Establishing a consistent vertical reference frame for the North Sea area

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1 Introduction

WP 3.11 of the INTERREG IV North Sea Region Program project BLAST aims at the development and application of a methodology for the unification of chart datums (CD) in the North Sea area and the connection with the onshore height systems with a spatial resolution and accuracy that meet current and future requirements of society. The main result are maps of the Mean Sea Level (MSL), the geoid, and LAT, which are consistent with the hydrodynamic constraints, and relations of these surfaces to the national vertical reference systems around the North Sea. In addition, the project will provide a properly calibrated hydrodynamic model with a well-defined reference surface, which will improve the quality of the modelled water levels.

1.1 Background and motivation

Lowest Astronomical Tide (LAT) has been adopted by the International Hydrographic Organization [International Hydrographic Organization, 2011, Technical Resolution 3/1919] as the vertical reference surface of hydrographic charts in the North Sea area (Chart Datum, CD). Therefore, observed water depths need to be corrected for the difference between the instantaneous sea surface and the LAT surface. This so-called 'water level reduction' [FIG Commission 4 Working Group 4.2 2006, Dodd and Mills 2011] can be done efficiently using modern Global Navigation Satellite Systems (GNSS) such as GPS or the forthcoming European system GALILEO. This requires, however, that the ellipsoidal heights of the LAT surface are known with respect to the GNSS ellipsoid.

LAT is defined as the lowest tide level to occur under average meteorological conditions and under any combination of astronomical conditions. Ellipsoidal heights of LAT can be derived from the analysis of a time series of water level observations, which may be acquired at tide gauges and offshore platforms equipped with GNSS receivers, GNSS buoys, or satellite altimeter crossovers. Among these techniques, only radar altimetry provides ellipsoidal heights of LAT with high spatial resolution, whereas the other techniques provide this information only at a sparse set of stations. On the other hand, the temporal resolution of data acquired at tide gauges, platforms, or GNSS buoys allows estimation of all relevant tidal constituents. The same applies to radar altimetry data in deep waters, but no longer in shallow waters. The reason is that in shallow waters the tidal spectrum contains many more relevant constituents, which cannot be resolved due to the poor temporal sampling of radar altimetry.

Therefore, in shallow waters, one determines the separation between Mean Sea Level (MSL) and the LAT surface using a shallow water tide model. Ellipsoidal heights of LAT are then obtained by subtracting the separation between MSL and LAT from the ellipsoidal heights of MSL. The latter are taken from radar altimetry data [e.g., Simon, 2001, Turner et al., 2010].

This procedure, however, cannot be applied in coastal regions and estuaries, since these regions lack reliable radar altimeter data (i.e., no reliable information about MSL is available there). The reason is that radar altimetry data are less accurate within a distance of 5-25 km from the coast. The width of this stroke depends on the radar altimeter data processing. So-called re-tracked radar altimetry data can already be obtained at distances 5-10 km from the coast, whereas the bulk of radar altimetry data available in data bases have been processed using a simpler waveform model, which is valid at a distance of 20-25 km from the coast.

Interpolation techniques are used to close this gap. They interpolate ellipsoidal heights of LAT inside the stroke using ellipsoidal heights of LAT at coastal stations and those derived from radar altimetry data outside the stroke [e.g., Turner et al., 2010]. Interpolation, however, increases the error budget, and accuracies of 5-10 cm are hardly achievable. Moreover, interpolation does not provide any meaningful results at some locations, e.g., between the Wadden island and in the estuaries.

Therefore, there is a need to develop a new approach for the realization of ellipsoidal heights of LAT, which provides a North-Sea wide coverage, without gaps, and sufficient spatial resolution, and accuracy.

1.2 LAT, geoid, and hydrodynamic model

When developing such an approach, one needs to remember that the LAT surface is not directly observable. Therefore, the pursued strategy followed in this project is to determine an intermediate reference surface in 3D, and to relate the LAT surface to this intermediate reference surface. In fact, there are two surfaces, which could serve as an intermediate reference surface: i) MSL and ii) the geoid. The latter is defined as an equipotential surface of the Earth's gravity field close to MSL. It is also a level surface. MSL and geoid differ by the so-called Mean Dynamic Sea Surface Topography (MDT). Ellipsoidal heights of MSL are provided by radar altimetry; ellipsoidal heights of the geoid (sometimes referred to as geoid heights) can be computed from gravity measurements. In this project we give preference to the geoid. The reason is two-fold: i) the geoid can be computed everywhere whereas the only technique available to provide information about MSL is radar altimetry, and the latter cannot provide reliable information about MSL everywhere on the North Sea down to the coast; ii) the geoid is also used as onshore vertical reference surface or can be related to the onshore reference surface. This allows to link the offshore vertical reference surface (LAT) to the one used onshore. This is an important property as it allows for instance to merge onshore and offshore data sets for coastal management. There is another aspect in favour of the geoid as intermediate reference surface. The natural reference surface of a hydrodynamic model is an equipotential surface of the Earth's gravity field. Hence, modelled water levels refer to an equipotential surface (and so does LAT computed from modelled water levels). By designing a suitable procedure, we can take care that this reference equipotential surface is identical to the geoid. In this way, we can guarantee that water levels of a hydrodynamic model refer to the geoid. In this way, the hydrodynamic model provides the separation between the (instantanteous and mean) sea surface and LAT, respectively, and the geoid. When adding this separation to the geoid heights, we obtain ellipsoidal heights of LAT and of the sea surface over the whole North Sea region down to the coast without gaps.

In principle, the geoid can be computed from gravity data. In practice, gravity data may not be available at sea with sufficient density and coverage. Therefore, they are often complemented by radar altimetry data. The basic idea is to subtract the dynamic topography signal (e.g., taken from a hydrodynamic model) from the observed sea surface heights, thus providing directly geoid heights. In reality the procedure is more complicated, which implies among others that geoid slopes are derived from radar altimetry data, combined with other land and marine data, and inverted into geoid heights (cf. Section 1.2). We believe that a North Sea geoid benefits from radar altimetry data. Using radar altimetry data for geoid computation implies that geoid and hydrodynamic model are not independent of each other. Water levels produced by a hydrodynamic model need to be referred to the geoid; vice versa, to compute the geoid, we need a properly vertically referenced hydrodynamic model. Therefore, LAT, geoid, sea surface and hydrodynamic model are closely related to each other. For more details we refer to Section 2.

Satellite radar altimeter data are considered as a key data set in high-resolution marine geoid modeling despite the fact that in some shallow waters like the North Sea ship borne gravity data are available. The latter, however, are often biased and of heterogeneous quality [Wessel and Watts 1988]. Using radar altimeter data for geoid modeling requires a correction for the dynamic topography, which is caused by tides, atmospheric forcing, and water density variations. Tidal corrections are usually provided by global tidal models, which do not perform well in shallow waters [e.g., Andersen and Knudsen 2000, Hwang et al. 2006, Sandwell and Smith 2009]. It is well-known that much better corrections can be obtained by using dedicated shallow water models.

In marine geoid computations from sea surface slopes, corrections for atmospheric forcing and water density variations are assumed to be negligible, at least in areas outside energetic mesoscale geostrophic currents. This assumption is motivated by comparing the magnitude of the sea-surface slopes associated with non-tidal water level variations with the expected precision with which slopes can be computed from observed along-track sea surface heights [Sandwell and Smith 2009]. The signal-to-noise ratio of surface slopes depends on the spectrum of surface slopes. Over wavelengths of tens to hundreds of kilometers, the signal to noise ratio may be more favorable, meaning that the corrections for atmospheric forcing and water density variations are not negligible anymore.

In contrast to the above reasoning, several authors [e.g., Brennecke and Groten 1977, Rapp and Yi 1997] have shown that applying a MDT correction to the observed SSHs has limited impact on the estimated gravity anomalies/geoid. Rapp and Yi [1997], who performed a number of experiments in the Gulf Stream region, provide two reasons to explain this expected lack of improvement. First, their computations were carried out after centering the data, so that the mean difference between the SSHs and the reference geoid was removed from the observations and this removed contribution was not restored to the computed anomalies. While the MDT is almost constant in the region they considered, this data centering effectively removed the MDT effect. The second reason is that they did not account for the seasonal variations in the Gulf Stream. This is especially relevant to the GM data from ERS-1 and GEOSAT, which are due to their spatially dense ground tracks of utmost importance for gravity field recovery, since for these data the seasonal variations cannot average out. Applying time-dependent corrections for the non-tidal contributions to the dynamic sea surface topography is also a must for the observed SSHs in the North Sea, which is known for its frequent storm surges. During a storm surge, sea level can be raised or lowered by several meters over a period ranging from a few hours to 2 or 3 days [Pugh 1996, Flather 2000]. Indeed, also these variations will not average out for the GM data from ERS-1 and GEOSAT.

In order to remove the long-wavelength SSH gradient components, Sandwell and Smith [2009] used an iterative approach; in each iteration, the long-wavelength signals are removed from the observations by subtracting the low-pass filtered differences (0.5 gain at 180 km) between the observed SSH gradients and the gradients of the geoid obtained in the previous iteration. In the initial run they used a geoid derived from the EGM96 spherical harmonic coefficients. Although they showed that this method provides good results, it is not very practical for the semi-enclosed North Sea. Indeed, while there is no data on land, the performance of this filter operation will decrease in coastal regions. In addition, unavoidably, also signal will be removed, i.e. radar altimeter data is only allowed to contribute to the determination of the geoid's short-wavelengths.

Therefore, in this study we will also consider the importance of the time-varying, non-tidal water level variation corrections for the computation of the marine geoid. Since the dominating acceleration term related to wind stress is ignored, the inverted barometer correction, as e.g. applied by Hwang et al. [2006], lacks accuracy in order to properly remove the sea level perturbations induced by atmospheric forcing in shallow water. The reason is that the dominating acceleration term related to wind stress is ignored. This also holds for the quasi-stationary corrections for the water level variations induced by water density variations derived from the Levitus climatological dataset [e.g., Hwang 1997]; according to Wunsch and Stammer [1998] their accuracy is limited to the 10-25 cm level. Both correction terms can be improved by using a shallow water hydrodynamic model. Here, we will use the same model as used to improve the astronomical tide corrections, which has the additional advantage that any non-linear interaction between the separate contributions (e.g. between tide and surge [Prandle and Wolf, 1978]) will be included as well. This means that rather than three separate corrections, one correction for the instantaneous water level will be applied that reduces the observed SSHs to geoid heights.

A hydrodynamic model that enables the reduction of the observed SSHs to geoid heights is also needed to derive the ellipsoidal heights of an area covering LAT surface in coastal waters and estuaries, where no reliable radar altimeter data are available and hence no area covering MSL model. In order to derive such a surface, two approaches can be used that differ from each other by the reference surface used as intermediate surface to which the modeled LAT values refer and, related to that, by the way the average meteorological conditions are accounted for.

- The first approach maintains the current procedure used to model LAT (Section 1.2.3), i.e. LAT is modeled relative to MSL, where MSL is taken to represent the average meteorological conditions. Here, the ellipsoidal heights of MSL are derived by adding the geoidal undulations to the modeled MDT expressed relative to that geoid.
- In the second approach, the geoid is used as intermediate reference surface, which requires to model LAT relative to that geoid. This can be achieved by an explicit modeling of the time-averaged meteorological conditions.

So, in the first approach, the hydrodynamic model is used to obtain the MDT, while in the second approach it is used to derive LAT relative to the geoid. In this study, the second approach is preferred. The reason for this is twofold: (i) this approach is conceptually clearer since the model's reference surface, as well as the geoid, is an equipotential surface [Hughes and Bingham 2008] and (ii) this approach allows for the inclusion of temporal variations in the definition of the average meteorological conditions. The error introduced by not taking into account that the model's reference surface is an equipotential surface, is shown by Prandle [1978], who showed that the contribution of the M_2 tide to MSL in the southern North Sea varies from -1 to 8 cm, with maximum values along the Dutch coast. While this contribution is also part of the observed MSL, an additional correction is required in case we compute the ellipsoidal heights of the LAT surface. The latter reason follows from a precise interpretation of the LAT definition that the average meteorological conditions in spring and those in fall should be included separately.

1.3 Objectives

In this project we develop a new methodology to determine ellipsoidal heights of LAT in the North Sea including coastal waters with sub-decimetre accuracy and to link LAT to the onshore height systems. This will be achieved by simultaneously modeling the geoid and exploiting a shallow water hydrodynamic model. The latter takes a key role in the realization of the ellipsoidal heights of an area covering LAT surface. First, it is needed to reduce radar altimeter SSHs to geoid heights, which are needed to realize an accurate high-resolution marine geoid. Secondly, it is needed to realize LAT relative to this estimated geoid. The main objective of this project is: to develop a procedure that enables to derive a consistent set of vertical reference frames; a marine geoid based on, amongst others, radar altimeter data combined with a shallow water flow model, and LAT relative to this geoid, modeled by a shallow water flow model. The developed procedure will be implemented and used to estimate a geoid, and to model the LAT and MDT surfaces.

The shallow water flow model used in this project is the extended Dutch continental shelf model version 5 (DCSM) described by Slobbe et al. [2012b] (the original DCSM model is described by Gerritsen et al. [1995] and Verlaan et al. [2005]). DCSM was developed to make tide and surge forecasts over the full nodal cycle in order to support operational management of the Eastern Scheldt storm surge barrier. The model is based on the WAQUA software package [Leendertse, 1967; Stelling, 1984] for depth-integrated flow. WAQUA includes the non-linear surge-tide interaction and is based on the depth-integrated shallow water equations. DCSM covers the area of the northwest European continental shelf to at least the 200 m depth contour, i.e. 12° W to 13° E and 48 °N to 62.3 °N, and has a horizontal resolution of $1/8° \times 1/12°$ (approximately 8×9 km) in longitude and latitude, respectively. This model is extended such that it includes the effect of horizontal water density variations on the modeled water levels.

2 Outline of the methodology

As shown in the introduction, an accurate realization of both the geoid and the LAT surface in the North Sea, requires the use of a shallow water hydrodynamic model. More specifically, the hydrodynamic model is needed to i) reduce the altimeter sea surface heights (SSHs) to geoid heights and, ii) model the LAT surface. Moreover, it was shown that modeling LAT relative to the geoid is conceptually clearer and allows the inclusion of the average seasonal variations of the meteorological conditions/MSL in the realization of LAT. Adding the separation between the geoid and LAT to the geoid heights provides the ellipsoidal heights of LAT. Obviously, this last step can only be applied if the model's reference surface can be identified with the geoid. The geoid, however, is not known, but has to be computed using the same hydrodynamic model. In this chapter, we present the flow chart of the procedure developed and implemented in this project i) to estimate a geoid, and ii) to obtain a hydrodynamic model that provides water levels relative to this geoid. Once this has been done, the hydrodynamic model is used to realize the LAT and MDT surfaces (relative to the geoid). The overall procedure can be divided in three phases: i) data pre-processing, ii) iterative geoid computation and vertical referencing of the hydrodynamic model, and iii) realization of the LAT and MDT surfaces. In the remainder of this chapter, these three phases will be briefly discussed.

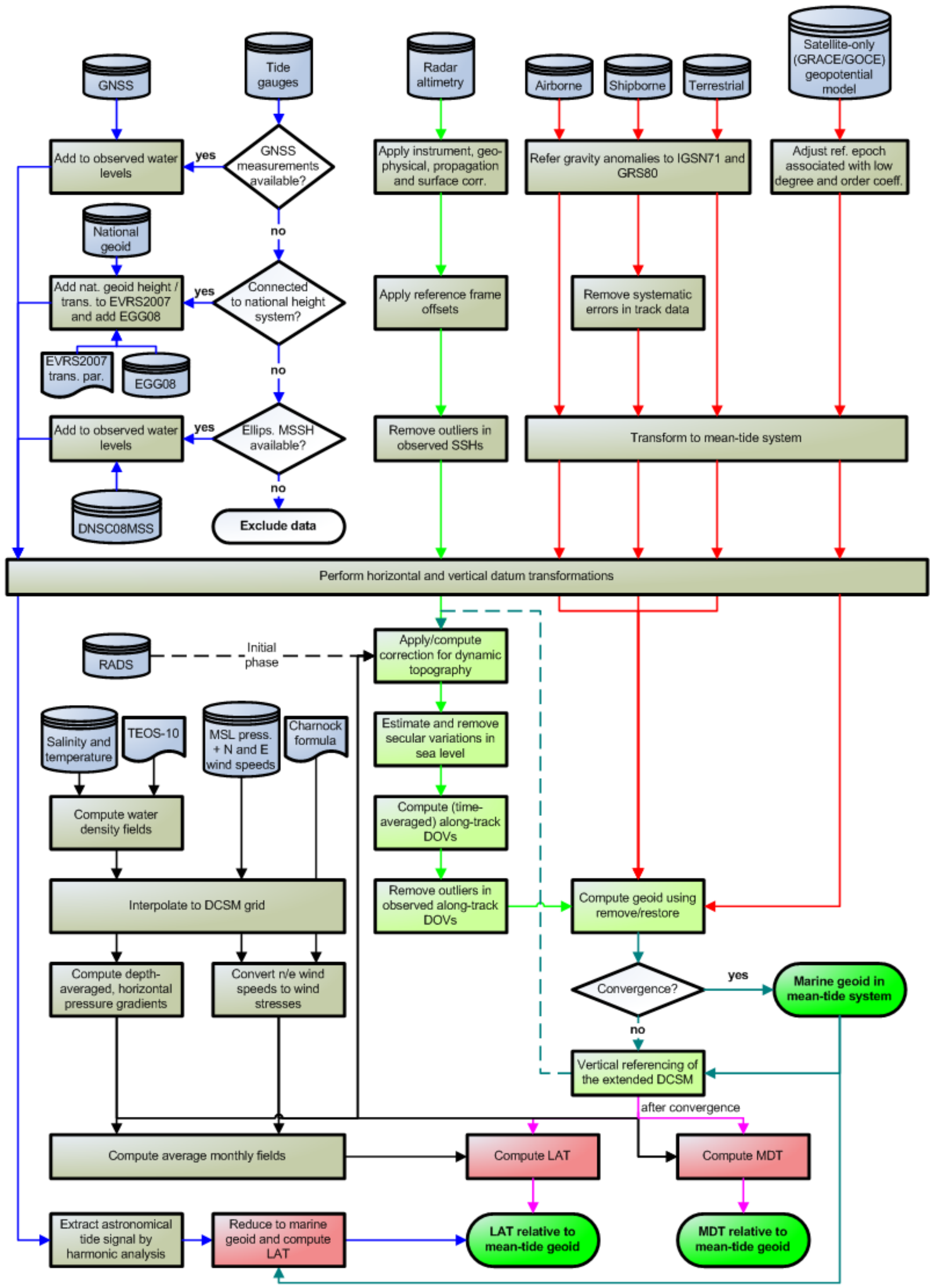

Figure 2.1: Flowchart of the procedure developed in this study. The different colors of the blocks refer to the different phases in the overall procedure: data preprocessing (green), iterative computation of geoid and vertical referencing of the hydrodynamic model (canary yellow), and realization of LAT and MDT (light coral). The different colors of the arrows refer to the different data flows: tide gauge data (blue), radar altimeter data (green), gravity data (red) and salinity, temperature and meteorological data (black). Grey is used in case the arrow refers to a non-data flow or when two or more data groups are involved.

2.1 Data pre-processing

The procedure, outlined in Fig. 2.1, starts with the preprocessing of the various data sets required to compute the geoid (satellite, airborne, ship borne, and terrestrial gravity data; radar altimeter data) and/or required as input for the hydrodynamic model (time-varying mean sea level pressure and north and east components of the 10-meter wind speed fields, time and depth-varying salinity and temperature fields, tide gauge data, and radar altimeter data). The preprocessing aims to harmonize the data and to prepare them for further processing.

First of all, we need to harmonize the horizontal and vertical datums to which the various data sets refer. The horizontal datum used in this study is the European Terrestrial Reference System 1989 (ETRS89) in combination with the GRS80 ellipsoid. Any height data (observed water levels at tide gauge stations and by radar altimeters/height of the gravity data points) are transformed to ellipsoidal heights in ETRS89 as well. Note that the way used to transform **tide gauge water levels** to ellipsoidal heights depends on the information that is available. If GNSS measurements are available, the transformation is straightforward. If not, we will add the national geoid height or we will transform the heights to the European Vertical Reference Frame 2007 (EVRF2007) after which we add the European Gravimetric Geoid (EGG08) heights. In case the tide gauge station is not connected to the national height datum, which applies to most offshore stations, the data are usually provided relative to MSL. Hence, ellipsoidal heights can be obtained by adding the ellipsoidal height of the MSL. In this study, the ellipsoidal heights of the MSL are derived from the mean sea surface computed by the Danish National Space Center (DNSC08MSS) that is not corrected for the inverse barometer effect [Andersen and Knudsen 2009]. The preprocessing of the tide gauge data is concluded by extracting the astronomical tide signal from the observed water levels, which is realized by a harmonic analysis.

Harmonizing the airborne, ship borne and terrestrial **gravity data**, furthermore involves harmonizing the gravity datums and, for ship borne gravity data in particular, the removal of other systematic errors [Wessel and Watts 1988] using e.g. a cross-over adjustment. For the satellite gravity data, we need to adjust the reference epoch of the low degree and order coefficients (C_{20}, C_{30}, C_{40}, C_{21} and S_{21}) to our reference epoch (1999-1-1). Note that all gravity data are supposed to be in the zero-tide system. While, however, the tide generating forces in the hydrodynamic model do not include any contribution at the zero frequency (i.e. modeled water levels refer to the mean-tide system), all gravity data are transformed to the mean-tide system in order to remain consistency. **Note that in this project, no terrain corrections are applied to the terrestrial gravity data, since no suitable digital elevation model was available. For the mountainous regions inside the computational domain, the gravity data are replaced by deflections of the vertical derived from EGG08.**

Besides gravity data, **radar altimeter data** are used to compute the geoid. The preprocessing of these data involves applying all instrument and propagation corrections (e.g. corrections for dry and wet troposphere refraction). In addition, we need to remove all contributions to the observed SSHs that will not be modeled by the hydrodynamic model i.e. the geophysical corrections for load, solid earth and pole tides and the surface correction for the electromagnetic bias. Reference frame differences among the various altimeters are accounted for by applying reference frame offsets to the data, using TOPEX as reference. The last step in the reprocessing involves a removal of outliers in the observed SSHs.

The **meteorological input** to the model is the (time-varying) mean sea level pressure and 10-meter wind speed fields in north and east direction. After interpolation to the DCSM model grid (GRS80 ellipsoid in the ETRS89), the next preprocessing step is to convert the wind speeds to wind stresses for which we make use of Charnock's relation. Only for the computation of the LAT surface, we also need to compute the average monthly variations in mean sea level pressure and wind stress. These variations represent the "average meteorological conditions" referred to in the definition of LAT [International Hydrographic Organization 2011, Technical Resolution 3/1919].

The processing of the 4D **salinity and temperature fields** starts with the computation of 4D water density fields for which we used the international thermodynamic equation of seawater 2010 (TEOS-10) [IOC, SCOR and IAPSO 2010]. After interpolating the water densities to the DCSM model grid, we computed the associated pressure gradients for each depth layer and integrated over the depth to obtain the depth-averaged horizontal pressure gradient fields. Similar to the preprocessing of the meteorological input, for the computation of the LAT surface we need to compute the averaged monthly fields.

2.2 The iterative phase

After preprocessing all data, the procedure continues with the iterative estimation of the geoid and the subsequent vertical referencing of the extended DCSM model. Vertical referencing refers to all steps needed to obtain a model that provides instantaneous water levels relative to the chosen geoid.

The first step is the removal of the dynamic topography signal from the altimeter SSHs. In the initial run, these corrections are derived from the Radar Altimeter Database System (RADS) as the sum of the ocean tide and dynamic atmosphere corrections. In all subsequent iterations, these corrections are replaced by instantaneous water levels relative to the latest estimated geoid obtained from the extended and vertically referenced DCSM model.

The next step is to estimate and remove secular variations in sea level. In general, these secular variations, induced by e.g. melting of glaciers and ice sheets, eustatic sea level changes and Glacial Isostatic adjustment, are not included in the modeled water levels. Hence, they need to be removed. As a result of these two steps, we obtain "geometric" geoid heights.

In order to suppress long-wavelength errors induced by, e.g., radial orbit errors, in the next step we compute the along-track derivatives of the observed SSHs, i.e. the along-track deflections of the vertical (DOVs). For the repeat orbit data we continue to compute the time-averaged along-track DOVs. In the last step before the actual geoid estimation, outliers in the derived along-track DOVs will be removed.

The geoid is computed using the classical remove-restore technique: all data are reduced for the contribution of a global gravity field; the reduced data are fitted to a Spherical Radial Basis Function (SRBF) representation using least-squares (Klees et al 2008); the geoid is computed as the sum of the geoid represented by the global gravity field and the SRBF representation. Note that the reduction of the data for the contribution of the global gravity field is part of the data preprocessing.

In the initial run, we continue with the vertical referencing of the extended DCSM model, after which we start the first iteration. In all iterations it will be evaluated whether the estimated geoid differs statistically from the one obtained in the previous run. If so, the extended DCSM model will be vertically referenced to the latest geoid and a new iteration starts. Otherwise, the extended DCSM model will be vertically referenced to this geoid and the iteration process stops.

2.3 The final phase

In the final phase of the procedure we compute the LAT and MDT surfaces relative to the estimated geoid using the extended and vertically referenced DCSM model. The MDT at each model grid point is computed as the time-averaged modeled water level over the entire simulation period (1984-2004). Here, the model is forced by tides, and meteo and water density variations.

The LAT at each model grid point is derived as the minimum water level over the 1984-2004 time series of modeled water levels. Here, the time-varying wind-stress and water density pressure gradient fields are replaced by their time-averaged values. In order to allow for the dominating seasonal variations in MSL these time-averaged values are not computed over the entire simulation period, but by averaging for each calendar month all available fields over this period. The obtained yearly time series have been used over the entire simulation period. The modeled LAT values are validated by LAT values derived from tide gauge records. These are derived as the minimum water level over the reconstructed astronomical tide signal where the tidal water levels are first reduced to the estimated geoid.

3 The estimated geoid

Test computations have shown that the corrections to the initial geoid after the first iteration are statistically not significant, i.e. the uncertainties in the computed geoid were higher than the corrections. This can be explained by the lack of access to a significant amount of classified gravity data. Therefore, we decided to skip the iteration and to use the state-of-the-art European gravity model EGG08 [Denker et al., 2008] as the initial geoid, referenced DCSM to EGG08, and computed and applied corrections to the radar altimeter data for the dynamic topography signal. In total, we used 164,840 along-track deflections of the vertical (DOVs) acquired by the ERS-1/2, Envisat, Geosat, GFO-1, Jason-1/2 and TOPEX/Poseidon satellites. Note that for the North Sea region, the ERS-1 data set comprises retracked radar data kindly provided by prof. P.A.M. Berry from the Earth and Planetary Remote Sensing Laboratory of the De Montford University in the UK.

Besides radar altimeter data, the data at sea used to compute the geoid include 578,670 ship borne free-air anomalies (FAA) and 8,360 airborne gravity disturbances. Systematic errors in the ship borne data are removed by adjusting the data acquired during a particular cruise to the altimetric-derived FAA obtained from the DTU10 global (on land EGM2008 is used) gravity field model (DTU10GRA) [Andersen et al. 2010, Andersen 2010], i.e. by removing the point wise mean difference between the observed and altimetric-derived FAA from the observed values.

In the non-mountainous regions of the terrestrial part of our computation domain (the Netherlands, Denmark, Belgium and parts of France and Germany) we used 148,004 observed FAA, in the remaining regions including those regions for which no terrestrial gravity data are available (Norway, Sweden, UK, Ireland, Farao Islands and parts of Germany), we use DOVs in northern and eastern direction derived from EGG08. The gravity field removed from the observations is the combined GRACE and GOCE satellite-only Delft Gravity Model release 1 (DGM-1S), which is complete to degree and order 250 [Hashemi Farahani et al., 2012a; Hashemi Farahani et al., 2012b].

To account for any mutual inconsistencies among the various data sets, we estimated one bias parameter for every airborne gravity survey data set, one bias parameter for all corrected marine gravity data, and one bias parameter for each terrestrial gravity data set. In total 62 biases have been estimated. The "optimal settings" of the SRBFs (bandwidth and density) are empirically derived over two test areas. In this way, reliable settings can be found, and the numerical costs are still affordable. In total we need 74,315 SRBFs in order to get a good representation of the residual geoid. The weight factors for the 5 observation groups (radar altimeter data, DOVs derived from EGG08, airborne, marine, and terrestrial gravity data) are estimated using the Monte-Carlo Variance Component Estimation technique of Kusche [2003]. Note that the weight factors are the averaged values of the factors estimated over three overlapping sub domains that comprise the entire computation domain.

3.1 Results and discussion

The huge number of data and unknown parameters require the development of an out-of-core solver to solve the normal equations. The result is shown in Fig. 3.1. In order to interpret the result, we computed the difference between the estimated geoid and the state-of-the-art geoid model EGG08 (cf. Fig. 3.2).

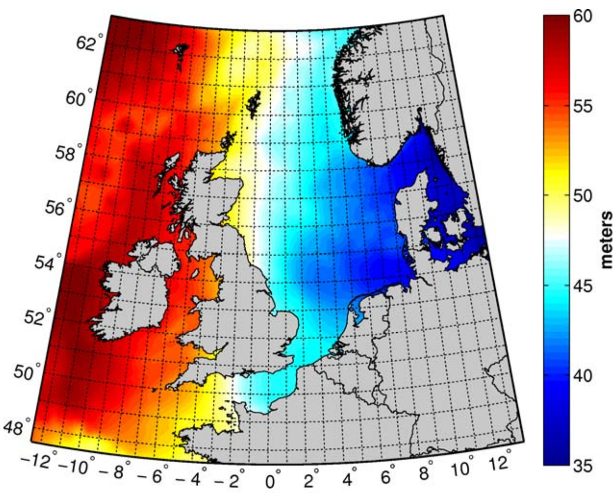

Figure 3.1: The estimated geoid.

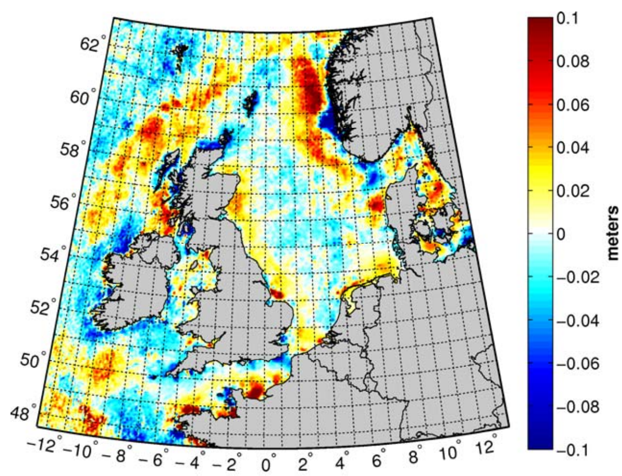

Figure 3.2: The differences between the estimated geoid and EGG08.

The differences are in the range of ± 10 cm. They are very small, not exceeding 2 cm in the middle of the North Sea. Figure 3.2 also reveals some remarkable features. First, we observe that the geographic plot of the differences is not smooth. This can be explained as follows: first, the along-track DOVs derived from radar altimeter data have not been smoothed using a low-pass filter (see below for the reason why this was not done). Second, apparently, the ship borne gravity data set still contains some outliers, which could not be removed in the outlier detection step. Examples for the latter are found in the North Sea just north to the Netherlands were the estimated residuals for some ship borne data points show large deviations, which points to outliers in the data. It should be noted, however, that there might still be some outliers in the radar altimetry data as well, which is suggested by the fact that most outliers show up in coastal waters where radar altimetry data are less reliable. The reason why the altimeter-derived DOVs were not smoothed was that already when applying a moderate smoothing the differences between the estimated geoid and EGG08 significantly increased along the shelf edge, i.e. where there is a strong gradient in the topography. Together with the differences observed along the Norwegian Coastal Current, these differences are the most pronounced features observed in Fig. 3.2. We know from our experiments with the DCSM model (see Slobbe et al. [2012b] and Section 4) that the model over/underestimates the strength of (parts of) these currents and hence the associated dynamic topography corrections applied to the radar altimeter data. Whether this fully explains the most pronounced features in Fig. 3.2 is not clear yet. Potential errors in EGG08 may also be responsible for (a part of) the differences in some areas.

The computation of a high-quality geoid for the North Sea region, which outperforms EGG08 is a huge task and worth to be the subject of a several-year research activity. Among others access to classified gravity data is necessary; the data editing procedure has to be refined, and carefully designed low-pass filters need to be applied to the radar altimetry data, probably with a spatially varying cut-off frequency. Furthermore, a global GRACE/GOCE gravity model may be combined with the other data sets in a statistically optimal way. Finally, terrain corrections may reduce the high-frequency content in the (residual) data and in this way may facilitate a better estimation of the residual field using SRBFs.

4 The mean dynamic topography

The MDT relative to EGG08 is obtained by averaging the modeled instantaneous water levels (time step of 10 minutes) over the period January 1, 1984 to January 1, 2004. The atmospheric forcing input to the model (time-varying mean sea level pressure and north and east components of the 10-meter wind speed fields) is obtained from the publicly available data derived during the interim reanalysis project ERA-Interim [Dee et al., 2011], performed by the European Centre for Medium-Range Weather Forecasts (ECMWF). Salinity and temperature fields used to compute the density variations and hence the depth-averaged baroclinic pressure gradients, are obtained from the Atlantic- European North West Shelf- Ocean Physics Hindcast performed by the Proudman Oceanographic Laboratory (POL), hereafter referred to as POL's hindcast [Holt et al., 2005]. The water levels prescribed at the open sea boundaries are derived as the sum of separate contributions that make up the full instantaneous water levels expressed relative to the geoid: the astronomical tide; surge and steric height. For more details about the experimental setup we refer to Slobbe et al. [2012b].

4.1 Results

The modeled MDT is shown in Figure 4.1. It shows well-known features like the piling up against the Danish and German coasts as a response to the prevailing wind direction and the abrupt rise of the MDT in the Kattegat that reflects the salinity front there, separating the brackish Baltic Sea water from the saline North Sea water, and the associated Baltic current [Ekman and Mäkinen, 1996]. For a detailed description of the MDT and the associated flow field we refer to Slobbe et al. [2012b].

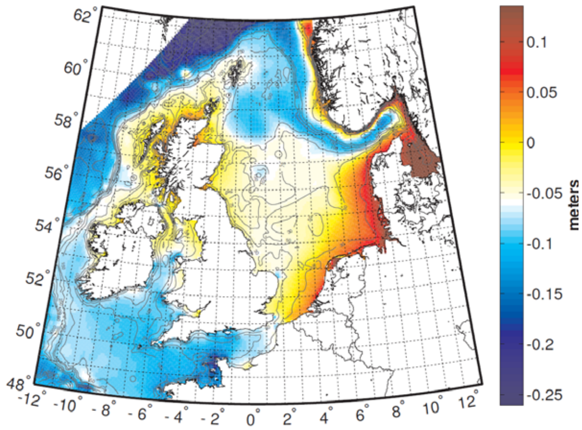

Figure 4.1: The mean dynamic topography relative to EGG08 (January 1, 1984 -January 1, 2004). The grey lines show the contour lines of the bathymetry.

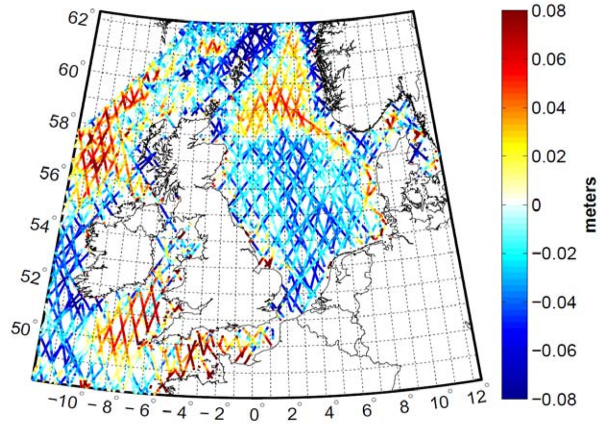

Figure 4.2: Time-averaged residuals at the locations of the radar altimeter data points.

4.2 Validation

In Slobbe et al. [2012b], the obtained MDT is validated using as reference i) mean water levels from POL's hindcast, and ii) differences between ellipsoidal heights of the instantaneous sea surface from radar altimeter satellites (ERS-2, Envisat, GFO-1, Jason-1 and TOPEX/Poseidon) and EGG08 geoid heights. We showed that the 2D model was able to reproduce POL's MDT and monthly mean water levels. The time-averaged differences (shown in Fig. 4.2) between the observed and modeled dynamic topography likely reveal geoid errors in the Celtic Sea and an overestimation of the inflow from the Atlantic Ocean along Shetland, the Fair Isle Current and of the Norwegian Coastal Current, but an underestimation of the inflow from the Atlantic Ocean along the western side of the Norwegian Trench.

5 The lowest astronomical tide

LAT relative to EGG08 is computed by forcing the model with the astronomical tides and the average monthly variations in the meteorological and steric contributions. The LAT value per grid point is derived as the minimum modeled water level over the entire time series. The average monthly variations in the meteorological and steric contributions are computed by averaging, for each calendar month, all available fields over a period of 20 years (1984-2004). The obtained yearly time series have been used for the twenty-year model run. In order to avoid jumps in the modeled water levels in the transition from month to month, the obtained time series are interpolated to 3-hourly values (the original temporal resolution of the meteo data) using a cubic spline interpolation. Here, we assigned the monthly means to the mid-epochs of each month. The meteorological and steric contributions to the tidal water levels prescribed at the open sea boundaries are obtained in the same way as was used to derive the average monthly variations. For more details about the experimental setup we refer to Slobbe et al. [2012c].

5.1 Results

The modeled LAT surface is shown in Fig. 5.1. Many of the features shown in this map are consistent with oceanographic expectations like coincidence of the locations where the modeled LAT values approaches zero with the well-known amphidromic system in the northwest European continental shelf. For a more detailed discussion we refer to Slobbe et al. [2012c]. The minimum and maximum differences between the LAT surface obtained in this experiment and the one obtained using the original approach [Slobbe et al., 2012c, Experiment I] are -15.81 and 5.86 cm, while the RMS is 5.49 cm. The largest differences are observed west of the Danish coast.

5.2 Validation

The obtained LAT surface is validated using LAT values at 92 onshore and 10 offshore tide gauges. Between 95 and 103 constituents are estimated depending on the station. The length of the time series varies between 2.5 years and 20 years. The set of constituents used to derive the observed LAT values is that set for which the RMS of the differences between the observed water levels and the reconstructed astronomical tide is lowest. The obtained differences between observed and modeled LAT reveal a mixture of errors in both model and observations. In Slobbe et al. [2012c], we found that systematic errors in the representation of the tidal amplitude dominate. After correction for all known systematic differences, the average difference between observed and modeled LAT equals -9.13 cm with minimum and maximum deviations of -53.64 and 15.11 cm, respectively. Using the standard deviations of the differences between observed and modeled tidal minima, we performed a test to evaluate the statistical significance of these differences. For three stations, the observed differences are statistically significant at the 99% confidence level. These are located in the Irish Sea where DCSM is known to perform less. In Slobbe et al. [2012c], we also showed that the choice of the set of tidal constituents for each tide gauge has a significant influence on the estimated LAT value at this station. Whether the sets of tidal constituents used in this study are a good choice needs to be investigated.

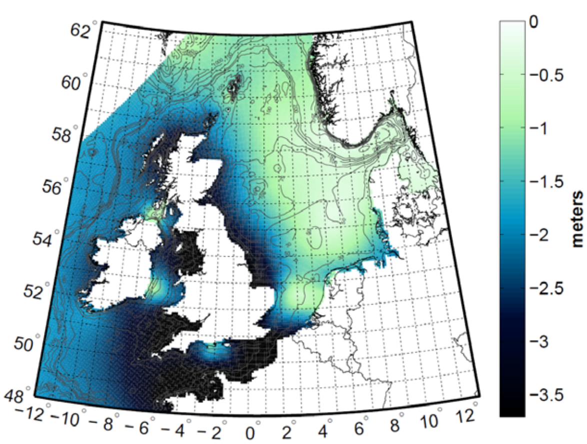

Figure 5.1: Lowest Astronomical Tide computed over the full simulation period January 1, 1984 - January 1, 2004. The grey lines in the map show the contour lines of the bathymetry.

6 Spectral inconsistencies

Computing the ellipsoidal heights of LAT as the sum of geoid heights and the modeled separation between the geoid and LAT is only allowed in case both data sets are spectrally consistent, i.e. both data sets have to cover the same spectral range. If this is not the case, the most obvious approach to achieve consistency is to apply a low-pass filter to the data set with the highest resolution. When, however, the LAT surface has the highest resolution, this approach fails in the coastal regions since the LAT signal is not defined on the continents. As part of the WP3.11, we evaluated whether the use of a bandlimited, spatially concentrated Slepian basis provides a solution to this problem. Since this problem recently attracted much attention in the context of determining the MDT as the difference between the altimetry-derived MSL and a gravimetric geoid [Losch et al. 2002; Bingham et al. 2008; Albertella et al. 2008; Albertella and Rummel 2009], we performed our study in this context as well.

Again, computing the MDT in this way requires that both data sets are spectrally consistent. In practice, it is quite common that the resolution of the geoid data is less than the resolution of the MSL data, hence, the latter need to be low-pass filtered before the MDT is computed. For this purpose conventional low-pass filters are inadequate, failing in coastal regions where they run into the undefined MSL signal on the continents. In our study published in Journal of Geodesy [Slobbe et al. 2012a], we computed Slepian functions for the oceans and parts of the oceans and compare the performance of calculating the MDT via this approach with other methods, in particular the iterative spherical harmonic approach in combination with Gaussian low-pass filtering, and various modifications. Based on the numerical experiments, we concluded that none of these methods provide a low-resolution MSL approximation at the sub-decimeter level. In particular, we showed that Slepian functions are not appropriate basis functions for this problem, and a Slepian representation of the low-resolution MSL signal suffers from broadband leakage. We also showed that a meaningful definition of a low-resolution MSL over incomplete spherical domains involves orthogonal basis functions with additional properties that Slepian functions do not possess. A low-resolution MSL signal, spectrally consistent with a given geoid model, is obtained by a suitable truncation of the expansions of the MSL signal in terms of these orthogonal basis functions. We computed one of these sets of orthogonal basis functions using the Gram-Schmidt orthogonalization for spherical harmonics. For the oceans, we could construct an orthogonal basis only for resolutions equivalent to a spherical harmonic degree 36. The computation of a basis with a higher resolution fails due to inherent instabilities. Regularization reduces the instabilities but destroys the orthogonality and, therefore, provides unrealistic low-resolution MSL approximations. More research is needed to solve the instability problem, perhaps by finding a different orthogonal basis that avoids it altogether.

Although the problem is not solved yet, the spectral inconsistencies between the modeled LAT and MDT surfaces on the one hand and the EGG08 geoid on the other hand are believed to be low. The reason is that at sea the spatial resolution of EGG08 is determined by the spatial resolution of the ERS-1 Geodetic Mission data, which is <10 km both in across- and along-track direction. This is approximately equal to the spatial resolution of the DCSM version 5 model (8×9 km).

7 Final VRF grids

The final VRF grids (MDT and LAT) are incorporated in the BLAST transformation tool [Einarsson, 2011]:

[1] MDT-surface, i.e. the MSL over the period 1984-2004 relative to EGG08 (in the zero-tide system). The MDT surface is modeled by the extended and vertically referenced DCSM model. The reference epoch is 1999-1-1. ***File: mss_delft.gri***
[2] LAT surface, i.e. LAT relative to EGG08 (in the zero-tide system). The LAT surface is modeled over the period 1984-2004 using the extended and vertically referenced DCSM model. ***File name: lat_delft.gri***

Since the MDT and LAT surface refer to EGG08, the transformation parameters for the land vertical datums are similar as those outlined by Strykowski [2010].

8 Dissemination of results

Published ISI journal papers:
Slobbe, D. C., Simons, F.J., Klees, R., 2012. The spherical Slepian basis as a means to obtain spectral consistency between mean sea level and the geoid. J Geod doi:10.1007/s00190-012-0543-x.

ISI journal papers under review:
Slobbe, D. C., Verlaan, M., Klees, R., Gerritsen, H., 2012. Obtaining instantaneous water levels relative to a geoid with a 2D storm surge model. Submitted to Continental Shelf Research.

Slobbe, D. C., Klees, R., Verlaan, M., Dorst, L. L., Gerritsen, H., 2012. Lowest Astronomical Tide in the North Sea derived from a vertically referenced shallow water model, and an assessment of its suggested sense of safety. Submitted to Marine Geodesy.

Conference proceeding paper (in preparation):
Slobbe, D. C., Klees, R., Verlaan, M., Dorst, L. L., Gerritsen, H., 2012. Lowest astronomical tide in the North Sea derived from a vertically referenced shallow water model, and an assessment of its suggested sense of safety. Submitted to Hydro12 – Taking Care of the Sea, Hydrographic Society Benelux, 13-15 November 2012, Rotterdam, The Netherlands.

Presentations (given c.q. submitted):
Slobbe, D. C., Klees, R., Verlaan, M., Dorst, L. L., Gerritsen, H., 2012. Lowest astronomical tide in the North Sea derived from a vertically referenced shallow water hydrodynamic model, and an assessment of the apparent safety. Meeting of the North Sea Tidal Working Group, Brest, France, 7-8 February, 2012.

Slobbe, D. C., Klees, R., Verlaan, M., Dorst, L. L., Gerritsen, H., 2012. Is there something like Lowest Astronomical Tide as Chart Datum? International Symposium on Gravity, Geoid, and Height Systems (GGHS2012), 9-12 October, Venice, Italy

Slobbe, D. C., Klees, R., 2012. The impact of high-resolution dynamic topography in the North Sea on the marine geoid derived from satellite radar altimeter data. International Symposium on Gravity, Geoid, and Height Systems (GGHS2012), 9-12 October, Venice, Italy

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