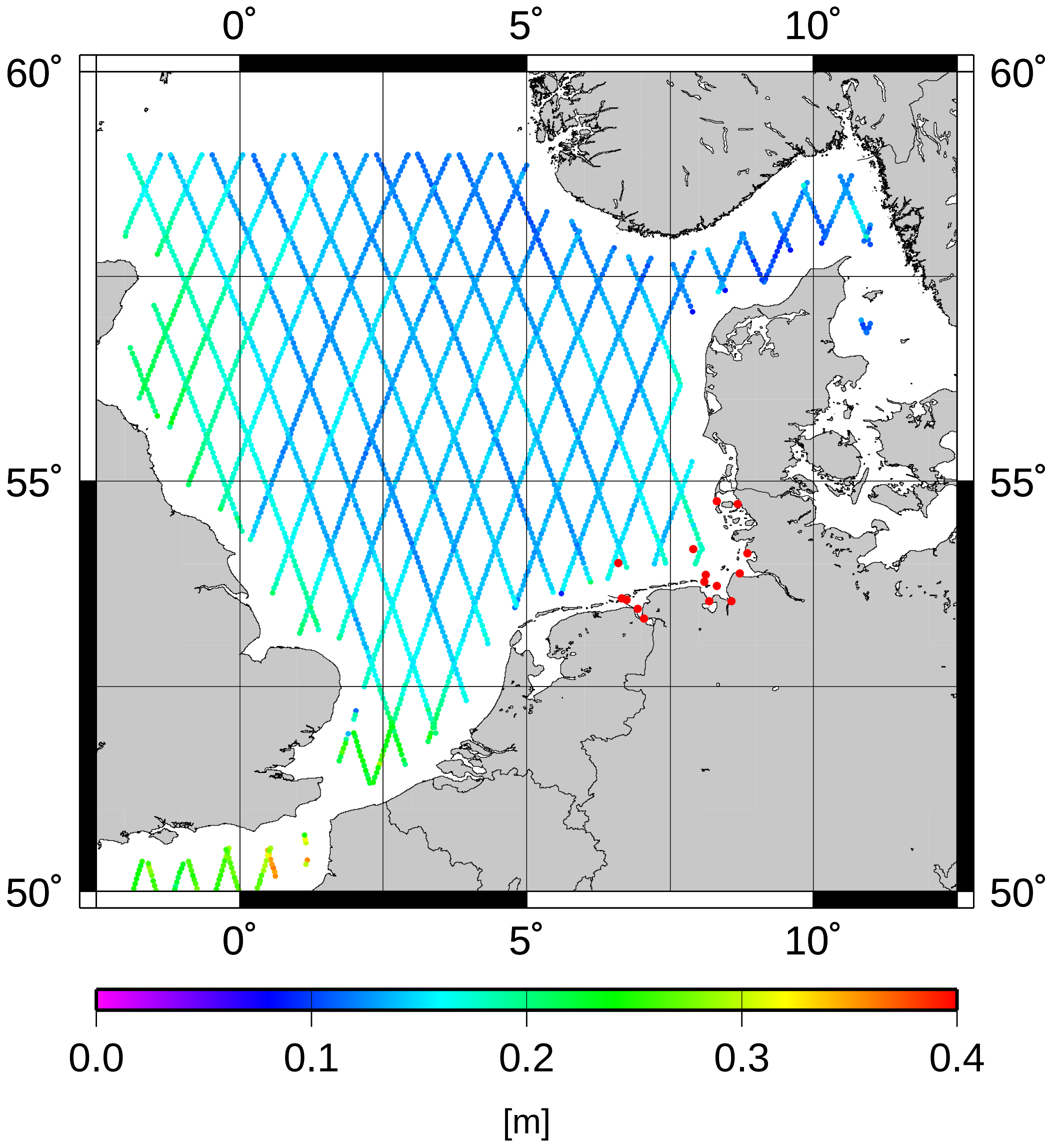

Figure 4. Standard deviation $\sigma_{N_i^{(a)}}$ of the altimetric-hydrodynamic water levels, obtained from averaging individual tracks

Table 2. RMS fit [m] of modelled water levels compared to tide gauge observations

	BSHcmod [m]	
TG	V.3 (2000.0-2008.0)	V.4 (2008.0-2010.0)
HOE2	0.22	0.17
TGDA	0.35	0.29
HELG	0.21	0.14
TGBU	0.35	0.23
FINO1	-	0.19
TGCU	0.26	0.18
LHAW	0.21	0.16
TGME	0.23	0.17
FLDW	0.33	0.27
BORS	0.21	0.14
TGBF	0.22	0.16
TGWH	0.44	0.27
TGBH	0.72	0.46
TGDU	0.28	0.17
TGKN	0.47	0.28

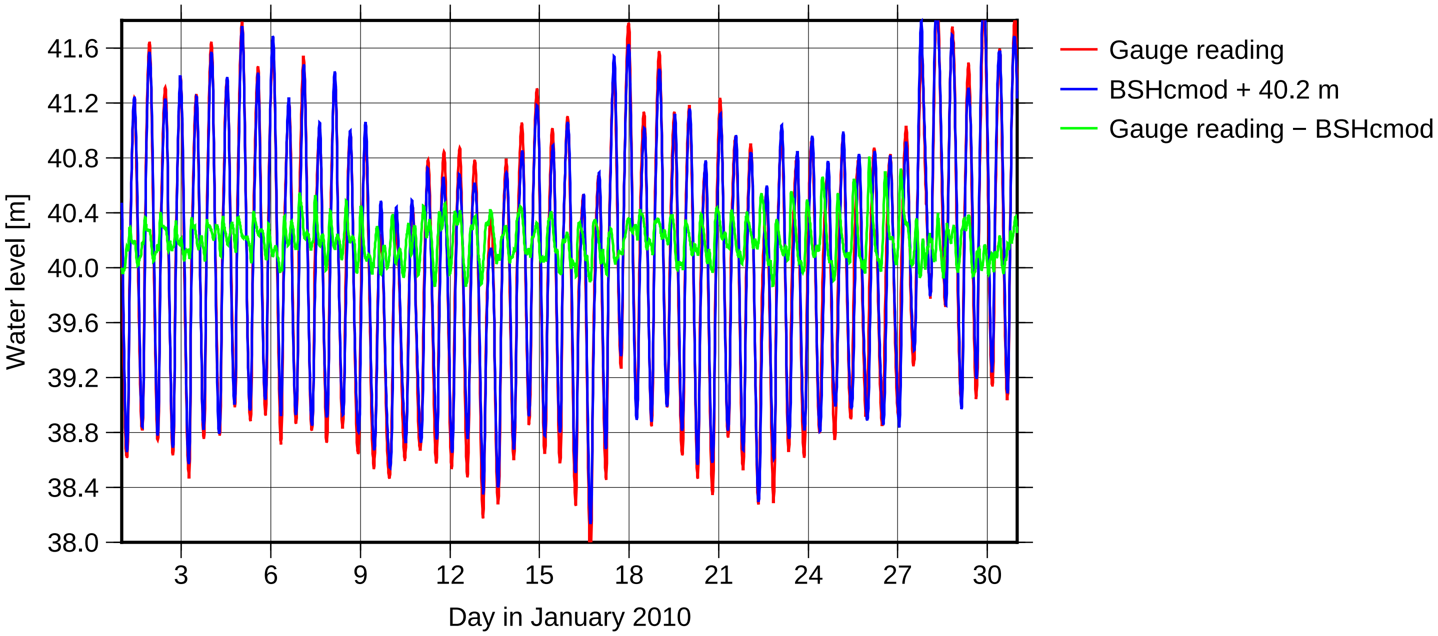

Figure 5. Modelled and observed water levels at TGBF

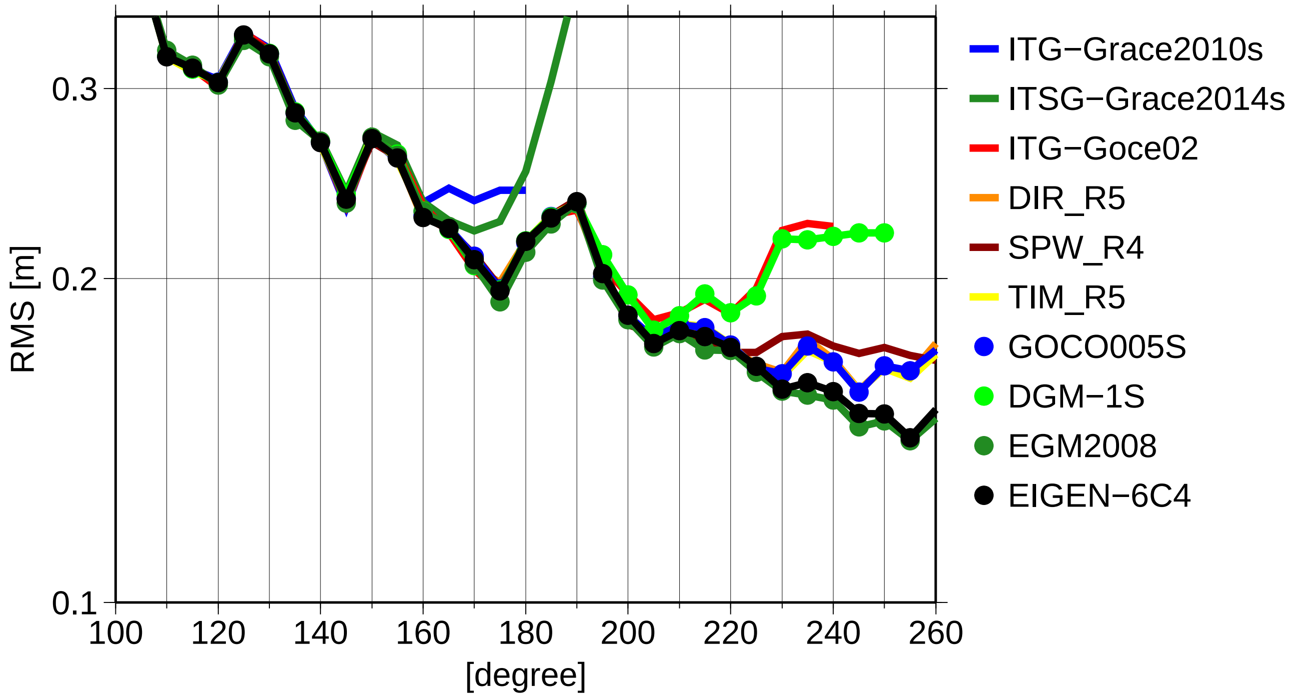

Figure 6. RMS misfit between gravimetric and altimetric-hydrodynamic geoid as a function of model truncation degree (without EGM2008 fill-up)

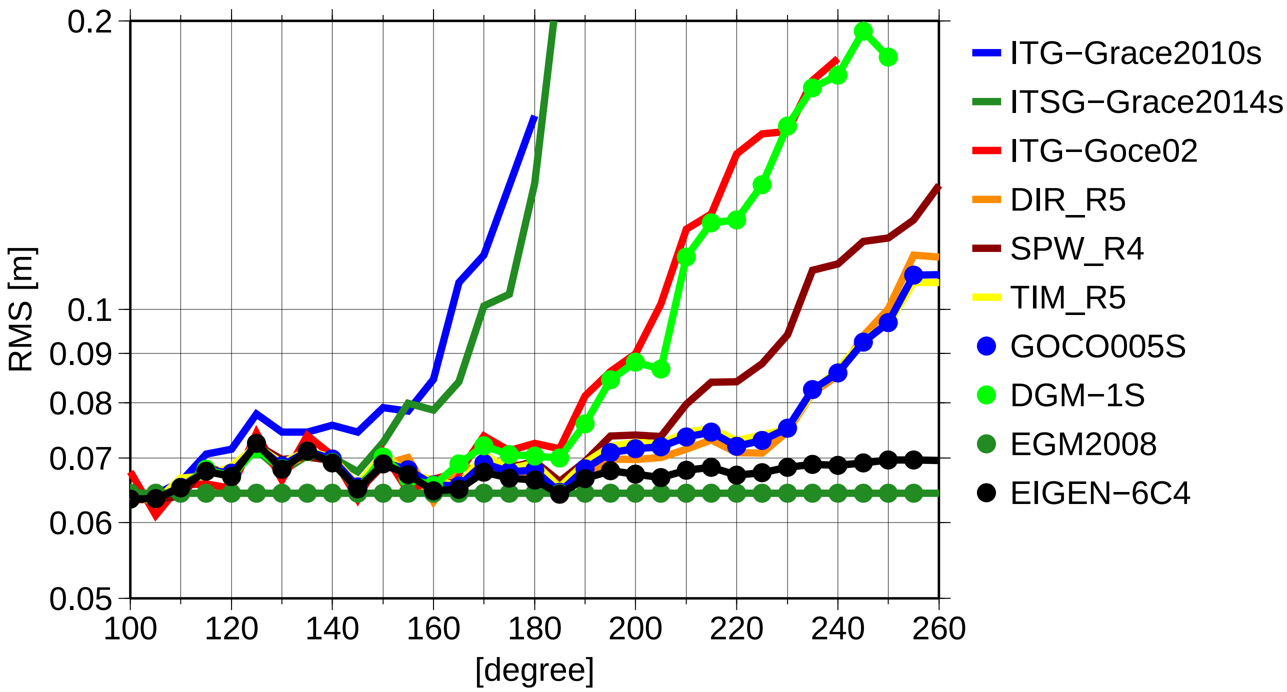

Figure 7. RMS misfit between gravimetric and altimetric-hydrodynamic geoid as a function of model truncation degree (with EGM2008 fill-up)

Table 3. RMS fit of regional geoid models

Model	RMS [m]	min [m]	max [m]
NLGEO2013	0.067	-0.229	0.277
EGG08	0.064	-0.242	0.293
GCG2011	0.034	-0.117	0.093