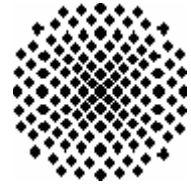




Universität Stuttgart
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Comparison of Mean Sea Surface Data For Oceanography

Studienarbeit im Studiengang

Geodäsie und Geoinformatik

an der Universität Stuttgart

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Selbstständigkeitserklärung

Hiermit erkläre ich, Naomi Zimmermann, dass ich die von mir eingereichte Studienarbeit zum Thema

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selbstständig verfasst und ausschließlich die angegebenen Hilfsmittel und Quellen verwendet habe.

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Preface

I would like to thank

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1. Introduction

The object of this paper is to compare different sets of data of the mean sea surface. For a better understanding definitions of different surfaces related to the topic of this paper are given in the second chapter. These are the Mean Sea Surface (MSS), the geoid and the Mean Dynamic Topography (MDT).

The data sets of the mean sea surface were acquired by satellite altimetry. Therefore an overview of this technique is given in the third chapter. At first the principle of satellite altimetry is explained, and then the measurements and calculations which are needed to acquire the sea surface height. Hereafter different error sources, which turn up at the altimeter, at the signal propagation in the atmosphere and at the sea surface are discussed. Finally a brief overview is given of how the raw data has to be processed to achieve a regularly spaced grid. This grid contains the height of the mean sea surface above a reference ellipsoid at a certain point where latitude and longitude are known.

In the fourth chapter an overview of the different data sets will be given. It will be explained which data sets were used for the calculations, the resolution of these sets and the coverage, i.e. between which latitudes the data is available. Furthermore the satellite missions on which the data sets are based are mentioned. They are TOPEX/POSEIDON (T/P), European Remote Sensing satellite 1 + 2 (ERS1, ERS2) and Geosat. Following this the data sets will be compared by subtracting them from one another. This new grid then contains the differences between two sets of mean sea surface data. The grid is then plotted, to see if there is a pattern in the differences. In the next step the root mean square of these differences is calculated. Hereby different sizes of a moving rms-square and step-size are applied.

In the fifth chapter an outlook is provided to a calculation of the Mean Dynamic Topography. It is calculated by subtracting a geoid from the mean sea surface. The geoid models used are based on measurements by GRACE (Gravity Recovery And Earth Experiment). Since this geoid is available only in harmonic coefficients, it is transformed into heights above a reference ellipsoid. There are two grids, one contains the mean sea surface data, the other contains the geoid data. These two grids can be subtracted from each other. An important problem can turn up here. If the wave length of the geoid used is too long, then features due to ocean bottom topography will be flattened. They therefore can turn up in the difference between the altimetric data, which measures these features and the geoid, which does not show them. If an ideal geoid was subtracted from the mean sea surface only oceanographic features would remain.

Keywords:

Geoid · Mean Dynamic Topography · Mean Sea Surface · Oceanography · Satellite Altimetry

2. Technical background

2.1 Definition of the Mean Sea Surface (MSS)

The mean sea surface is the height of the sea surface above a certain reference ellipsoid. This surface is an average sea surface, based on measurements over several years. The data is obtained through satellite altimetry. The altimeter measurements have to be corrected for the effects of the ocean and solid earth tides and variations of the sea surface due to atmospheric pressure loading.

Measurements are only available on the satellite ground tracks. This radar signal illuminates a circular area at the sea surface. This is the so-called footprint. Because only one measurement is taken for the whole footprint this means that the measured height is an average (Seeber, 2000).

Because measurements are only available on the satellite ground tracks the data still has to be processed to form a regularly spaced grid. Of each grid point latitude, longitude and height are known. The sea surface itself consists not only of these grid points, but also contains the area between these points (Zandbergen, 1990).

2.2 About the geoid models

The geoid is an equipotential surface. Geoid undulations are the differences between the geoid and a suitable reference ellipsoid. These undulations can range between -105 m and + 85 m over the oceans. They originate in the variable ocean bottom topography and are also due to the inhomogeneous distribution of mass inside the earth (Chelton et al., 2001). The earth's gravity field also changes with time due mainly to the lunar and solar tides, but also due to other forces, e.g. post-glacial rebound or the varying distribution of mass within the atmosphere and ocean (Chelton et al., 2001).

Since a geoid, which constantly varies with time is difficult to use for calculations, geoid models are introduced. These are defined to be time invariant. The models are mostly presented by a harmonic expansion (Zandbergen, 1990). They are available in harmonic coefficients.

The two geoid models used in this paper are the EIGEN-GL04S1 and the EIGEN-GRACE02S based on GRACE (Gravity Recovery And Climate Experiment) measurements. Both geoid models were calculated by using gravity data based on satellite measurements only. This is important if geoid models are to be used for oceanographic purposes. The two models are developed until degree and order 150.

2.3 Definition of the Mean Dynamic Topography (MDT)

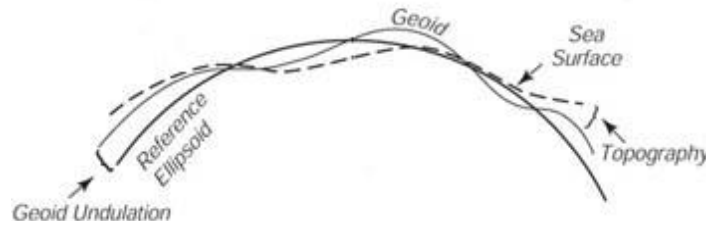


Figure 1 Mean Dynamic Topography

http://oceanworld.tamu.edu/resources/ocng_textbook/chapter03/Images/Fig3-13s.jpg

The mean dynamic topography (MDT) ζ consists of the mean sea surface minus the geoid:

$$\zeta = h - N \quad (1)$$

h - mean sea surface

N - geoid undulations

If there were no forces acting on the sea surface except for gravitation and the centrifugal force introduced by the rotation of the earth, then the sea surface would coincide with the geoid, an equipotential surface (Chelton et al., 2001).

The mean dynamic topography should not contain the ocean tides, solid earth tides and variations of the sea surface height due to atmospheric pressure loading. Therefore these quantities have to be removed from the mean sea surface, which is needed to calculate the mean dynamic topography.

Another problem in the calculation of the mean dynamic topography is, that the geoid also has to be exact over short wave-lengths. A long wave length geoid is based on coefficients up to degree and order 150. The shortest wave lengths show the features due to the topography and the variations in the crustal density (Cazenave, 1995). If the MDT is calculated with a long-wavelength geoid the variations of the sea surface due to the ocean bottom topography overlay oceanographic features. This means the factor which limits extraction of oceanographic features due to large or meso-scale ocean circulation is the choice of geoid models. The measurements of the mean sea surface by altimetry are exact enough for oceanographic purposes. With current altimeter missions 1 to 2 cm are available (Seeber, 2003).

3. About altimetry

3.1 Principle

The basic principle of altimetry is easy to understand. From a satellite a radar pulse is transmitted toward nadir and reflected by the sea surface. The travel time between emission of the radar pulse and reception at the altimeter is measured. From this time t the distance, here called range R , between sea surface and satellite can be obtained by using the simple equation

$$R = 1/2 c t \quad (2)$$

where c is the speed of light in vacuum (Chelton et al. 2001). Now the height of the satellite above a certain reference ellipsoid is needed. This height can be calculated by Precise Orbit Determination. If the distance to the sea surface is subtracted from the height of the satellite the result is the mean sea surface (Chelton et al. 2001).

3.2 Quantities

To calculate the mean sea surface above a reference ellipsoid two quantities are needed:

- the height H of the satellite above a certain reference ellipsoid
- the range R between the satellite and the sea surface

By subtracting the range from the height the mean sea surface h can be obtained (Chelton et al., 2001):

$$h = H - R \quad (3)$$

The quantity measured directly by the altimeter is the range R . However there still need to be some corrections applied to this measurement.

The height H can be obtained by Precise Orbit Determination.

In the next sections the acquisition of these quantities is elaborated.

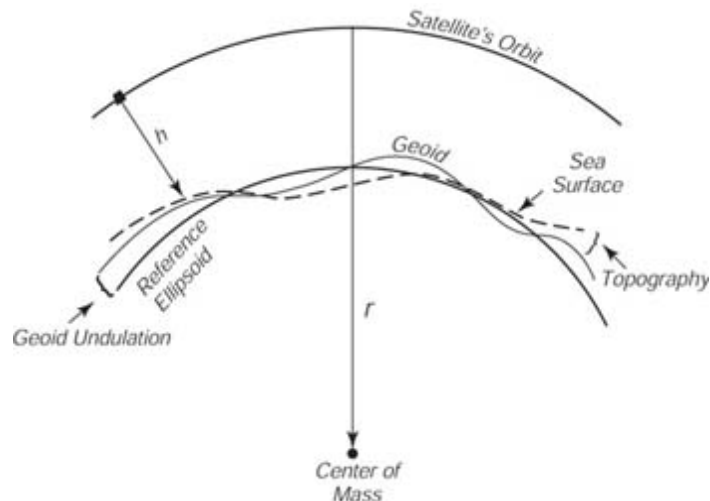


Figure 2 Principle of Altimetry
http://oceanworld.tamu.edu/resources/ocng_textbook/chapter03/Images/Fig3-13s.jpg

3.3 Precise Orbit Determination

The height of the satellite above a reference ellipsoid is achieved by Precise Orbit Determination (POD). This location is described by the exact 3-dimensional coordinates of the satellite's center of mass at a certain time. POD uses mathematical models and precise observations of position or velocity of the satellite (Chelton et al., 2001).

One method to achieve a precise orbit is through dynamic modelling. With this method the orbit of the satellite is calculated using an initial starting point, a gravity model and other estimated parameters which are then combined in a mathematical model. With this model the orbit can be predicted. There are two main error sources, which limit the accuracy of dynamic modelling of the orbit. One is that the gravity model is not known exactly enough. The other is the modelling of atmospheric drag, which is the slowing down of the satellite due to friction in the atmosphere. The extent to which this drag influences the velocity depends on the air density, and therefore on the altitude of the satellite. The air density also depends on solar effects and on changes in the geomagnetic field. Another problem is that the satellite's center of mass can change due to consumption of fuel during the mission (Chelton et al., 2001). Because there are additional errors in the initial parameters this first calculated orbit will not be sufficiently exact (Chelton et al., 2001). Therefore direct observations of the satellite are taken into account. These observations are used for mathematical models as well. In these models the motion and orientation of the satellite are inserted and also the motion of the observing station, which can be a ground- or satellite-based tracking station. The motion of a ground-based station is due to the rotation of the earth and to surface deformations (Chelton et al., 2001). Since the observations are also achieved by measuring with frequencies which are influenced by the ionosphere, the model should also contain corrections for atmospheric refraction (Chelton et al., 2001). For further processing the observations are compared with the calculated observations. To minimize the differences a least squares method is applied iteratively until the desired accuracy is achieved (Chelton et al., 2001).

With Topex/Poseidon mainly two types of observations were used: Satellite Laser Ranging (SLR) and Doppler Orbitography and Radiopositioning Integrated by Satellite (DORIS) (Chelton et al., 2001). An experimental GPS receiver was also installed. The observations from the receiver were however not taken into account for the determination of the orbit, but were used to figure out whether satellite tracking by GPS meets the requirements of POD (Chelton et al., 2001).

In SLR an optical pulse is transmitted from a ground station, is reflected at the satellite and then received at the ground station. Since a position of the satellite's center of mass is required, the difference between the position of the reflector and the satellite's center of mass has to be taken into account (Chelton et al., 2001). An advantage of SLR is that the frequencies in the optical spectrum are less influenced by the ionosphere and the water vapor content. The disadvantages are that it only works for clear skies and that there are too few ground stations (Chelton et al., 2001).

DORIS is a network of one-way ground-to-satellite tracking stations with Omnidirectional Beacons (ODB). The satellite carries a receiver, which measures the Doppler shift. With this the range-rate (rate of change of range) relative to the beacon can be calculated. To eliminate the errors due to ionospheric refraction the signal is transmitted on two frequencies. An advantage of DORIS is that it does not depend on clear skies. The global coverage is better, because the satellite can receive DORIS signals along most of its orbit, since the beacons transmit in all directions. DORIS does not measure absolute positions. In combination with SLR it provides a high orbit accuracy estimation (Chelton et al., 2001).

GPS, a further possibility of location determination, provides continuous 3-dimensional coordinates of the satellite. For Precise Orbit Determination with GPS a mathematical model would not be necessary, since the orbit could be determined purely geometrically (Chelton et al., 2001).

3.4 Range R

The principle seems quite easy. A radar pulse is sent toward nadir, is reflected at the sea surface then received at the altimeter. The time t between transmission and reception is measured and multiplied by $c/2$ where c is the speed of light in vacuum (Chelton et al., 2001). So then the range is

$$R = 1/2 c t \quad (4).$$

3.4.1 Determination of time t

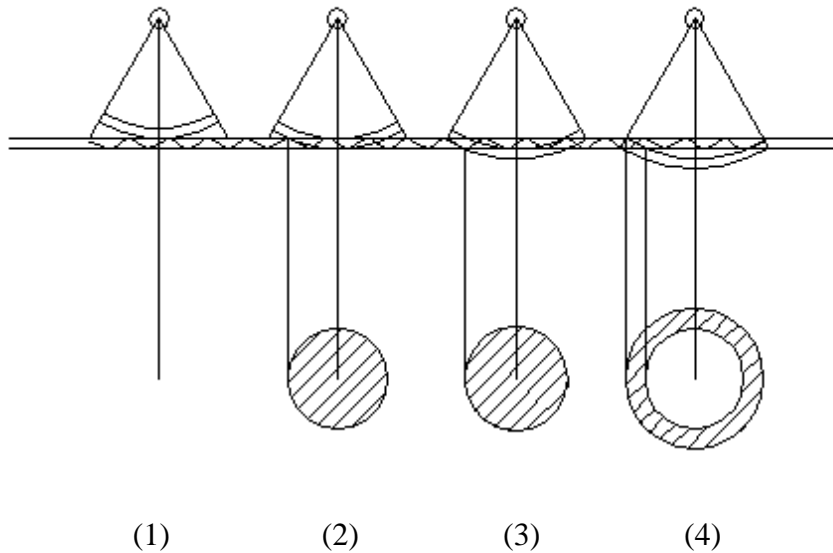


Figure 3 Pulse propagation from satellite to sea surface

When the signal first is transmitted there is no returned signal until the leading edge of the pulse is reflected by the wave crests at the sea surface (Fig. 3 (1)). From then on a returned signal can be measured. The illuminated area on the sea surface is now an expanding circle (Fig. 3 (2)), the greater the circle, the more power is received at the altimeter (Fig. 3 (3)). When the trailing edge of the pulse is reflected at the plane of wave troughs, the circle becomes an expanding ring, as shown in the picture above (Fig. 3 (4)). Figure 3 shows the propagation of the pulse for four different times t .

It can be shown (Chelton et al., 2001) that the area of this ring remains the same, which means that now the power received at the altimeter stays constant, except for the power lost by scattering at the sea surface. The returned waveform now looks like this (Fig. 4).

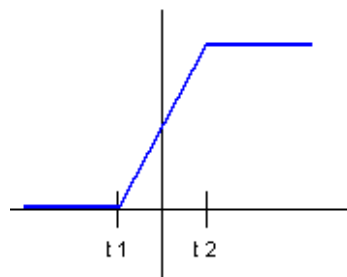


Figure 4 Returned waveform

Time t_1 is the time when the leading edge of the pulse is reflected by the sea surface and the altimeter at first receives a signal. Time t_2 is the time, when the trailing edge is reflected by the sea surface and the expanding circle becomes an expanding ring, from which the backscattered power stays the same (Zandbergen, 1990). The midpoint of the slope of this returned waveform is the time $t_{1/2}$ with which the range R can be calculated. This is a simplification of the returned waveform since the larger the ring becomes the more power is lost from scattering (Chelton et al., 2001). This waveform can also be used for other measurements. The width of the slope can be taken for an estimate of the significant wave height, and the height of the slope for a surface roughness estimate (Sandwell, 2001).

3.4.2 Corrections

3.4.2.1 Corrections at the altimeter

At the altimeter several corrections need to be applied:

- the Doppler-shift error
- the acceleration error
- the oscillator drift error
- the pointing angle
- the antenna feed bias
- timing error

Doppler-shift error

From the altimeter a so-called chirp is transmitted. This is a frequency-modulated pulse, which means the frequency of the transmitted signal changes with time. The pulse is reflected at the sea surface and analyzed depending on the frequency in order to obtain time t . If this frequency changes due to a different phenomenon than the intended frequency modulation an error in the estimate of the two-way travel time is introduced. A reason for this frequency change is the Doppler shift because of the relative velocity of the satellite compared to the sea surface. The Doppler-shift can easily be calculated now and from this shift the corresponding error in the estimated time can be derived. (Chelton et al., 2001)

Acceleration error

The adaptive tracking unit is used to extract the desired information e. g. the two-way travel time from the returned waveform. Connected to this unit is a device, which estimates range rates (rate of change of the range). This device does not measure the relative acceleration between sea surface and altimeter. If there is a non-zero acceleration, this leads to an incorrect estimate of travel time. (Chelton et al., 2001)

Oscillator drift error

The altimeter measures the travel time of the pulse by counting oscillator cycles. If the frequency of the oscillator is not known correctly, the estimation of the travel time contains an error. To compensate the drift of the oscillator, it has to be calibrated at least once per week. (Chelton et al., 2001)

Pointing angle

A further error is introduced when the pointing angle of the altimeter is not determined correctly. This leads to an error in the estimated time, and therefore to an error in the range estimate. (Chelton et al., 2001)

Antenna feed bias

It is important to know the difference between the satellite center of mass and the antenna phase center, because the range is measured from the antenna phase center, but the satellite orbit is determined with respect to the satellite's center of mass. This correction parameter can be measured before the launch of the satellite. However the satellite's center of mass may change during the mission, due to fuel consumption. This introduced error should also be taken into account. (Zandbergen, 1990)

Timing error

This error occurs if the time of the range observation is not determined correctly. To calculate the mean sea surface height at one point the position of the satellite at a certain time has to be determined along with the range measured at the same time. If the time of the range estimation is not correct, the range is not applied to the correct corresponding position of the satellite, which would result in an incorrect estimation of the mean sea surface. (Zandbergen, 1990)

3.4.2.2 Corrections at the signal propagation through the atmosphere

Microwave signals are refracted when passing through the atmosphere. In the ionosphere the refraction depends on the frequency, in the troposphere it does not. The higher the frequency, the lower the refraction is in the ionosphere. However the atmospheric absorption in the troposphere increases if the frequency is higher than 30 GHz (Seeber, 2003).

The signal is influenced by the atmosphere at the following levels:

- the ionosphere
- the dry troposphere
- the wet troposphere

denomination	frequency
P-band	220 - 300 MHz
L-band	1 - 2 GHz
S-band	2 - 4 GHz
C-band	4 - 8 GHz
X-band	8 - 12.5 GHz
Ku-band	12.5 - 18 GHz
K-band	18 - 26.5 GHz
Ka-band	26.5 - 40 GHz

Table 1 Radar bands from Seeber (2003)

Ionosphere

The ionosphere is a dispersive medium. This means that the velocity of the signal depends on the frequency. The signal is delayed because of the free electrons in the ionosphere. How much the signal is influenced depends on the total electron content (TEC) of the ionosphere. The TEC depends on the solar activity and the geomagnetic field, which means it depends on the time of day, and the geographic location. In higher latitudes and near the equator, the TEC is higher (Seeber, 2003).

There are different methods of correcting the error introduced by the ionosphere. One is to install a dual frequency altimeter onboard the satellite. From the difference in the arrival times of the signals the TEC can be calculated. Topex/Poseidon e.g. has a dual frequency altimeter which transmits at 13.6 and 5.3 GHz (Seeber, 2003). Another possibility is to install a device which measures the TEC directly. The correlation between signal attenuation and TEC is well known (Seeber, 2003). A third possibility is to obtain the measurements of the TEC from other satellites, which have measuring devices installed or which transmit at two different frequencies. However, the total electron content must be then adjusted for the position of the satellite where these measurements are needed (Chelton et al., 2001). There are also models to calculate the signal propagation delay e.g. a model by Klobuchar (1987), but measurements with two frequencies tend to be more exact. With knowledge of the total electron content the range correction can be calculated (Seeber, 2003).

Troposphere

In the troposphere the refraction does not depend on the frequency. The refractivity depends on the air pressure, the partial pressure of water vapor, the pressure of dry gas and the temperature as in Hopfield (1969, 1971). It is usually not possible to measure the refractivity along the signal path directly. Therefore models are introduced, which rely on measurements of meteorological data acquired at the surface (Seeber, 2003). However, there are very few places on the open ocean, where the data can be measured directly.

Dry troposphere

The dry tropospheric refractivity depends mostly on the atmospheric pressure at the nadir point. Since there are very few points where the atmospheric pressure at the sea level can be measured directly, one has to find different means of obtaining these observations. The atmospheric pressure could be calculated from weather prediction models for example. The refraction which occurs due to the dry part of the troposphere can be modelled quite well (Seeber, 2003).

Wet troposphere

The wet tropospheric refraction depends mostly on the water vapor content and by about two orders of magnitude less on the cloud water droplet density. The contribution of the wet tropospheric refraction to the total tropospheric refraction is about 10% (Seeber, 2003). However it is more difficult to model than the dry tropospheric contribution. That is why a radiometer to measure the water vapor content has been installed onboard all altimeter satellites since SeaSat. An exception is Geosat. To acquire the content of water vapor in the atmosphere, Geosat used observation data from meteorological satellites (Seeber, 2003). The signal attenuation introduced by rainfall can not yet be calculated with sufficient precision. Therefore data which is influenced by rainfall is not used for range estimation (Chelton et al., 2001).

3.4.2.3 Corrections at the sea surface

Following factors influence the range measurement:

- the sea state bias
- the slope induced error
- the ocean tides, solid earth tides and atmospheric pressure

Sea state bias

The total sea state bias consists of the electromagnetic bias and the skewness bias. These biases occur because the pulse transmitted by the altimeter is reflected by specular reflectors (wave facets at right angles to the direction of propagation of the pulse) and not by the mean sea surface (Chelton et al., 2001). The electromagnetic bias occurs, because more energy is reflected from the wave troughs than from the wave crests due to the fact that wave troughs tend to be flatter than wave crests (Chelton et al., 2001). The electromagnetic bias depends on the significant wave height. This height is acquired by measuring the heights of waves in a wave field and taking the average of the top third (e.g. the 33 highest waves out of a wave field containing 99 waves and calculating the average of these 33 waves). This is referred to as $H_{1/3}$. This bias increases if the waves are higher (Chelton et al., 2001). This bias is the difference between the mean sea surface and the mean scattering surface of specular reflectors (Chelton et al., 2001). The mean is the sum of the heights of the specular reflectors divided by their number (Hamburg, 1970).

The skewness bias occurs because the pulse transmitted by the altimeter is reflected by specular reflectors. This means, that the range between the actual sea surface and the altimeter is not measured, but rather the range between the median height of these specular reflectors and the altimeter. This surface is called electromagnetic sea level. The skewness bias is the difference between the mean scattering surface of specular reflectors and the median scattering surface. It does not depend on the significant wave height (Chelton et al., 2001). The median is the middle height value, if the heights of the specular reflectors were arrayed by size. This skewness arises because the distribution of the height values of the specular reflectors is not symmetric (Hamburg, 1970).

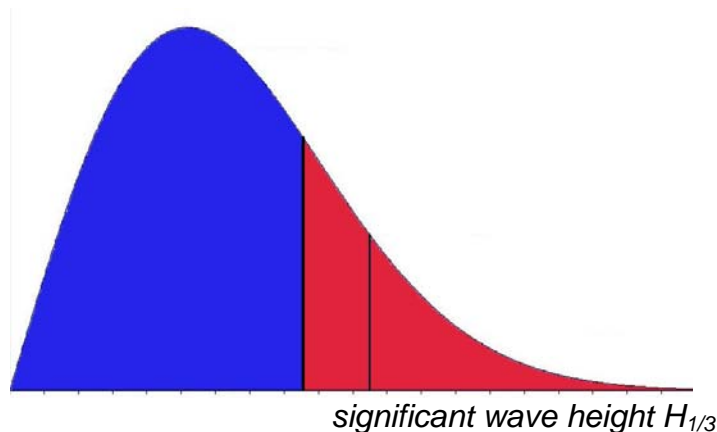


Figure 5 Wave height
http://www.vos.noaa.gov/MWL/aug_05/Images/nws_2.jpg

Slope induced error

If the altimeter measures at a slope, the signal is not reflected exactly at the nadir point, but at some distance to it. This results in an error in the range estimate. It is difficult to correct this error, since the slopes which show up on the sea surface have to be known beforehand (Zwally and Brenner, 2001).

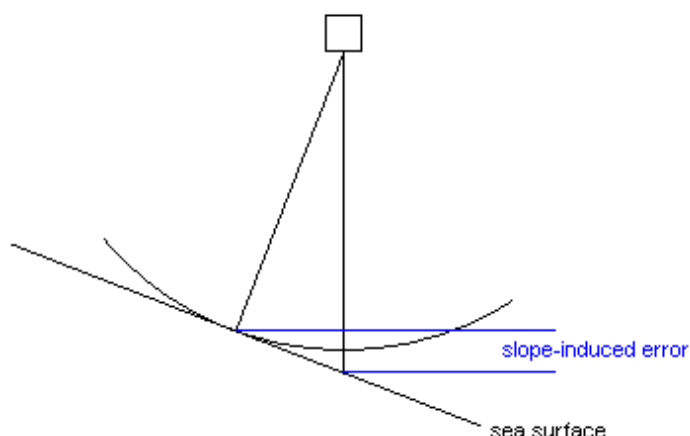


Figure 6 Slope induced error

Ocean tides, solid earth tides, atmospheric pressure

Another correction which has to be applied is the difference between the instantaneous and the mean sea surface due to the solid earth tides, ocean tides and atmospheric pressure loading. These corrections are seen as differences from the mean sea surface and have to be removed before the gridding process. This correction is different from those mentioned above as it does not depend on errors introduced by altimetric measurement.

The ocean tides mainly depend on the gravitational forces from the moon and the sun. Because the motion of sun and moon relative to the earth are known quite exactly a tidal-generating potential can be estimated. This tidal potential can be developed in a harmonic expansion. The exact coefficients for this potential were developed by Doodson (1922). The six coefficients with the largest amplitudes are sufficient to calculate a relatively exact tidal potential. From this potential a tidal model can be derived. So far the best tidal models have an accuracy of 2 to 3 cm (Chelton et al., 2001; Seeber, 2003). Because the repeat periods of the altimeter satellites mostly exceed several days and the tides occur once or twice a day (diurnal or semi-diurnal), these variations are not resolved. Ocean tides have to be removed, if the mean dynamic topography is to be used for estimation of ocean circulation. (Chelton et al., 2001)

The solid earth tides consist of the tides introduced by moon and sun and the tidal upload due to the pressure which the ocean tides exert on the earth's crust. These have to be removed from the altimeter data along with the ocean tides. (Seeber, 2003)

The pressure of the atmosphere influences the sea surface. The variations of the sea surface due to pressure can reach 10 - 20 cm. To obtain the pressure field estimates at certain areas, weather prediction models are used. However the uncertainties in these estimates are large enough to introduce an error of about 2 to 4 cm in the height. The inverted barometer effect limits the accuracy of the sea surface height. (Seeber, 2003)

3.5 Gridding

For most applications for which altimeter data is used, not only data measured directly along the satellite ground track is needed, but a regularly spaced grid containing the mean sea surface heights. To obtain this grid the altimetric data is interpolated and smoothed with different algorithms. It is important to know how these algorithms affect the resulting calculated sea surface. If the signals are smoothed too much, some smaller signals might be lost. Which algorithms are to be used depends on the orbit configuration of the altimeter satellite as well as on the purpose for which the data is to be used (Zandbergen, 1990). One possibility for the mapping of the grid points is the least squares minimization. However this method requires much computer power. Other methods calculated with less effort prove to be nearly as exact (Zandbergen, 1990).

It is hardly possible to map a grid with data from just one altimeter satellite, since the orbit configuration only allows short repeat periods or closely spaced data points. This means that the data either has a high temporal resolution whereas the ground tracks have greater distances from each other or the data has a high spatial resolution, but a long repeat period (Chelton et al., 2001).

To observe oceanographic features which change over several weeks to several months a ground track repeat period exceeding 20 days is unfavorable (Chelton et al., 2001). Topex/Poseidon has a 10 day repeat, whereas Geosat and ERS have repeat periods of 17 and 35 days. Topex/Poseidon is therefore better suited to observe oceanographic features (Chelton et al., 2001). The mean sea surface data sets discussed in this paper are based on measurements by several altimeter satellites e.g. Topex/Poseidon, ERS1 and ERS2, and Geosat.

Another process in gridding is the insertion of data in the overland gaps. Here no data from the altimeter satellite is available. Usually the areas over land are substituted by a geoid. For the boundary between the altimeter data and the geoid a smoothing function is applied (Zandbergen, 1990).

4. About the data sets

4.1 Description of the data sets

For this paper three sets of data of the mean sea surface are presented. The first data set is the CLS01 MSS. The second data set is the KMSS04. The third data set is the WHU2000. All these sets have a global coverage between 80° N and 80° S, since this is about the limit to which latitude altimetric data is available. The resolution of the data sets is 1/30°, or 2' x 2'. In the following passage the satellite missions on which the different mean sea surface data sets are based are described.

Satellite missions	Mean Sea Surface
Topex/Poseidon	CLS01 MSS, KMSS04, WHU2000
ERS - 1/2	CLS01 MSS, KMSS04, WHU2000
ERS 1 geodetic phase	CLS01 MSS
GFO	KMSS04
Geosat	CLS01 MSS
Geosat GM	KMSS04
Geosat ERM	WHU2000

Table 2 Satellite missions for Mean Sea Surface data

The CLS01 Mean Sea Surface is based on a Topex/Poseidon 7-year mean profile, a ERS-1 and ERS-2 5-year mean profile, a Geosat 2-year mean profile and the two 168-day periods of the ERS-1 geodetic phase. The T/P ellipsoid was used as reference surface.

The KMSS04 data calculated by the DNSC is likewise based on these satellite missions, whereby the data acquired by Topex/Poseidon from 1993 - 2001, ERS2 from 1995.5 - 2001.5, Topex/Poseidon TDM from 2002 - 2003, GFO (Geosat Follow-On) from 2000 - 2001, ERS-1 GM (Geodetic Mission) from 1994 - 1995 and GEOSAT GM from 1985 - 1986 was used. The WHU2000 data was calculated with 7 year Topex/Poseidon data, Geosat ERM (Exact Repeat Mission), ERS-1 and ERS-2 data. The T/P ellipsoid is the reference ellipsoid for this data as well.

4.2 Two different areas in the North Atlantic

From these global sets of data two smaller areas in the North Atlantic were extracted. The first area lies in the region of the Gulf stream off the East coast of North America. The exact coordinates are 280 - 300° longitude, latitude: 25 - 45° N. This area was chosen because it is expected to be more turbulent due to the Gulf Stream. The second area is in a region where less movement is expected since it lies in the middle of the North Atlantic Ocean. The coordinates of this area are 310 - 330° longitude, latitude: 25 - 45°.



Figure 7 Two areas in the North Atlantic
http://www.mygeo.info/karten/physische_weltkarte_cia_2007.jpg
Physical Map of the World April 2007

4.3 Error files for one of the surfaces

By plotting the error file of the first data set, a typical pattern of satellite ground tracks can be seen. The points measured directly along-track from the satellite are most exact. The regions between these ground tracks have a greater error, since they have to be calculated from the surrounding measured points. At the interpolated grid points the error can be as high as 0.1 m.

Near the coastal areas the data tends to be less reliable. So if data sets are compared, the coastal areas should not be taken into account. As can be seen on the plot (Fig. 8) the maximum errors occur near the coast and over islands.

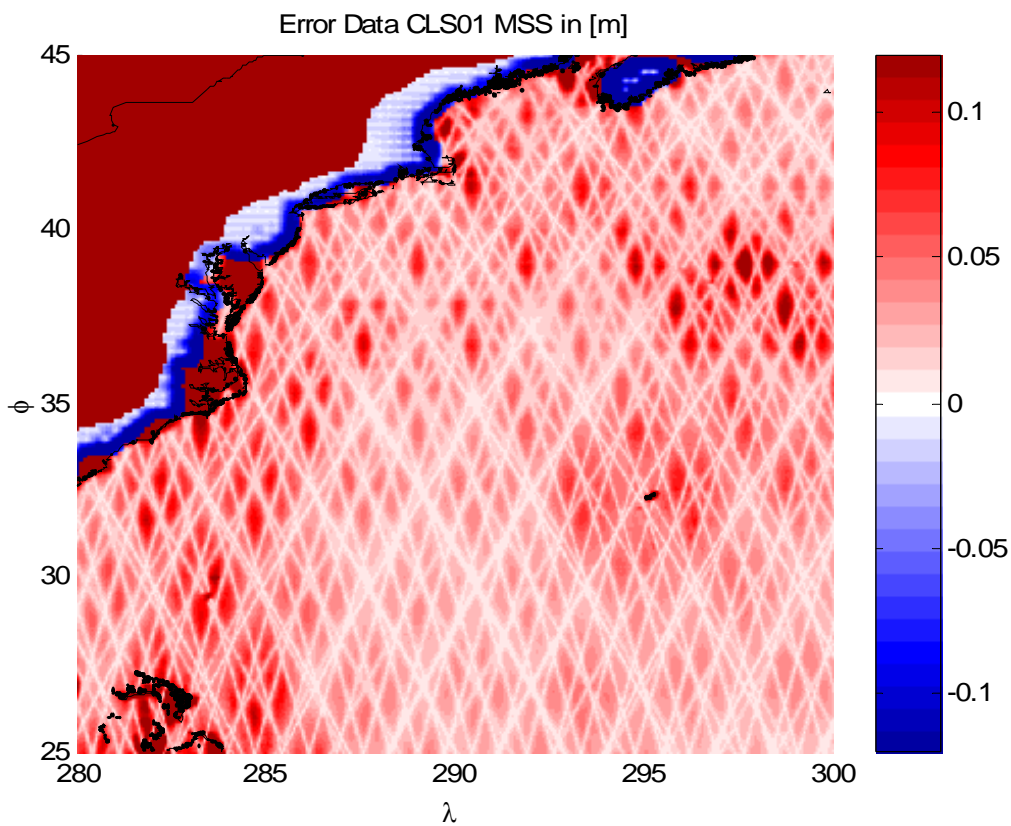


Figure 8 Error data for the first area (Longitude: 280 – 300°)

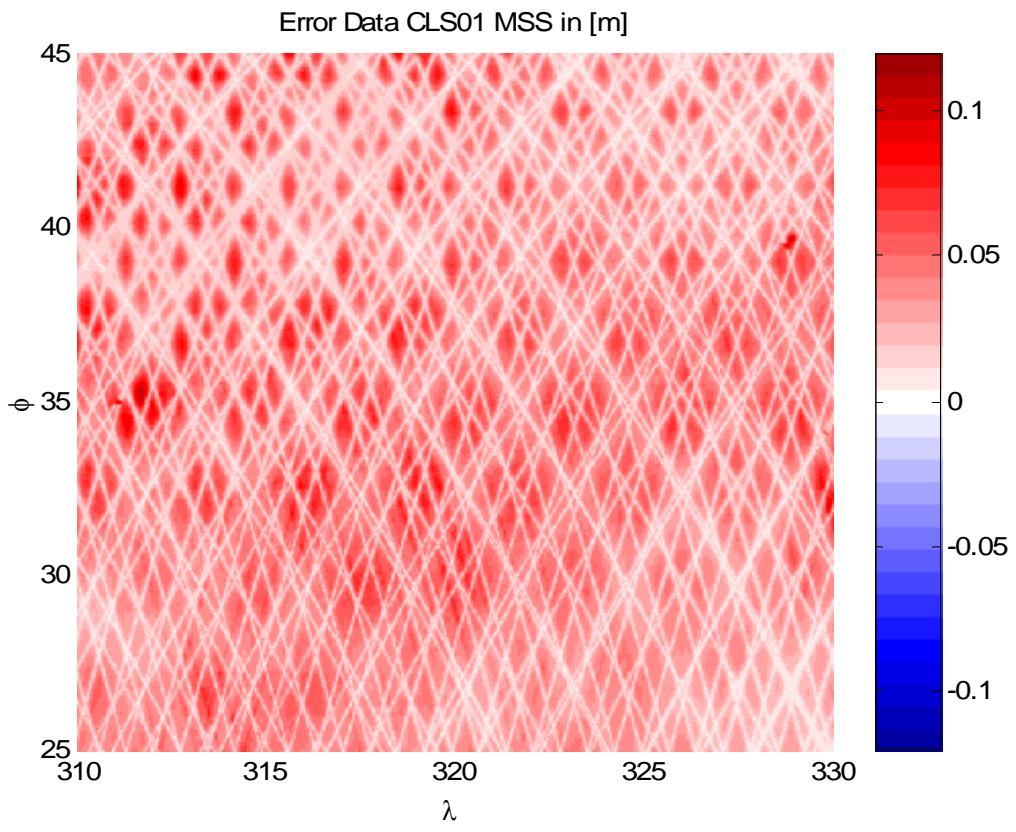


Figure 9 Error data for the second area (Longitude: 310 – 330°)

4.4 Comparison of the three data sets

At first the KMSS04 is subtracted from the CLS01 MSS. There was no problem with different resolutions, because both data sets have a resolution of $1/30^\circ$. With the data set from the DNSC there is a program provided with which the grid can be interpolated. There are two different interpolations possible: a linear interpolation or a spline interpolation. This interpolation program was not needed in this case.

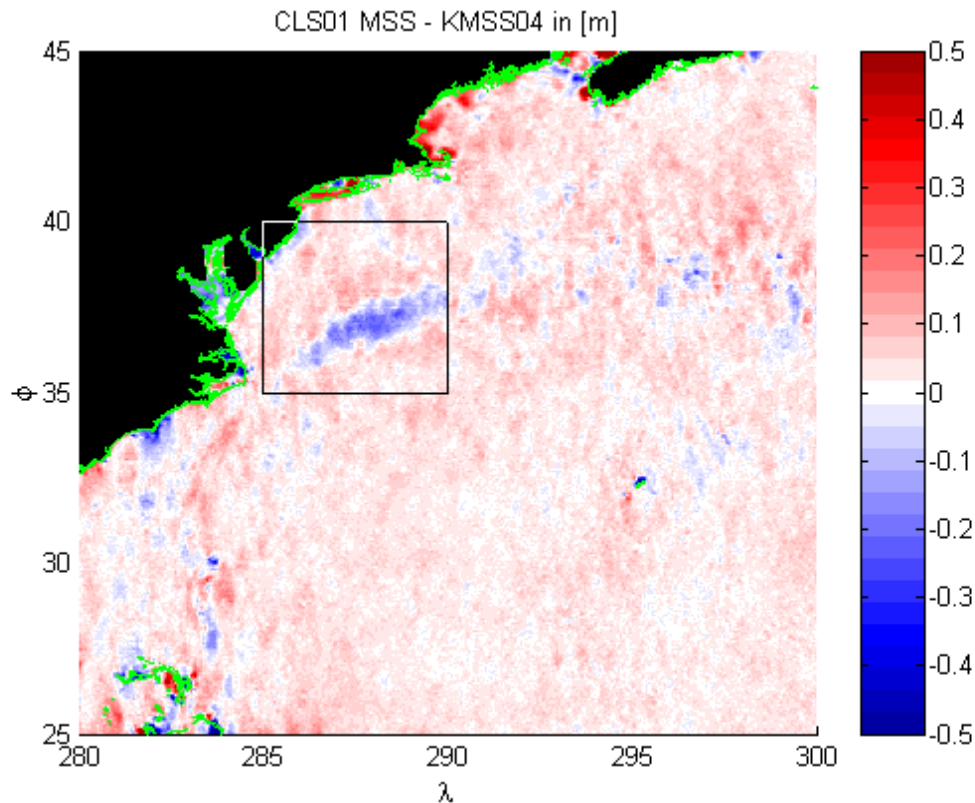


Figure 10 Difference between CLS01 MSS and KMSS04

Figure 10 shows the differences between two sets of mean sea surface data, the CLS01 MSS and the KMSS04. As was expected the areas near the coast differ most, since here the data from the altimeter is not reliable, firstly because there can be land in the altimeter footprint. There is no altimeter data available over land. Secondly tides can not be modelled sufficiently exactly near the coast. The tidal range here is much greater than in the open ocean.

The standard deviation for this area is 0.06 m. The maximum difference is 1.03 m, the minimum difference is -2.5 m. However, these differences show up near the land. Over land areas a geoid was used to continue the sea surface. Different methods of calculating the transition or the use of different geoids lead to larger differences near coastal areas. Satellite altimetry is not suitable for these areas. These larger differences can also be seen around islands.

The area in the square is however unusual. It is far enough from the coast not to be influenced by the problems mentioned above. The data in the region of 30-35°N lat and about 280 - 285° longitude differs by almost 0.3 m. The reason for this could be that the data was acquired at different times. Consequently this difference could have been introduced by the Gulf Stream, since it meanders continuously. Ocean currents can introduce a difference from the marine geoid up to 1 m over several months.

The CLS01 MSS is used here as the reference surface. Therefore both the KMSS04 and the WHU2000 are subtracted from this surface. The next plots show the region in the square, where the data sets differ most. Here the difference is up to about 0.3 m.

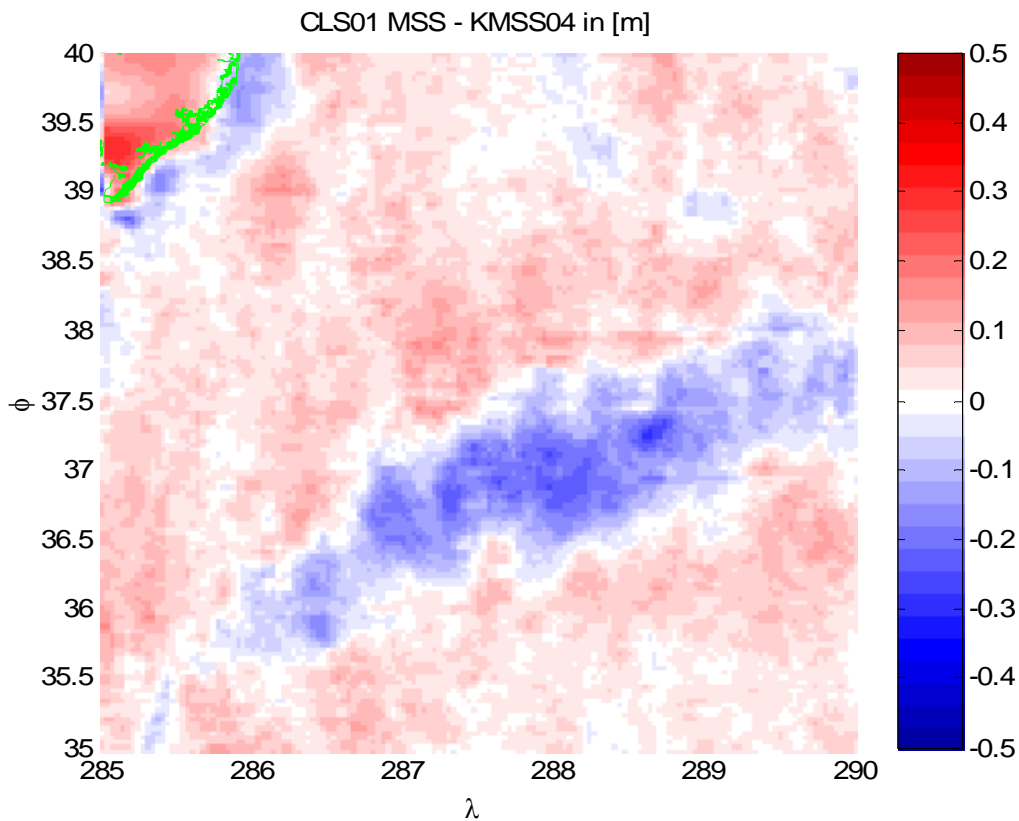


Figure 11 Difference between CLS01 MSS and KMSS04 (small area)

The following plot shows the difference between the WHU2000 data and the CLS01 MSS data.

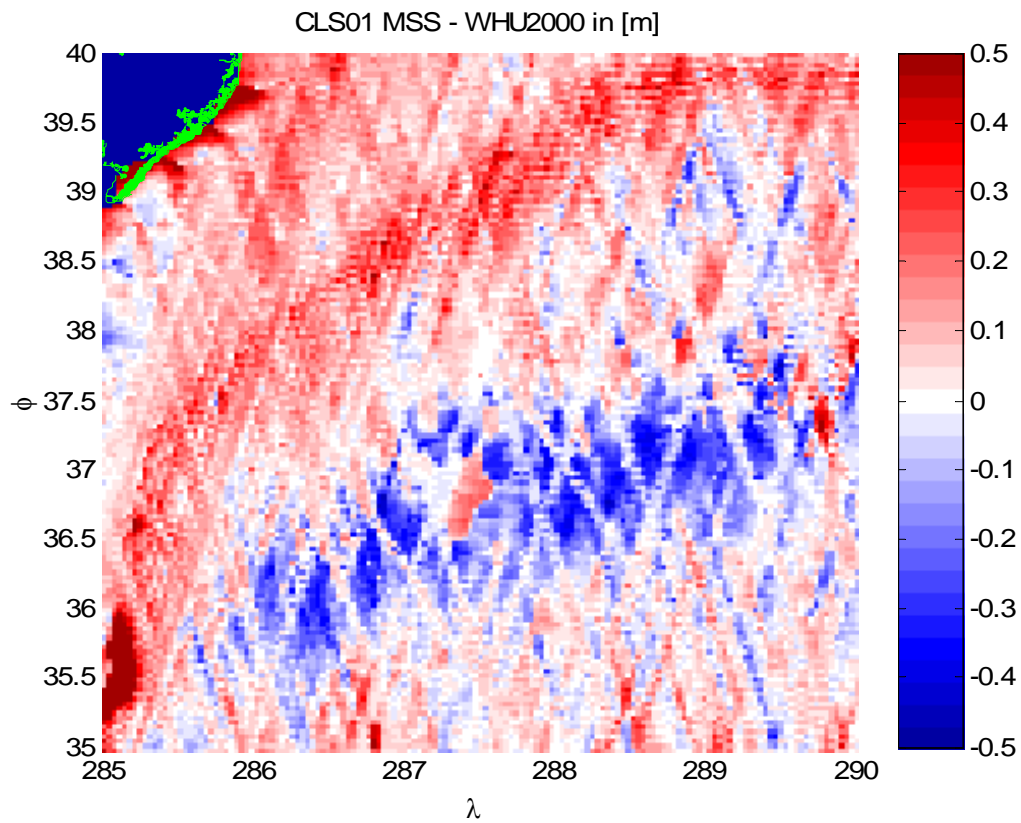


Figure 12 Difference between CLS01 MSS and WHU2000 (small area)

The difference between the CLS01 MSS and the WHU 2000 shows unusual features. Noticeable here too is a region where the sets differ more, by about -40cm (blue area), however the sets differ in other regions as well by about 40 cm (red areas).

There seem to be features which have the same patterns as satellite ground tracks. Along these tracks the differences seem to be lower. The reason for this could be that the data was processed differently or the grid points were interpolated with other methods. Also different initial data sets could have been used.

The second area to be analyzed is an area more in the middle of the North Atlantic between 35 - 45° N lat and 310 - 330° longitude. The differences between the sea surface data sets are less distinct here. The second area was chosen because less turbulence was expected here than in the first area. As can be seen on the plot (Fig. 13), this assumption proves to be correct. The only irregularity here turns up near part of the Azore islands (circle in blue area). As said before the data acquired near islands is not reliable.

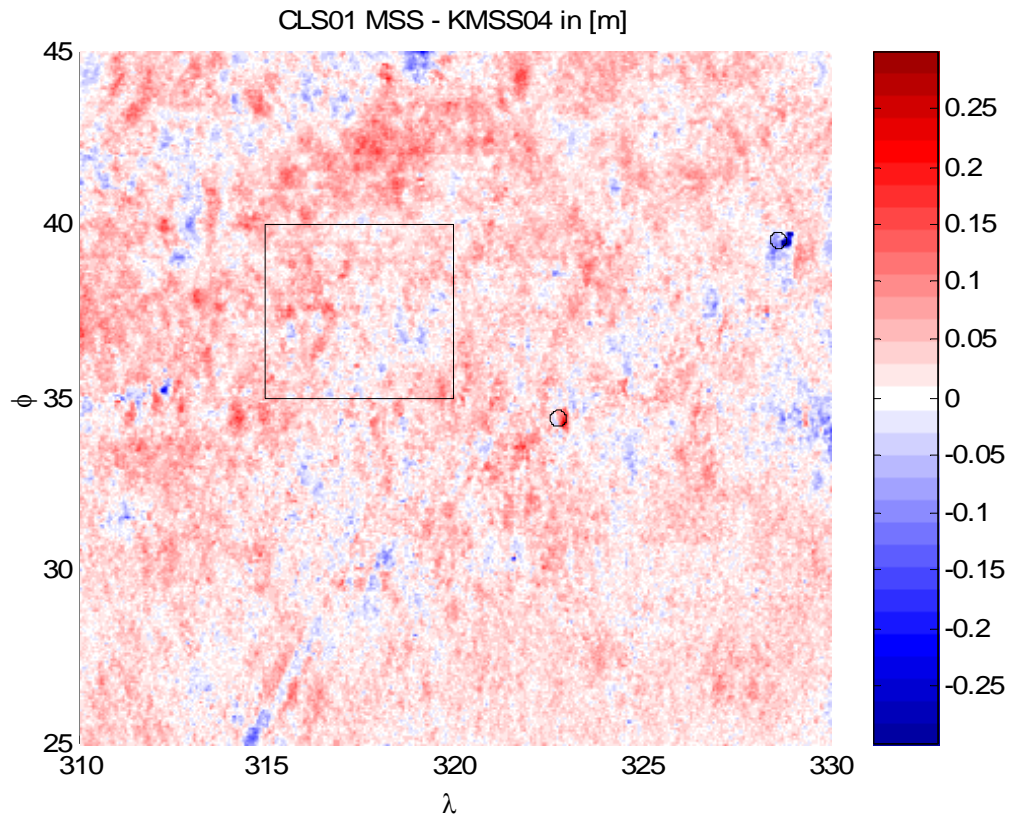


Figure 13 Difference between CLS01 MSS and KMSS04

The standard deviation of the difference plotted above is 0.03 m. The small island areas seen at 40° latitude and 329° longitude (part of the Azores) were not eliminated before the calculation of the standard deviation. Since they only cover a small area, their influence is not noticeable if the standard deviation for a large area is calculated. In this area the data sets differ only by about ± 0.25 m, without the Azores.

Figure 14 shows the area in the square shown above, though here the WHU2000 data is subtracted from the CLS01 MSS data.

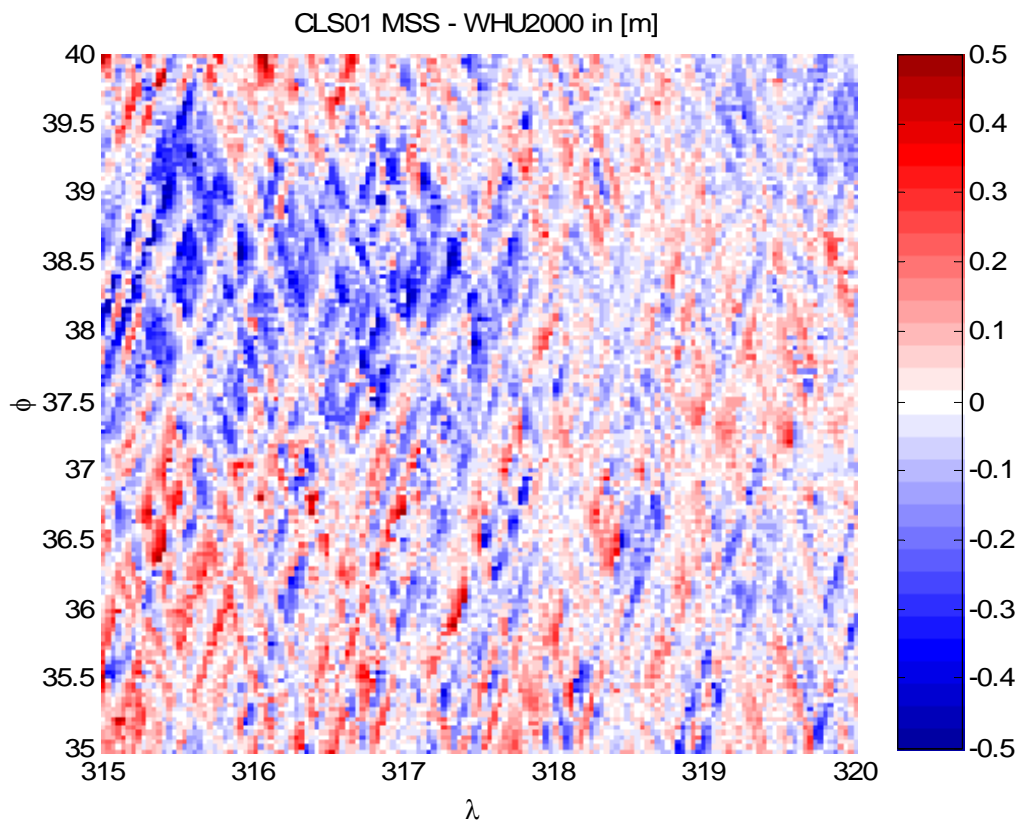


Figure 14 Difference between CLS01 MSS and WHU2000 (small area)

The standard deviation for this area is 0.12 m. The data sets differ up to ± 60 cm. Though along certain tracks as seen in the first area the errors are less.

4.4.1 Summary of statistics

Standard deviations (in [m]):

	CLS01 MSS - KMSS04	CLS01 MSS - WHU2000
area: 285 - 290° lon 35 - 40° lat	0.07	0.15
area: 315 - 320° lon 35 - 40° lat	0.03	0.12
area: 280 - 300° lon 25 - 45° lat	0.06	-
area: 310 - 330° lon 25 - 45° lat	0.03	-

Table 3

Minimum and maximum differences in [m]

	CLS01 MSS - KMSS04		CLS01 MSS - WHU2000	
	min	max	min	max
area: 285 - 290° lon 35 - 40° lat	- 0.28*	0.28*	- 0.46*	1.47*
area: 315 - 320° lon 35 - 40° lat	- 0.10	0.19	- 0.60	0.67
area: 280 - 300° lon 25 - 45° lat	- 2.50*	1.03*	-	-
area: 310 - 330° lon 25 - 45° lat	-0.65*	0.27	-	-

Table 4

The (*) are not reliable.

The land area was not taken into account in the calculation of the maximum and the minimum. However the area near the land is also not reliable. This is why some maxima or minima in the difference are over 1 m. These differences could have occurred near the coast or islands.

The only maxima and minima which are reliable, are those which were acquired in the second area, in the North Atlantic. The greatest negative difference (-0.65 m) is the Azores (See Fig. 13). The white areas show where the two data sets coincide.

4.5 Calculation of the root mean square

The root mean square is given by

$$rms = \sqrt{\frac{\sum_i^n d_i^2}{n}} \quad (5)$$

where d are the differences and n is the number of grid points

The root mean square is now calculated for different sizes of the rms -square and different step sizes, by which this square is shifted.

The size of the square for the root mean square calculation has to be carefully chosen, especially if the data contains land area. The land area of course influences the calculations. The method uses a moving rms . For this method a square is placed in the upper left-hand corner. The rms of this square is calculated. Then the square is moved to the right by a certain number of degrees and the procedure is repeated.

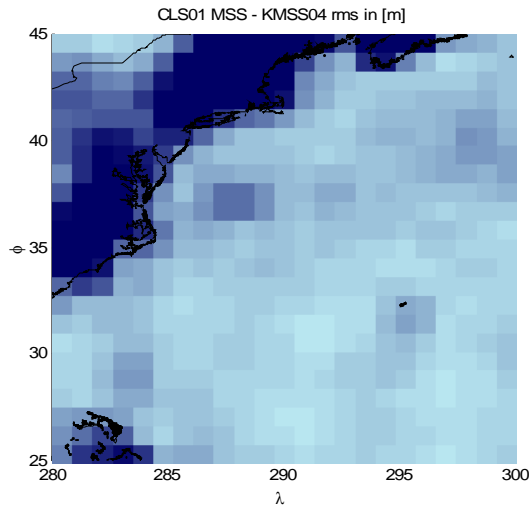


Figure 15

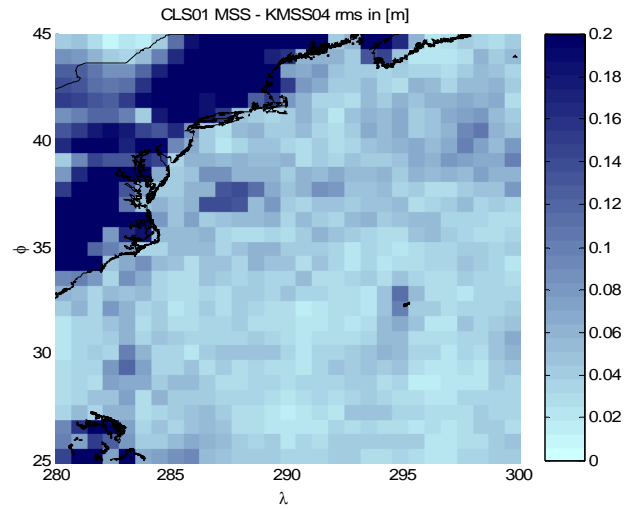


Figure 16

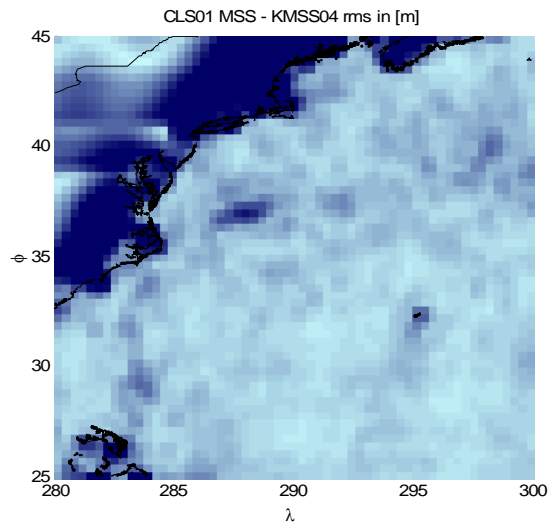


Figure 17

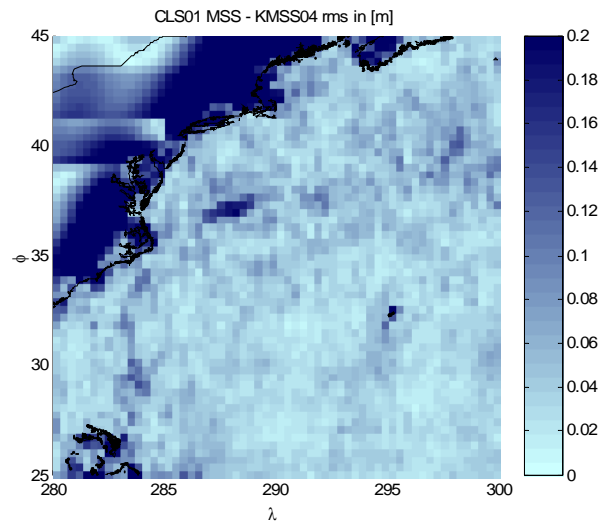


Figure 18

In Figure 15 the size of the square was chosen to be quite large. The influence of a small island like Bermuda is seen clearly. Therefore a reduction of the *rms* square size would be advisable, and is applied in the next plot (Fig. 16). But even here the influence of the island can be seen clearly.

	size of <i>rms</i> area	step size
Fig. 15	51/30° x 51/30°	25/30°, overlapping
Fig. 16	21/30° x 21/30°	20/30°, not overlapping
Fig. 17	21/30° x 21/30°	10/30°, overlapping
Fig. 18	11/30° x 11/30°	10/30°, not overlapping

Table 5

The root mean square clearly shows the difference which could be seen in the first plot of differences (Fig. 10). This irregularity (Longitude: 285 – 290°, Latitude: 35 – 40°) could be due to the more turbulent area caused by the presence of the Gulf Stream. If the data used for the calculation of the mean sea surface is acquired over different periods of time, and the Gulf Stream shifts during this time, this can explain the greater differences found in the area. The *rms* also shows that regions around small islands should be treated carefully, since here the differences are larger.

Minimum and maximum *rms* in [m]

CLS01 MSS - KMSS04

	min	max	Square	step size
Fig. 15	0.03	0.11	51/30° x 51/30°	25/30°
Fig. 16	0.02	0.13	21/30° x 21/30°	20/30°
Fig. 17	0.02	0.18	21/30° x 21/30°	10/30°
Fig. 18	0.02	0.18	11/30° x 11/30°	10/30°

Table 6

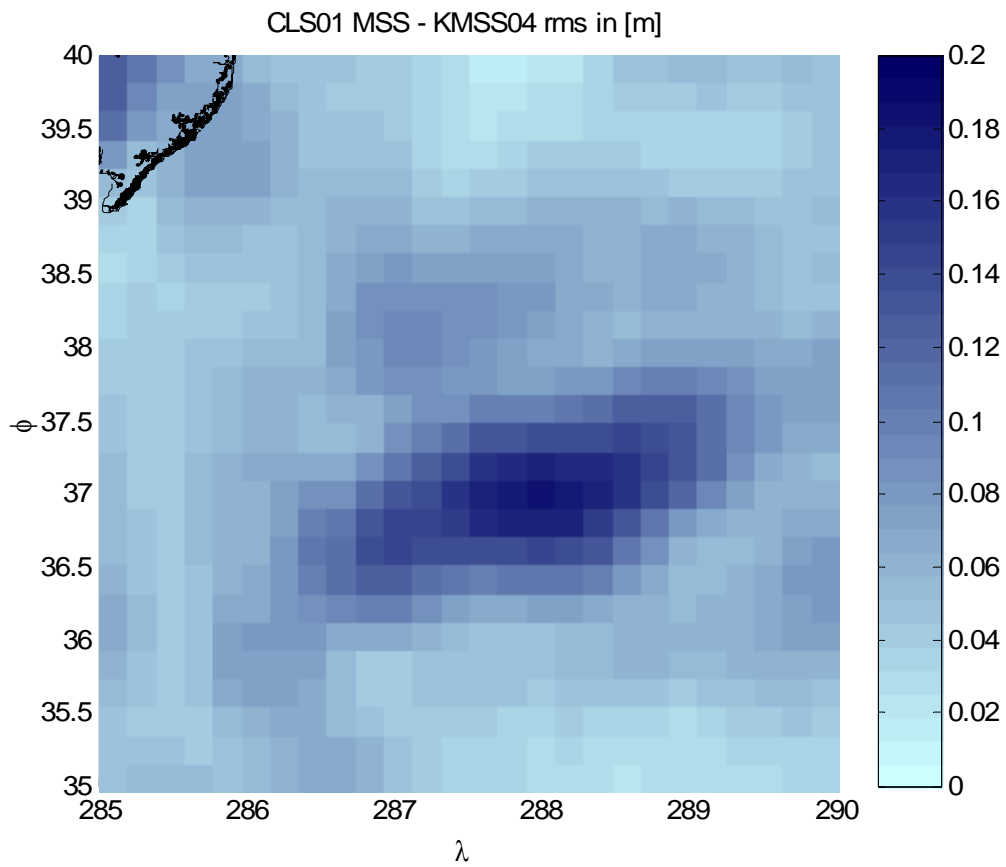


Figure 19 Size of *rms* area : $21/30^\circ \times 21/30^\circ$ step size: $5/30^\circ$, overlapping

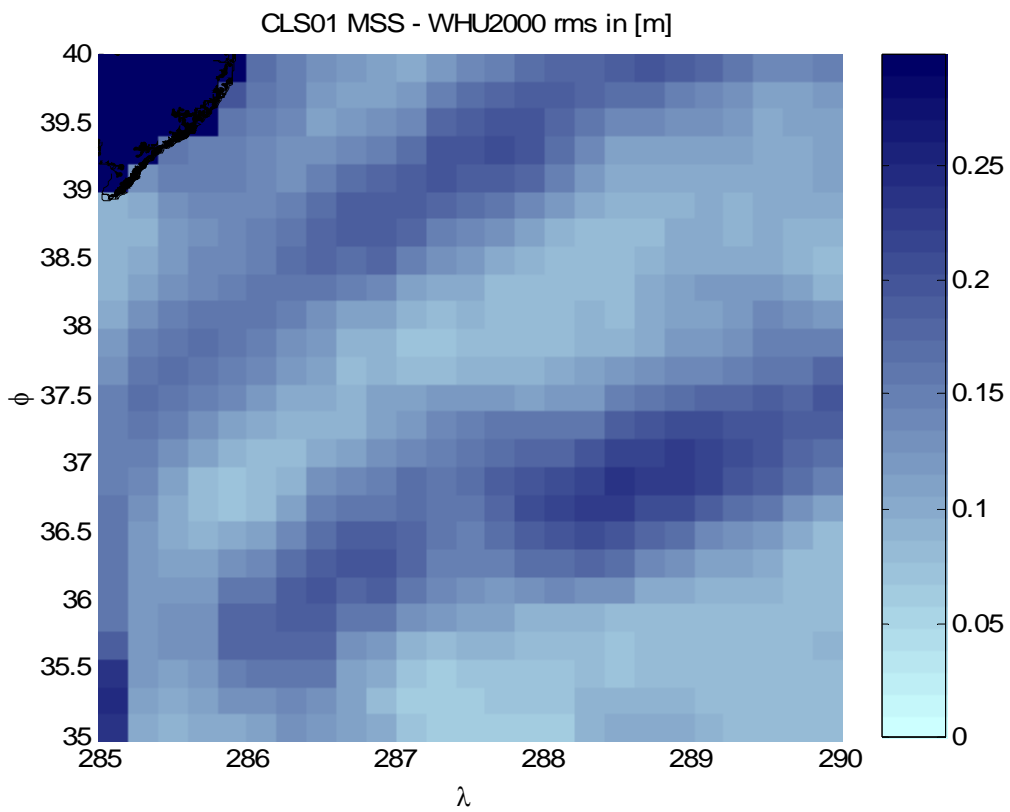


Figure 20 Size of *rms* area : $21/30^\circ \times 21/30^\circ$ step size: $5/30^\circ$, overlapping

The two plots above show the root mean square of the smaller area. The first plot shows the root mean square difference between the CLS01 MSS and the KMSS04. The area with the greater root mean square difference can be seen clearly.

The second plot shows the root mean square difference between the CLS01 MSS and the WHU2000. Here the area with greater root mean square difference is on a slightly different position. This could also be due to the meandering of the Gulf Stream. The darker area above might be influenced by the presence of land in the proximity.

Minimum and maximum *rms* in [m]

	min	max	
Fig. 19	0.02	0.18	CLS01 MSS - KMSS04
Fig. 20	0.06	0.23	CLS01 MSS - WHU2000

Table 7

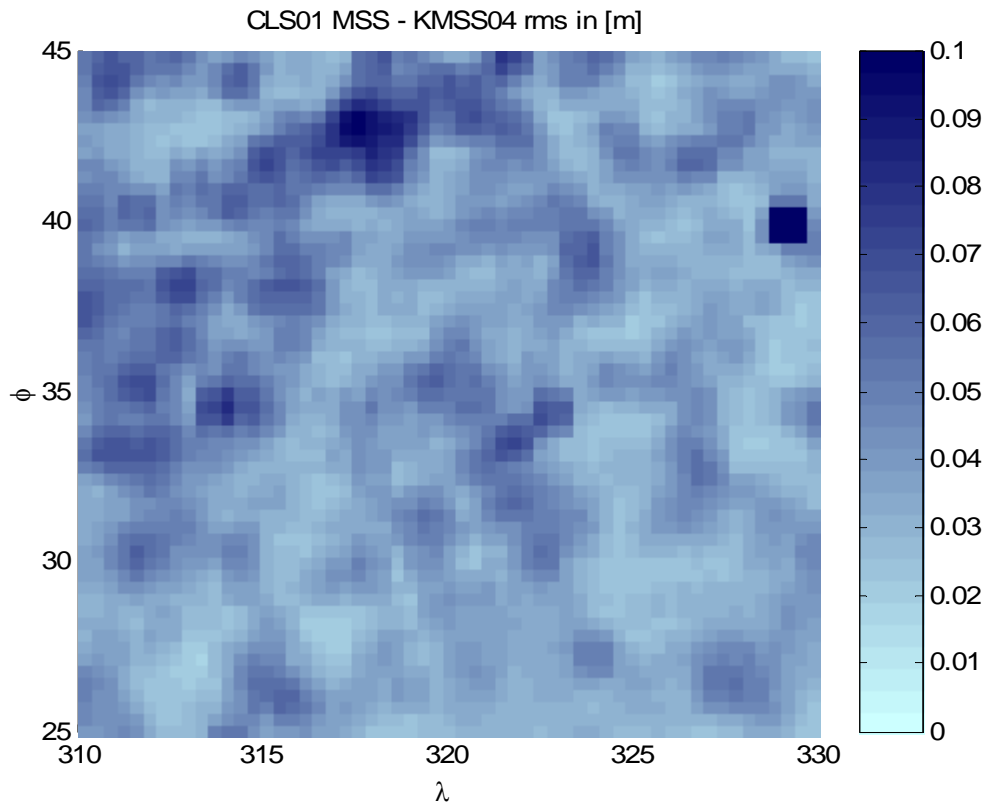


Figure 21 Size of *rms* area : $31/30^\circ \times 31/30^\circ$ step size: $10/30^\circ$, overlapping

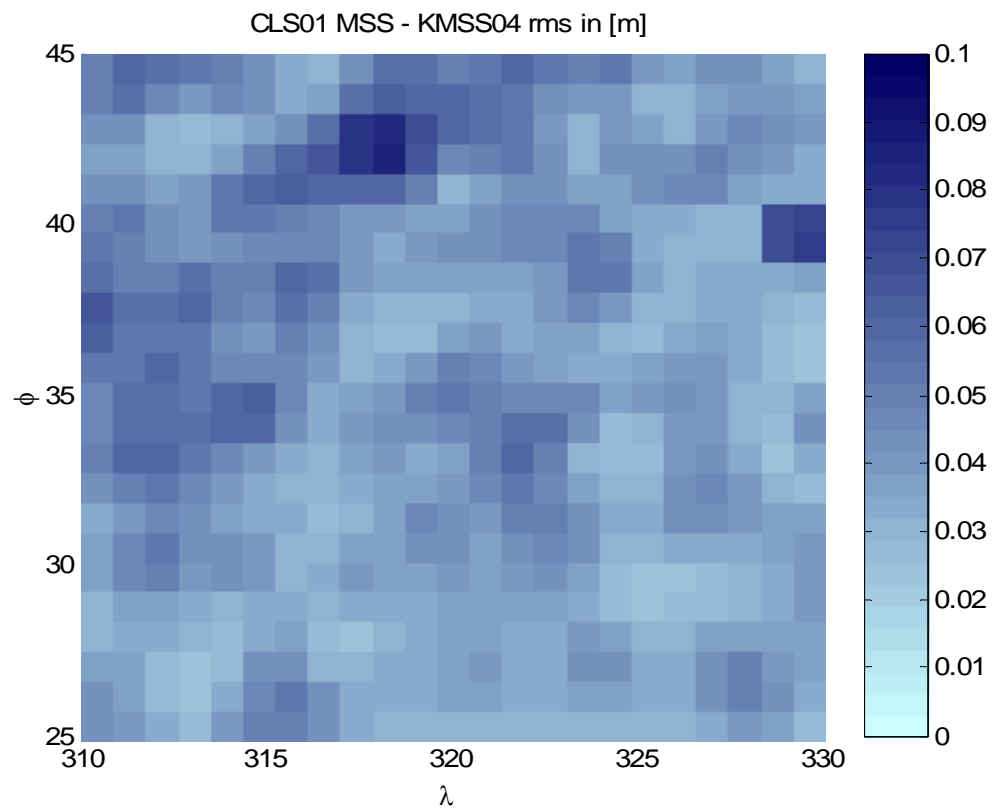


Figure 22 Size of *rms* area : $51/30^\circ \times 51/30^\circ$ step size: $25/30^\circ$, overlapping

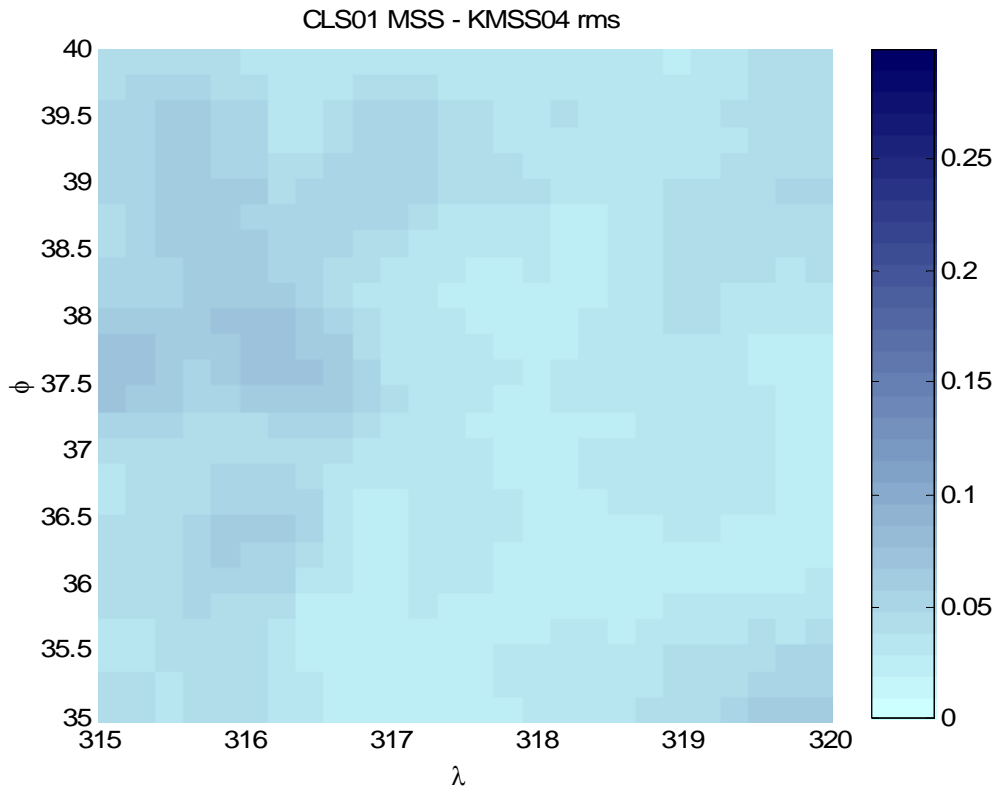


Figure 23 Size of *rms* area : $21/30^\circ \times 21/30^\circ$ step size: $5/30^\circ$, overlapping

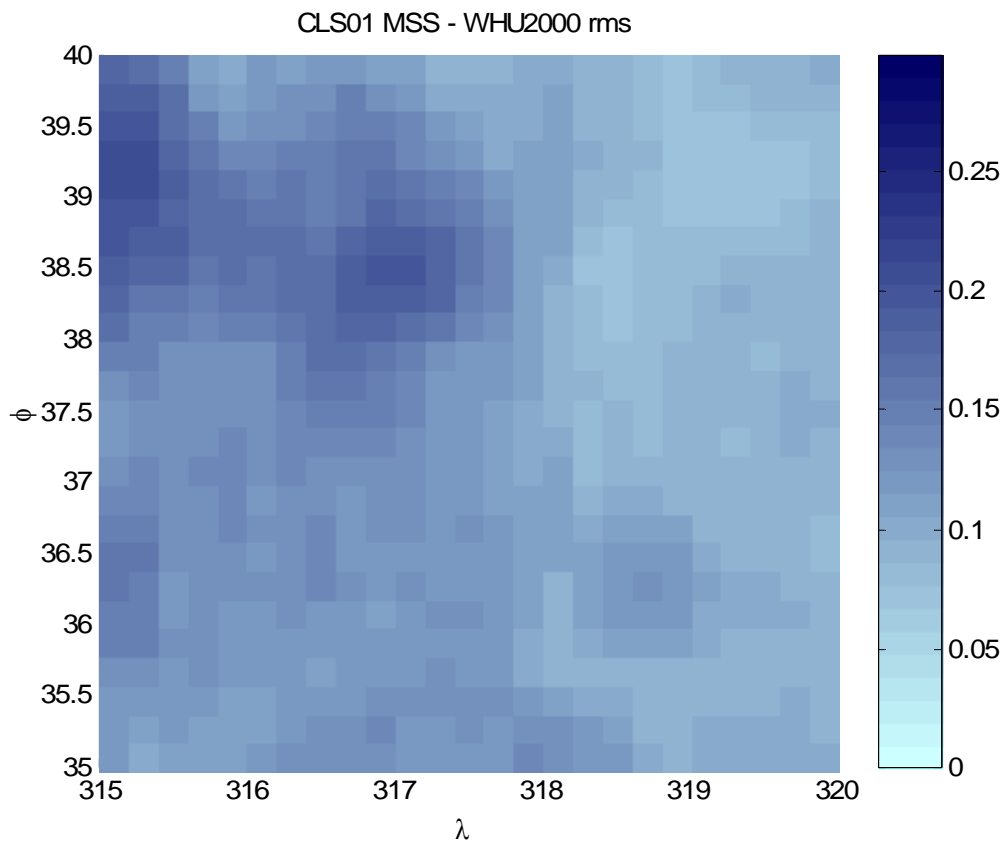


Figure 24 Size of *rms* area : $21/30^\circ \times 21/30^\circ$ step size: $5/30^\circ$, overlapping

In the second area the calculation of the *rms* shows no unusual features, expect for the island area, which influences the surrounding squares, especially if large *rms*-squares are chosen. In the middle of the ocean the data sets differ less from one another. There are no large ocean currents or large scale ocean circulation present.

What can be shown, is that data sets acquired in areas with more ocean movement differ more than data sets acquired in less turbulent areas. The reason for this could be that large scale ocean currents meander and therefore change over several months. If the data was acquired over different periods of time, then greater differences occur.

As can be seen here the root mean square of the difference between CLS01 MSS and KMSS04 are smaller than the *rms* of the difference between CLS01 MSS and WHU2000 (Table 9). The same color scale boundaries were used for both plots.

Minimum and maximum <i>rms</i> in [m]		CLS01 MSS - KMSS04		
	min	max	square	step size
Fig. 21	0.02	0.08	51/30° x 51/30°	25/30°
Fig. 22	0.02	0.06	31/30° x 31/30°	10/30°

Table 8

Minimum and maximum <i>rms</i> in [m]			
	min	max	
Fig. 23	0.02	0.07	CLS01 MSS - KMSS04
Fig. 24	0.07	0.21	CLS01 MSS - WHU2000

Table 9

5. Outlook: Calculation of the Mean Dynamic Topography

This chapter contains an outlook to the calculation of the MDT. For the calculation of the MDT two different geoids were used. The first is the EIGEN-GL04S1 and the second, newer one is the EIGEN-GRACE02S. Both geoid models are developed until degree and order 150.

Both these geoids are based on satellite-only gravity measurements. This is important if they are to be used for oceanographic purposes. The satellite-only geoid model is achieved through inversion of satellite orbit perturbations by downward continuation of the model (Tapley and Kim, 2001).

The geoid N is given by Cazenave (2001)

$$N(R, \varphi, \lambda) = R \left[\sum_{l=2}^{\infty} \sum_{m=0}^l ((c_{lm} \cos m\lambda + s_{lm} \sin m\lambda) P_{lm}(\sin \varphi)) \right] \quad (6)$$

R mean radius of the earth
 φ, λ latitude, longitude
 c_{lm}, s_{lm} Stokes' coefficients
 P_{lm} normalized Legendre functions

(Cazenave, 2001). Sets of c_{lm}, s_{lm} are given in EIGEN-GL04S1 and EIGEN-GRACE02S. The MDT is calculated by subtracting the geoid from the mean sea surface. The geoids are presented in harmonic coefficients. To be able to subtract the two surfaces from each other, there are two possibilities. Either the mean sea surface is developed into harmonic coefficients, or the harmonic coefficients are transformed into heights above a reference ellipsoid. In this case the second method was applied. The resolution of the calculated geoid surface was $1/10^\circ$. Because a resolution of $1/30^\circ$ is needed an interpolation program had to be written. This program interpolates the geoid grid linearly. Then the geoid grid was subtracted from the mean sea surface grid.

A problem with this method is the different spectral contents of the geoid and the mean sea surface. The geoid is exact over long wavelengths, e.g. the coefficients which describe the spherical shape of the earth's gravity field. The short wavelength geoid shows the features resulting from topography and crustal density variations (Cazenave, 1995). The problem is, that the higher the degree of the coefficients the higher the cumulative error on the geoid (Cazenave, 2001). The mean sea surface which is derived from altimetry currently has an error of 1-2 cm (Seeber, 2003). The error at degree 10 for a geoid model is 10 cm (wave length of about 4000 km) (Cazenave, 2001). The errors are nearly as large as the difference between mean sea surface and geoid, which means the error is nearly as large as the MDT which is to be calculated. This means this method of calculating the MDT is not ideal. Other methods like the development into spherical harmonic coefficients of the sea surface should be examined to see whether they would lead to better results.

The results of method, where the harmonic coefficients of the geoid are transformed into heights above a reference ellipsoid are shown in the four plots (Fig. 25 - Fig. 28) seen below. The first two plots show the first area, the second two show the second area.

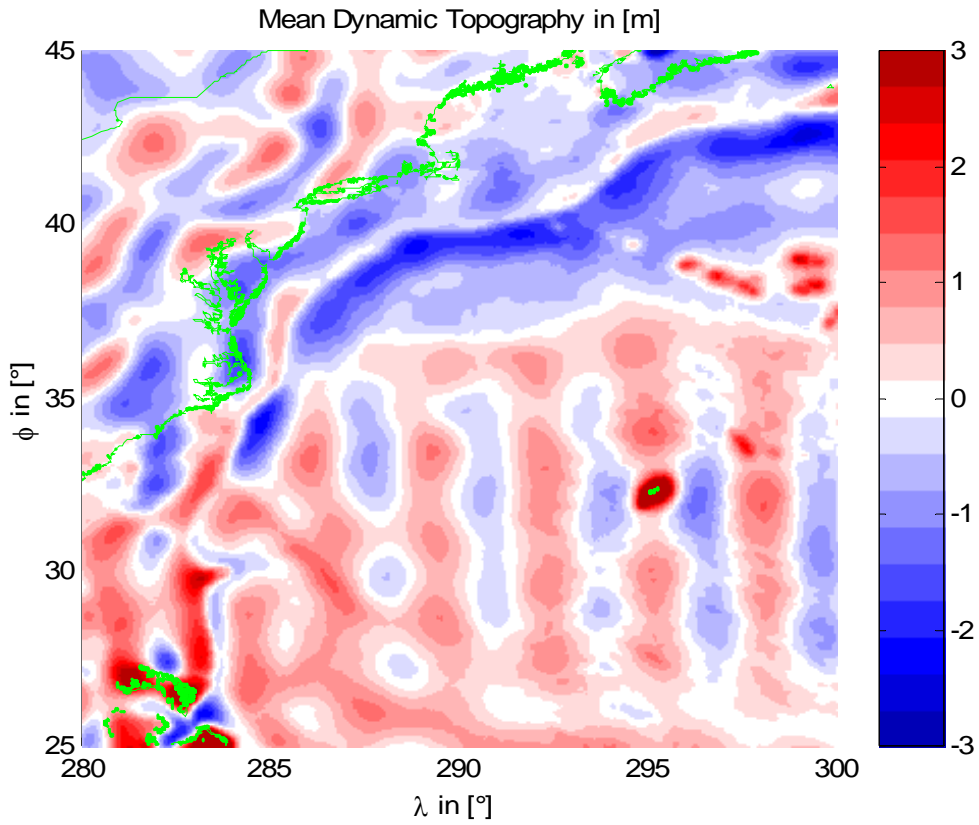


Figure 25 MDT based on EIGEN-GL04S1 and CLS01 MSS

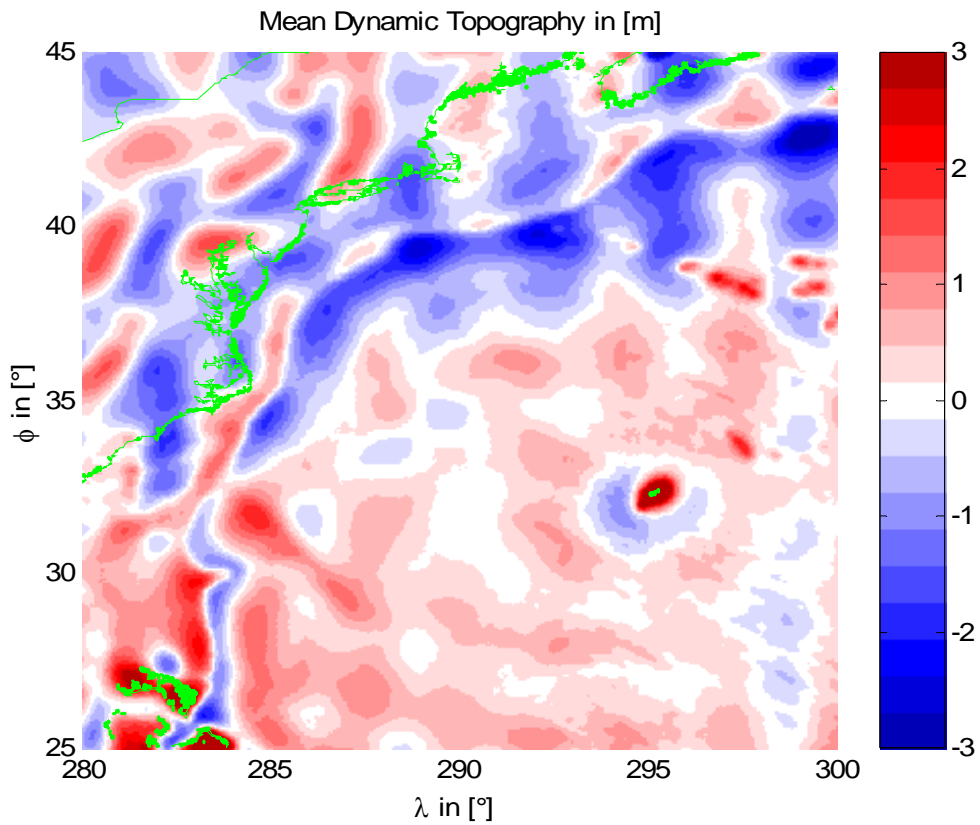


Figure 26 MDT based on EIGEN-GRACE02S and CLS01 MSS

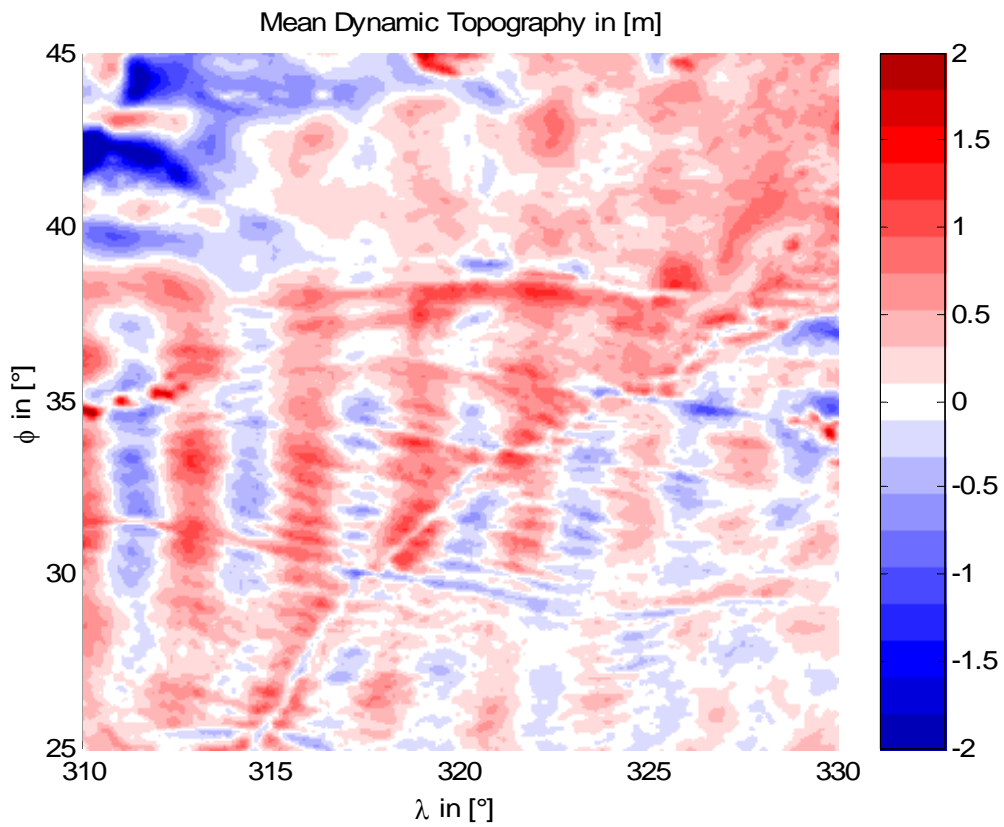


Figure 27 MDT based on EIGEN-GL04S1

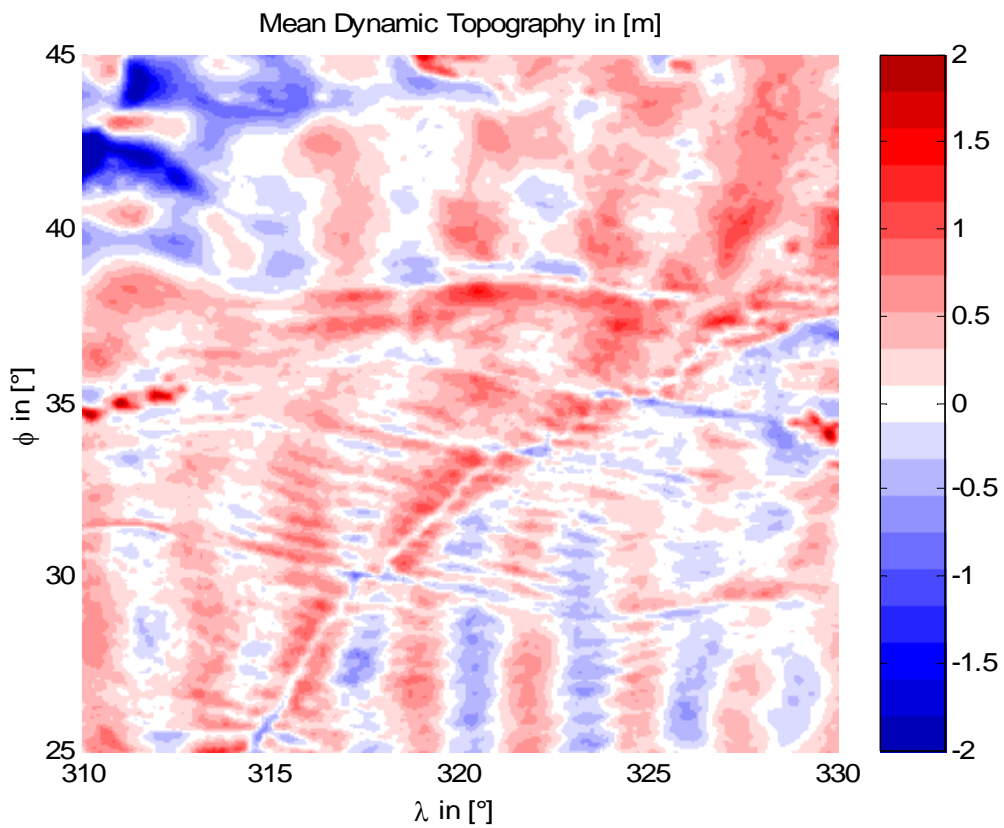


Figure 28 MDT based on EIGEN-GRACE02S

On the plots shown above different features can be clearly seen. The GRACE – error patterns are clearly visible. These artefacts have nothing to do with the mean dynamic topography, but are due to the measurements for the calculation of the geoid. In the area where the greatest differences occurred at the comparison of two sets of mean sea surface data, there is also a greater difference to be seen in the mean dynamic topography. This could be due to the shelf, but also to the presence of the gulf stream. Here once more Bermuda can be seen clearly. This is probably because of the inexact or lacking altimeter measurements.

In the second area the GRACE – error patterns can also be noticed clearly. There are some rift areas noticeable, especially the Mid - Atlantic Ridge. No oceanographic features are clearly visible. This is probably because of the long wave-length geoid. On this geoid the features due to the ocean bottom topography are flattened. They therefore turn up in the difference between the altimetric data, which measured these features and the geoid, which does not show them. If an ideal geoid was subtracted from the mean sea surface only oceanographic features would remain. The mean sea surface data is exact enough for oceanographic purposes. These geoids are exact over larger areas, but not over the areas which would be useful for oceanographic means.

6. Conclusion

In this paper the mean sea surface and the mean dynamic topography were discussed, along with satellite altimetry.

The topic was to compare different sets of data of the mean sea surface. In the second chapter an overview of the different surfaces relating to the subject was given: the mean sea surface, the geoid and the mean dynamic topography.

In the third chapter the use of altimetry to acquire data for the calculation of the mean sea surface was described.

In the fourth chapter three different sets of data for the mean sea surface were compared: the CLS01 MSS, the KMSS04, the WHU2000. The CLS01 MSS was used as a reference surface, therefore the other two surfaces were subtracted from it. Two areas were extracted from the data. The first area is near the coast of North America. The second area shows the middle of the North Atlantic. The calculated differences were plotted along with the coastlines. In the first area a larger difference was to be seen. A reason for this could be the Gulf Stream. Since the Gulf Stream meanders, data taken from different times of observation could show greater differences.

In the second part of the third chapter the root mean square was calculated. Here this larger difference could also be seen. The plot of the root mean square calculation between CLS01 MSS and WHU 2000 shows the greater differences in a slightly different area. This again could be due to the meandering Gulf Stream. The differences and also their root mean square showed that areas near the coast should be treated carefully. These coastal areas are not very exact. The reasons are that the tidal range here is very wide and irregular, and therefore difficult to model. Also there is no altimeter data available over land. Therefore the sea surface has to be continued over land through a geoid for example. In the second area there were no unusual features to be seen. The root mean square showed that the difference between the data sets of the sea surface in the middle of the ocean is not very large. Near islands the data sets can be less exact and show greater differences.

In the fifth chapter the mean dynamic topography was calculated. Here there were some interesting features to be seen: the GRACE error pattern, but also features due to the ocean bottom topography. This means that the geoid used has a wave length which is too long. The mean sea surface shows variations due to the topography, the geoid does not. Therefore these features can be seen in the difference. However, no oceanographic features can be seen in the middle of the North Atlantic. There is no large scale circulation here and the geoid is not exact over short wave lengths. In the first area the Gulf Stream possibly could be seen. Large ocean currents can introduce a difference from the geoid of up to 1 meter. Altimetry achieves results which can resolve shorter heights. Altimetry is sufficiently exact for large scale ocean circulation studies. The problem can be traced to the geoid. It is exact over long wave lengths, but not for the shorter wave lengths needed. This is why only large ocean currents can be noticed. A short wave length geoid as should be obtained from GOCE will be better suitable for studies of the ocean circulation.

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Current Version: KMS04 - 2004
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