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THE STABILITY OF RMS *Lg* MEASUREMENTS AND THEIR POTENTIAL FOR ACCURATE ESTIMATION OF THE YIELDS OF SOVIET UNDERGROUND NUCLEAR EXPLOSIONS

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ABSTRACT

Data on underground nuclear explosions have recently become available from modern digital seismic stations installed within the Soviet Union and China. Observations of root mean square (rms) *Lg*-wave signals for Soviet underground nuclear explosions at the Shagan River Test Site in East Kazakhstan show that the relative amplitudes of the rms signals at stations in Norway, the USSR, and China are very similar for different explosions, the standard deviation of the differences being only about 0.03 in logarithmic units (i.e., magnitude units).

This is consistent with earlier observations comparing NORSAR and Graefenberg array data, and the observed scatter is significantly lower than has been reported for *Lg* data from Nevada Test Site explosions. In view of the excellent correspondence found by Nuttli (1986) and Patton (1988) for *Lg* versus yield at Nevada, this indicates that rms *Lg* has a potential for yield estimation with very high accuracy at Shagan River.

Our study has shown that: (a) selected stations in the USSR and China, situated at regional distances, provide a much improved signal-to-noise ratio of the *Lg* phase for events at Shagan River, as compared to NORSAR array data; (b) the scaling of rms *Lg* amplitudes between different-sized events recorded at the same single station site appears to be consistent with that of NORSAR, indicating a remarkable degree of precision in single station measurements of *Lg* signal; (c) rms *Lg* amplitude measurements for the best of these stations may be made at 1.5 to 2.0 magnitude units lower than at NORSAR or Graefenberg, allowing a much lower threshold for *Lg*-based yield determination; and (d) the *P*-wave detection capabilities of these single stations do not match those of the NORESS and ARCESS arrays; thus, teleseismic signals continue to be important for detection of small nuclear explosions.

Our conclusion is that *Lg* signals appear to provide an excellent basis for supplying estimates of the yields of nuclear explosions even down to below 1 kt when such signals are recorded at high-quality, digital in-country seismic stations, and when calibrated by access to independent (nonseismic) yield information for a few nuclear explosions at the test sites of interest. In the context of monitoring a low-yield threshold test ban treaty, it will, in addition, be important to take into consideration various environmental conditions in the testing area, such as the possible presence of cavities, and to devise appropriate procedures for on-site observations in this regard.

INTRODUCTION

We report our observations of root mean square (rms) *Lg*-wave signals for Soviet underground nuclear explosions at the Shagan River Test Site in East Kazakhstan. We show that the relative amplitudes of the rms signals, at stations in Norway, the USSR, and China, are very similar for different explosions. Thus, if we consider only well-recorded explosions (i.e., requiring that rms *Lg* be at least 1.5 times the rms level of noise preceding the *P* arrival), our basic observation is that rms *Lg* amplitudes at pairs of stations are in excellent agreement, the standard deviation of the differences being only about 0.03 in logarithmic units (i.e., magnitude units).

This observation indicates that a seismic measure of source size can be estimated with unprecedented precision from observations of *Lg* waves at a single station. (*P*-wave amplitudes, for example, as measured to obtain m_b, show significantly greater scatter.) We refer to such indications of precision of rms *Lg* as "stability."

Quantitative studies of *Lg* began much later in seismology than such studies of *P*, *S*, and teleseismic surface waves because *Lg* waveforms are in general more complex than those of other phases. *Lg* waveform modeling typically does not yet achieve the quality of fit between synthetics and data that has been attained with more conventional phases. It is, therefore, somewhat surprising to find that potentially the most precise estimator of seismic source size may be one based on a phase as complex as *Lg*.

In this paper, we are principally concerned with developing those properties of rms *Lg* that are pertinent to making accurate estimates of the yield of Soviet nuclear explosions, particularly at the Shagan River Test Site. For Shagan River explosions with $m_b > 5.5$, *Lg* signals at NORSAR alone were found to provide magnitude estimates that indicated stability comparable to and possibly better than those obtained from *P* waves recorded on a large worldwide network (Ringdal, 1983). Underlying this conclusion are the assumptions, articulated by Nuttli (1973), that the magnitude of seismic sources can usefully be assigned at "long-period" or "short-period," and that short-period magnitudes can be estimated either with *P* waves, or, in many circumstances, with *Lg* recorded at periods around 1 sec. We use m_b to denote short-period magnitude in general and m_b (*P*) or m_b (*Lg*) where it is necessary to indicate the wave type used for measurement.

We report the first analyses of rms *Lg* signals of Soviet nuclear explosions recorded within the USSR. We used data recorded at four in-country stations installed in the summer of 1988 by the Incorporated Research Institutions for Seismology (IRIS), under an agreement negotiated with the Soviet Academy of Sciences. What is important about these stations is that they have been allowed to run even during times when the Soviets were conducting underground nuclear explosions at weapons test sites, and for the first time this in-country data have routinely become available for analysis in the West. Using these four high-quality digital stations installed within the Soviet Union by IRIS and one installed by the British [GAM, The BSVRP Working Group, (1989)] located near the IRIS Garm station, we confirm that the stability of rms *Lg* is present at distances about 1,500 to 3,000 km from Shagan River, and can be used for explosions much smaller than those observed teleseismically. Specifically, we show an example for one of these IRIS stations, ARU (installed in 1988 at Arti in the Urals), indicating that the improvement in signal-to-noise ratio is such as to permit rms *Lg* to be used for yield estimation of explosions down to about m_b 4.0. A similar performance is found for the station GAM. We note that, according to the magnitude-yield relations presented by Vergino (1989a), m_b 4.0 would correspond to a yield well below 1 kt for nuclear explosions conducted under typical tamped conditions.

We further analyze rms *Lg* signals from Shagan River explosions recorded at two stations of the China Digital Seismograph Network (CDSN). These stations, which have sampling rates of 20 Hz and operate in a triggered mode, are at Urumqi (WMQ) and Hailar (HIA), at a distance of 950 and 2900 km, respectively, from Shagan River. Stability of rms *Lg* is again confirmed, and it appears that WMQ, if set to record continuously, could provide rms *Lg* for yield estimation down to m_b 3.5.

As part of this project to investigate *Lg*, we also address the excellent *P*-wave detection capability of the NORESS and ARCESS arrays (See Ringdal, 1990). We point out the advantages of combining the excellent *detection* capability of these teleseismic arrays with the potentially superior *yield-estimation* capability of in-country stations, for purposes of both detecting and estimating the yields of small nuclear explosions.

To place our new results in context, the next section reviews earlier studies describing the promise and problems of using *Lg* signals. This review is followed by a description of our analysis of the Soviet and Chinese data.

Review of Previous Studies of *Lg*

Lg waves are seismic waves that are observed to propagate across continental paths. They were first described by Press and Ewing (1952) from earthquakes in California that were observed at Palisades, New York, shortly after seismographs were installed at what then was called the Lamont Geological Observatory. The following characteristics were noted for what these authors called "surface shear waves":

1. initial period about 0.5 to 6 sec
2. sharp commencements
3. amplitudes larger than any conventional phase for continental paths at distances up to 6,000 km
4. observed for continental paths only, being gradually eliminated as the ocean path increases beyond 100 km
5. group velocity (near onset) around 3.5 km/sec, decreasing to below 2 km/sec for periods above 10 sec
6. inverse dispersion at distances greater than about 20° (i.e., frequency decreases for later times in the wave train).

Press and Ewing found that earthquakes as small as magnitude 4.7, at a distance of about 35°, consistently displayed the above properties. In remarking that amplitudes were "larger than any conventional phase," they were presumably comparing *Lg* to body waves that arrive more-or-less as isolated pulses, and/or to single-mode surface waves that could be identified with a particular dispersion curve.

Press and Ewing noted properties of the three components of ground motion that indicated another type of continental surface wave, which they called *Rg*, was also being observed with large amplitudes. It had group velocity about 3.05 km/sec and the characteristic retrograde elliptical particle motion of a Rayleigh wave.

The reason Press and Ewing labeled these waves *Lg* and *Rg* was that the speeds and some features of the commencement of the observed signals were similar to those predicted theoretically for Love and Rayleigh short-period surface waves in a granitic layer (i.e., for waves at periods shorter than periods seen in conventional teleseismic surface waves). They attempted quantitatively to show that *Lg* consists of *SH* waves multiply reflected within a superficial sialic layer. However, as noted by them and by Lehmann (1953), the idea of such a layer was quickly abandoned (although use of the names *Lg* and *Rg* has persisted) because:

1. The observed duration of the wave train was much longer than that indicated by Love-wave calculations in a superficial granitic layer.
2. *Lg* was recognized (even in these earliest papers) as having particle motion in vertical and radial directions, as well as in the transverse direction of conventional Love waves.

3. *Lg* was found to be strong in some earthquakes that originated below the proposed layer and thus at depths unfavorable for exciting *SH* multiples that propagate to great distances.

The basic observation that short-period *Lg* has considerable vertical and longitudinal motion was noted in these earliest studies, but not explained except to point out that a plate floating on a fluid substrate would retain *SV* multiples that arrived concurrently with *SH* out to great distances.

In retrospect, we may say that Press and Ewing identified what is still recognized as the defining properties of *Lg* waves. But, for many years after these properties were discovered, little progress was made in explaining them quantitatively in terms of synthetics. In contrast, the smaller amplitude "conventional phases" (body waves and teleseismic surface waves) have been synthesized more and more successfully. Quantitative fits to travel times and waveforms, including normal mode synthesis, have become standard methods for obtaining detailed information about Earth structure, and about earthquake and explosion sources.

However, the fact that *Lg* can be "larger than any conventional phase" carries its own imperative, whether or not it is a wave that can be fully explained with models of Earth structure and theories of seismic source and wave propagation. For decades, *Lg* (and *Rg*) have, therefore, of necessity been studied empirically by those scientists and engineers whose work inclines to a study of the largest seismic motions. Examples of such empirical work include the many uses of Richter local magnitude, M_L, comparative studies of areas of perceptibility of earthquakes in different continental regions, the related subject of how amplitudes of the largest seismic waves vary with epicentral distance, and studies of small magnitude events when only *Lg* may be apparent above noise levels.

Much pioneering work on *Lg* waves was done in the 1970s and 1980s by Otto Nuttli of St. Louis University. Thus, Nuttli (1973) proposed that, "since *Lg* represents a higher mode wave traveling with minimum group velocity," it would be appropriate to relate amplitude (A) and distance (Δ) via:

$$A = K[\Delta^{-1/3}](\sin\Delta)^{-1/2}e^{-\gamma\Delta} \tag{1}$$

where K is governed by the source strength, and γ is the spatial decay rate due to nongeometrical attenuation. This formula is the stationary phase approximation appropriate for frequencies, f, near a minimum in group velocity, U, and

$$\gamma = \pi f/(QU) \tag{2}$$

where $1/Q$ is a dimensionless measure of attenuation. For values of Δ small enough that $\sin\Delta$ is approximately proportion to Δ (i.e., when sphericity of the Earth can be ignored), the geometrical attenuation described by equation (1) is given by a factor of $\Delta^{-5/6}$. Nuttli (1973) claimed that the Richter local magnitude scale, M_L, developed for the Western United States, was based on waves that could be interpreted via equation (1), but with γ values about ten times higher than the γ values appropriate to the use of equation (1) in fitting observed amplitudes for *Lg* waves in Eastern North America.

With the goal of defining a magnitude scale for source strength at short periods, based on *Lg* observations that are corrected for path-dependent attenuation, he described in detail (Nuttli, 1973, 1986a) a three-step procedure to obtain what he called an m_b (*Lg*) value for an earthquake or an explosion of interest. The three steps were as follows:

1. γ was estimated for a particular source-receiver path
2. equation (1) was used to predict an amplitude at one particular distance (he chose Δ corresponding to 10 km for reference)
3. magnitude was assigned via the formula

$$m_b(Lg) = 5.0 + \log[A(10 \text{ km})/110]$$

where $A(10 \text{ km})$ is the amplitude, in microns, resulting from step 2.

Nuttli's method is based on a mix of phenomenological properties of observed signals and theories of *Lg* propagation. Nuttli specified in detail his procedures for estimating γ. He used a method described by Herrmann (1980) in which the tendency of signal to move to lower frequencies in later portions of the *Lg* wave train is used to obtain Q values. Q itself is taken to have a power-law dependence upon frequency. A key assumption of Nuttli's method, namely that geometrical decay of *Lg* amplitudes is described essentially by a factor $\Delta^{-5/6}$, has subsequently been given some support by calculation of synthetics in layered crustal structures (e.g., Campillo *et al.*, 1984).

In order to improve the consistency of m_b (*Lg*) estimates resulting from different stations at different distances from the same event (this is the quality referred to as "stability" in the present paper), the measurement that Nuttli actually made from seismograms (short-period WWSSN vertical components) was based on the third largest amplitude in the time window corresponding to group velocities of 3.6 to 3.2 km/sec.

For 22 nuclear explosions below the water table at NTS, Nuttli (1986a) showed that his $m_b(Lg)$ values, using only three WWSSN stations in the Western United States, were remarkably well correlated with the logarithm of announced yield. He proposed a best-fitting line through this magnitude-yield data, from which magnitudes had a standard deviation of only about 0.05. Patton (1988) developed computer-automated measures of *Lg* amplitude aiming at reproducing Nuttli's NTS results. Patton measured *Lg* amplitudes from digital seismograms in two ways—by using the third-largest peak and by computing the rms amplitude in the *Lg* time window—and found very little difference (around 0.01 magnitude units) in the amount of scatter about regression lines using the two measures. However, he found that standard deviations from best-fitting $m_b(Lg)$ − log(yield) relations were low, 0.07 to 0.08 magnitude units, only if explosions were restricted to subregions of NTS (Pahute Mesa, northern Yucca Flat, southern Yucca Flat).

Based on the success in estimating yields for NTS explosions, Nuttli proceeded to apply the same magnitude-yield relation, together with *Lg* signals recorded at analog WWSSN stations in Eurasia, to estimate the yields of nuclear explosions at three Soviet test sites (Nuttli 1986b, 1987, 1988). For the period 1978 to 1984, after the 150 kt Threshold Test Ban Treaty had gone into effect, his yield estimates for Shagan River explosions included 20 that exceeded the threshold, including one (5 December 1982) estimated by Nuttli to be about 300 kt. While acknowledging the pioneering work involved in these studies, it is clear that the generally low signal-to-noise ratios and the problematic data quality of these analog recordings made very precise measurements impossible to attain, a fact also recognized by Nuttli himself. Also, at the teleseismic distances for which Nuttli had *Lg* data (1,900 to 4,400 km), yield estimates based on absolute measures of ground motion that have to be extrapolated back to 10 km are a severe test of the validity of equation (1) [even if equation (1) is appropriate] and are very sensitive to errors in γ. Overestimating γ by 10 to 15 per cent would result in yield estimates about two times too high.

In the first of a number of *Lg* studies undertaken by the NORSAR staff during the 1980s, Ringdal (1983) analyzed digital NORSAR *Lg* data of selected Semipalatinsk underground nuclear explosions. He found that, when using NORSAR rms *Lg* instead of *P* waves recorded at NORSAR to estimate source size, it was possible to eliminate effectively the magnitude bias relative to worldwide m_b observed at NORSAR between Degelen and Shagan River explosions. The method consisted of averaging log (rms) values of individual NORSAR channels, filtered in a band of 0.6 to 3.0 Hz in order to enhance *Lg* signal-to-noise ratio. Ringdal and Hokland (1987) expanded the data base and introduced a noise compensation procedure to improve the reliability of measurement at low SNR values. They were able to identify a distinct *P–Lg* bias between the Northeast and Southwest portions of the Shagan River Test Site, a feature that was confirmed by Ringdal and Fyen (1988) using Graefenberg array data. Ringdal and Marshall (1989) combined *P*- and *Lg*-based source size estimators to estimate the yields of 96 Shagan River explosions from 1965 to 1988, using data on the cratering explosion of 15 January 1965 as a reference for the yield calculations.

Recent developments have permitted access to high-quality digital data from sites significantly closer to Shagan River and, in addition, some information on yields at this test site has become openly available. This obviates the need to make distance corrections to absolute measures of *Lg* ground motion amplitude for purposes of yield estimation at this site. Thus, the focus of this paper will be on using rms *Lg* measurements to investigate the stability of this measure for fixed station source combinations.

Data Analysis for Shagan River Nuclear Explosions

Recently, data have become available from seven stations located within the Soviet Union and China for explosions in the Semipalatinsk area (see Tables 1 and 2, and Fig. 1). These stations are comprised of the IRIS stations (Given and Berger, 1989), the CDSN stations, and the Garm station operated by the British as described previously. This new data allows the comparison of the stability of the rms *Lg* measurement technique for stations at various distances. In particular, we will compare *Lg* amplitudes of events recorded at the close-in stations with *Lg* recorded at NORSAR, and *P*-wave detectability at NORESS.

The seismograms from our data set were all processed in a manner similar to that used for the NORSAR recordings. The processing is illustrated in Figure 2. Figure 2a represents a well-recorded event of magnitude m_b (P) = 5.9, whereas

TABLE 1

Seismographic Station Locations

Station	Latitude	Longitude	Elevation (m)
WMQ	43.821°N	87.695°E	970
HIA	49.267°N	119.742°E	610
ARU	56.40°N	58.60°E	250
GAR	39.00°N	70.32°E	1300
KIV	43.95°N	42.68°E	1206
OBN	55.10°N	36.60°E	160
GAM	39.00°N	70.19°E	1300

TABLE 2

VERTICAL COMPONENT STATION VALUES

No.	Date	m_b	NAO Lg	WMQ Lg	HIA Lg	ARU Lg	GAR Lg	KIV Lg	OBN Lg	GAM Lg
1	87171	6.03	3.012	3.851	2.189	—	—	—	—	—
2	87214	5.83	2.911	3.693	2.072	—	—	—	—	—
3	87319	5.98	3.014	3.870	2.298	—	—	—	—	—
4	87347	6.06	3.133	3.907	2.352	—	—	—	—	—
5	87361	6.00	3.086	3.851	2.334	—	—	—	—	—
6	88044	5.97	3.082	3.911	—	—	—	—	—	—
7	88094	5.99	3.103	3.925	2.307	—	—	—	—	—
8	88125	6.09	3.084	3.958	—	—	—	—	—	—
9	88258	6.03	3.014	3.827	2.224	4.142	3.802	3.014	3.342	3.184
10	88270	3.8	—	—	—	2.215	—	—	—	1.196
11	88317	5.20	2.307	3.104	—	3.429	3.165	—	—	2.521
12	88352	5.80	2.846	3.636	1.947	3.935	—	—	3.191	3.034
13	89022	6.0	3.005	—	—	4.075	—	—	—	3.161
14	89043	5.90	2.836	3.619	1.921	3.891	—	—	3.228	2.923
15	89189	5.60	—	—	—	3.562	3.326	2.609	2.823	—
16	89292	5.9	2.834	—	—	3.942	—	—	3.208	—

Magnitudes (m_b) and log rms Lg values for vertical components at stations NORSAR, WMQ, HIA, ARU, GAR, KIV, OBN, and GAM for 16 explosions analyzed in this study. Note that the IRIS stations (ARU, GAR, KIV, and OBN) have been normalized to a constant gain level to adjust for response changes. The values for the three stations (WMQ, HIA, and GAM) reflect unadjusted count values of the raw seismograms.

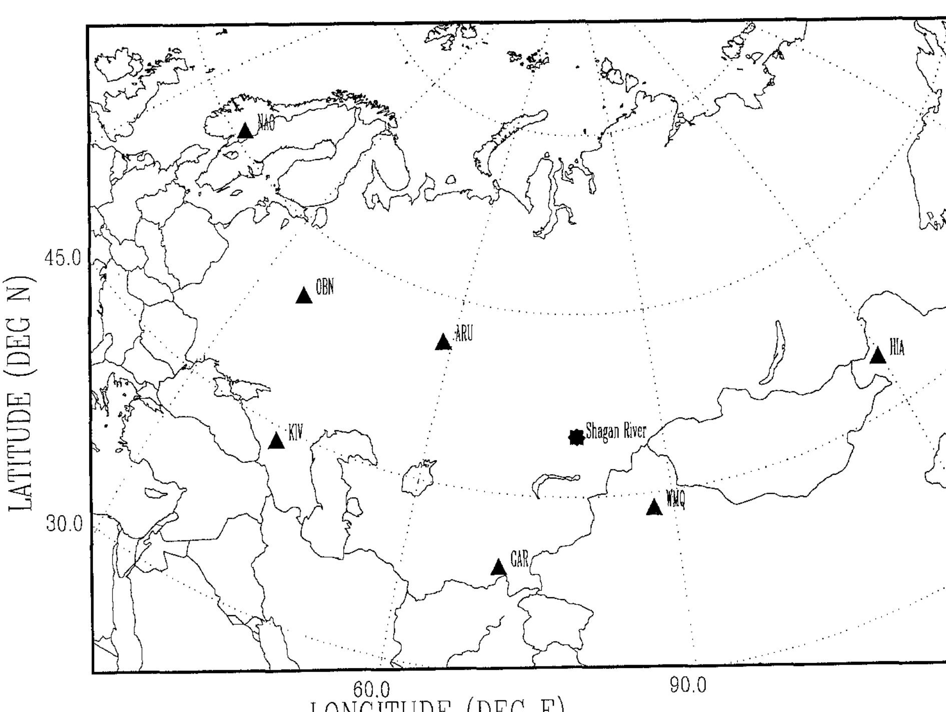
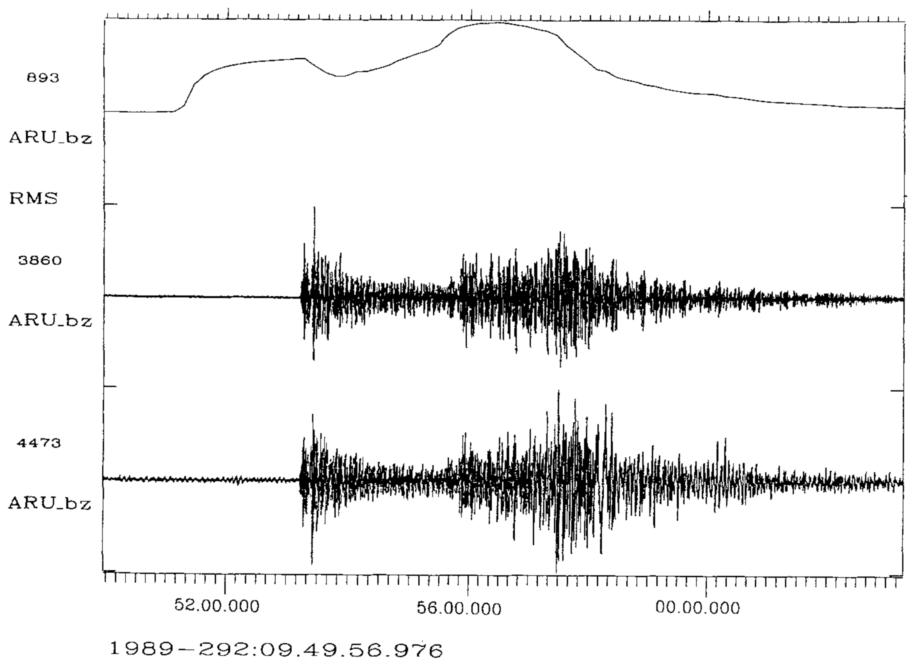


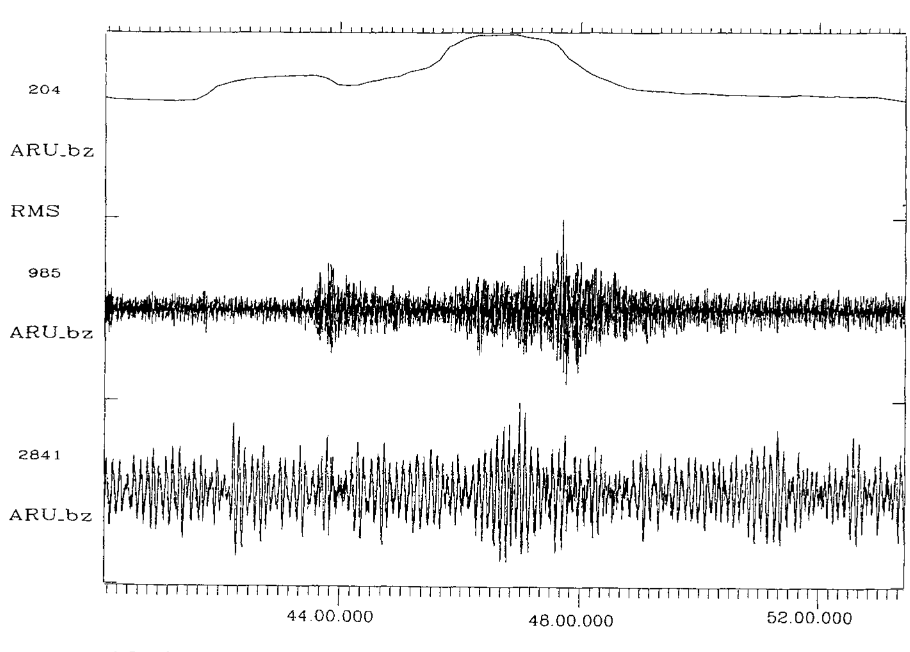
FIG. 1. Map indicating the locations of the Shagan River Test Site, the IRIS and British stations in the USSR, the NORSAR array in Norway, and the stations WMQ and HIA in China. The NORESS array is collocated near the NORSAR array, and station GAM is collocated near the GAR station.

m_b 5.9 Event at Semipalatinsk



(a)

m_b 4.9 Event at Semipalatinsk



(b)

FIG. 2. Example of recordings from two Soviet nuclear explosions at the IRIS station ARU. (a) An m_b 5.9 event at Shagan River on 19 October 1989 illustrating a good SNR, and (b) an m_b 4.9 event at Degelen Mountains to illustrate the improvement in SNR by bandpass-filtering in the range 0.6 to 3.0 Hz. For each of the events, we show the unfiltered trace (bottom), the filtered trace (0.6 to 3.0 Hz) (middle), and the 120-sec window rms measure (top) as a function of time.

Figure 2b presents an event of magnitude $m_b(P) = 4.9$, each as recorded at station ARU. The bottom trace for each event in Figure 2 is the observed data. These seismograms illustrate the broadband character of the typical recordings from modern digital seismometers, where the response is flat from about 5 Hz to well below the frequencies of interest for *Lg* waves (to between 30 and 100 sec period for these stations). We first bandpass filter the seismograms shown in the bottom trace in the frequency band from 0.6 to 3 Hz to produce the bandpassed version in the center of each plot. This is clearly necessary to enhance the *Lg* waves relative to the long-period microseisms in Figure 2b and higher frequency *P* and *Sn* coda, as well as to allow comparison to analyses of short-period data.

An rms trace, shown on the top of each plot, is then computed where each point of the trace represents the rms amplitude measure for the subsequent time window. We then measure the rms amplitude for the window centered on the phase of interest. In this respect, we did not use a fixed-group velocity window for analysis, but rather for simplicity, the same length window of 120 sec was chosen for all distances and centered near the 3.5 km/sec group velocity arrival time. The rms measure of *Lg* was read for the particular 120 sec window for all recording stations (and individually for all components of recording). Again for simplicity, the largest value of the rms trace was chosen as the amplitude measurement as long as the window is still centered near the 3.5 km/sec group velocity. Likewise, an rms measurement of the noise preceding each event arrival was calculated and applied as a correction term for calculating the *Lg* amplitude measure as originally defined by Ringdal and Hokland (1987). In contrast to NORSAR, the Soviet and Chinese stations are single site stations, so no averaging of vertical component measures was possible. However, these stations do record three components that may be averaged. We thus computed both individual component rms data as well as average values, but our results were inconclusive as to whether reduced scatter could be achieved in this way. In this paper, we present results based on vertical components only.

Examples of the IRIS recordings are shown in Figure 3 for the JVE event of 14 September 1988. Again, in this figure, are the unfiltered three-component data along with bandpass-filtered versions in the frequency range from 0.6 to 3.0 Hz. Above each filtered trace, we show a 120 sec window rms measure of the amplitude. The first striking feature of the three-component seismograms is that the horizontal instruments consistently exhibit a larger amplitude for the *Lg* phase than the verticals. The closer stations, ARU and GAR, at a distance near 1,500 km, show this *Lg* phase as the largest amplitude, while stations OBN and KIV at a distance nearer to 2,900 and 2,800 km, respectively, have the *P* phase as the largest amplitude. Station KIV has no discernible *Lg* phase for this explosion, presumably because *Lg* does not propagate efficiently in the crustal structure associated with the Caspian Sea.

The CDSN stations at WMQ and HIA also show excellent *Lg* recordings of Semipalatinsk explosions, as illustrated by the examples in Figure 4. Note in particular the dominance of the *Lg* phase at HIA as the largest recorded phase even at the distance of 2,900 km for this azimuth.

Figure 5 compares the signal-to-noise ratios (SNRs) (defined as rms *Lg* signal to pre-*P* rms noise in the 0.6 to 3.0 Hz band) for stations at various distances, using five large explosions. The range in magnitude (m_b) is from 5.2 for the event on day 317 of 1988 to 6.1 for the JVE event on day 258 of 1988. The event on day 317 indicates the minimum for which rms *Lg* was measured at NORSAR at a

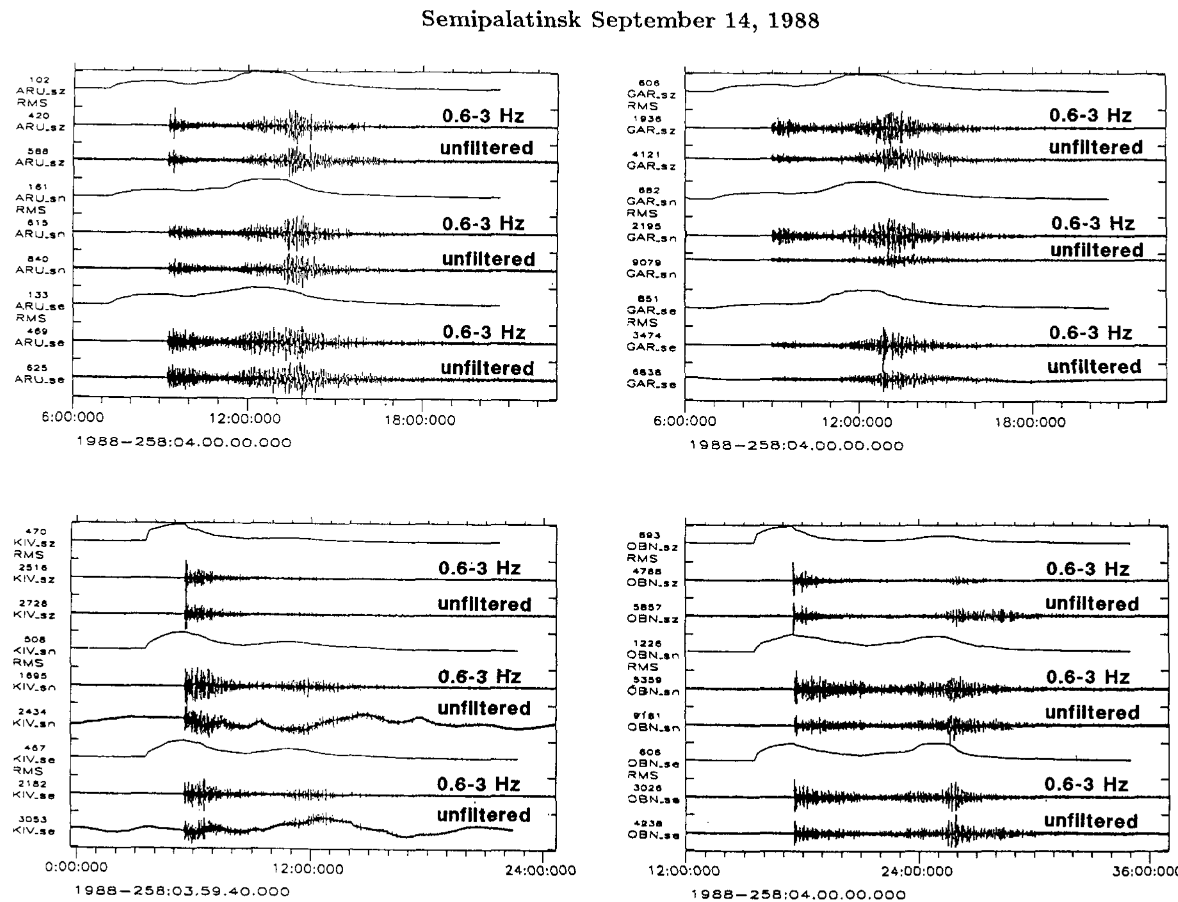


FIG. 3. Plots of the data recorded on the four IRIS stations located in the USSR for the Soviet JVE explosion of 14 September 1988. For each of three components at each site, we show the unfiltered trace (*bottom*), a filtered version in the band of 0.6 to 3.0 Hz (*middle*), and the 120-sec window rms amplitude measure (*top*) as a function of time.

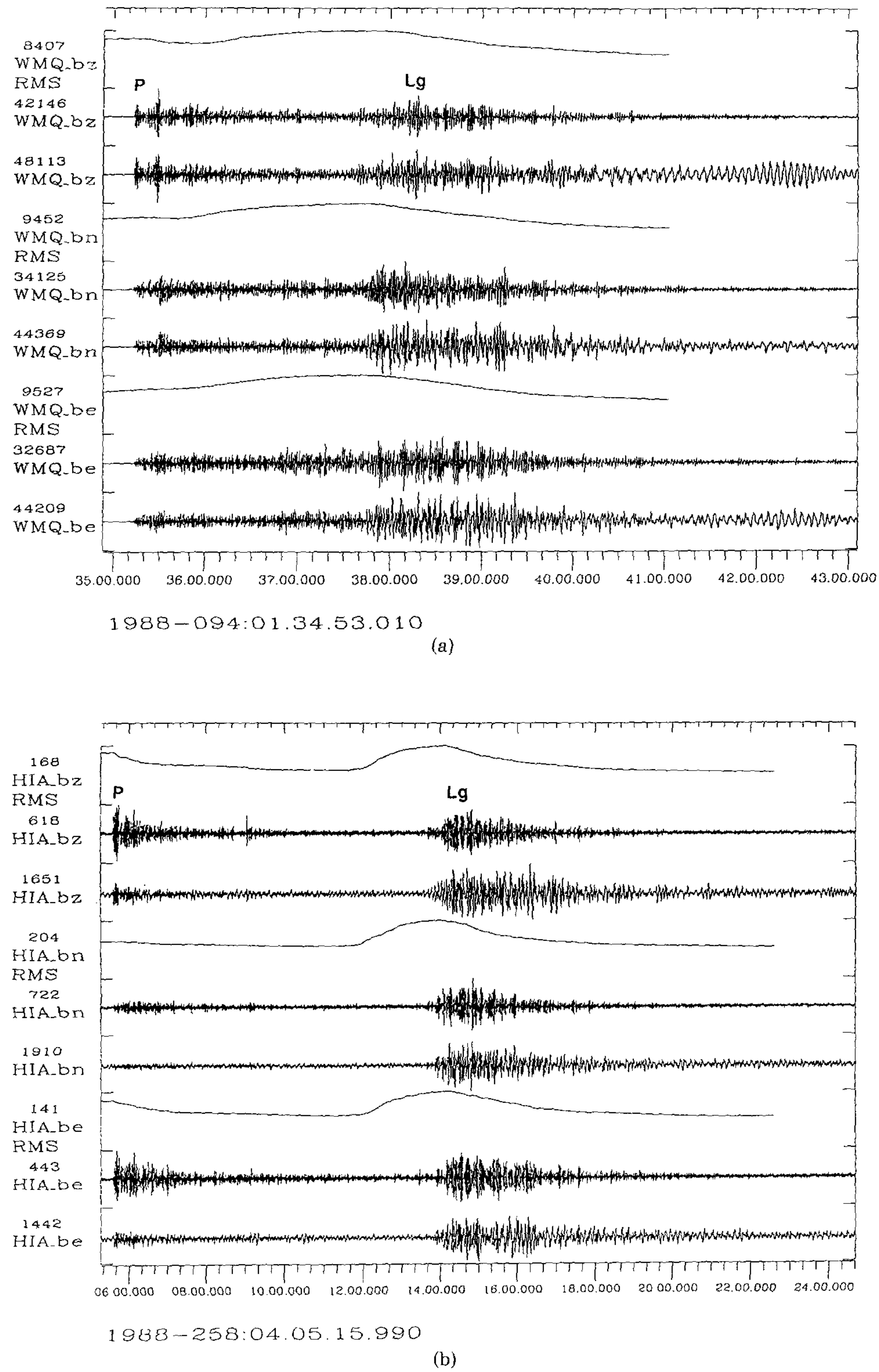


FIG. 4. Example of recordings from two Soviet nuclear explosions at the two CDSN stations. (a) 3 April 1988 at station WMQ and (b) 14 September 1988 at station HIA. For each of the three components, we show the unfiltered trace (*bottom*), the filtered trace (0.6 to 3.0 Hz) (*middle*), and the 120-sec window rms measure (*top*) as a function of time.

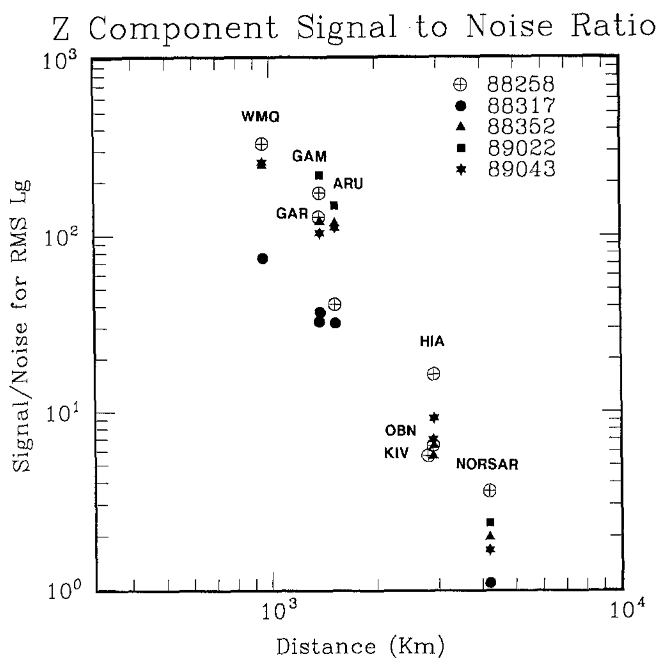


FIG. 5. Graph showing the variation of the SNRs (log rms *Lg* over log rms noise) among the four IRIS stations, the NORSAR array, the CDSN stations WMQ and HIA, and GAM. Epicentral distance to the Shagan River Test Site is plotted along the horizontal axis.

distance of about 4,200 km with an SNR of about 1.1. For this same event, an SNR of about 30 is observable at ARU and GAR at a distance of about 1,500 km and about 80 at WMQ at a distance of 950 km. Again, the event at day 258 of 1988 in Figure 5 (shown with the open circle around a plus sign) shows an SNR gain of nearly 100 between NORSAR with an SNR of 3.5 and WMQ with an SNR of 331. (It should be noted that the low SNR for this event at ARU is because this event was only recorded on the low-gain channel, which does not adequately resolve the background noise.) It is noteworthy that WMQ shows the best SNR for all the events. The figure suggests that WMQ, if set to record continuously, would be able to give *Lg* measurements for events close to two magnitude units smaller than the NORSAR threshold of approximately 5.5. Unfortunately, there were no low-magnitude events for WMQ in our data base, so we have not been able to confirm this hypothesis. We do, however, show an example of an $m_b(P)$ 3.8 explosion, whose *Lg* signal was recorded by ARU (see below).

In order to investigate the stability of the rms *Lg* amplitudes observed at the Soviet and Chinese stations, the amplitudes were compared with NORSAR amplitudes for common events. Since the instrument response of the different IRIS stations was changed several times, and was different at different stations (each being different from that of a NORSAR station), we decided to convert all measurements of IRIS stations to the equivalent gain of a typical short-period instrument in the 0.6 to 3 Hz range. The CDSN stations and station GAM had a constant gain throughout the recording period of this study, so no gain adjustment was required.

For comparison of actual measurements of rms *Lg* amplitudes between NORSAR and four of the new stations (ARU, GAM, WMQ, and HIA) for all common events, we plot in Figure 6 data for the vertical components of rms *Lg*. A straight line has been fit to the data for each of the four stations and a measure of the misfit is given

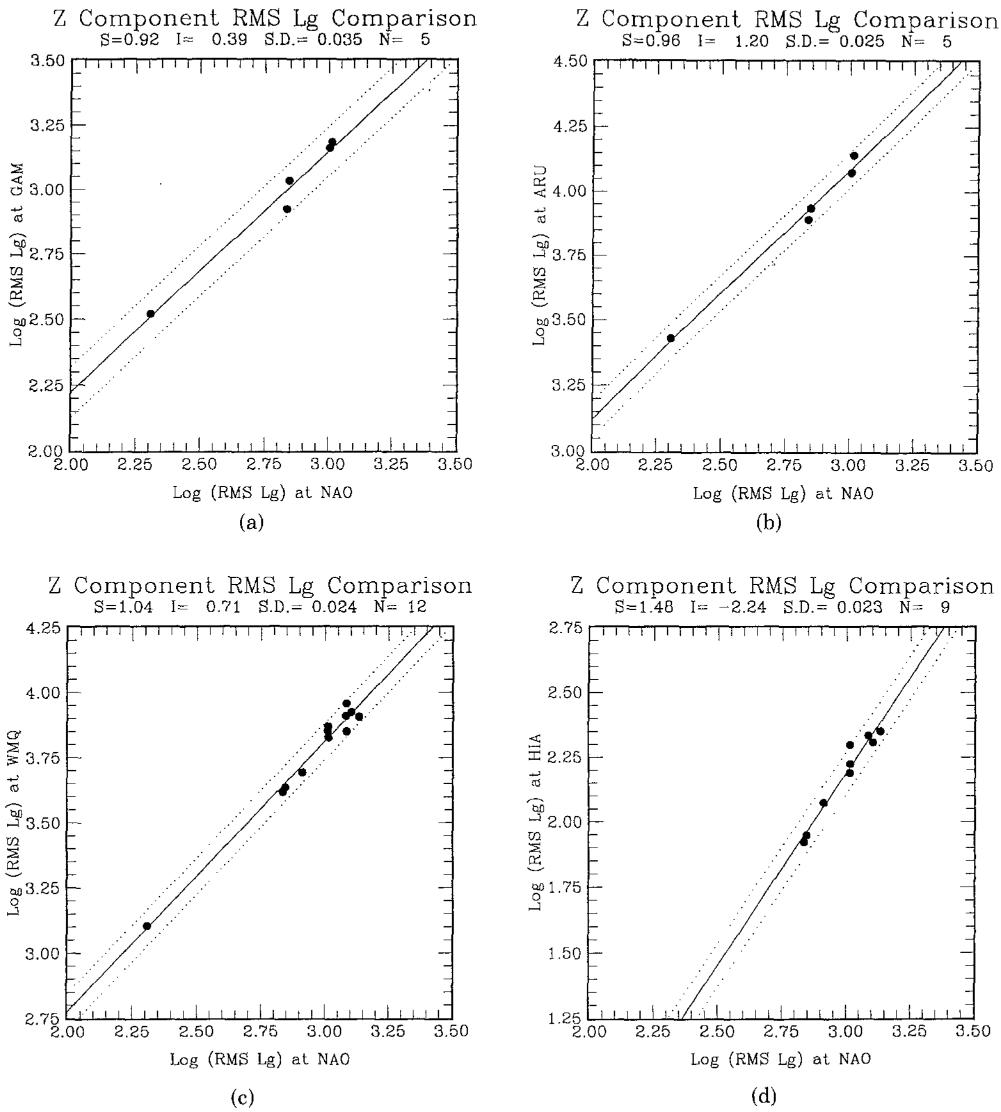


FIG. 6. Comparison of log rms *Lg* at NORSAR with rms *Lg* measurements obtained at four of the close-in stations. (a) GAM with a fitted slope of 0.92 and an orthogonal rms misfit of 0.035 magnitude units, (b) ARU with 0.96 and 0.022, (c) WMQ with 1.03 and 0.024, and (d) HIA with 1.48 and 0.023. The dotted lines correspond to ± 2 S.D.

by an orthogonal standard deviation (dotted line in figure corresponds to two standard deviations).

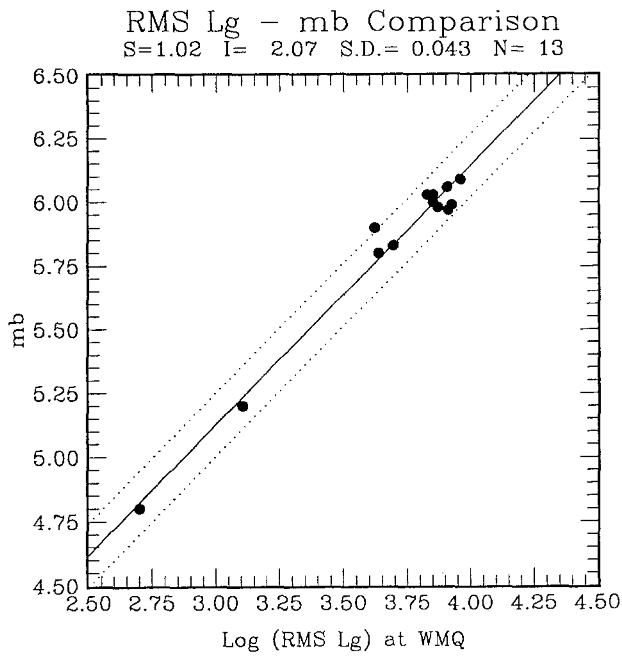
Figure 6 (a to c) shows the comparison of GAM, ARU, and WMQ versus NORSAR log rms (*Lg*) estimates for all common events. The slopes of these plots are 0.92, 0.96, and 1.03, respectively, with orthogonal standard deviations of the misfits being only 0.035, 0.022, and 0.024 units.

Figure 6d shows a comparison of HIA and NORSAR log rms (*Lg*) estimates. In this case, the slope of the least-squares linear relationship (1.48) is significantly different from unity, and we note that a similar observation was also made by Ringdal and Marshall (1989) when comparing NORSAR and Graefenberg *Lg*. We will not go into any detail discussing possible underlying physical reasons for this variability in slopes. For our purpose, the important point is to note that the scatter of the relationship is still very small; the orthogonal standard deviation relative to the straight line fit is 0.023, which compares very closely to the results found for the other station pairs. Although not shown on the figure, the fit between HIA versus WMQ log rms (*Lg*) values again gives a least-squares slope (1.36) that is significantly different from unity. Once more, however, the scatter is very small, with an orthogonal standard deviation of 0.028 units. We thus find essentially the same scatter for all data when comparing different station pairs, and this confirms the excellent stability of the rms *Lg* estimates when considering a suite of explosions within the limited source region of the Shagan River area.

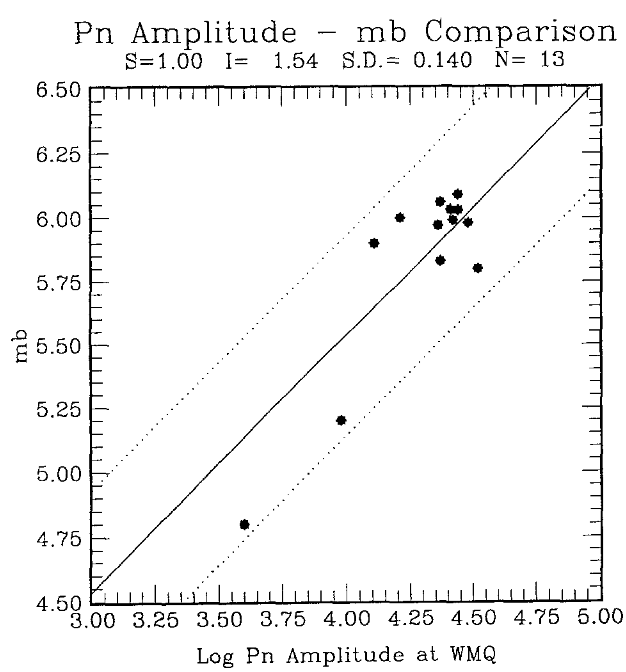
In Figure 7a, we plot the rms *Lg* amplitude at WMQ against worldwide $m_b(P)$ magnitudes for all recorded events at Shagan River. The slope is 1.02 and the orthogonal standard deviation is 0.044. This scatter is also quite small, but it must be noted that only one event from the northeast part of Shagan is in the data base. Thus, we cannot assess whether the $m_b(Lg)$ versus $m_b(P)$ bias earlier found for this subregion (Ringdal and Marshall, 1989) is also present when measuring *Lg* at WMQ. For comparison, we have also plotted in Figure 7b the same $m_b(P)$ estimates against the logarithm of the largest *Pn* amplitude at WMQ measured within the first 5 sec of the first arrival. Here, we see a much larger scatter for the single station than for the rms *Lg* amplitudes. This is consistent with previous studies of teleseismic *P* at single stations. For example, Lilwall *et al.* (1988) found a typical standard deviation of 0.12 m_b units when comparing single station m_b to worldwide m_b for a set of Shagan River explosions.

Figure 8 illustrates the capabilities of the ARU station to record an $m_b(P)$ 3.8 event from the Shagan River Test Site on day 270 (September 26) of 1988. [This magnitude is based on the NORSAR $m_b(P)$ of 4.3 with an assumed regional correction of 0.5 units for comparison to worldwide m_b estimates, and therefore must be considered somewhat uncertain.] The unfiltered broadband trace at ARU essentially shows no signal for this event; however, the bandpass-filtered trace clearly shows energy arriving that can be identified as *Lg* with an SNR of about 2. (Similar SNR was obtained for the recording at GAM for this event.) This SNR is near the lower limit of about 1.5 for allowing reliable rms *Lg* estimates at a single site. In an attempt to enhance the detectability of other phases, the vertical component of ARU was filtered in several passbands. Even considering frequency bands up to the Nyquist frequency of 10 Hz, we found no additional enhancement of the *P* phase or other phases. (It may be noted that ARU is at a distance within a shadow zone for *P* waves from seismic sources in East Kazakhstan.) In comparison, the NORESS array is clearly capable of detecting the *P*-wave arrival with an SNR of nearly 30, as illustrated in Figure 9 and the ARCESS array also shows a clear *P* detection for this event. Thus, even though the ARU station may not be capable of detecting an event of this size in an automatic fashion, regional arrays such as NORESS and ARCESS can correctly detect the event while the analysis of the *Lg* phase at a much closer station can provide an estimate of the rms *Lg* magnitude suitable for giving independent information on explosion yield.

Figure 10 illustrates the stability of the rms *Lg* amplitudes by comparing GAM and ARU. These stations are chosen because they are the only pair for which we have *Lg* recordings of the $m_b(P)$ 3.8 event shown in Figure 8 and so illustrate the stability of measurement covering a span of two full magnitude units. Here, we again have a slope of very nearly one still with an orthogonal standard deviation of only 0.026 logarithmic units (i.e., magnitude units).



(a)



(b)

FIG. 7. (a) Comparison of log rms Lg at WMQ to worldwide m_b magnitude. Standard deviation is 0.044 orthogonal to the line. (b) Plot showing the WMQ log Pn amplitude measured within the first 5 sec of the Pn arrival against worldwide m_b. The slope of the straight line has been fixed to 1.0. The orthogonal standard deviation is 0.140. The dotted lines correspond to ±2 S.D.

Semipalatinsk September 26, 1988

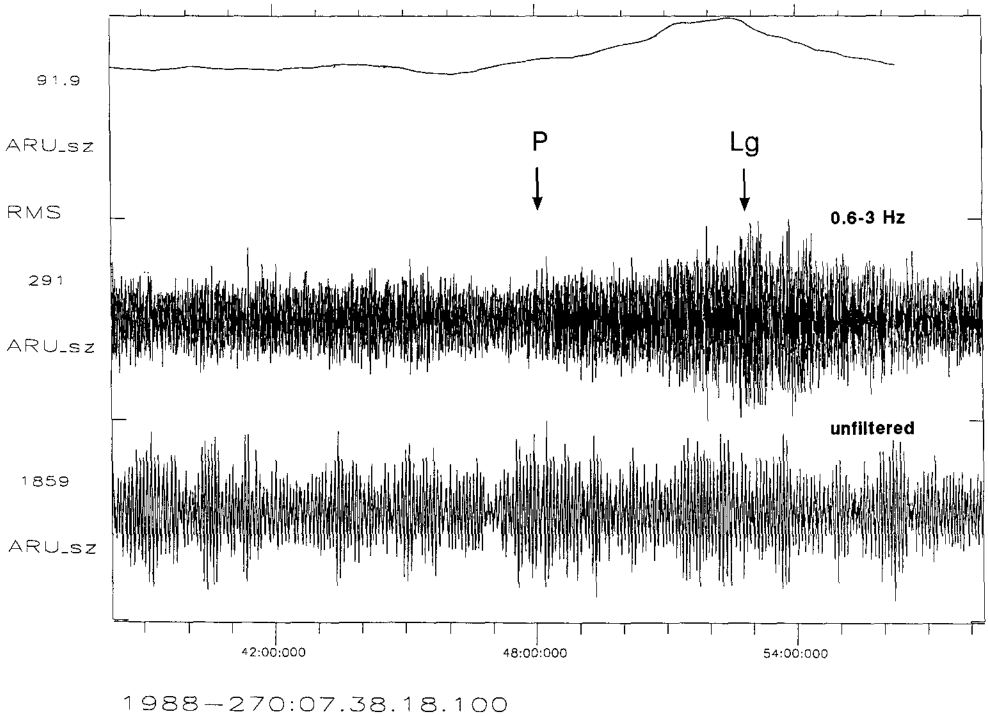


FIG. 8. The ARU vertical component seismogram for the m_b 3.8 explosion on 26 September 1988. The lower trace is the unfiltered seismogram, the middle trace is the bandpass-filtered seismogram between 0.6 and 3.0 Hz, and the upper trace is the rms amplitude as a function of time.

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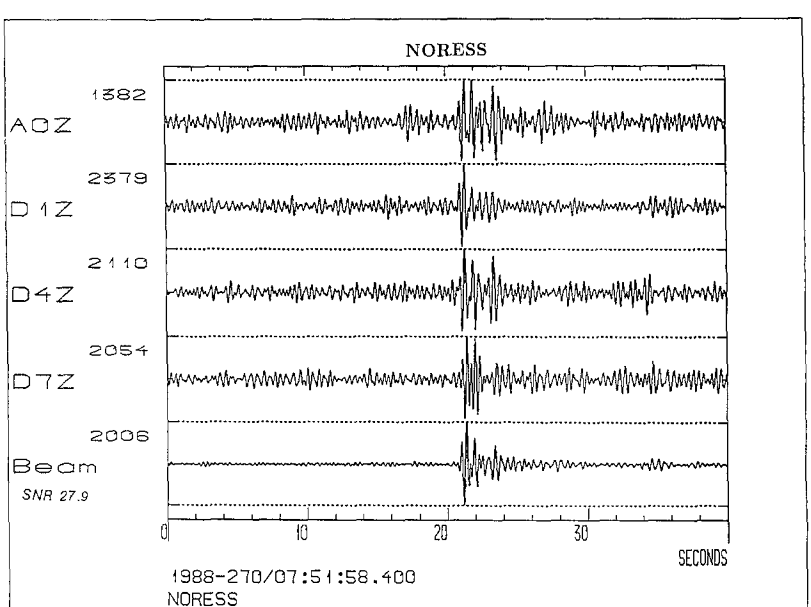

FIG. 9. Example of four vertical component seismograms from the NORESS array in Norway for the m_b 3.8 explosion on 26 September 1988. Shown on the bottom trace is the beam formed by steering toward the explosion site. Note the large improvement in SNR on the beam.

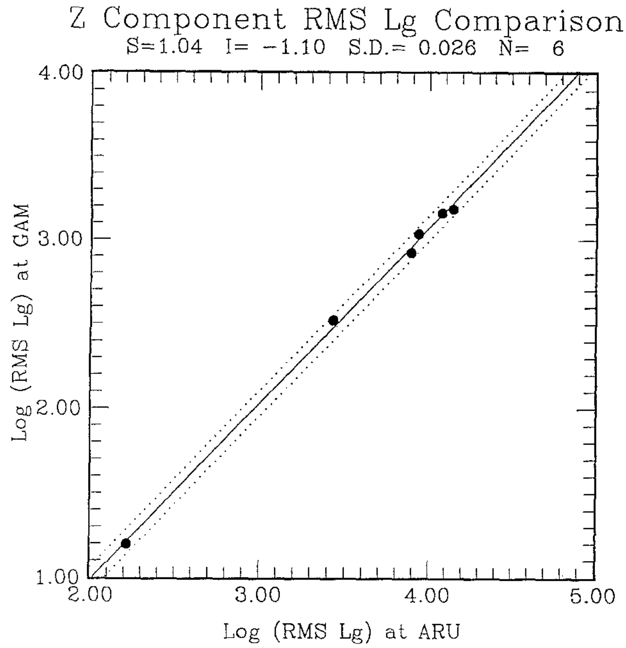

FIG. 10. Comparison of log rms *Lg* measurements at ARU and GAM. The slope of the line is 1.04, and the standard deviation of the misfit of the line to the data is 0.026 orthogonal to the line. The dotted lines correspond to ±2 S.D. Note the remarkable stability of measurement between the two stations over two full magnitude units.

DISCUSSION

A heuristic explanation for the superior stability of *Lg*, as compared to the stability of *P*, lies in the difference in the nature of the sampling of the seismic source for each of these phases. *P* waves for each source-station pair sample only a very limited portion of the focal sphere and are susceptible to focusing and defocusing. To get an improved average using *P* waves, it is necessary to use many stations around the globe and even when using a teleseismic network, only a relatively small part of the focal sphere will be sampled. But *Lg* waves are composed (for each source-station pair) of multiple rays that sample a larger portion of the focal sphere, and therefore, the Earth is doing the averaging for us.

In demonstrating that a single station can provide rms *Lg* measurements with a precision (one standard deviation) of about 0.03 magnitude units at Shagan River, we note that several issues are raised in considering how best to use such measurements for yield estimation.

For example, there are general questions concerning how to define $m_b(Lg)$: can we carefully define an $m_b(Lg)$ scale that is indeed a property of seismic sources, and then establish a procedure by which m_b on this scale can be estimated by measurements made with one or more stations in a seismograph network? One way to proceed would be to define $m_b(Lg)$ as the measurement made in a particular way with a particular seismographic network. The $m_b(Lg)$ for a particular seismic event could then be directly measured (to the extent that the full network supplied data) or, instead, estimated if only a subset of the data were available (e.g., from only a limited number of stations).

Fortunately, in many projects in which a suite of seismic events is under study, an accurate estimate of absolute $m_b(Lg)$ values is not needed. Rather, one may only need estimates of the relative $m_b(Lg)$ values. The key quality needed is precision of measurement; absolute levels are unimportant or may be derived from separate information. This is the situation in making yield estimates based on seismic data for a suite of underground nuclear explosions at a particular test site, if independent (perhaps nonseismic) information on the yield of some of the events is made available. This information can be used to calibrate in absolute terms a seismic amplitude scale that may be defined uniquely for a particular source region and for a particular network of stations. In this context, in claiming that the stability of rms *Lg* is excellent, we mean that relative magnitudes of explosions in the same region can be estimated very accurately from one or two stations that record *Lg* if the SNR level is high enough.

However, for other purposes, we recognize that there is a need to work with absolute rather than relative $m_b(Lg)$ values. For sources and receivers at any location on the same continent (*Lg* does not propagate across oceans), the need eventually is to understand how to make path corrections to rms *Lg* measurements, for purposes of assigning $m_b(Lg)$ as a characteristic directly of source strength. It is clear that such corrections will depend on both source and receiver locations, and not merely on the scalar distance. (As noted, Nuttli did begin the process of making specific path corrections by making a correction for Q effects.) Obtaining accurate path corrections depending on four spatial coordinates (depth is a separate issue), whether determined empirically for each path or by predictions based on data from a coarse grid of sources and receviers, is certain to be a complex procedure. However, it is likely, too, to be associated with discovery of much new information about continental crustal structure. Our point here is that the precision of rms *Lg* measurements presents new challenges and new opportunities.

Assigning absolute levels of $m_b(Lg)$ for nuclear explosions at a fixed test site and for a fixed network is a far simpler task, one that we have addressed in this paper without special comment. While we have not discussed the problem of converting rms *Lg* to a magnitude value, this is a relatively straightforward task, implying calibration to a given magnitude scale. Presuming that magnitude in this sense, and yield, are related at the test site by a best-fitting line in the form

$$m_b(Lg) = a + b \cdot \log(\text{yield}),$$

it is clear that the scatter of points about this line is controlled by two factors. One is the precision with which $m_b(Lg)$ can be measured (e.g., 0.03 at a single station, as shown in this paper for the Shagan River area). The second is the additional uncertainty caused by variability of coupling from nuclear yield into *Lg* signal, a key issue that at present we are not in a position to resolve.

Assistance in addressing the second issue would come from open availability of yield information for some explosions at test sites of interest, preferably for the same explosions whose seismic signals were recorded at high-quality digital stations. We note that yields are not currently announced at the world's two main test sites (the Nevada Test Site and Shagan River), and yields announced at these sites for explosions in the past (Springer and Kinnaman, 1971, 1975; Bocharov *et al.*, 1989; Vergino, 1989a, b) were for the period prior to 1973 when few digital stations were in operation. However, preliminary indications, from a study of the four Semipalatinsk explosions for which there is both an announced yield from Bocharov *et al.* (1989) and an rms *Lg* signal measurement at NORSAR, are that rms *Lg* correlates well with log (announced yield) (Ringdal, 1989). This comparison can be used, for example, to estimate the yield of the Joint Verification Experiment (JVE) explosion, conducted at Shagan River on 14 September 1988. From NORSAR *Lg* signals alone, the resulting estimate would be about 110 kt. As yet, the yield of this explosion, as determined from nonseismic measurements made on site, has not been announced, other than that it met the provisions of the JVE agreement between the United States and the USSR, and thus that it was indeed between 100 and 150 kt (Robinson, 1989).

An important advantage of the rms *Lg* method is its ease of use in combination with the robustness of the results. Thus, it makes essentially no difference where one uses a 2-min window or one based on a range of *Lg* group velocities (which would give a window about 40 sec at ARU for the range of group velocity used in NORSAR analyses). Also, the choice of filter band is not critical as long as the band enhances the main part of the *Lg* energy and is kept fixed in the analysis of different events. Our choice of a 0.6 to 3.0 Hz passband has been made in order to be consistent with previous NORSAR analyses.

Conclusions

This study has demonstrated that rms *Lg* amplitudes estimated from stations within the Soviet Union and China for Shagan River explosions show excellent consistency with NORSAR rms *Lg* estimates. This has several important implications:

1. rms *Lg* appears to be a stable source size estimator when computed at widely distributed stations and would therefore provide a reliable magnitude estimate once the proper correction term has been estimated for each station.
2. The stations studied (notably ARU, GAM, and WMQ) can be used to estimate *Lg* magnitudes for Shagan River explosions of much lower yield than is possible using the more distant NORSAR and Graefenberg arrays. Our analysis indicates that the SNR improvement allows rms *Lg* estimates to be made down to approximately m_b 3.5 at WMQ, compared to a threshold of about m_b 5.5 at NORSAR. An important precondition for WMQ is that it be set to provide continuous recording, rather than the triggered recording currently used.
3. Although single stations do not offer the increased stability obtained through array averaging, this is partly compensated by the higher SNR ratio, which means that modest noise fluctuations will be insignificant for the *Lg* measurements. Also, a possibility of decreasing scatter of magnitude estimates through averaging the three components of each station exists. Our initial analysis indicates that such an approach could be useful, but it may be necessary to determine correction terms for each component individually.
4. As more data (and possible additional stations) become available, a data base will be developed that will enable us to compute network averages, based on individual station data "calibrated" to NORSAR $m_b(Lg)$. This would facilitate both obtaining improved uncertainties of future explosions and maintaining a comparison to historic data. The calibration would best be done using direct, independent yield information, thus permitting reduced uncertainties in yield estimation (using seismic methods) for future explosions.
5. The *P*-wave detection capabilities of these single stations do not match those of the NORESS and ARCESS arrays; thus, teleseismic signals continue to be important for detection of small nuclear explosions.

It would be desirable to develop a theoretical basis to allow correction for attenuation of the *Lg* phase. Extension of the study to other nuclear explosion sites will also be an important topic. Of particular interest here is to study further the possible differences between the Shagan River and Degelen Mountains region.

Our studies confirm that *Lg* magnitude estimates of Semipalatinsk explosions are remarkably consistent between stations widely distributed in epicentral distance and azimuth. It thus appears that a single station with good SNR can provide $m_b(Lg)$ measurements with an accuracy (one standard deviation) of about 0.03 magnitude units. Therefore, *Lg* signals appear to provide an excellent basis for supplying estimates of the yields of nuclear explosions even down to below 1 kt, when such signals are recorded at high-quality digital, in-country seismic stations, and when calibrated by access to independent (nonseismic) yield information for a few nuclear explosions at the test sites of interest. In the context of monitoring a low-yield threshold test ban treaty, it will, in addition, be important to take into consideration various environmental conditions in the testing area, such as the possible presence of cavities, and to devise appropriate procedures for on-site observations in this regard.

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