An Assessment of the Accuracy of GSN Sensor Response Information

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INTRODUCTION

The Sumatra-Andaman earthquake of 26 December 2004 provides a special opportunity to validate the accuracy of sensor sensitivities reported for the IRIS Global Seismic Network (GSN). A goal of the GSN is to publish instrument responses to an accuracy of 1% in amplitude and 1° in phase. This earthquake excited long-period free oscillations such that modes with low decay rates could be observed above ambient noise for many weeks after the event. By examining the subset of modes least sensitive to short wavelength structure, one may test the reliability of published station response information. We investigate here the amplitude of the radial mode $0_5^0$, which is predicted to be uniform globally, by using two different techniques.

The IRIS Global Seismic Network (Butler et al., 2004) is now a fully operational scientific facility routinely used to investigate Earth structure and seismicity in many regions of the world where instrument coverage was previously very poor. The principal goal of the GSN, to place high-performance instruments at 2,000 km spacing around the planet, has largely been achieved except for some parts of the oceans. Several agencies, including the USGS National Earthquake Information Center, NOAA Tsunami Warning Centers, and United Nations Comprehensive Test Ban Treaty Organization, now rely routinely on GSN seismic data.

Agencies and individual researchers throughout the world depend upon the accuracy of the metadata provided to them through the IRIS Data Management System (http://www.iris.edu/). The IRIS DMS in turn distributes response information supplied by the GSN network operators, the USGS Albuquerque Seismic Lab (ASL) and the University of California at San Diego’s Project IDA (IDA). Both network operators conduct tests to measure the instrument response of the equipment they deploy at GSN sites.

The Sumatra-Andaman earthquake of 26 December 2004 affords a rare opportunity to assess independently the accuracy of these metadata. This earthquake, the largest in forty years, excited long-period free oscillations well above the ambient noise level at all GSN stations (Berger et al., 2004; Park et al., 2005). The lowest frequency radial oscillations are only weakly coupled to other modes by rotation and laterally varying Earth structure, so their amplitudes and phases should be approximately uniform globally (see for example Park, 1990).

To test the validity of this approximation, we performed a full Galerkin coupled-mode calculation following the method of Park and Gilbert (1986) for modes $0_5^0$ through $1_7^1$, a total of 287 singlets. The strongest coupling to $0_5^0$ was from modes $1_2^2$, $0_5^0$, and $1_7^1$, and all were found to be weak with an amplitude perturbation of less than one part in one thousand. Therefore, variations in the amplitude and phase of modes, such as $0_5^0$, at GSN stations can be attributed to inaccuracies in the reported instrument response rather than to perturbations caused by coupling to other modes through rotation or structure.

We first describe the procedures employed by the network operators to calibrate seismic instruments at GSN stations. We then describe two methods used to measure the amplitude of $0_5^0$ and their results.

CALIBRATION

System calibration of an IRIS GSN station is accomplished by measuring the response of the sensor(s) and data logger in separate steps. Both measurements must be made by a skilled technician and are generally done only during installation, upgrade, or repair. The final outcome is a description in SEED format describing the instrument response (see Ahern et al., 1993, Appendix C). GSN systems employ a standard pole-zero format to represent the analog stages, which consist of the sensor and anti-aliasing amplifier filter unit:

$$ G(f) = S_d A_0 \prod_{m=1}^{N} \frac{(f - p_m)}{(f - r_m)} , $$

where $p_m$ and $r_n$ represent the poles and zeros, respectively. Conventionally, the leading factor $A_0$ is chosen to normalize the series at a canonical frequency $f_0$, so the scalar $S_d$ contains all information about the stage’s sensitivity.

The data logger’s sensitivity is determined by imposing a precisely known potential on the input terminals of the digitizer or any buffer electronics that may be placed between the sensor and digitizer, and then measuring the resulting output digital counts (Figure 1). This test yields the combined gain of the anti-aliasing filter (G2) and the DAS sensitivity (G3). Although this procedure may be performed in the field, the electronics are stable enough that digitizer calibration is usually performed only in the laboratory prior to shipment.
Figure 1. Calibration scheme for IRIS GSN systems. To calibrate the digitizer and ancillary electronics, a known voltage is inserted at B and measured at C. To determine the shape of the seismometer’s response, a random binary or other broadband, white-noise signal is fed into the calibration circuit of the seismometer at A and the output is recorded at C.

The performance characteristics of the mechanically delicate seismometers may change during shipment and set-up, however, and a sensor calibration is always performed in situ. The procedure and theory of calibration are explained in detail by Berger et al. (1979). A random binary signal or other broadband white-noise process is fed to the calibration coils of the seismometer, and the output is recorded for analysis at the network operator’s data collection center. The shape of the sensor frequency response is found by fitting a perturbed function of the nominal system to the cross-spectrum of the output and the (known) input (Fels and Berger, 1994). This test determines the poles and zeros $p_m$ and $r_m$ in the equation above for the seismometer (G1), and the anti-aliasing amplifier/filter (G2), represented in Figure 1. In newer systems, the input signal is digitized on a separate channel and recorded for later analysis. In practice, this method should yield calibrations to an accuracy of better than 1% in amplitude and 1° in phase.

The random binary test can be used to determine the shape of a sensor’s response as a function of frequency but not the sensor’s overall sensitivity. Both IRIS GSN network operators use the sensitivity values provided by the manufacturer from measurements made at the factory. These may be determined using shake tables or other specialized devices. These tests establish $S_d$ for the generator constant of the seismometer (G1).

The stability of the system response is quite good. Illustrated in Figure 2 is the small relative change in amplitude and frequency response for the KS-54000 seismometer deployed at station KAPI (Kappang, Indonesia) between successive calibrations in October 1999 and January 2002. The change in amplitude and phase between these two tests is at the level of uncertainty in the calibration procedure. Reports of a few cases where the system’s response varies with time (Ekström, personal communication) are currently being investigated.

**MEASUREMENT OF $_0S_0$**

We first obtained all low-rate data (the VHZ-00 channels in SEED nomenclature) from the principal vertical sensor at all operational GSN stations for the period beginning with the Sumatra-Andaman event itself and ending with the second major Sumatran event on 28 March 2005. The most useful data proved to be from the first 75 days following the initial quake. The time series from all stations had gaps, and at many, the background noise level varied with time.

We used two methods to determine the amplitude of $_0S_0$ for comparison across the network. The first is a variant of a technique proposed by Thomson (1982). We divided each station’s time series into segments four days long and computed an individual amplitude estimate by applying a frequency transformation $z(t)$ to the time series segment $x(t)$,

$$z(t) = x(t)e^{-i\omega_0 t}e^{-\frac{t}{2Q_0}}$$

where $\omega_0$ and $Q_0$ are the mode’s frequency and attenuation factor. This operation shifts the frequency of the mode to be
measured to DC and **compensates for attenuation**. The main properties of $\omega_0$ have been studied for some time (Knopoff et al., 1979; Riedesel et al., 1980; Zürn et al., 1980; Masters and Gilbert, 1983), so the values of $\omega_0$ and $Q_0$ were known *a priori*. In this study we used a frequency 7 mHz lower than the 0.814664 MHz value reported by Riedesel et al. (1980) and a $Q_0$ of 5,400. The perturbation in frequency was necessary to explain phase shifts observed in this study and will be the subject of a separate report (Davis and Masters, in preparation). We then tapered the transformed $n$-point series with 4π discrete spheroidal wave functions $v_k$ as advocated by Thomson (1982):

$$Y_k = \sum_{0}^{n-1} z(t) \cdot v_k(t) .$$

By selecting only the first five tapers and choosing a four-day record length, we effectively filter out energy from all other well excited modes at nearby frequencies, including $s$-waves. The weighted estimate for the modal amplitude $\mu$ is then

$$\mu = \sum_{k=1}^{5} U_k \cdot Y_k / \sum_{k=1}^{5} |U_k|^2 ,$$

where

$$U_k = \sum_{0}^{n-1} v_k(t) .$$

and the variance of the result may be estimated from the residual of the fit:

$$\sigma^2 = \frac{\sum |Y_k - \mu U_k|^2}{\sum |U_k|^2} .$$

An example of the results for a single station, TLY (Talaya, Russia), is shown in Figure 3. The measurements of displacement amplitude and phase remain stable until day 70 for this station. At stations not as quiet as TLY, this scattering behavior begins earlier. Note that this amplitude measurement includes a correction for the putative instrument response but **not for the free-air perturbation**, an effect induced by motion in a variable gravity field. This latter correction should be made before comparing these observations with recordings made by other types of instruments. (See Chapter 10 of Dahlen and Tromp, 1998, for a more complete explanation.) If the free-air gravity correction is applied, the value of this amplitude measurement will decrease by $2g/r_o^2$ (where $g$ is the local acceleration of gravity and $r_o$ is the Earth’s radius), or about 12%.

In tabulating the results for the entire network, we reviewed the pattern of scatter at each station and selected a cut-off point when the scatter began to increase substantially. The summary of all stations of the GSN having two or more measurements is shown in Figure 4. For this set, the mean displacement amplitude is 57.7 microns and phase, 65.8°. This latter value was used
A Figure 3. Measurements of the amplitude (o) and phase (x) of \( \theta S_0 \) observed at station TLY (Talaya, Russia) made using the first method applied to time segments four days in length. The scatter of the data increases starting at day 70. Similar scatter of data was observed at other stations beginning earlier or later, depending upon the signal to noise ratio at each location.

by Park et al. (2005) to constrain the duration of the Sumatra-Andaman event.

A second method to measure the excitation of amplitude of \( \theta S_0 \) and other modes observed over the network is with a receiver stripping technique (Masters et al., 2000). The time series for an isolated multiplet at the \( j \)th receiver, \( u_j(t) \), can be represented as

\[
  u_j(t) = \sum_{k=1}^{2l+1} R_{jk} a_k(t) e^{i \omega_0 t},
\]

where \( l \) is the angular degree of the mode, \( R_{jk} \) describes the motion of the \( k \)th singlet at the \( j \)th station, \( a_k(t) \) contains the effects of rotation, ellipticity, and heterogeneity as well as the singlet excitation, and \( \omega_0 \) is the degenerate frequency of the multiplet (Woodhouse and Girnius, 1982). For multiple receivers, the above equation can be written in a vector format as

\[
  u(t) = R \cdot a(t) e^{i \omega_0 t}.
\]

The receiver strips \( b(t) \) of an isolated multiplet are defined as (Masters et al., 2000)

\[
  b(t) = R^{-1} u(t) = a(t) e^{i \omega_0 t}.
\]

This operation collapses observations at various stations to \( 2l+1 \) strips which contain information about splitting and source mechanism, i.e., it requires no knowledge of the source properties. The receiver strip of the mode \( \theta S_0 \) obtained from the Sumatra-Andaman event is shown in Figure 5.

The receiver strip estimates, \( b_{est} \), can then be used to assess the accuracy of instrument responses. The spectra of individual stations are predicted by

\[
  u_{est} = R \cdot b_{est}(t),
\]

which can be compared to observed spectra \( u(t) \) to identify mismatches in amplitude and phase due to an incorrect instrument response or timing errors (Masters et al., 2000). An example
of this is shown in Figure 6 for station WRAB (Warramunga, Australia). The predicted spectrum, computed based upon data from many GSN stations, fits the observed spectrum in phase well but not quite as well in amplitude. This misfit, 15% for this station, is most likely due to an incorrect value for the WRAB instrument’s sensitivity (G1). The scalar value needed to compensate for the mismatch was tabulated for all GSN stations for which a well recorded signal of $oS_0$ was available.

The two methods are compared in Figure 7. The distribution of points indicates that both methods identify the same stations at the extremes of the distribution, and the general trend demonstrates consistency between the two techniques. When measurement uncertainty is taken into account, many stations fall nominally close to the center, which indicates that the published instrument response is correct on average.

**DISCUSSION**

The amplitude measurements made here indicate that the calibration of the IRIS GSN is extremely good but has not yet met the goals set for operational standards. By one measurement technique, 16% of station amplitude estimates fall greater than two standard deviations from the network mean. For phase, this value is 11%. The stations on the tails of the distribution (Figures 4 and 6) should be targeted for greater scrutiny in the future.

There is no systematic difference observed between the principal broadband sensors deployed in the GSN. The extrema in the distribution shown in Figure 4 contain observations from both STS1 and KS-S4000 seismometers, and there does not appear to be a cluster of data for either, which might indicate a systematic miscalibration.

One way to verify calibration of normal mode excitation independently is to measure the amplitude and phase of tidal signals. Modern models of ocean tidal loading (see for example Agnew, 1995) utilize satellite altimetry information to improve their modeling capabilities and can reliably predict the amplitude of a tidal line to 1% accuracy at most points on the Earth. Examining tidal lines would also permit one to estimate changes in a sensor’s calibration over time.

The results reported here apply only to the vertical sensors of the GSN. The second measurement method can be applied to modes recorded on both vertical and horizontal instruments. Future studies can therefore generalize this investigation to all three components of the principal seismometers used at GSN sites.

Superconducting gravimeters also recorded $oS_0$ very well following the Sumatra-Andaman quake. A sampling of these data indicate the GSN mean is about 4% larger than measurements at several superconducting gravimeters thought to be calibrated to better than 0.5% (Widmer-Schnidrig, personal communication). Resolving these and other inconsistencies poses an interesting challenge to the GSN station operators as they begin to fine-tune the performance of this vital scientific instrument.

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Figure 7. Comparison of results based upon the two methods described in the text. Observations from Method 1 are expressed as percentage deviation from the mean of the data set, 57.7 microns. Observations from Method 2 are expressed as the ratio (in terms of %) of synthetic to observed spectrum.

REFERENCES


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