Geophysical Research Letters
NOV 2 2006; 33 (21) NIL_10-NIL_13
http://dx.doi.org/10.1029/2006GL027452

An edited version of this paper was published by AGU.

Comparison of in situ bottom pressure data with GRACE gravimetry in the Crozet-Kerguelen region

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Abstract:

Two time series of deep ocean bottom pressure records (BPRs) in between the Crozet Islands and Kerguelen are compared with GRACE (Gravity Recovery And Climate Experiment) equivalent water heights. An analysis of the correlation is performed for four time series: 1) monthly averages of the equivalent water height at the Crozet Islands, 2) the same near the Kerguelen Islands, 3) the mean of the two preceding series and 4) the difference between the two locations expressed in terms of geostrophic transport. We find that smoothed GRACE solutions are strongly correlated with the BPR data with correlation coefficients in the order of 0.7–0.8. Consequently GRACE measures real oceanic mass variations in this region.

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Introduction

The GRACE mission (Gravity Recovery And Climate Experiment), launched in 2002, continuously measures the Earth's gravity field. Temporal variations in gravity are caused by the redistribution of masses originating from sources such as the atmosphere, ocean, hydrology, ice-sheets or the solid earth (eg. *Tapley et al.* [2004]).

Compared to the gravity signal of continental hydrology the contribution of the ocean is in general much weaker (*Wahr et al.* [1998]). Detecting these variations from GRACE therefore represents a major challenge, but would be of great value for climate studies and validation of ocean models.

The validation of the GRACE data using independent in situ bottom pressure measurements, has not been accomplished so far. *Kanzow et al.* [2005] already compared in situ bottom pressure in the tropical northwest Atlantic ocean. However, no significant correlation between GRACE and the BPR data was found in that analysis. In this article we compare GRACE fields, processed by CNES/GRGS (Toulouse, France), with two time series of in situ bottom pressure in the southern section of the Indian Ocean.

Several large-scale oceanic phenomena justify the deployment of the BPRs. This region is characterized by the presence of a merged set of strong fronts north of the Crozet Islands and Kerguelen accounting for around 75% of the total Antarctic Circumpolar Current (ACC) transport (*Park et al.* [1993]). Furthermore, *Park et al.* [1993] suggested that through the passage between Crozet and Kerguelen a deep western boundary current is flowing northwards into the Crozet Basin.

In this study it is demonstrated that GRACE has the ability to measure temporal variations of ocean bottom pressure in the Crozet-Kerguelen region.

Data sets

Bottom Pressure Recorders

The analysis is performed using data of two BPRs which were deployed at approximately 4000 m depth at (47.12 ° S, 54.90 ° E) and (48.83 ° S, 61.28 ° E), respectively. These positions, separated by 510 km, are on either side of the saddle point in between the Crozet plateau and the Kerguelen plateau as can be seen from the bathymetric contours in figure 1. The period covered February 2004 until February 2005 with a temporal resolution of 15 minutes.

The BPRs effectively measure the mass of the overlying water column plus that of the atmosphere. Hence, mass fluctuations in the ocean and atmosphere above the BPR induce pressure variations. Through the Coriolis force, differences of pressure anomalies measured at different locations are linked to geostrophic current velocity changes. Thus, large-scale mass transports can be measured by a set of at least two BPRs. When assuming geostrophy, the variation of the mean vertically averaged transport between the two stations can be derived from *Kanzow et al.* [2005] [eq. 2].

$$\delta\overline{V} \quad = \quad \left|\frac{1}{\rho_w f}H(\delta p_2 - \delta p_1)(\vec{e}_z \times \vec{e}_{1,2})\right| \qquad (1)$$

In which $\delta\overline{V}$ is the change of the mean vertically averaged geostrophic transport. The change in pressure at station 1 and 2 is denoted by δp_1 and δp_2, ρ_w is the density of the seawater. The unit vector $\vec{e}_z$ points in the zenith direction and the vector $\vec{e}_{1,2}$ connects the two stations. The Coriolis parameter f is taken constant using a latitude of 48°, which is the average latitude of the BPR stations. The vertical scale height H is set to 4000 m, the average depth of the two BPRs. This choice is justified since the flow is predominantly barotropic in regions where the ACC is present (*Hughes et al.* [2003]).

For this study, the BPR time series were de-tided by applying a harmonic fit of 73 tidal constituents on the data. Then, the BPR data is smoothed with a 30 day running mean centered on a 10 days time axis, according to the same weighting scheme that is applied to the GRGS-GRACE solution.

GRACE data

The GRACE data used is processed by CNES/GRGS in France (*Biancale et al.* [2005]). They consist of 30-day solutions of the mean gravity field expressed in Stokes coefficients up to the degree 50 every 10 days. The relative weights applied for the three consecutive ten-day intervals of each monthly solution are 0.5/1/0.5. LAGEOS satellite laser ranging data is used to increase the accuracy of the lower degree coefficients, mainly for degree 2. For the higher degree coefficients, effectively from degree 30 and onward, the solution is constrained toward the static gravity field. The effective spatial resolution is therefore approximately 666 km (*Llubes et al.* [2006]). The data has been corrected for ocean tides using the FES2004 model (LEGOS Toulouse) as well as earth tides (according to IERS Convention 2003). Aliasing of high frequency atmospheric and ocean variability has been taken into account by using ECMWF 3-D atmospheric pressure fields and a barotropic ocean model, MOG2D (see *Carrère and Lyard* [2003]). However, in our case, monthly averages of the atmospheric fields and the barotropic ocean are added back to the solution as explained below, since those effects are measured by the in situ bottom pressure recorder.

The procedure to obtain the bottom pressure at the BPR locations is as follows. We first subtract a static gravity model, EIGEN-GL04S (GRGS/GFZ) from the monthly GRACE solution. Degree 1 coefficients derived from a geo-center motion model from *Crétaux et al.* [2002] are added back as recommended in *Chambers et al.* [2004]. The gravity field is then smoothed to remove noise of the higher degree coefficients and converted to bottom pressure using a similar equation as from *Wahr et al.* [1998]:

$$\Delta p_{bott}(\phi,\lambda) = \frac{ag\rho_e}{3}\sum_{l,m}\frac{2l+1}{1+k_l}W_l P_{lm}(\cos\phi) \left[\Delta C_{lm}\cos(m\lambda) + \Delta S_{lm}\sin(m\lambda)\right] \qquad (2)$$

Here $\Delta p_{bott}(\phi,\lambda)$ is the change in bottom pressure at the geographical location with colatitude ϕ and longitude λ. ΔC_{lm} and ΔS_{lm} are the fully normalized stokes coefficients of the gravity field relative to the static gravity field (EIGEN-GL04S). The associated normalized Legendre polynomial of degree l and order m is denoted by $P_{lm}(\cos\phi)$. The load Love numbers k_l are calculated following *Han and Wahr* [1995]. Symbols a, ρ_e and g are the Earth's mean radius, mean density and mean gravity respectively. W_l denotes the Gaussian smoothing weight factor.

Finally, the smoothed averaged atmospheric and barotropic ocean model is added back. The GRACE derived bottom pressures for the two BPR locations are used to calculate $\delta\overline{V}$ as in eq. 1. We used various smoothing radii to investigate the effect it has on the solution.

Bottom pressure comparison with GRACE

Here, a correlation analysis on the derived GRACE solution and the equivalent BPR set is performed. We compare the four time series for 1) bottom pressure near Crozet, 2) bottom pressure near Kerguelen, 3) the mean pressure of both locations representative for the midpoint and 4) the change of vertically averaged mean geostrophic transport through the BPR section.

Figure 2 shows the four above mentioned time series of the averaged BPRs, GRACE and the monthly averaged ocean and atmosphere models. The first three series show strong similarities in signatures and magnitudes. Standard deviations of the BPR series are comparable to those of the GRACE series. Furthermore, the approximate signal to noise ratio, the ratio of the variance of the GRACE series and the variance of the difference between the BPR and GRACE series, are 1.5, 1.2, 1.8 respectively.

Figure 2 (d) displays a much weaker agreement between the series of $\delta\overline{V}$. The BPRs measure a significant variation in transport through the section in the order of 20 Sv whereas GRACE and the models show only little variation. Consequently, the signal to noise ratio is 0.37. The cause for this is the smoothing process applied to the GRACE solutions. As the separation between the stations is small compared to the smoothing radii used, the GRACE solutions at the two locations are correlated. When considering differences, the associated signals in common would be subtracted decreasing the true variation of the signal. Although the performance of GRACE is less than the first three series it appears to perform better than the atmosphere/ocean model alone.

Figure 2 (a-c) shows a large discrepancy in the first few months between GRACE and the BPR series. The ocean and atmosphere model also display this feature. This discrepancy could partly be caused by a possible initial drift in the BPR measurements not uncommon for deep ocean pressure recorders (*Vassie et al.* [1994]) and indicated by relative drifts of the two pressure channels in the BPRs, which occur in the first month. Additionally, it could also reflect some physical phenomena not properly reproduced by the barotropic ocean model and the atmosphere model. This will result in aliasing of unmonitored high frequency ocean or atmosphere phenomena in the GRACE solution.

Table 1 lists the correlations of GRACE with the BPRs. Already for smoothing radii around 800 km strong correlations exist in the order of 0.8. The high correlation levels support the discussion above. In particular, the correlation is greatly increased when removing the first two months of the time series. This effect is strongest for the Crozet position .

The confidence intervals at the Kerguelen position are slightly tighter and high correlations are reached for smaller radii than Crozet. This leads to the conclusion that GRACE represents the true bottom pressure somewhat better at the Kerguelen BPR. Possibly, due to the smoothing process, the GRACE solution near Crozet is more sensitive to contamination by signals from the nearby fronts.

For all four time series the monthly averages of the MOG2D and the ECMWF data show a weaker correlation with the BPR series. Relative to GRACE and the BPRs the models display a considerable drift. The good correlations found earlier can therefore be contributed to the GRACE data and are not due to the ocean and atmosphere models only.

The good agreement between the series suggests that the BPRs measure predominantly a large-scale signal. A regional plot in Fig. 1 shows such a large-scale signal overlying the Kerguelen plateau. This signal could correspond to the observation of *Meredith and Hughes* [2004] who suggested that wind curl anomalies around the plateau caused Ekman flow onto the Kerguelen plateau increasing the overlying mass. Furthermore, the averaged BPR time series at the two location correlate with each other in the order of $0.45 - 0.6$, which also illustrates that the in situ measurements are representative for large-scale signals such as seasonal signals or possibly variations in the ACC.

Conclusion

The in situ bottom pressure data and GRACE data are in good agreement, both at each individual location as well as in terms of the mean bottom pressure in the center. The geostrophic transport variations show a weaker resemblance, because of the spatial correlation introduced in the GRACE smoothing process. The high correlations demonstrate that GRACE is able to detect changes in bottom pressure accurately in space and time and that GRACE has the potential to measure actual ocean mass transport variations. However, the results of this studies are valid for the Crozet-Kerguelen region only, and do not necessarily apply elsewhere in the global ocean.

The good agreement seems to be due to coherent large-scale mass variations in this region. The steep slopes in the bathymetry and the general position of the Kerguelen plateau might contribute to the enhancement of the currents due to topographic steering of the circulation.

Furthermore, the positioning of the BPRs turned out to be beneficial for the study of large-scale bottom pressure variations. Whether the present results can be generalized for other BPR records remains to be seen but we suspect that the best results will be found for in situ observations which are sufficiently remote from strong sources of land hydrologic signals, such as those from the Amazon basin which affected the study by *Kanzow et al.* [2005]. Additionally, deployment at higher latitudes is advantageous due to the increased accuracy of the GRACE solution associated with the denser groundtrack pattern and due to the expected increase in barotropic contribution to the bottom pressure (*Kanzow et al.* [2005]).

Acknowledgments. We would like to thank IPEV (Institut Polaire Francais Paul Emile Victor) for allowing the deployment and recovery of the BPRs from the vessel Marion Dufresne. Furthermore we thank Peter Foden from Proudman Ocean Laboratory and Olivier Peden from IFREMER for assuring the technical success of the BPR operation. Finally, the valuable contribution of two anonymous reviewers was strongly appreciated.

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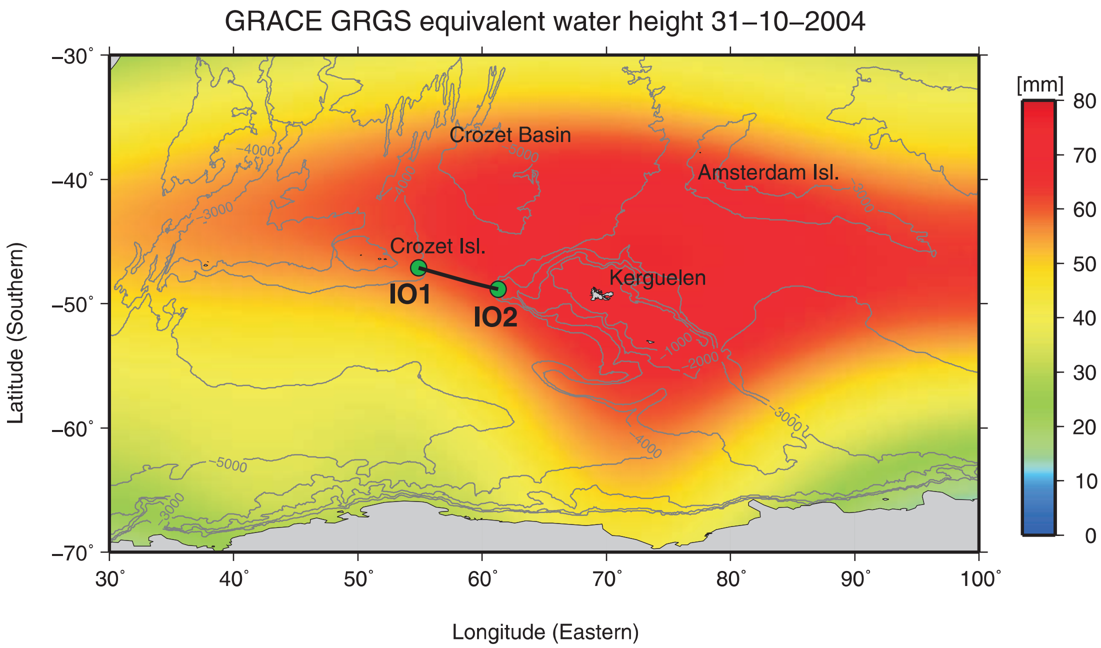
This preprint was prepared with AGU's LaTeX macros v5.01, with the extension package 'AGU^{++}' by P. W. Daly, version 1.6b from 1999/08/19.

	Crozet BPR (IO1)		Kerguelen BPR (IO2)		Mean at Midpoint		Vert. aver. transport	
$r_{\frac{1}{2}}$ [km]	ρ	95% low.	ρ	95% low.	ρ	95% low	ρ	95% low
0	0.29(0.41)	0.04(0.20)	0.56(0.57)	0.35(0.20)	0.68(0.78)	0.44(0.60)	-0.09(0.01)	-0.44(-0.48)
600	0.55(0.76)	0.23(0.56)	0.69(0.76)	0.47(0.60)	0.72(0.89)	0.43(0.77)	0.33(0.33)	-0.04(-0.11)
800	0.61(0.81)	0.29(0.62)	0.68(0.79)	0.43(0.64)	0.71(0.90)	0.40(0.77)	0.62(0.58)	0.31(0.20)
1000	0.64(0.83)	0.32(0.64)	0.68(0.80)	0.41(0.66)	0.71(0.91)	0.39(0.79)	0.76(0.72)	0.53(0.43)
1400	0.65(0.84)	0.33(0.66)	0.68(0.81)	0.42(0.67)	0.73(0.92)	0.41(0.83)	0.83(0.79)	0.67(0.58)

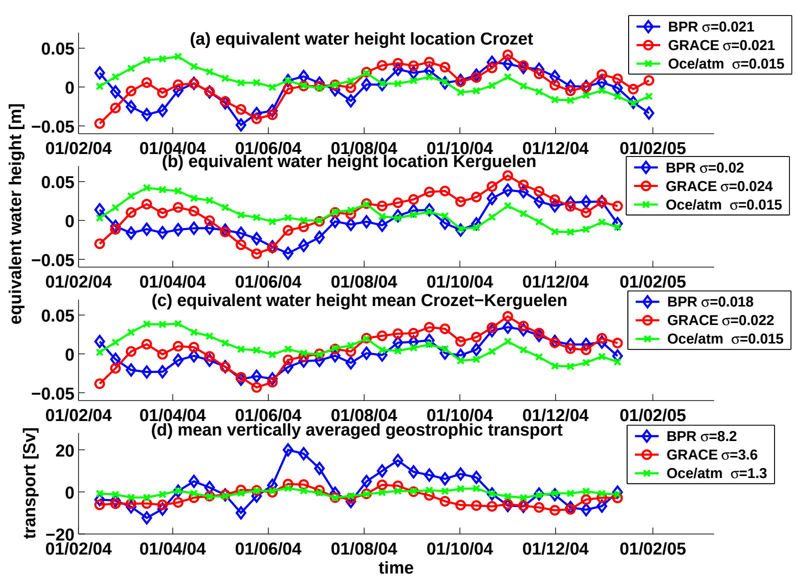
Table 1. Correlation coefficients, ρ, with their corresponding lower bounds of the 95% confidence intervals, yielded by the comparison of the local bottom pressure for the BPR stations, the mean at the midpoint and vertically averaged geostrophic transport vs. GRACE-GRGS (smoothed for given radii). The values between brackets represent a subset of the BPR data, which excludes the first two months. Confidence intervals are obtained by a bootstrapping method (percentile method).

Figure 1. GRACE equivalent water height field for the solution of 31-10-2004. Note the strong anomaly at the Kerguelen plateau. Superimposed are the bathymetric contours and the positions of the BPR deployments denoted by IO1 and IO2.

Figure 2. Equivalent water height at the BPR location near Crozet (a), near Kerguelen (b), at the midpoint (c) and the change of the mean vertically averaged transport through the section bounded by both BPRs (d). Equivalent water height and geostrophic transport are given for 1) the averaged BPR data (blue diamonds), 2) GRACE (red circles) and 3) sum of barotropic ocean model (MOG2D) and ECMWF pressure fields (green crosses). The smoothing radius used is 800 km. Standard deviations are denoted by σ.



GRACE GRGS equivalent water height 31−10−2004
Crozet Basin
Amsterdam Isl.
Crozet Isl.
Kerguelen
IO1
IO2
[mm]
Latitude (Southern)
Longitude (Eastern)



(a) equivalent water height location Crozet
BPR σ=0.021
GRACE σ=0.021
Oce/atm σ=0.015
(b) equivalent water height location Kerguelen
BPR σ=0.02
GRACE σ=0.024
Oce/atm σ=0.015
(c) equivalent water height mean Crozet−Kerguelen
BPR σ=0.018
GRACE σ=0.022
Oce/atm σ=0.015
(d) mean vertically averaged geostrophic transport
BPR σ=8.2
GRACE σ=3.6
Oce/atm σ=1.3
equivalent water height [m]
transport [Sv]
time