

Topic	The Russian K-class system, its relationships to magnitudes and its potential for future development and application
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1 Abstract

The paper aims at reviewing the development and use of the energy class (K-class) system since the late 1950s in the former Soviet Union (FSU) for quantifying the size of local and regional earthquakes, at comparing its results with current full-fledged estimations of released seismic energy and energy magnitude via the integration of teleseismic broadband velocity P-waveforms and at assessing the potential of modernized and standardized K-class determinations as a valuable complement to current teleseismic procedures. K-class in its original preliminary version was proposed by Victor Bune. His version was radically improved by Tatiana Rautian to constitute a rapid and simple means of estimating the radiated seismic energy E_S from an earthquake. K-class was defined as $K = \log_{10} E_S$ with E_S in units of Joules. In practice, a graphical method was used based on the maximum horizontal (for S-wave) and vertical (for P-wave) amplitudes which was calibrated to independently determined energy estimates in Soviet Central Asia, and was used directly or with modifications as a standard throughout the FSU. The empirical relationships between K and classical magnitude scales vary somewhat between areas, but are, on the whole, consistent with magnitude-energy relationships proposed in the past fifty years, as well as with energy magnitude M_e . We show that the concept of the K-class can continue to be applied, with modifications, through the use of modern recording, filtering and data analysis tools and applied to regions outside the FSU as well as to the extensive historical data set from the FSU.

2 The development of the *K*-class system and methodology

After the devastating magnitude 7.4 Khait earthquake of 1949, the Complex (Interdisciplinary) Seismological Expedition was deployed by the former Geophysical Institute of the Soviet Union (USSR) in the Garm region of Tadjikistan to study its aftershocks and related phenomena. No specifications for seismic stations, procedures for measurements, standards for data processing, or necessary documentation had been established at that time. Thus, members of the expedition developed their own procedures. Due to the large number of events and limited computational and data analysis tools available, relatively simple, primarily graphical, methods were required.

In its original preliminary version, "Energy class K" was proposed by Bune (1955) with the intention to create a tool for absolute source energy calibration. It was based on integration of squared trace and theoretically would be a big advance with respect to the determination of magnitude *M*. Yet, in practice, it was unworkable with its hand-made integration on band-limited photographic records; it also used an imperfect way to reduce recorded wave energy to *E_s*. To make the *K* scale workable in mass processing, Rautian (1958, 1960) made a critical step by returning to the measurement of peak-amplitude instead of integrating squared velocity amplitudes, thus developing a method to estimate earthquake energy that could be implemented by technical staff with only limited training, as summarized below. To calibrate her scale, Rautian performed hand integration of trace and *A/T* measurements in parallel for a large training data set. She also radically improved the source energy estimation procedure. A more detailed description of its development and calibration is presented in Rautian et al. (2007).

If seismic energy (*E_S*) radiates uniformly, then at very short distances in a homogeneous Earth, $E_S = 4\pi r^2 k\varepsilon$; where ε is the total energy density that crosses normal to a surface of a circular wave front with radius, *r*, per unit area, and *k* is a coefficient which accounts for a diversity of effects such as the Earth's surface topography and the ratio of the measured component to the full displacement vector. Since the short-period seismographs deployed in early years in and around Garm recorded displacement both amplitude and frequency would need to be measured to estimate ε ; however, visual "spectral analysis" of a seismogram by technicians was impractical and resulted in considerable scatter.

Both the terminology and the analysis of waves used in developing the *K*-class system have some ambiguity as they were selected more for operational purposes. *A* stands for the displacement amplitude of a specific oscillation, in recent usage generally the first arrival, and *A_{max}* for the largest amplitude in the whole wave group belonging to a specific type of seismic phase. *T* refers to the period measured for a specific pulse with amplitude *A* or *A_{max}*, whichever is being referred to, within the duration, τ , of an arrival as measured from approximately the start to the end of oscillations with $A > A_{max}/2$. In practice, *A_{max}* was, and is still, used in calculating and calibrating *K*.

Generally, ε depends on the amplitudes, frequency content, and signal duration of the analyzed seismic phases; however, observations at Garm showed that signal duration was controlled primarily by epicentral distance (due to the multiple scattering of waves in the structurally complex Earth medium in this mountainous area), whereas the frequency content in a wave group was largely controlled by the energy released by the earthquake and by its hypocentral distance (Rautian, 1960), due to the energy and stress-drop dependent shift of the corner frequency of the radiated source spectrum on the one hand, and frequency-dependent

attenuation on the other hand (see Sections 3). Thus, by measuring amplitude, which was easy to measure, alone, it was found that energy could be estimated to within a factor of 2 to 3, with proper regional calibration (i.e., by accounting in a rough manner for the combined impact of diverse parameters that influence the measured values). Moreover, to reduce the scatter and smooth the effect of the source radiation pattern, the sum of the **maximum** amplitude of the P wave on the vertical component (A_P) and of the greater S-wave on one of the horizontal components (A_S) was chosen, with P being Pg and S being Sg at local and shorter regional distances (usually < 400 km), whereas at larger regional distances, the measured arrivals could be Pg and Lg. This yielded a pragmatic solution. Operationally, especially in the continental regions of eastern Russia, Pg is the commonly measured P phase with very few Pn amplitudes ever picked. At distances of greater than 700-800 km (or less in central Asia and Baikal), Sg (Lg) is often the only amplitude measured.

To calibrate the amplitude measurements to energy, Rautian (1960) estimated the energy of a large number of earthquakes, predominantly from the northern Garm region, by visually measuring frequencies and amplitudes on particularly clear records, and using many other simplifying assumptions. In brief, the energy density, ε , was determined by summing all visually separable $(Af)^2$, which is proportional to the ground motion velocity squared, over the measured pulse duration τ , where f is the measured frequency of a given cycle with displacement amplitude A . A correction, k , to account for frequency-dependent magnification of the seismograph, the use of only one record component instead of the complete vector, surface effects, etc., was estimated and applied. The dependence of $k\varepsilon$ on distance, r , was observationally determined and then used to normalize the estimate of ε to $r = 10$ km. Then, $\log E_S(10 \text{ km}) = \log 4\pi k\varepsilon$. Further, the amplitude measurements, $[A_P(r) + A_S(r)]$, were also normalized to $r = 10$ km using an amplitude-distance relationship.

The measurements were made on short-period instruments, initially of type VEGIK, with a seismometer eigenperiod $T_s = 0.8$ s, in common use in the 1950s and 1960s, later amended or replaced by Kirnos seismographs of type SKM and SK-III with slightly longer eigenperiod ($T_s = 1.5$ s and 2 s, respectively), followed by the more medium to long-period Kirnos seismographs of type SK ($T_s = 10$ s) and SKD ($T_s = 20$ s). The approximate average displacement response curves of all these seismographs, compared with the response of the short-period Benioff seismographs in common use in the United States in these years, all normalized to a maximum magnification of 1, are depicted in Figure 1.

Amplitude measurements were typically made at periods < 3 s, i.e. for crustal S waves with a wavelength typically less than 10 km. As stated above, the K scale was calibrated to the energy flow through a focal sphere of radius $r = 10$ km only. Such a sphere will, on the average, only enclose earthquake ruptures up to a magnitude of about 6. Therefore, the K-classification is principally suitable only for scaling local and regional earthquakes up to magnitudes of about 6 to 6.5 or $K \leq 14$ to 15 (see sections 4 and 5 on magnitude-K relationships). For larger earthquakes which extend far beyond the size of the reference sphere and which release their maximum of seismic energy at periods $\gg 3$ s the amount of released seismic energy will necessarily be underestimated (see Chapter 3, average seismic source spectrum and “saturation effect”).

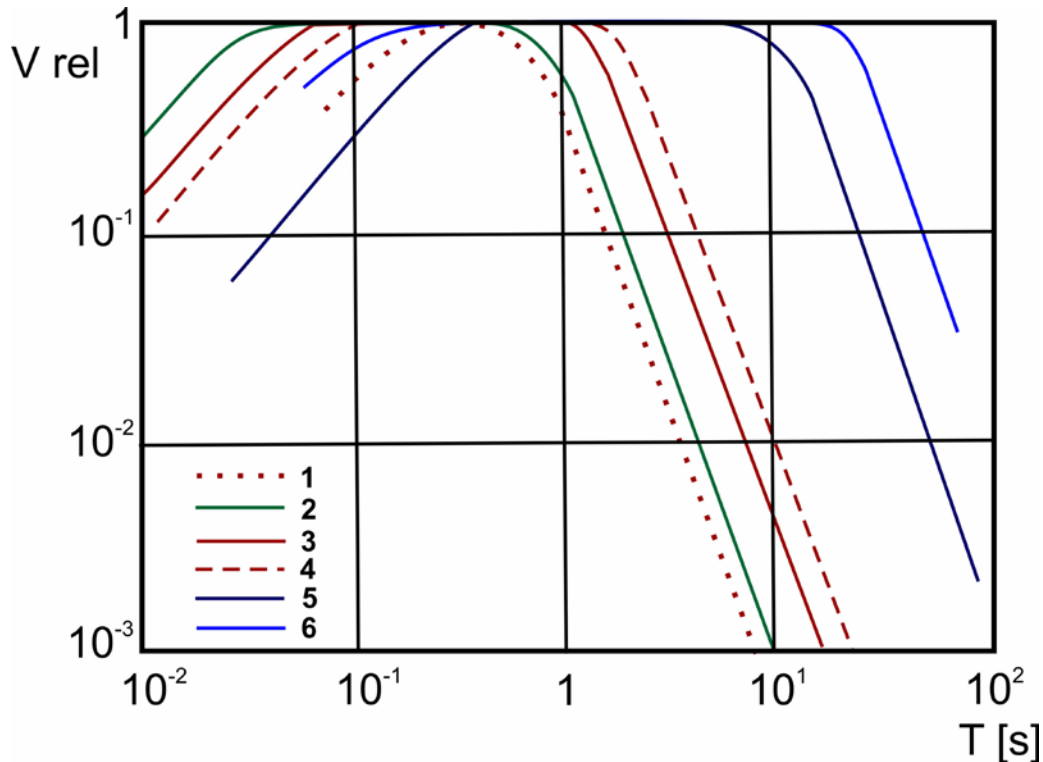


Figure 1 Normalized average displacement response curves of typical classical Soviet seismographs in comparison with the short-period US Benioff seismograph: **1** – Benioff; **2** – VEGIK with $T_s = 0.8$ s; **3** – SKM with $T_s = 1.5$ s; **4** – SK-III with $T_s = 2$ s; **5** – SK with $T = 10$ s and **6** – SKD with $T_s = 20$ s.

For amplitude measurements on VEGIK records, the following empirical calibration formula was derived:

$$\log E_S (\text{at } 10 \text{ km in Joules}) = 1.8 \log [A_P(10 \text{ km}) + A_S(10 \text{ km})] + 6.4. \quad (1)$$

In terms of trivial physics, formula (1) looks a bit strange, because in the case of wave propagation in a homogeneous medium one would expect $\log E_S \approx 2 \log A_{rms} + \log(\text{duration}) + \text{const}$. But, according to formula (1), the waveform duration would then be $\propto A^{-0.2}$, i.e., it would decrease with amplitude, contrary to average observation. Yet, empirically, in real Earth, deviation from an ideal coefficient 2 may be correct and have also been found in other empirical magnitude-energy relationships, e.g., the relationship published by Gutenberg and Richter (1956) between $\log E_S$ and body-wave magnitude m (with coefficient 2.2), and the one given by Richter (1958) between $\log E_S$ and surface-wave magnitude M_s (with 1.5).

Moreover, also jumping back from considering in the original design of the scale ε , i.e., the total energy density crossing the circular wave front with radius r per unit area, to using the peak amplitude recorded at seismic stations instead, is significant and conceptually problematic. But in practical terms this step was unavoidable with hand measurements and, therefore, does not deserve any criticism.

Formula (1) yields an amplitude sum, $A_P(10 \text{ km}) + A_S(10 \text{ km})$, of 100 microns for $\log E_S$ of 10. The shape of this calibration curve with distance was similar to that of Richter's (1935) local magnitude M_L but included deviations that were later explained as possibly being due to

refractions from the Moho or other arrivals, as well as to changing spectral content of waves with distance.

For simplicity in practical use, the log-distance scale was adjusted to make the amplitude-distance curves straight lines, as in Figure 2. Thus while the distance axis appears to be logarithmic, close examination of it shows that the horizontal scale was adjusted to incorporate the deviations.

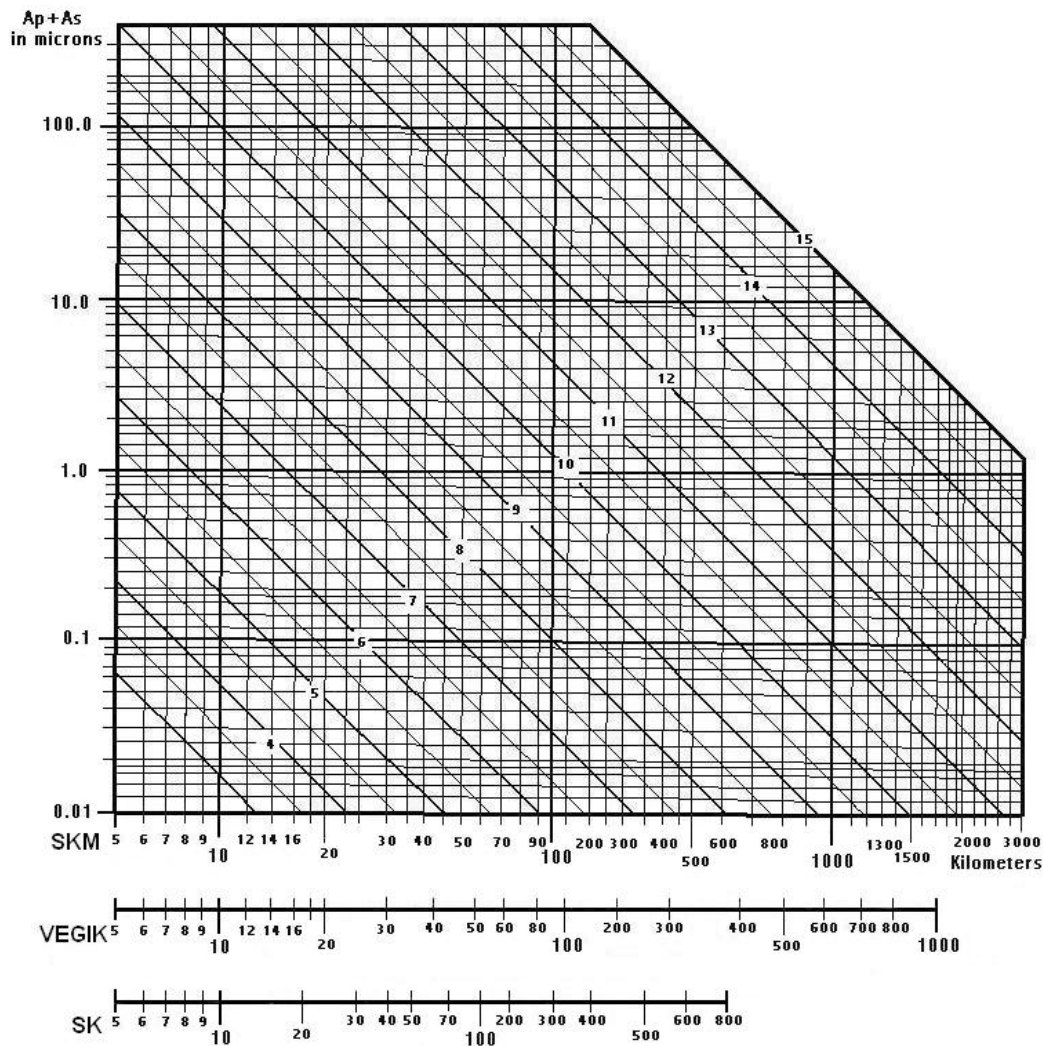


Figure 2 The Rautian (1964) K-class nomogram calibrated for SKM, VEGIK, and SK seismometers. Note that the horizontal distance scales are not truly logarithmic; see text for explanation. Vertical axis: sum of P and S-wave amplitudes in microns (modified from Figure 2 in Rautian et al., 2007, *Seism. Res. Lett.*, **78**(6), p. 582, © Seismological Society of America).

Since the calibration to E_s depends on the response of the seismograph, the calibration curves should have been adjusted as new instruments with different responses were deployed. However, the derivation of new calibration formulas for K for newly introduced instrumentation would have required a tremendous amount of empirical and theoretical work.

Moreover, the most commonly measured frequencies of local earthquakes (2-5 Hz) were such that the differences in instrument response did not appear to significantly change the results. Therefore, the same nomogram was used and only the distance scales were adjusted for different instruments (Rautian, 1964). Figure 2 shows the respective nomograms for calibrating the sum of P and S amplitude readings on short-period Kirnos SKM and VEGIK as well as on medium-period SK records, which continue to be used until now. The mismatch between the K-scales for different instruments reflects the fact that their passbands probe different parts of the source spectrum.

SKM seismographs became the main instrument for K-determination. They had typically a seismometer eigenperiod, T_s , around 1 to 2 s, a galvanometer period, T_g , around 0.3 s, and a damping of the seismometer and galvanometer chosen so as to produce a displacement-proportional response in the frequency range between some 0.5 and 20 Hz (see Figure 1).

Today, digital broadband seismograms could easily be accommodated by filtering such records to simulate the records with instrument responses of either the VEGIK, SKM or SK instruments. On Kamchatka, this is now routine procedure, i.e., converting BB digital velocity traces to 1.2s-instrument galvanometric traces of pre-digital era, and calculation of K values from S-waves. We are unaware, however, whether digital recordings in the FSU are nowadays routinely filtered everywhere before measuring parameters such as K – to emulate one of the historic instruments so as to remain compatible with the original definitions and baselines.

Because of the many simplifying assumptions made and uncertainties involved, the scale was called “energy class” instead of “energy” (see also below). Using the Russian spelling класс, the scale was defined as $K = \log E_s$, with E_s in Joules. The first outline of this scale was presented in Rautian (1958) and updated in Rautian (1960). The scale had its problems, but was easy to apply with the facilities and personnel of the time. K-class was adopted to quantify the size of earthquakes throughout the former USSR by 1961 and continues to be used today. Outside the USSR, K-class was used extensively only in Mongolia and in Cuba..

The standard deviation in K-class estimates between stations, due to site effects, station distribution with respect to the radiation pattern, heterogeneities in the crust, using only one horizontal component, etc., was empirically determined to be about 0.35 log units. No site corrections were originally used; however, they were later added by some networks.

In current practice, networks use only the maximum horizontal displacement amplitudes for K determination, usually at hypocentral distances up to about 600 km (see Figures. 2 and 4 to 6), since the amplitude of S_g (or L_g) dominates that of P_g . S_g and L_g are not differentiated and simply referred to as S_g , even at distances up to 1500 km where L_g is often still observed on the craton. These arrivals typically have dominant periods below 2.5 s. Operationally, the typical range of periods measured for determining K-class values today is 0.2 to 1.1 s. (see Figure 3). P-arrivals are seldom used in continental eastern Russia; S is used in place of S_g in the Kuriles and S and P are independently used on Kamchatka.

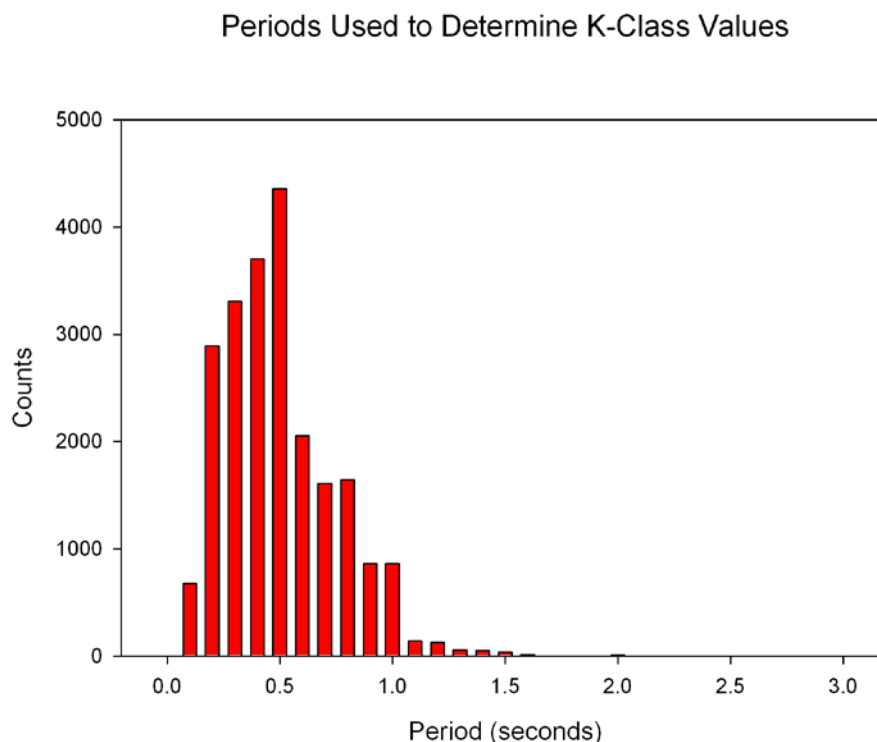


Figure 3 Average histogram of periods at which P and/or S-wave amplitudes have been measured by different local networks in the FSU for estimating the K values of mostly local earthquakes between 1978 and 2009.

In summary, K estimates, being based on short-period P- and S-wave amplitudes with periods less than 2 s, and wavelengths typically less than 10 km, are suitable only for scaling local and regional earthquakes with magnitudes up to 5.5 or 6 at most, corresponding to $K \approx 14$ to 15. For larger earthquakes, the amount of released seismic energy will necessarily be underestimated (“saturation”).

3 Regional variations of the K-class method

K-class (or more correct, K-magnitude) was adjusted regionally in the FSU to significant degrees. This is indispensable, as for other short-period local/regional magnitude scales applicable in different seismo-tectonic environments (e.g., ML and mb_Lg; see IS 3.3 and Table 2 in DS 3.1).

As noted, the routine K-class calculations ignored the specific spectral content of the individual displacement records, which was observed to differ randomly between events in the same area, and systematically between events in different tectonics settings. Thus, when calculated from displacement only, K is more closely related to the logarithm of the seismic moment, M_0 , than to $\log E_S$ as calculated by using velocity spectra.

The ratio between these two parameters may vary regionally, depending on the seismotectonic environments such as continental crust, subduction zones, deep seismic zones, oceanic ridges, etc. and thus also on the predominant type of source mechanism (e.g., Choy and Boatwright, 1995; Choy et al., 2006; Kanamori and Brodski, 2004; Bormann and Di

Giacomo, 2011). On a global scale the variability reaches about 3 orders of magnitude. It is mainly related to the differences in stress drop, or “apparent stress” (Wyss and Brune, 1968), but also to variable rigidity in the yielding earthquake source volume (e.g., Houston, 1999; Polet and Kanamori, 2000) and thus to “fault maturity” (see IS 3.5). In this context one should note that both stress drop and rigidity are considered to be constant in M_0 calculations.

Another useful related parameter is “earthquake hardness S_M ”. This term was originally introduced by N.V. Shebalin to describe the deviation of macroseismic intensity (see Chapter 12) from its average value at a given M_s , which are closely related to differences in the high-frequency content of the seismic wave field in the shaken areas. Thus, S_M is quite relevant for describing variations of the E_S/M_0 ratio too. The correspondence between K and M_0 was found on FSU territory to be best in crystalline rocks and regions of thrust tectonics ($S_M \approx 0$) and worst (significantly negative S_M) for strike-slip events within strongly fractured rocks with low-frequency source spectra. S_M was found to be significantly positive for intermediate-depth events.

With respect to the Kurile Islands, Fedotov used in 1963 S-wave amplitudes only and replaced the proper distance scale (as used, e.g., in Rautian, 1964) by S-P time (as in Figure 4). His scale was usable for earthquakes in a wide depth range of the Kurile Islands environment. The K-Fedotov63 was improved in 1968, and additionally an analog P-wave K-scale was proposed. Both Fedotov K-scales are still in use on Kamchatka, but their calibration curves and also the way of their absolute calibration are radically different from the Rautian K scale (for details see Fedotov, 1972). This applies, more or less, to other regional K-class formulas as well. Note that on Kamchatka it is routine practice to use besides the two Fedotov K scales also a coda-class Kc. The latter is very efficient when S-wave amplitudes are off scale on photographic records, or out of the dynamic range on a digital channel (see Lenzikov and Gusev, 1991).

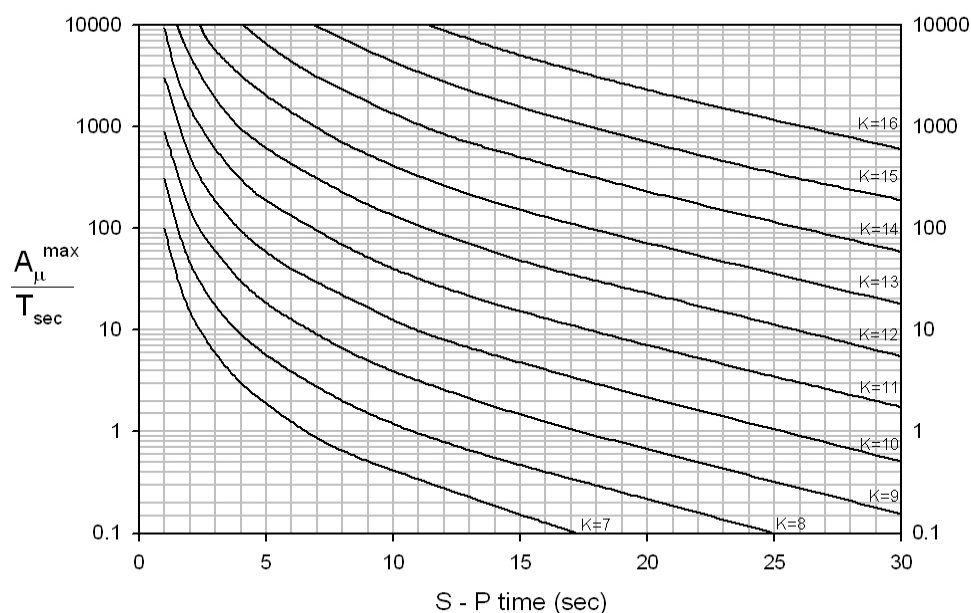


Figure 4 The 1967 Kurile nomogram for calibrating A/T with A measured in micron (μ) and T in seconds on SK records. The horizontal axis is linear in differential S-P travel time, which is proportional to hypocentral distance, in this diagram up to about 250 km (copied from Figure 5 in Rautian et al., 2007, *Seism. Res. Lett.*, **78**(6), p. 585, © Seismological Society of America).

The Kamchatka and Sakhalin regional networks are both located in subduction zone settings with very different velocity structure, L_g propagation, and a stronger attenuation, than in the continental regions. Thus, they developed their own K scales (see Rautian et al., 2007) based on using only A_{Smax} , but normalized by the period, T , and thus being a measure of the maximum S-wave ground motion velocity. The calibration techniques were similar and different calibration relationships were developed for shallow and for intermediate depth earthquakes. The Sakhalin network nomogram, used in the Kurile Islands for scaling SK records (see Figure 4), and Kamchatka nomograms for scaling SKM records (see Figure 5) were revised several times as both networks were expanded and more data was acquired. In the compilation by Kondorskaya and Shebalin (1977), the 1961 scale (Fedotov et al., 1964; K_F) and 1967 scale (Solov'ev and Solov'eva, 1967; K_S) were converted to Rautian (1960) K-class values using $K = K_F + 0.6$ and $K = K_S + 1.7$.

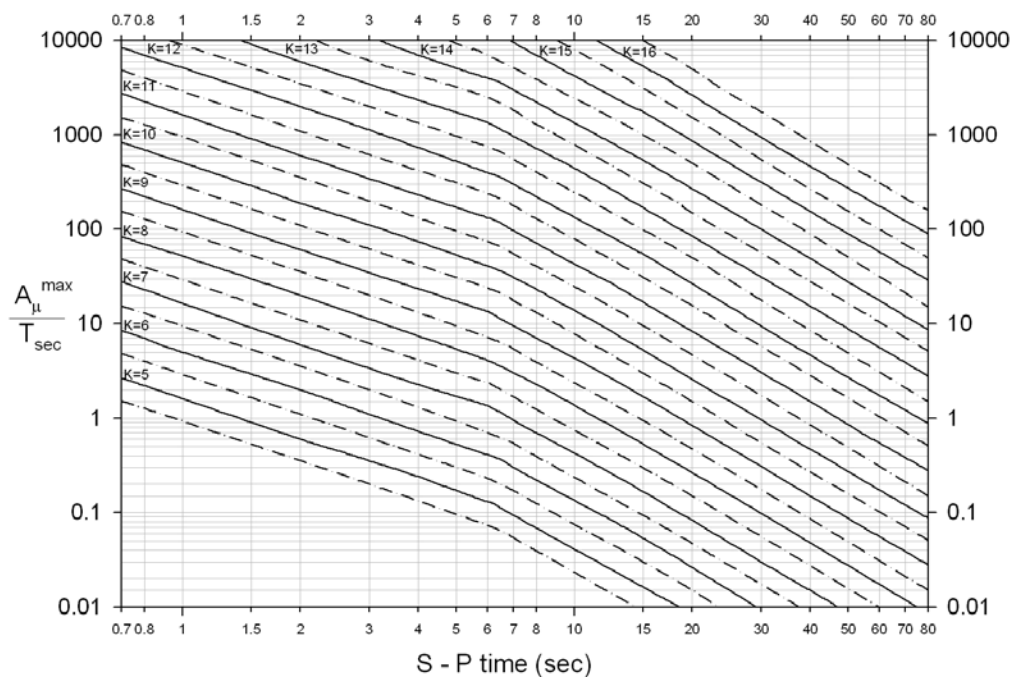


Figure 5 The current Kamchatka nomogram for scaling A/T measured on SKM records. Dashed lines represent limits of integer determinations. Vertical axis amplitude is in microns (μ), period, T , in seconds. The horizontal axis is logarithmic in S-P travel time, and corresponds to hypocentral distances up to about 640 km (copied from Figure 6 in Rautian et al., 2007, Seism. Res. Lett., **78**(6), p. 585, © Seismological Society of America).

A K-class nomogram was also developed for the Crimea (Pustovitenko and Kul'chitsky, 1974), although it is unclear how extensively it was used. Recently, Adilov et al. (2010) published K-class nomograms for Dagestan utilizing A_P only, A_S only, and $A_P + A_S$ (Figure 6), using data from 1994-1997. As elsewhere, A_S dominates and therefore a reasonable approximation can be obtained from A_S data alone, especially at lower K values. According to Kondorskaya et al. (1981) the definition of K on the nomograms gives satisfactory results to within ± 0.5 K, which corresponds to about 0.3 m.u. (magnitude units).

The Baikal (Solonenko, 1977) and Caucasus (Dzhibladze, 1971) networks also calculated regional attenuation curves and K nomograms for their regions; however, in both cases, the results were fairly close to those of the Rautian (1958) formulation and therefore these regional nomograms were not put into use. K has also been back-estimated from various magnitudes and, specifically in the northern Caucasus, from duration magnitudes for estimating a duration K (Gabsatarova, 2002).

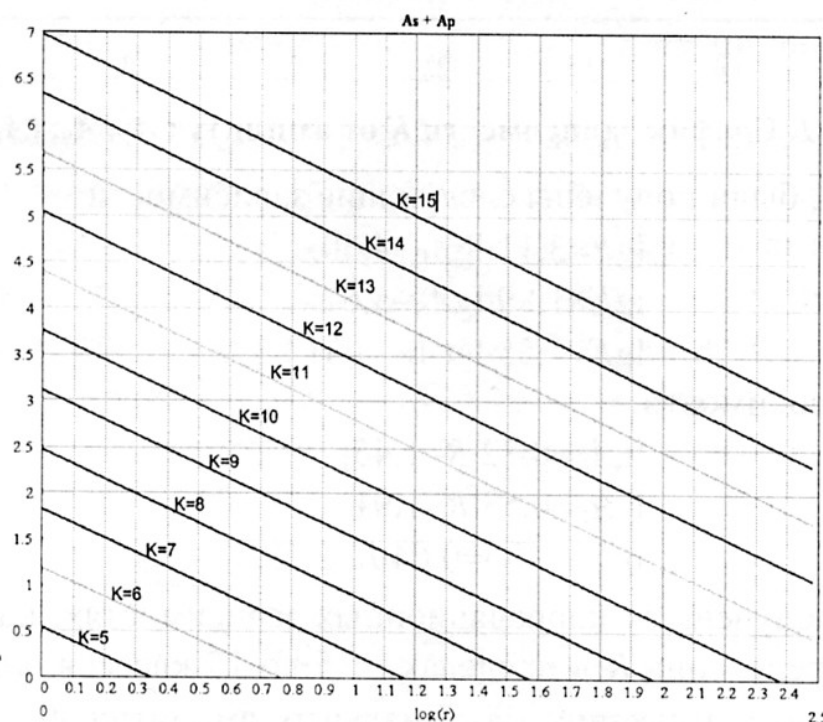


Figure 6 The Dagestan nomogram for SKM seismometer records. Vertical axis is in $\log A$ in microns (μ) and the horizontal axis is logarithmic in distance (r) in km, here up to 300 km. (copied from Figure 4 in Adilov et al., 2010),

In summary, we realize a need to derive regional calibration relationships for short-period K estimates, aimed at classifying mainly local and regional earthquakes down to very small size. If properly done, they would allow us to compensate for the unavoidable and often very significant variations in local/regional wave propagation conditions, thus making data derived by different networks, for events in different regions, compatible. This is the same situation as for the classical local magnitude scale M_l , which is applied in the same frequency range as the K scale. Meanwhile numerous regional M_l calibration relations have been published (see DS 3.1). Yet, one of the disadvantages of the K class is currently the inconsistency in which some of these regional formulas have been derived and scaled (e.g., using only P or S or the sum of P+S amplitudes, only A or A/T, and different types of responses for deriving or applying such relationships). Thus there is an urgent need for standardization of procedures compatible with modern recording and processing techniques. In conjunction with related international efforts for standardization of the procedures for the calculation of the most common magnitude measurements are needed that ensure the global compatibility of values of the same type, and the stability of the relationships between complementary types of parameters describing earthquake size or strength (see, IS 3.3 specifically, and Chapter 3 in general).

4 Relationship of K class to magnitude scales

(Note: In the following we use two types of magnitude nomenclature:

a) the NMSOP generic nomenclature which generally writes magnitude symbols not in Italics and without subscript and prefers to write the local magnitude in general as MI instead of the original Richter (1935) M_L or the new IASPEI (2011) standard ML ; b) the nomenclature for the IASPEI (2005 and 2011) approved standard magnitudes mb, ML, Ms_20 and Ms_BB.

Magnitude, M, began to be used throughout the USSR for larger earthquakes starting in 1955. The so-called “Prague-Moscow magnitude” formula for surface-waves (Vanek et al., 1962) became the accepted standard in the annual *Earthquakes in the USSR (Zemletryaseniya v SSSR)* starting with the 1962 compilation and became the international IASPEI standard in 1967. This magnitude, Ms_BB when written in recent IASPEI standard nomenclature (see below), was computed using the ratio $(A/T)_{max}$ of the whole surface-wave train in a wide range of periods between 3 s and 30 s (more recently extended to 60 s), originally as recorded by medium-period (SK) and the more long-period (SKD) Kirnos type broadband displacement seismographs (see Fig. 1) in the distance range between 2° and 160° (today measured directly from very broadband velocity meters; see Bormann et al., 2009 and IS 3.3).

Since local and regional networks in the FSU primarily used short-period instruments, the calculation of K class for stronger events with longer period energy was not possible. Thus K class was generally calculated for smaller events, while magnitude was calculated for larger events; however, both were potentially calculated for magnitudes between 4 and 5.5 (corresponding to K class 11 to 14).

The primary difference between K class and M is that K class is technically calibrated to energy. Empirically, the 20 s Ms, now standard Ms_20, has also been scaled to seismic energy by Richter (1958) using the log Es-mB and mB-Ms relationships in Gutenberg and Richter (1956) and, more recently, by Choy and Boatwright (1995) based on direct Es estimates from velocity broadband records (see discussion below). The “Prague-Moscow” Ms, equivalent to current standard Ms_BB (see IS 3.3), is presumed to be even more directly related to seismic energy than Ms_20 since it measures $(A/T)_{max}$ over a much wider period range. As originally developed, by using amplitudes of both P and S waves, K-class utilized a somewhat better estimate of the overall ground motion, thus making it less dependent on the effects of the focal mechanism. On the other hand, longer period surface wave measurements will be less affected by attenuation and scattering in heterogeneous media and by local site effects.

As noted above, since A_S dominates over A_P , many K values calculated throughout Russia today are only based on A_{Smax} , e.g., in Kamchatka and Sakhalin, so that they should be viewed as being in fact an equivalent to MI and are considered as such in the ISC Bulletin.

Since Es is proportional to A^2 , assuming constant T, the relationship between K and MI should be of the form

$$MI = c + 0.5 K \quad (2)$$

where c is a constant. However, variations in attenuation and radiated energy spectra, source depth, and tectonic styles will cause differences both between, and within, regions. Conceptionally, formula (2) is disputable because signal duration is not taken into account, yet still it may work in practice reasonably well.

According to Rautian (1960), early studies for Tadzhikistan yielded

$$M = (M_L) = -2.22 + 0.56 K = -2.22 + 0.56 (\log E_S) = 5.62 + 0.56(K - 14) \quad (3)$$

in the K class range $4 \leq K \leq 13$.

In this context it is worth mentioning that Berckhemer and Lindenfeld (1986) found that when calculating energy from broadband velocity records, $\log E_S \sim 2 M_L$, which corresponds precisely to (2), and is close to (3), the original empirical estimate by Gutenberg and Richter (1956) for Southern California ($\log E_S \sim 1.92 M_L$), and to the much more recent relationship derived by Kanamori et al. (1993) based on US TERRAScope data:

$$\log E_S = 1.96 M_L + 2.05 \quad (4a)$$

which could also be written as

$$M_L = 6.1 + 0.51 (K - 14). \quad (4b)$$

Since K class is calculated from short-period instruments, like mb and M_L , it saturates at around $K \approx 15$ to 17 or $M \approx 6$ to 7, unlike M_s , which saturates around 8-8.5 (Kanamori, 1983), corresponding to $K \approx 18$ -19.

To determine the variations between regions of the empirical relationship between M and K class, Rautian et al. (2007) tabulated M and K class values reported for 1970-1997 for each of the seismic regions used in *Earthquakes in the USSR*, and its successor publications, and compared them with mb and M_s values reported in the catalog of the International Seismological Centre (ISC). Because K class and M are both independent variables with their own uncertainties, they used an orthogonal regression which minimized the sum of the squares of the distance to the regression line in both the ordinate and abscissa directions. Because K class tends to be calculated for smaller events, and magnitude for larger events, and both are calculated using different methods (see above, for K, and Chapter 3 and IS 3.3 for M_s) there is some inconsistency and scatter in the different magnitude – K class ($\log E_S$) relationships (for more details see Rautian et al., 2007; also Sadovsky et al., 1986, and Ryznichenko, 1992). All regressions in Rautian et al. (2007) and in Table 1 were standardized to the general form

$$M = c + s (K - 14). \quad (5)$$

By using $K - 14$, the sign on c is always positive and makes comparisons clearer in the range $9 \leq K \leq 14$ where both K and M are likely to have been calculated.

The mb regressions (see Figure 7a and for details Table 1) are generally similar, close to an average relationship of

$$mb = 5.41 + 0.43 (K - 14) \quad (6)$$

in the K class range of interest ($9 \leq K \leq 14$), except for the Crimea, Sakhalin, Kurile, and Kamchatka. The Crimea, Sakhalin, and Kamchatka regions have higher constants and slope values than the other regions, while Kurile region has a similar slope, but a larger constant. The different constants suggest differences in attenuation conditions, and different slopes point to different amplitude-distance-source-depth conditions and relationships than those on which the magnitude and K class formulas are based. This clearly applies to the three Far Eastern regions (Sakhalin, Kurile, Kamchatka), which have different tectonic settings in

subduction zones with a different and more complex velocity structure and attenuation conditions. This explains why their K formulas differ significantly from the rest of the former USSR. It is interesting that the Kurile regression differs from both Sakhalin (which administers the Kurile network) and Kamchatka (which is tectonically similar).

For Central Asia, where Rautian (1958) developed the K class scale, Rautian (1960) gave the relationship $mb = 5.48 + 0.55 (K - 14)$ and Rautian et al. (2007) calculated $mb = 5.53 + 0.449 (K - 14)$; both are close to the average relationship given in (6).

If one replaces K by $\log E_S$ in equation (6) and solves for $\log E_S$ in order to make it comparable with classical $\log E_S$ -M relationships, one obtains

$$\log E_S = 2.326 mb - 1.42 \tag{7}$$

which has a slope remarkably similar to the original Gutenberg and Richter (1956) relationship

$$\log E_S = 2.4 mB - 1.2 \tag{8}$$

although Eq. (8) is meant to be used for the medium-period broadband mB measured by Gutenberg at periods between 2 s and 20 s (primarily between 5 s and 10 s) and not for the short-period mb. . However, for equal values of mb and mB between 5 and 7, Eq. (8) yields energy values 4 to 5.5 times larger than from Eq. (7),. Moreover, for equal events, broadband mB is generally larger than mb (on average by about 0.3 – 0.5 m.u. between mb = 5 and 7 and this difference increases further with magnitude. Thus, event estimates of E_S in this magnitude range are about 8 to 17 times larger when based on mB than those based on mb.

Table 1 Regressions between K class and magnitudes*

Region	mb	Ms	Me	N(Me)	Me r^{2**}
Carpathians	5.54 + 0.397 (K-14)	5.92 + 0.661 (K-14)			
Crimea	6.20 + 0.699 (K-14)	Insufficient Data			
Caucasus	5.60 + 0.391 (K-14)	6.02 + 0.782 (K-14)	6.49 + 0.547 (K-14)	8	0.61
N-Caucasus			5.23 + 0.635 (K-14)	3	1.00
Kopetdag	5.53 + 0.467 (K-14)	5.71 + 0.781 (K-14)	5.15 + 0.672 (K-14)	5	0.93
Central Asia	5.53 + 0.449 (K-14)	5.36 + 0.594 (K-14)	5.73 + 0.218 (K-14)	15	0.10
Altai - Sayan	5.47 + 0.482 (K-14)	5.37 + 0.633 (K-14)	5.79 + 0.517 (K-14)	5	0.59
Baikal	5.23 + 0.434 (K-14)	5.54 + 0.828 (K-14)	5.13 + 0.474 (K-14)	5	0.60
Yakutia	5.49 + 0.427 (K-14)	5.55 + 0.539 (K-14)	Insufficient data		
Northeast	5.33 + 0.445 (K-14)	Insufficient Data	Insufficient data		
Amur	5.01 + 0.394 (K-14)	5.10 + 0.755 (K-14)			
Sakhalin	7.25 + 0.669(K _S -14)	7.57+0.773 (K _S -14)	Insufficient data		
Kurile***	6.30 + 0.460(K _S -14)	6.56+0.642 (K _S -14)	6.93 + 0.729 (K-14)	18(39)	0.85
Kamchatka	6.11+ 0.552 (K _F -14)	6.47+0.838 (K _F -14)	6.26 + 0.657 (K-14)	52	0.66

* mb and Ms correlations from Rautian et al. (2007).

** r^2 is Pearson's correlation coefficient squared.

*** see Rautian et al. (2007) for calculation methodology; for N(M_e) the number of bins is given, the number of total data points is in parentheses.

Ms regressions by Rautian et al. (2007) are more variable (Figure 7b, Table 1), reflecting the fact that the methodology and frequencies used for calculating mb and K are much closer than those for determining Ms. Further, one has to distinguish between Ms as generally reported by NEIC, measured in a the narrow range of periods between 18 and 22 s at teleseismic distances between 20° and 160° only, and the broadband Ms measured today at periods between 3 s and 60 s and in a wider distance range between 2° and 160°. Although both these values of Ms are calculated according to the “Prague-Moscow” IASPEI standard formula $M_s = \log(A/T) + 1.66 (\log \Delta^\circ) + 3.3$, the former, now termed Ms₂₀, is mostly used in western countries and at all stations and centers reporting to the NEIC, whereas the latter, now termed Ms_{BB}, is the more correct application of the “Prague-Moscow” formula and has been used ever since its development in the FSU and allied countries, as well as by China. The orthogonal Ms-K relationships tabulated by Rautian et al. (2007) use Ms values published by the ISC, which are based on a mix of both types of data, depending on what has been reported to it.

The slopes of Rautian et al. (2007) regressions vary from 0.5 to 0.8 with a mean regression of

$$M_s = 5.52 + 0.702 (K - 14), \quad (9)$$

excluding the Carpathians, which are based on very limited data, and the subduction zone networks of the Russian Far East, which have c values > 6.4 and used different methodologies for calculating K as noted above. In general, however, most of the curves are fairly close to each other in the range $10 \leq K \leq 14$ (Figure 7b).

In this context we also draw the attention to the non-linear relationships between K-Fedotov and Mw (see Figure 9) as well as the linear relationships between K-Fedotov and the short-period Obninsk mPVA = mSKM, published by Gusev and Melnikova (1991) and Gusev (1992). Note that mb(NEIC, ISC) is on average smaller by about 0.18 magnitude units than mPVA = mSKM, because mb is mostly measured on records with responses similar to that of the Benioff instrument with a slightly lower seismometer eigenperiod and a narrower bandwidth than SKM.

Russian regional events listed in western seismicity catalogs, including the ISC, are reported with magnitudes calculated from K class using locally derived regressions, although this is not always clearly stated, resulting in uncertainty as to which the primary determination was. Moreover, there have been procedural changes which affect how the magnitudes were calculated from K class which bias reported values. For examples see Rautian et al. (2007).

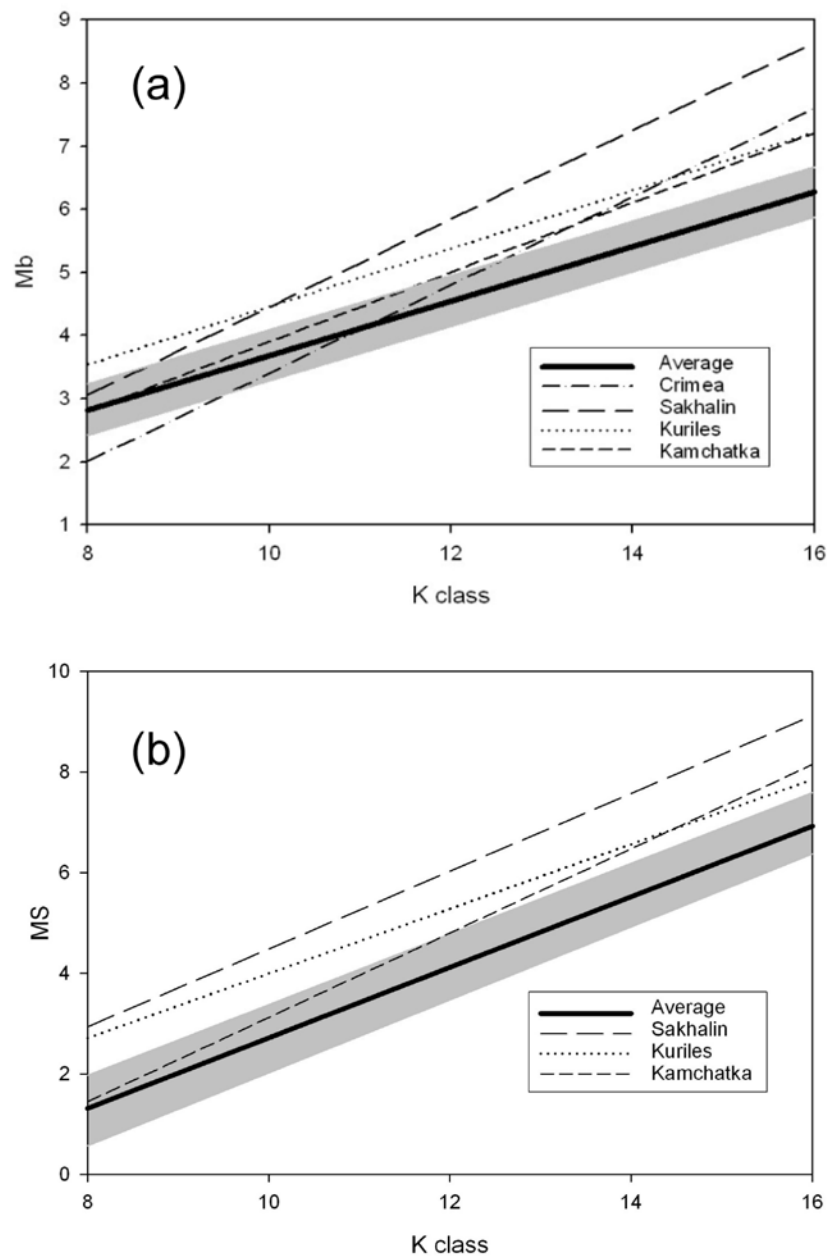


Figure 7 Comparisons of linear orthogonal regressions between K class and (a) mb and (b) Ms for various regions of the FSU. Most regressions fall within the shaded region; the mean of those regressions is shown as black line and corresponds to the formulas (6) for mb and (9) for Ms, respectively. Outlier regressions for Sakhalin, Kuriles and Kamchatka are shown and labeled separately (copied from Figure 8 in Rautian et al., 2007, *Seism. Res. Lett.*, **78**(6), p. 586, © Seismological Society of America).

Replacing K by $\log E_S$ in (9) one obtains

$$\log E_S = 1.424 M_S + 6.14. \quad (10)$$

The slope is similar but with a very different constant relative to the commonly cited relationship $\log E_S = 1.5 M_S + 4.8$, derived (see below) from the relationships published by

Gutenberg and Richter (1956). The constant in (10) is even more different than that in the relationship $\log E_s = 1.5 M_s + 4.4$, derived by comparing NEIC M_{s_20} with E_s measurements obtained by direct integration and Earth model scaling of squared P-wave velocity amplitudes (Choy and Boatwright, 1995). However, almost all data calculated by Choy and Boatwright (1995) are from M_s values between about 5.5 and 8.5, i.e., strong earthquakes, at which K values tend to saturate or are already saturated. This might explain the very large constant in (10) which results in an underestimation of M_s as derived from K for weak earthquakes ($K < 10$). This likely could be corrected by correlating K not with the M_{s_20} dominated ISC M_s but rather with M_{s_BB} , which is applicable in the near regional range also for periods much smaller than 20 s. Bormann et al. (2009) showed that, on the average, $M_{s_20} = 3.5$ is equivalent to $M_{s_BB} = 4$. Such a revised K - M_{s_BB} correlation is strongly recommended in the future when more M_{s_BB} data, now a IASPEI standard (see IS 3.3), become available on a global scale, the FSU included. This will allow a direct correlation of data determined by these two independent approaches with different instruments. Regardless, in the range of K between about 12 and 16, M_s estimates determined using (9) are reasonable, with errors likely to be smaller than about 0.5 magnitude units.

Finally, we refer to the extremely important compilation of data by Riznichenko (1992), comparing K and M with $\log M_0$ in relation to stress drop variations by more than three orders from 0.01 to more than 10 MPa (Figure 8).

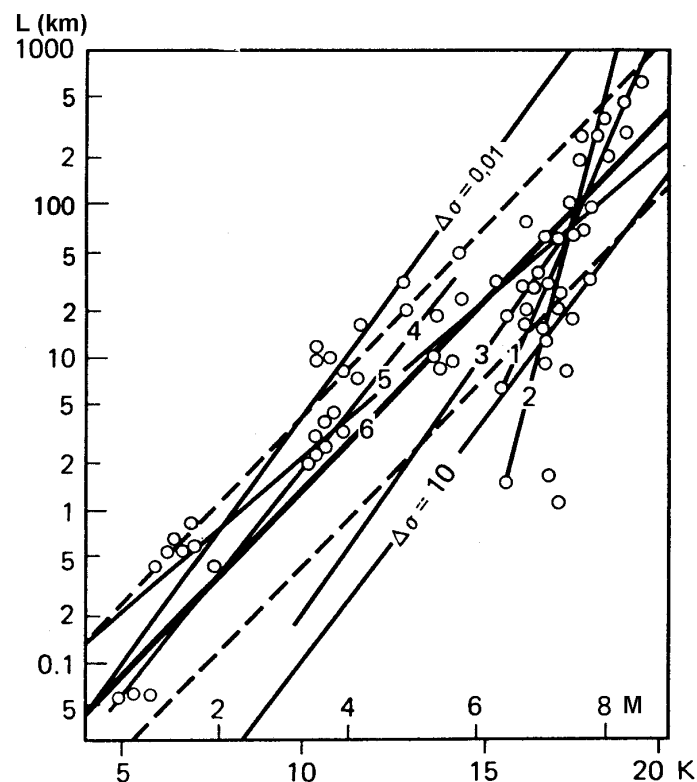


Figure 8 Correlation of source length L with magnitude M and energy class K according to data from various sources (e.g., curve 1 by Tocher, 1958, curve 2 by Iida, 1959; curve 6 average by Riznichenko, 1992; curves 3 to 5 from other authors quoted in Riznichenko, 1992). Thin straight lines: the associated stress drop, $\Delta\sigma$, are given in MPa; broken lines mark the limits of the 68% confidence interval with respect to the average curve 6 (modified from Riznichenko, 1992, Fig. 3; with permission of Springer-Verlag).

The values of M given in this diagram on the abscissa, along with $K = \log E_s$, correspond, depending on the distance and event size, to either M_L (for values $\ll 4$ to about 6), mb (for > 4 to about 5.5), mB (between about 5.5 to 7.5) or M_s (> 6 to about 8.5). The M and K values given on the abscissa correspond to the following average “global” relationships between K and M :

$$\log E_s \approx K = 1.8 M + 4.1, \quad (11)$$

$$M = 0.556 K - 2.3 \quad \text{or} \quad (12)$$

$$M = 5.5 + 0.556 (K - 14) \quad (13)$$

with the understanding as to which approximate type of specific magnitude, M , this applies to in different magnitude ranges. The range of uncertainty is on the order of about 0.3 magnitude units. If one compares Eq. (13) with the Eq. (4b) for M_L , (6) for mb and (9) for M_s , then one realizes that the constants in all these relations are within a few tenths, that the slope in (13) is very close to that in (4b) for M_L , and in between, the slopes are 0.43 for the mb - K and 0.70 for the M_s - K . The agreement between Eqs. (4b) and (13) could likely be further improved by correlating K specifically with M_s _BB, using values determined in the local and regional distance range, typically measured at periods between 2 s and <15 s. mb , however, is generally not as good an estimator for earthquake size and energy release for magnitudes above 5.5, for which mB _BB is better suited. Yet, since mb is measured in the teleseismic range only, few values are available for event magnitudes below 3.5. Therefore, in the local and regional distance range, K and M_L are better estimators of the released seismic energy for smaller earthquakes, and M_s _BB for stronger earthquakes.

When working with small earthquakes, it is often useful to relate K values to M_w which is practically a standard reference magnitude at present. Direct determination of M_w meets with an obstacle for small earthquakes: their flat low-frequency part of the displacement spectrum is rarely observed with an acceptable signal-to-noise ratio. Thus M_0 or M_w are impossible to determine en masse on a routine basis, and catalogues of small earthquakes inevitably contain K (or M_L) as a main magnitude value. To convert these catalogue values to M_w at least approximately, an adequate M to M_w relationship is a must. .

Figure 9 gives non-linear relationships between different K scales, M_L , mB , and variants of mb , on one side, and M_w , on the other side. The mb values are based on records with short-period Benioff or WWSSN-SP instruments, respectively equivalent simulation filter responses (NEIC practice) and mb SKM is from SKM-3 instrument (Obninsk practice). The $mb_n(M_w)$ trend reflects mb values obtained through older NEIC practice with narrow time window of amplitude measurement; see figure caption for details. Figure 9 is partly based on average relationships published by Gusev and Melnikova (1991) and Gusev (1992), extended here to the $M_w = 2$ -5 range. Various other sources were also used; the most significant ones are Hanks and Boore (1984) and Tan and Helmberger (2007).

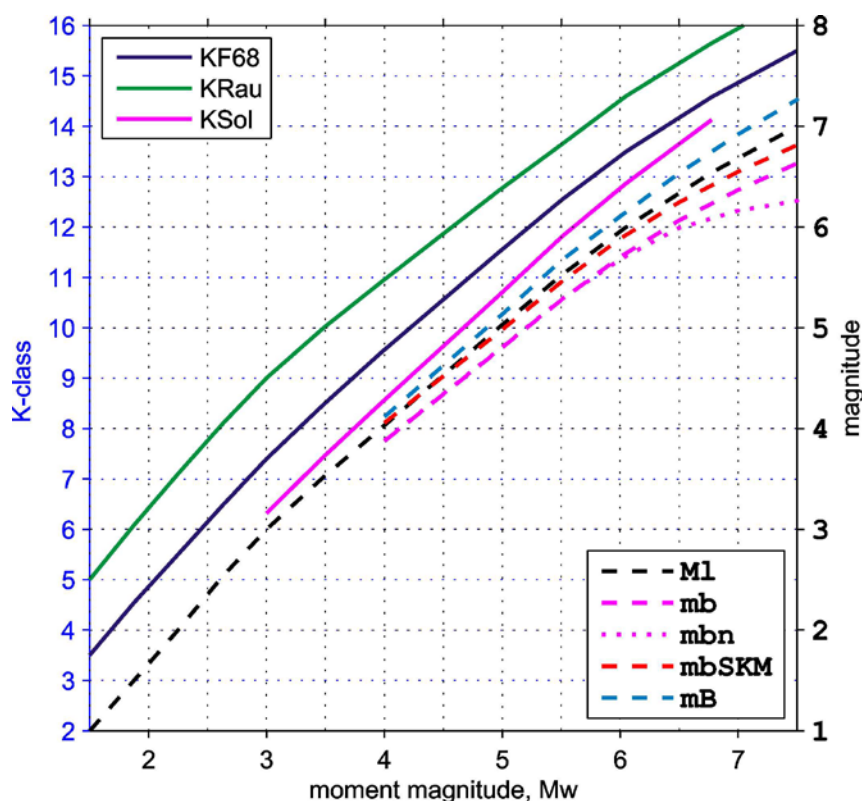


Figure 9 Various K-classes and magnitudes vs. moment magnitude M_w . The left ordinate scale is K-class. Corresponding relationships (solid curves) are given for Fedotov (1968) S-wave K-class used on Kamchatka, Soloviev (1967) K-class used at Sakhalin for Kurile Islands, and Rautian (1960) K-class used in Central Asia and, in modified form, for many other continental areas of the FSU. The right ordinate scale is magnitude. Corresponding relationships (broken and dotted lines) are given: for classic local magnitude M_l ; for variants of mPV measured on short-period Benioff or WWSSN-SP type of records within a narrow measurement time window of 5-10s after the P-wave first arrival, denoted mb_n ; its new IASPEI standard variant measured within an “unlimited” time window between the P and PP first arrival (see IS 3.3), denoted mb ; the similar Russian mPV variant measured on slightly longer-period SKM-3 records, denoted mb_{SKM} (with $mb_{SKM} \approx mb + 0.18$ on average); and for medium-period, traditional m_B of Gutenberg.

Note that the relationships M_l vs. M_w or K vs. M_w vary significantly between regions and even sub-regions. Therefore, the curves in Figure 9 can serve only as a preliminary reference. The “saturation effect” for body-wave magnitudes is clearly seen on Figure 9. Yet, it should be kept in mind that true saturation takes place only for mb_n that scatters around the $mb_n = 6.4$ level at $M_w \geq 7.5$. In contrast, mb values based on amplitude readings within “unlimited” P-wave measurement time windows such as IASPEI standard mb (Bormann et al., 2009), its forerunner applied by Houston and Kanamori (1986), and the Obninsk mb_{SKM} (Gusev and Melnikova, 1991) keep growing up to M_w around 9, attaining then values of about 7.4 to 7.6.

Note that recently, the Kamchatka network began to publish regional M_l directly calculated from the Fedotov (1968) S-wave K-class via the conversion relationship

$$M_l = 0.5K_F - 0.75. \quad (14)$$

This relationship is in agreement with the related curves of Figure 9. Interestingly, (14) is slope-wise identical with the equivalent formula (4b) derived above for California, although the constants differ by 0.29 m.u.. Accordingly, for equal $K = \log E_S$, Eq. (14) would yield about 0.3 m.u. smaller M_I values than (4b), hinting to regional differences in attenuation, the latter being larger in Kamchatka.

5 Future development and use of K for energy-related earthquake classification

The paper by Rautian et al. (2007) reviewed the K classification of earthquakes mainly to

- explain the methodology of its development and modifications over time in different seismotectonic regions of the FSU,
- provide the most recent relationships between K and the m_b and M_s magnitudes as published by the International Seismological Centre, and
- give recommendations - sometimes cautionary - for the interpretation and use of the extensive K -class data available in current Russian and past FSU catalogs in terms of common magnitude scales, so as to make seismicity data for the vast territory covered by the K class system compatible with those of the rest of the world.

However, in this section, we would like to go one step further and look at the K -class system not only from a historical and data compatibility perspective, but in terms of the merits of this approach in general.

It is interesting to note that Gutenberg and Richter (1956) published the first semi-empirical energy-magnitude relationship, based on rather meager data, relating energy in Joules (J), to body-wave magnitude:

$$\log E_S = 2.4 m_B - 1.2 \quad (15)$$

Combining (15) with the Gutenberg and Richter (1956) body-wave to surface-wave magnitude conversion relationship,

$$m_B = 0.63 M_s + 2.5 \quad (16)$$

the often quoted Gutenberg-Richter magnitude energy relationship

$$\log E_S = 1.5 M_s + 4.8 \quad (17)$$

is obtained. Hanks and Kanamori (1979) based their scaling of seismic moment M_0 to M_s , and thus the definition of their moment magnitude scale, M_w , on this relationship. Also the Russian authors of the K -class system calculated E_S in Joules from this equation, even though it is based on 20 s surface waves, which are not representative for the assessing the high-frequency related damage potential of earthquakes or the total seismic energy release of smaller earthquakes with corner frequencies well above 0.1 Hz. Gutenberg and Richter (1956) published only Eq. (15), relating E_S to m_B , which Gutenberg measured in a wide range of periods less than 20 s.

K was supposed to be an estimate of $\log E_S$, and therefore should have been determined from the (A/T) spectrum. However, because of difficulties in conducting proper spectral analyses and integrating analog records in the past, all K formulations have been based on measuring

only A_{max} . Gutenberg and Richter (1956) realized that the body-wave scale would yield stable magnitude estimates only by normalizing the A_{max} value in the P-wave train by its period, T , which varied in a fairly wide range between some 2 s and 20 s even in records from the limited broadband responses of the medium-period seismographs of that time. Now, modern force-feedback broadband sensors have direct velocity proportional responses at least in the range between about 0.05 and 40 s, or wider, and allow the direct measurement in the entire P- or S-wave train, of the maximum velocity, V_{max} , which is related to the maximum seismic energy released by the seismic source at the corner frequency of its displacement spectrum (see Bormann, 2011). Thus, we can expect that the single amplitude magnitude

$$mB_BB = \log(V_{max}/2\pi) + Q(h, \Delta), \quad (18)$$

where $Q(h, \Delta)$ is the empirically determined Gutenberg and Richter (1956) calibration function, which depends on focal depth, h , and distance, Δ . $Q(h, \Delta)$ is available in tabulated and diagram form (see DS 3.1) and accounts for the cumulative effects of the diverse, and theoretically not yet sufficiently well known or tractable in detail influences of wave propagation. Thus, mB is, in fact, a reasonably good estimator for E_S . This has been confirmed by Di Giacomo, Saul and Bormann (see Bormann, 2011, and Figure 10).

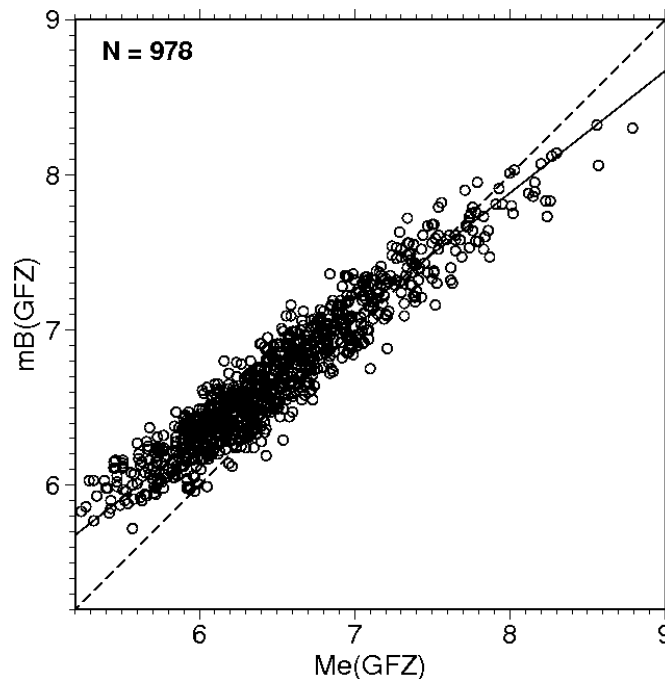


Figure 10 The relationship between automatically determined values of mB , equivalent to standard mB_BB , and Me , according to procedures developed at the GFZ German Research Centre for Geosciences by Bormann and Saul (2008) and Di Giacomo et al. (2010). The dashed line is the 1:1 relationship and the solid line the orthogonal regression relation (19). (Figure by courtesy of D. Di Giacomo, 2011).

In the range $5.5 < mB_BB < 8.5$ the data in Figure 10 satisfy the following orthogonal relationship between mB_BB and the energy magnitude $Me = (\log E_S - 4.4)/1.5$:

$$mB_BB = 0.79 Me + 1.58. \quad (19)$$

Eq. (19) permits the estimation of M_e (and thus E_S) by using very simple and easily to automate mB_BB measurements (see Bormann and Saul, 2008) with a standard deviation $\sigma = 0.18$ magnitude units, about the best average of direct M_e measurements. This “error” corresponds to an approximately 95% (2σ) uncertainty in E_S estimates by a factor of less than 4, which is close to the factor of 2-3 given by Kanamori and Brodsky (2004) for E_S estimates via the integration of very good broadband seismic data.

Choy and Boatwright (1995) and Boatwright et al. (2002) suggested a method of calculating energy magnitude, M_e , based on broadband digital data, which has been applied by the U. S. Geological Survey. Therefore, we have attempted orthogonal regressions between K and M_e from the U. S. Geological Survey database for a limited number of regions in which data were available (see Table 1) using the same regression method as in Rautian et al. (2007). As noted above, in most cases, the number of events is small (< 5), because M_e is calculated only for larger events (generally $M_e > 5.5$) and K for smaller local and regional events (generally $K < 14$). Only for the region of Kamchatka enough data have been available to plot K as a function of M_e (USGS) (see Figure 11). Despite the large data scatter, the agreement between the orthogonal K - M_e regression relationship and the $\log E_S$ - M_e relationship by Choy is surprisingly good, for constant M_e being within 0.15 to 0.25 units of K , i.e., just a factor of 1.4 to 1.8 larger in E_S than from the teleseismic estimate, or, for equal K , about 0.1 to 0.2 m.u. smaller in related M_e as compared with the Choy $\log E_S$ - M_e relationship.

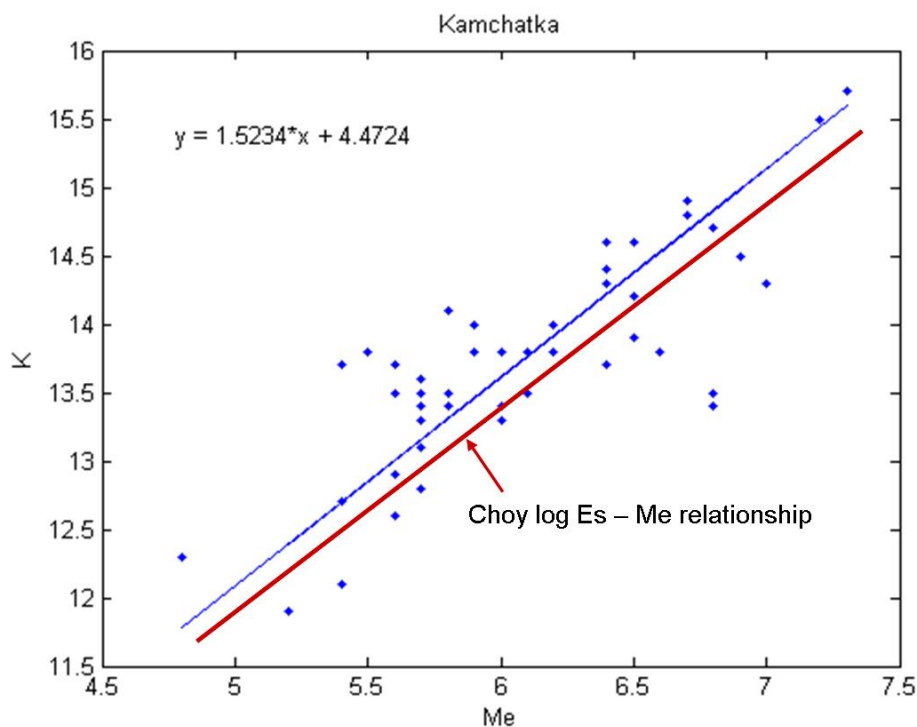


Figure 11 Orthogonal regression relationship between K measured for Kamchatka earthquakes and the related teleseismic E_S - M_e (USGS) relationship.

For other regions, with much less data, the relationships between K and M_e are so far less clear and more variable, also with respect to level differences. Central Asia data are the most varied and the regression is biased by a series of low M_e , but relatively high K , events (see Table 1). This may hint to rather high-frequency waves radiated from high stress drop earthquakes, which are not well covered by teleseismic observations.. As the data are

concentrated in a narrow range around $M_e = 5.5-6.0$ and $K = 14-15$, this regression has a very low correlation coefficient. Other relationships are more stable, e.g., for the Kurile Islands and the Caucasus, yielding for K values between 12 and 15.5 M_e estimates that differ -0.1 to $+0.6$ m.u. from the values derived via the Choy relationship $M_e = (\log E_S - 4.4)/1.5$. The latter (for derivation see IS 3.5 and Bormann and Di Giacomo, 2011) can be written in terms of K as.

$$M_e = 0.667 K - 2.933 = 6.40 + 0.667 (K-14). \quad (20)$$

For magnitudes below 5.5, however, global M_e is rarely measured because of the small signal-to-noise ratio of P waves in teleseismic broadband records as well as the difficulties in including frequencies higher than 2 Hz into the complex model calculations.

In their conclusions, Rautian et al. (2007) correctly state that currently used methods for estimating seismic energy and energy magnitude, M_e , that are based on global models, commonly make simplified theoretical assumptions with respect to Earth structure, attenuation, scattering, site amplification, source radiation pattern, and/or other local or regional geologic factors that preclude their use at non-teleseismic distances.. Especially for small earthquakes recorded at local and regional distances, no easy practical and reliable methodology is currently available. On the other hand, although K is “ideologically” an energy scale, its practical implementation and use so far is more of a magnitude-like nature, with K being associated with energy only by correlation (Kondorskaya et. al., 1981). More truly energy-related K scales would require the consistent measurement of maximum ground motion velocity instead of displacement.

Despite the drawback of current K scales, the agreement in trend and level of the relationship between the (A/T) derived K values for earthquakes in Kamchatka (see Figure 5) and teleseismic M_e (USGS) is very promising. It lets us expect that a K -type procedure could be developed, which allows to estimate E_S and M_e for large numbers of local and regional earthquakes sufficiently reliably in a similarly simple, interactive or easy to automate way as via teleseismic mB_{BB} . Therefore, we recommend to further develop the K class system for events with magnitudes below 6, down to magnitudes of about 2, which are recordable with good signal-to-noise ratio at regional and local distances with standardized high-frequency responses. They should have a velocity-proportional pass-band of not less than about 5 octaves relative bandwidth, covering the frequency range between about 0.4 to 10-20 Hz). This would allow to measure in this frequency range directly the largest event velocity amplitudes. This eases both the derivation of local energy-related K calibration functions as well as later routine K measurements for estimating E_S and M_e of local earthquakes.

On the other hand, we realize from Figure 7 that the orthogonal regression relationships between K and teleseismic m_b , respectively M_s for earthquakes in the Russian subduction zone environments of Kamchatka, Kurile Islands and Sakhalin differ systematically, partially depending on K , from those derived for other intra-continental regions in FSU states. The differences in M_s (for equal K) reach up to one and for m_b up to two magnitude units. This is not acceptable and hints, as mentioned already above for Kamchatka, to significant differences in the absolute calibration of the various regional K scales. Thus there is an urgent need to develop standards for both calibration in continental (shallow earthquakes) and subduction zone environments (with also deep and intermediate depth earthquakes) which yield K data that are really compatible in a wide range of K between the different seismotectonic environments and for larger events also with teleseismic M_e data.

Although there is no more need to measure period specifically for deriving the ground motion velocity, local records in a wide range of magnitudes and from different seismotectonic settings are likely to show a great variability of frequencies at which V_{max} is measured. The higher the measured frequencies the more the heterogeneity of the Earth medium and of the surface topography at local and regional scale will impact the attenuation effects. They may have to be compensated for by frequency-dependent and maybe even source–distance and azimuth-dependent local and regional calibration functions.

To derive and use such calibration functions goes well beyond traditional K formulas and their application. It will require years of careful and complex empirical data collection and analysis. However, digital data recording, filtering and processing in a standardized way eases this task. Yet, for assuring the better compatibility of future K estimates the new calibration functions have to be scaled so that they agree with original definitions and values at a suitable reference distance, in a similar way as regional MI calibration functions in other countries have to be scaled to the standard Southern California MI calibration function (see DS 3.1 and IS 3.3) in order to make them mutually compatible. This degree of compatibility has obviously not yet been achieved with respect to the various regional K-scales currently in use. E.g., in Figure 9, most of the difference between KF68 and KSol has no seismological reason. Rather, it is due to an error in calculating the absolute level of the $(A_{max}/T)(\Delta, h=40\text{km})$ curve of Fedotov (1963). This makes the K-Sol values by 0.6 units lower than they should be (see Fedotov, 1972, p63). Moreover, the slight differences in slope of these two curves are most likely due to the difference of seismometer periods used, which are 0.7 s for KSol vs. 1.2 s for KF68. Minor differences between the regional K-scales may also originate from different assumptions made by the scale authors on the input parameters used in calculating the wave energy density on the reference sphere, e.g., Fedotov (1972) used $v_s = 3 \text{ km/s}$, whereas Rautian used $v_s = 3,5 \text{ km/s}$. Such details need to be checked and agreed upon in order to assure an unambiguous absolute calibration as well as standard (simulated) instrument responses and measurement procedures in order to assure interregional K-value compatibility. Only then it will be possible to interpret differences in measurement values that are beyond the range of measurement errors in seismological terms only.

Although most of the seismic energy is radiated in the S-wave group, teleseismic procedures for estimating E_S generally use the first arriving P waves. The reason is that S-waves from strong earthquakes with large rupture durations of several 10 seconds or even minutes are often superimposed by later secondary, phases. This may bias the results of full waveform integrations in a wide range of periods and rupture durations. Moreover, theoretical corrections for propagation and attenuation effects in a wide range of periods are more difficult to account for S waves than for P waves. But these objections against the use of V_{Smax} instead of V_{Pmax} are less relevant for the analysis of weak earthquakes in the regional and local distance range. Earthquakes with magnitudes below 6 have average rupture durations of less than 6 s. Recorded at distances typically less than 1000 km, they have clearly developed distinct later Sg/Lg groups that are well separated in time from the earlier and usually much weaker P-wave phases. This eases significantly proper phase identification and parameter measurement both in automatic and interactive modes.

Finally, another approach, considered by Rautian et al. (2007) to be the most appropriate way to get seismic energy along with seismic moment, apparent stress, etc., is to use the later part of the coda for source spectral estimation (Rautian and Khalturin, 1978). This method had been originally developed for analog records with multi-bandpass filtering and could easily be

adapted as well for digital instrumentation and processing techniques now available. A certain step in this direction is the routine use of short-period coda-class Kc on Kamchatka mentioned above.

Acknowledgments

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