Search for the Translational Triplet of the Earth's Solid Inner Core by SG Observations at GGP Stations

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Abstract A very large data set of 21 high precision tidal gravity observation series recorded with superconducting gravimeters (SG) at 14 stations in the Global Geodynamics Project (GGP) network are used to search for the translational triplet of the Earth's solid inner core (Slichter mode). The original observations at each station are pre-processed carefully at the International Centre for Earth Tides with an identical remove-restore technique. The time series of the tidal gravity residuals are obtained after subtracting synthetic signals and pressure influence. The power spectral densities for 8 long-term and 13 short-term series are estimated using 12000h Parzen window function based on the FFF technique, then the product spectral densities are calculated using a stacking technique. In order to detect effectively any weak geodynamical signals from the interior of the Earth, the remaining pressure effect is further removed. The final results show several significant common spectral peaks but not where previous studies had recently located the Slichter triplet. Most of the peaks probably correspond to the five- and six-diurnal non linear tides and it is probably a mere coincidence if the periods determined theoretically for the Slichter modes, fit three of the common peaks. However, we shall not be able to confirm them as the spectral peaks will be obscured by the non-linear tides generated either by oceanic effects or instrumental non-linearities. The study shows also that the detection of the dynamic effects of the solid inner core motion and other general core modes is a very complicated task. We estimate the detection limit on the gravity signals induced by the core mode to be 0.7 nGal level when the global SG observations are stacked simultaneously. The further investigation depends on the establishment of an acceptable theoretical model and the accumulation of long-term global SG data.

1、 Introduction

The translational oscillations of the Earth's solid inner core, which include the equatorial prograde, axial and equatorial retrograde translations near the centre of the Earth, are the fundamental free modes of the Earth that usually called the Slichter triplet or Slichter mode if one use a non-rotate Earth model [*Slichter,* 1961]. The predominant components of these eigenmodes are degree 1 spheroidal displacement vectors (called also the eigenvectors). However other kind of spheroidal and toroidal displacements (with same order but different degree) are coupled with them due to Earth's ellipticity and rotation [*Smith,*1974]. By using a generalized spherical harmonic expansion technique, the translational triplet was theoretically studied by *Smith* [1976] and the eigenvectors were expressed as sum of 1st degree spheroidal and 2nd degree toroidal displacement vectors and (m=-1, 0 and 1). They were expected as 4.916h, 4.441h and 4.055h for equatorial prograde, axial and equatorial retrograde translational components when adopting a DG597 Earth model. This study showed that the influence of stratification in fluid outer core and elastic property of the solid inner core on eigenperiods to be significant. To overcome the uncertainty due to the convergence of the generalized spherical harmonic expansion, the theoretical studies were developed using a finite element technique based on so-called subseimic approximation [*Smylie and Rochester,* 1981] and a variation approach [*Smylie et al.* 1992a, 1992b]. As a result, the eigenperiods of the Slichter modes were predicted as 3.581h, 3.766h and 4.015h for Earth's model CORE11 and 2.603h, 2.702h and 2.824h for model 1066A. However in their computation, the Brunt-Väisälä frequency, describing the stability of fluid stratification, is taken a constant in whole fluid outer core. This hypothesis may be not obviously reasonable and may lead to some discrepancies in the numerical results. On the other hand the theoretical studies of *Smith* [1976] and *Crossley et al.* [1999] indicated that the magnitudes of gravity variation, due to core modes, may be at one nGal level only which falls inside the SG precision range [*Richter,* 1987]. The inertial waves induced by Earth's fluid outer core were studied by *Aldridge and Lumb* [1987], the possible discovery of these inertial waves using SG spectra at Brussels were investigated by *Melchior and Ducarme* [1986].

The modern SG manufactured by GWR has many advantages over a spring gravimeter, such as high-sensitivity, high-sampling rates, high-precision (10^{-12} g), long term stability and wide dynamics range [*Warburton,* 1985; *Goodkind,* 1991; *Crossley et al.,* 1999; *Ducarme and Sun,* 2001]. SGs provide us with an important tool for the detection of super-weak signals arising from dynamic behaviours in the Earth's deep interior. The International Centre for Earth Tides (ICET) has been collecting the SG data for more than 20 years and is also the data centre for the Global Geodynamics Project (GGP) network since 1997 [*Ducarme et al.,* 2002]. The global data set is now equivalent to more than 100 years of observations. By using these data, we hope to study the behaviours of the Earth's interior dynamics [*Ducarme and Sun,* 2001; *Sun et al.,* 2002a]. Based on the eigen-parameters of the Earth's resonance at certain periods, some significant constraints on physical properties can be obtained, such as density or viscosity of the Earth's deep interior [*Smylie,* 1992a; *Smylie,* 1999; *Smylie and McMillan,* 2000].

Past studies have shown that it is difficult to detect experimentally general core modes, including the Slichter modes for the following reasons: (1) there is no well-accepted theoretical set of eigenfrequencies due to our imperfect knowledge of core properties, such as density, (2) gravity signals from the core are relatively weak and (3) ground-based gravity observations are heavily contaminated by some known and unknown systematic background noise. In order to overcome the difficulties, the stacking technique for the SG data from various stations may be the most effective way to enhance common harmonic signals.

The first multi-station experiment was done by *Cummins et al.* [1991] based on stacking the IDA data from the LaCoste-Romberg spring gravimeters; no core modes signals could be identified clearly due to the poor sensitivity of the instruments. Four long-term SG data sets from Europe stations were stacked by *Smylie et al.* [1993a], who claimed that three weak resonance signals with central periods as 3.5820h, 3.7665h and 4.0148h could be identified. These periods were in agreement with those computed theoretically using the sub-seismic approximation [*Smylie et al.,* 1992a]. Using SG observations at six globally distributed stations and a more sophisticated stacking technique, further claimed identification of the inner core translational triplet was obtained by *Courtier et al.* [2000]. However, when using SG observations at stations Strasbourg/France and Cantley/Canada, the observational studies of *Hinderer et al.* [1995] and *Jensen et al.* [1995] did not find the spectral peaks related to the Slichter triplet claimed by *Smylie et al.* [1993a, 1993b] in their product and cross-spectrum. They suggested the use of simultaneous long-term SG data distributed globally. Therefore the main purpose of present study is to search for the translational modes of the Earth's inner core using SG data from the GGP network.

2、 Data Preparation

Twenty one SG time series with a total of 755,866 hourly values at 14 global distributed stations are used in this study. The observation series are divided into two groups, group 1 (G-I) for 8 long-term series and group 2 (G-II) for 13 short-term series in which most of them are starting from the GGP period. The station information including the names, observation periods, atmospheric gravity admittances and coefficients of long-term instrument drift are given in Table 1. The data sets used include 7 stations in Europe (Brussels, Membach, Metsahovi, Potsdam, Strasbourg, Vienne and Wetzell), 3 stations in Asia (Esashi, Matsushiro and Wuhan), 2 stations in northern American (Boulder and Cantley) and 2 stations in southern hemisphere (Canberra and Syowa).

During data pre-processing, the original observations are pre-processed carefully at the International Centre for Earth Tides (ICET) with identical remove-restore technique. The same procedure is applied on one minute sampled data using T-Soft [*Vauterin,* 1998]. Anomalous signals such as spikes, steps and large-amplitude ascillations caused by large earthquakes are carefully removed, by an interactive procedure. Missing data due to short power interruption are interpolated using a synthetic tidal gravity model. A low-pass filter is used to decimate one minute sampled data into hourly observations. The gravimetric tidal parameters are determined with ETERNA3.4 software [*Wenzel,* 1998] using the high precision tide-generating potential developments with 1200 waves developed by *Tamura* [1987]. Gravity residuals are obtained by subtracting synthetic tidal gravity from hourly observations. The pressure admittances C ($nm\cdot s^{-2}\cdot hPa^{-1}$) (see Table 1) are determined by a regression technique between gravity residuals and station pressure. The instrument drift is simulated using a quadratic polynomial (see polynomial coefficients in Table 1).

As a matter of fact, the gravity residual series *Res(t)* can be expressed as

(1)

where *Obs*(t) and *Pr*(t) are original tidal gravity records and station pressure, δ_k and $\Delta\varphi_k$ the amplitude factor and phase difference of the k^{th} tidal wave group to be determined, α_k and β_k the initial and final index of the k^{th} tidal wave group in tide-generating potential, A_i, ω_i and φ_i are theoretical amplitude, angular frequency and initial phase.

Table 1. SG gravity residual series used in the present study

Station	Observation Period	C	a_0	a_1	a_2
G-I					
Brussels1/Belgium	1982-06-02~1986-10-15	-3.428	5477.74	5.1615×10^{-3}	-2.27187×10^{-7}
Brussels2/Belgium	1986-11-15~2000-09-20	-3.428	8261.31	-4.3336×10^{-3}	3.95268×10^{-8}
Membach/Belgium	1995-08-04~2000-05-31	-3.428	-1068.06	8.6815×10^{-3}	-7.62658×10^{-8}
Potsdam/Germany	1992-06-30~1998-10-08	-3.500	42.6904	1.2907×10^{-2}	-1.25675×10^{-7}
Strasbourg/France	1987-07-11~1996-06-25	-3.000	-752.296	3.7677×10^{-2}	-3.09172×10^{-7}

Boulder/U.S.A.	1995-04-12~2001-03-29	-3.240	4100.57	1.2123×10^{-2}	-6.78259×10^{-8}
Wuhan/China	1988-11-17~1994-01-04	-3.840	-270.722	7.5364×10^{-3}	-4.17849×10^{-7}
Cantley/Canada	1989-11-07~1993-08-17	-3.000	-1416.08	-2.5717×10^{-1}	2.18132×10^{-6}
G-II					
Brussels/Belgium	1997-07-01~2000-09-21	-3.428	8065.87	1.4342×10^{-2}	-3.33212×10^{-8}
Boulder/U.S.A.	1997-07-01~1999-06-30	-3.240	4298.51	1.1964×10^{-2}	-1.44745×10^{-7}
Cantley/Canada	1997-07-01~1999-09-30	-3.000	-530.54	7.0119×10^{-3}	-3.58618×10^{-7}
Canberra/Australia	1997-07-01~1999-12-31	-3.002	3271.41	3.4538×10^{-3}	-6.42189×10^{-9}
Esashi/Japan	1997-07-01~1999-12-31	-3.145	3161.77	-3.0466×10^{-3}	2.58302×10^{-7}
Matsushiro/Japan	1997-07-01~1999-12-31	-3.334	2334.66	4.6403×10^{-2}	-4.85485×10^{-7}
Membach/Belgium	1997-07-01~2000-05-31	-3.428	-909.798	-4.9404×10^{-6}	1.37124×10^{-7}
Metsahovi/Finland	1997-07-01~2000-06-30	-3.810	-1854.78	2.89513×10^{-2}	-2.69912×10^{-7}
Strasbourg/France	1997-07-01~1999-07-31	-3.000	2.56017	1.53173×10^{-3}	3.06189×10^{-8}
Syowa/Antarctic	1997-07-01~1998-12-31	-3.920	-1914.72	-8.6945×10^{-3}	-2.89515×10^{-7}
Vienna/Austria	1997-07-01~1999-06-30	-3.220	-4995.73	4.39883×10^{-3}	-1.24985×10^{-7}
Wetzell/Germany	1996-07-28~1998-09-23	-3.484	2353.82	-2.8184×10^{-1}	-3.79209×10^{-7}
Wuhan/China	1997-12-20~2000-08-31	-3.498	3111.63	4.89498×10^{-3}	-1.43076×10^{-7}

3、Estimation of the Power and Product Spectral Density

Based on the Fourier transform technique, we estimate the spectrum for a given tidal gravity time series follows the methods of *Smylie*, [1993a]. However a suitable averaging method should be selected in order to increase reliability of the results. In our case, the observations with total length T are split into several segments, each of which has a common length M with 75% overlap. Hence the number of the data segments is given as

. (2)

The Fourier transform of the n^{th} segment may be written as

, (3)

where ω denotes the angular frequency and $w(t)$ is a chosen window function with a common length M and i the imaginary number. The Fourier transform of the full residual series can be evaluated by averaging over the results of all the κ segments, expressed as

, (4)

where and are the amplitude and phase. Therefore the estimation of the power spectral density can then be conveniently deduced as

, (5)

where I is the normalisation factor given as

(6)

When using N different time series, if the Fourier transform of one series is denoted as , and its power spectral density expressed as (with i=1, 2, ..., N), then the product spectral density (PSD) estimate can be defined as

(7)

This expression is equivalent to the cross spectral density estimation given by *Hinderer et al.* [1995]. Based on equation (5), the individual power spectral densities for all 21 residual series are obtained. The selected length for common segment is taken as M=12000h and the Parzen window is chosen as the time domain window. The Product Special Density (PSD) signals for both G-I and G-II, stacked based on equation (7), are then computed. As a convenient comparison, the PSD estimations for station barometric pressure are also computed.

It is seen that the plots of the PSD estimate for barometric pressure for G-I and G-II are very similar, showing similar properties of the station pressure. Two significant spectral peaks relating to solar heat tides (S5 and S6) in sub-tidal band are noted. The analysis shows that although the atmospheric pressure signals are removed from original tidal gravity records with a single coefficient, however, it is not sufficient to remove them entirely due to the frequency dependence of the admittance [*Merriam*, 1993; Crossley 1995, *Sun,* 1995; *Sun,* 1997; *Kroner and Jentzsch* 1999]. Moreover S5 and S6 waves are coherent harmonic signals at planetary scale that is not true for the background noise. This is the reason why the corresponding influences of the S5 and S6 signals also appear in the gravity product spectrum. In order to detect effectively the weak geodynamical signals, it is necessary for us to remove further the remaining pressure signals. Using a cubic polynomial, the mean station background noises are simulated in sub-tidal band.

The ratios of gravity and of pressure product spectral amplitude are taken as the frequency dependent pressure-gravity coefficients for remaining correction for S5 and S6 wave. The linear function between pressure and gravity product spectral amplitude is adopted to fit the coefficients in order to remove further the remaining barometric influence for other wave band. After such correction, the final PSD estimates for both G-I and G-II are shown in Figure 1.

Comparing upper and lower curves in Figure 1, it can be seen that the mean background noise is much lower in G-II than that in G-I. It confirms the high quality SG observations during the GGP period. From Figure 1, there are no obvious peaks relating to the Slichter triplet claimed by *Smylie* [1992b], *Smylie et al.* [1993a, 1993b] and *Courtier et al.* [2000]. However several significant spectral peaks denoted as SP1, SP2, ..., and SP8 are found for both G-I and G-II. What are the sources of these common spectral peaks? In order to find answer, the PSD estimates for both pressure and tidal gravity residuals in G-II at stations inside and outside Europe are also carried out, it is optimal that the similar peaks are situated. This implies that 8 common peaks do not arise from the pressure influence, local background noise and systematic observation errors, it is possible the expression of some hidden common harmonic signals. Besides the 8 common spectral peaks, there exist also two obvious peaks near the SP1 in G-II.

In order to check the effectiveness of a stacking technique in above study, a verification test has been performed. Signal of amplitudes (0.5, 0.7 and 0.9 nGal), at a known frequency, are injected into each gravity residuals series in G-II, then the PSDs for each series are stacked. The selected frequency is located in the central part of our 8 common peaks, i.e., 0.23 cpd. The results show that no common peak appears when injecting a signal with an amplitude 0.5 nGal. However, when injecting a signal with an amplitude 0.7 nGal, the signal to noise ratio is about 1.15 and the peak is marginally identifiable. When using a signal with an amplitude 0.9 nGal, the signal to noise ratio reaches 1.25 and an clear peak is found. This test shows that the stacking technique can effectively detect a common signal with amplitude of one nGal, distributed globally around the world. Therefore the 8 peaks found in our study are approved.

Figure 1. Final product spectral density (PSD) estimates for G-I (upper) and G-II (lower) after further pressure correction. Three vertical solid lines correspond to the locations of the Slichter modes (claimed by *Smylie et al,* 1993, and *Courtier et al.,* 2000). The vertical dashed lines are the locations of the common significant peaks.

4. Identification of the spectral peaks

Now let us determine the resonance parameters, including the central frequencies and periods, quality factors and resonance strengths. As *Smylie* [1992a, 1992b], the eigen-resonance frequency near the spectral peaks are simulated by a harmonic oscillator given as

, (9)

where $s(f)$ is the PSD estimates at frequency f, A resonance strength in nm·s^{-2}·cph$^{-1/2}$, f_0 central frequency of corresponding spectral peak and Δf damping interval of the resonance frequency.

Therefore the central period T and quality factor Q can then be estimated as and . The resonance parameters with their error estimation for total of 10 common spectral peaks for both G-I and G-II are determined (Table 2).

From Table 2, one can see that the discrepancy of resonance parameters determined for both groups is very small. Taking an average, the central periods of spectral peak are then obtained numerically. The periods for peaks SP1, SP4 and SP7, as examples, are given as 4.93441 ±0.00186h, 4.42734 ±0.00159h and 4.09380 ±0.00082h. They are surpassingly similar to those of the translational modes, computed theoretically by *Smith* [1976] for model DG579 as 4.916h, 4.441h and 4.055h. The discrepancies are 0.4%, -1.4% and 1.0% respectively. Therefore it is necessary for us to analysis if these common spectral peaks come from the translational oscillation movement of the solid inner core. The SP7 has two large neighbour peaks SP6 and SP8 situated symmetrically which may relate to the eigenfrequeny splitting of the well-known free spheroidal and toroidal oscillation modes due to the Earth's rotation and ellipticity [*Dahle,* 1968, 1969]. A similar splitting appears near SP1 in G-II. These facts could be in favour of the identification of the Slichter triplet with the peaks SP1, SP4 and SP7. However the amplitude of the SP4 is very weak compared with that of SP7 and even of the SP1. It is at the detection limit of gravity signal of 0.7 nGal. Moreover the peak SP7 is located in the frequency band of the six-diurnal tides. In Table 4 we give the main constituents identified for example in the oceanic tides at Oostend in Belgium of *Melchior et al.* [1967]. It is interesting to note that the two main peaks in Figure 7 corresponds to the two largest amplitudes in Table 3 and that peak SP8 is significantly weaker. The wave 2MK6 is probably mixed up with 2SM6 and the wave MSK6 with 2SM6. At the left of the Figure 7 we are in the frequency band of the fifth-diurnal tides, which are not observed in Oostend (Table 4)

Table 2. Resonance parameters of the solid inner core determined when stacking the SG data

	G-I				G-II			
	f_0/cph	T/h	Q	A	f_0/cph	T/h	Q	A
SP1	0.20244	4.93969	94	0.98293	0.20288	4.92907	103	0.56581
	(± 0.00009)	(± 0.00212)	(± 14)	(± 0.02250)	(± 0.00007)	(± 0.00160)	(± 19)	(± 0.01703)
SP2	0.21435	4.66530	121	0.87720	0.21451	4.66185	88	0.51704
	(± 0.00003)	(± 0.00062)	(± 10)	(± 0.00998)	(± 0.00007)	(± 0.00157)	(± 17)	(± 0.01275)
SP3	0.22021	4.54122	96	0.75371	0.22034	4.53841	69	0.43480
	(± 0.00017)	(± 0.00346)	(± 46)	(± 0.03481)	(± 0.00014)	(± 0.00287)	(± 25)	(± 0.01333)
SP4	0.22592	4.42633	100	0.74564	0.22582	4.42835	108	0.44698
	(± 0.00005)	(± 0.00088)	(± 13)	(± 0.01145)	(± 0.00012)	(± 0.00229)	(± 31)	(± 0.01678)
SP5	0.23608	4.23589	66	0.68372	0.23574	4.24199	142	0.42847
	(± 0.00020)	(± 0.00363)	(± 20)	(± 0.01105)	(± 0.00007)	(± 0.00133)	(± 33)	(± 0.01764)
SP6	0.24154	4.14012	184	0.81952	0.24153	4.14036	200	0.52575
	(± 0.00007)	(± 0.00116)	(± 41)	(± 0.04121)	(± 0.00007)	(± 0.00117)	(± 42)	(± 0.03009)
SP7	0.24421	4.09493	162	0.81406	0.24434	4.09269	248	0.59393
	(± 0.00006)	(±0.00092)	(± 29)	(± 0.02427)	(± 0.00004)	(± 0.00072)	(± 35)	(± 0.02634)

	G-I				G-II			
	f_0/cph	T/h	Q	A	f_0/cph	T/h	Q	A
SP8	0.24755	4.03952	180	0.78873	0.24722	4.04494	133	0.43971
	(± 0.00004)	(± 0.00064)	(± 32)	(± 0.01862)	(± 0.00007)	(± 0.00119)	(± 25)	(± 0.01624)
S5	0.20833	4.80000	289	1.68925	0.20835	4.79957	335	1.14333
	(± 0.00003)	(± 0.00065)	(± 33)	± 0.06622)	(± 0.00003)	(± 0.00057)	(± 39)	(± 0.04596)
S6	0.24998	4.00027	145	0.93283	0.24987	4.00214	205	0.62452
	(±0.00006)	(± 0.00090)	(± 20)	(± 0.03001)	(± 0.00004)	(± 0.00069)	(± 26)	(± 0.02216)

The wave M5 may explain the very large peak in G-I. The waves 2MO5 and 2SO5 correspond to the small peaks symmetrical with respect to the wave SP1 in G-II. The wave 2SK5 is mixed up with wave S5. The peaks SP2 to SP5, situated between the five- and six-diurnal bands, are not so easily explained but they are generally weaker and close to the detection level illustrated in Figure 8. In the oceanic tides the sixth diurnal tides are called "shallow water components" and are due to non-linear interactions among the main tidal constituents close to the coasts. These generated modes explain why the fifth-diurnal tides are not observed in Oostend as the diurnal tides are very weak in this area.

Fifth-diurnal tides can be produced indeed in places, like China sea, where both diurnal and semi-diurnal oceanic tides are present. As many GGP stations used in this study are close to the sea we could imagine an oceanic tidal loading due to the shallow water components. However these components are restricted to the coastal areas and their loading efficiency is thus very weak. A more evident way of generating fifth- and sixth-diurnal components is the existence of non-linearities in the instruments and their electronics. For example wave M5, which is a clear non-linear term, is large in G-I where we have a majority of old superconducting instruments of lower quality.

Table 3: Six-diurnal tides identified observed in oceanic tides at Oostend harbour

and identified in the PSD estimates [Melchior et al., 1967].

Wave	Cycle/hour	Period (h)	Ampl.(cm)	identification
2MN6	0.240022	4.1663	3.7	
M6	0.241534	4.1402	6.8	SP6
MSN6	0.242844	4.1179	1.8	
2MS6	0.244356	4.0924	7.0	SP7
2MK6	0.244584	4.0886	1.9	
2SM6	0.247178	4.0457	1.3	SP8
MSk6	0.247406	4.0419	1.0	

Table 4: five-diurnal tidal frequencies observed in PSD estimates

Wave	Cycle/hour	Period	identification
2MO5	0.199753	5.0062	G-II
M5	0.020128	4.9683	G-I
2MK5	0.202803	4.9309	SP1
2SO5	0.205397	4.8686	G-II
2SK5	0.208447	4.7974	

5、 Conclusions and Discussion

The PSD estimates for both G-I and G-II of the SG tidal gravity residuals at 14 GGP stations distributed globally are estimated accurately by using the FFT and a stacking technique in order to confirm the triplet translational resonance of the Earth's solid inner core. The final results indicate that there are no obvious signals in our PSD estimates related to Slichter modes at the frequencies claimed by *Smylie et al.* [1993a, 1993b] and *Courtier et al.* [2000].

Instead, it is found that there exist several common significant spectral peaks in the sub-tidal band, which are not present in barometric pressure series. Unhappily the main peaks are located in the fifth- or sixth-diurnal bands and coincide with the expected frequencies of the non-linear tides. It is probably a pure coincidence if the central periods of three peaks are estimated accurately at 4.93438, 4.42734 and 4.09269, periods in agreement with the eigenperiods of the Slichter modes computed theoretically by *Smith* [1976]. We can conclude that even if the values computed by Smith are correct we shall not be able to confirm them as the spectral peaks will be obscured by the non-linear tides generated either by oceanic effects or instrumental non-linearities.

We can also wonder why the peaks detected by *Smylie et al.* [1993] and *Courtier et al.* [2000] are not visible in our study. Perhaps the Slichter modes are not always exited and we should then apply techniques derived from the wavelet analysis. Another possibility is also that increasing the extension of the network, without taking into account the geographical phase distribution of the Slichter modes, is degrading the coherence of the signal.

Of course, to detect experimentally the dynamic effects of the solid inner core motion and other general core modes is a very complicated task so far. The detecting limit of surface gravity variations is at the level of 0.7 nGal or more when high-precision SG gravimeter observations recorded at globally distributed stations are used simultaneously. This is nearly the upper limit on the signals induced at the Earth's surface by the geodynamics of the Earth's interior. On the other hand, there is not any convincing theoretical model of the core modes for a realistic Earth so far. Therefore the further investigation depends on the establishment of a reasonable and acceptable theoretical model, to provide the spectral properties of the core modes, and on the further accumulation of the global SG gravity data via the GGP network since they can provide us with more accurate spectral resolution. With that respect it is encouraging to see the improvement in signal to noise ratio in the G-I and G-II.

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