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EMPIRICAL SCALING OF STRONG EARTHQUAKE GROUND MOTION - PART II: DURATION OF STRONG MOTION

V.W. Lee
Civil Engineering Department
University of Southern California
Los Angeles, California, U.S.A.

ABSTRACT

Definitions of duration are reviewed, with emphasis on the one, which is used in most studies. In this definition, the duration is related to the rate of energy input into structures. Next, the physical parameters of the earthquake source and propagation path, and regional geologic and local soil characteristics, are discussed, as those influence the duration of strong motion. Finally, scaling models are illustrated, obtained from regression analyses of strong motion recordings in the western USA.

KEYWORDS: Duration, Energy, Rupture, Path and Regional Effects

INTRODUCTION

The duration of strong earthquake ground motion determines the rate of energy input into a structure, and should be considered in all analyses of linear and non-linear structural response. It has an important role in the analysis of liquefaction (Trifunac, 1995) and of permanent displacements of soils, and in the procedures and algorithms for probabilistic assessment of structural response to earthquakes. For example, studies of the statistical distribution of peaks in structural response (Udwadia and Trifunac, 1974; Amini and Trifunac, 1985; Gupta and Trifunac, 1991) require that the duration of shaking is specified. These studies determine the probabilities of exceedance of given levels of displacement, shear force or overturning moment, for a given number of times, at any level of a multistory building. The duration of strong shaking is also required for generation of site-specific artificial accelerograms (Lee and Trifunac, 1985, 1987; Wong and Trifunac, 1979).

The importance of the duration of shaking for non-linear response has been recognized, but it is still not used in the building design codes. The fatigue and non-linear yielding effects are thus ignored, or are considered in a simplified way. Clear and direct definitions of duration, and scaling models relating it to the earthquake source parameters (Trifunac et al., 2001), the propagation path characteristics, and the regional geologic and local soil conditions at the site, are required to incorporate the duration in analyses and design of structures.

In this paper, selected empirical regression equations of the duration of strong earthquake ground motion, in terms of earthquake magnitude, M, epicentral distance, Δ, site intensity I_{MM}, and the region specific propagation parameters, will be illustrated. A variety of such models were reviewed elsewhere (Trifunac and Westermo, 1976a, 1976b, 1977a, 1977b, 1978, 1982; Westermo and Trifunac, 1978, 1979; Novikova and Trifunac, 1993a, 1993b, 1993c, 1994a, 1994b, 1994c, 1995a, 1995b, 1995c; Novikova et al., 1993, 1995; Trifunac and Novikova, 1994, 1995). In this paper, we emphasize new, recent models, which use detailed characterization of the recording site conditions.

DEFINITIONS OF STRONG MOTION DURATION

The first studies of strong motion duration did not present its quantitative definition or dependence on magnitude and epicentral distance. In studies that followed, the duration was defined as the time interval between the first and the last time when the acceleration exceeds the level of 0.05g (bracketed duration, Page et al. (1972)), or the time interval during which 90% of the total energy is recorded at the station (Trifunac and Brady, 1975). Kawashima and Aizawa (1989) refined the definition of bracketed duration, and introduced normalized duration (elapsed time between the first and the last acceleration excursion greater than μ times the peak acceleration). Mohraz and Peng (1989) introduced the structural frequency and damping into the definition, and used a low-pass filter for computing the duration.

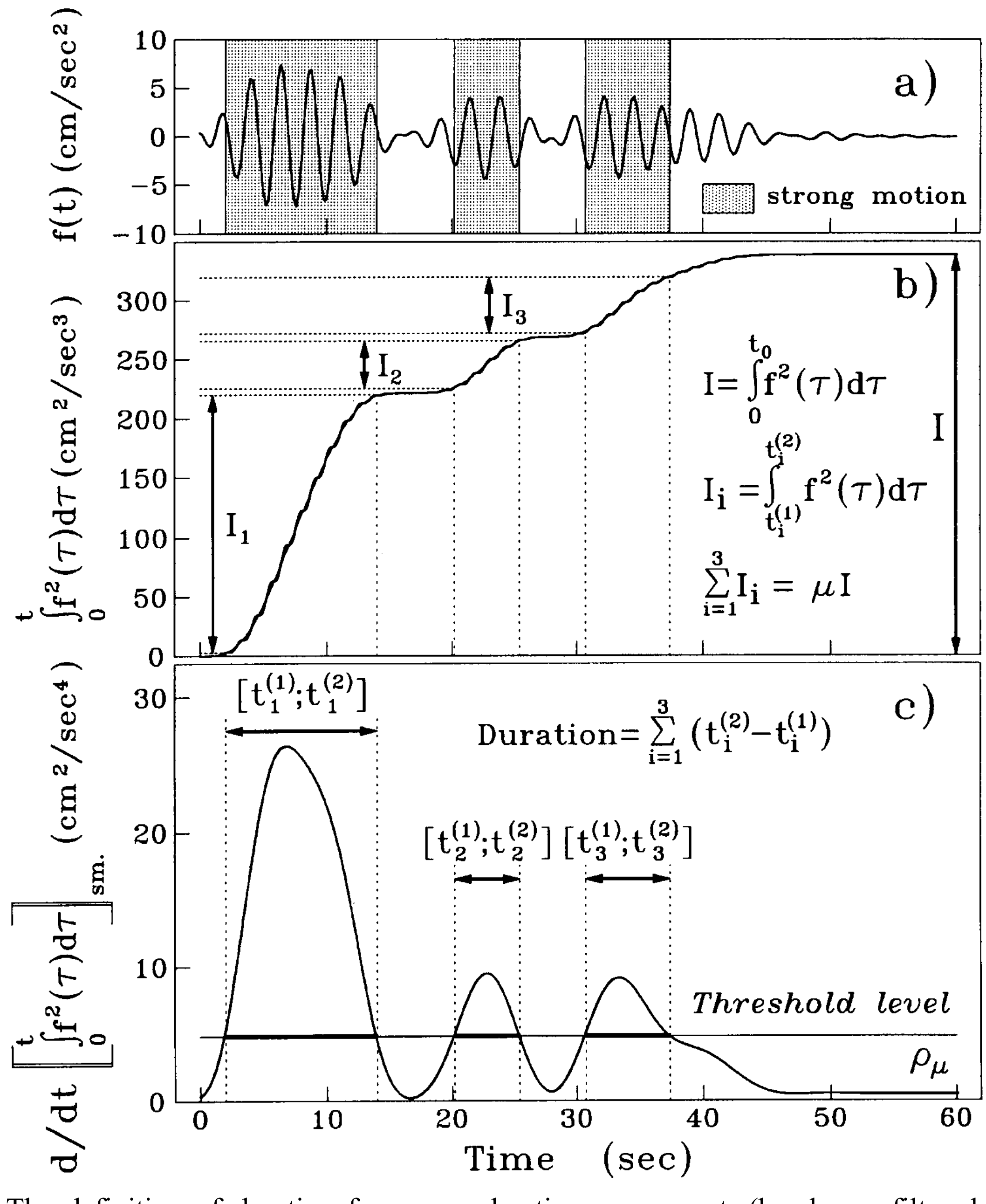

Fig. 1 The definition of duration for an acceleration component, (band-pass filtered, central frequency 0.37 Hz): a) time history $f(t)$ (the strong motion intervals are shaded); b) $\int_0^t f^2(\tau)d\tau$; c) the derivative of the integral of $f^2(t)$ after smoothing and the threshold level ρ_μ (the time intervals giving contribution to duration with $\mu = 0.9$ are highlighted)

The studies of Trifunac and Westermo (1977a, 1977b, 1982), Westermo and Trifunac (1978, 1979), and Novikova and Trifunac (1993a, 1993b, 1993c, 1994a, 1994b, 1994c, 1995a, 1995b, 1995c), use a definition that is related to the energy input in structures (Trifunac et al., 2001). This energy is proportional to $\int_0^\infty a^2(t)dt$. Next, the seismic wave energy radiated from the earthquake source is proportional to $\int_0^\infty v^2(t)dt$, where $v(t)$ is the ground velocity. In the statistics of the peak amplitudes of a random function, $f(t)$, the expected value of the peaks depends on the number of peaks, N, which is proportional to the duration of strong motion, and on $\left[(1/T)\int_0^T f^2(t)dt\right]^{1/2}$, the root-mean-square of $f(t)$. A common feature in the above functionals is the integral of the form $\int f^2(t)dt$, a monotonically increasing function of time, and proportional to the work during response. It is natural then to associate the duration of strong motion with the time interval during which most of this work (e.g., 90 percent) is realized.

The studies of Novikova and Trifunac (1993a, 1993b, 1993c, 1994a, 1994b, 1994c, 1995a, 1995b, 1995c) consider dependence of duration on frequency. This is done by filtering the signal into 12 frequency bands (also called channels), with central frequencies from f_0 = 0.075 Hz to f_0 = 21 Hz. For each channel independently, the strong motion duration is evaluated, and it is analyzed how it depends on the earthquake parameters, on the propagation path characteristics, and on the local conditions at the site. Figure 1 illustrates how the duration of a function $f(t)$ is calculated for a frequency channel $(f(t)$ can be ground displacement, velocity or acceleration). The duration is the sum of the lengths of the strong "pulses" of $f(t)$. The beginning and end of the pulses is determined from the condition that the integral $\int_0^\infty a^2(t)dt$ gains 90% of its final value (Figure 1). This condition is equivalent to the time derivative of the same integral being greater than a threshold level (the duration is evaluated using smoothed $\int_0^\infty f^2(\tau)d\tau$). The duration evaluated by this procedure directly from the recorded motion is referred to as "observed" duration.

One feature of the above definition is that only the strong pulses contribute to the duration. Other physically related definitions (Trifunac and Brady, 1975; McCann and Shah, 1979) consider the strong motion interval to be continuous. Another feature is that the duration evaluated by the above procedure is not related to absolute levels of motion. The knowledge of the frequency-dependent duration in this "relative" sense, combined with the information about the Fourier spectral amplitudes (Trifunac, 1991, 1993, 1994a, 1994b, 1994c), provides a complete description of strong motion.

SCALING PARAMETERS

The duration of strong ground motion may be represented by the sum,

$$dur = \tau_0 + \tau_\Delta + \tau_{\text{region}} \tag{1}$$

where, dur is the total duration of acceleration, velocity or displacement, τ_0 stands for the duration of the rupture process at the source, τ_Δ represents the increase in duration due to propagation path effects, and τ_{region} describes the prolongation effects caused by the geometry of the regional geologic features and of the local soil at the recording site.

1. Duration of Rupture, τ_0

The duration of the rupture process, τ_0, depends on the amount of released energy (magnitude M), the geometry of the ruptured area, fault length and width (L and W), the speed of the rupture process (dislocation velocity, v), and on the shear wave velocity in the medium, β (Trifunac and Novikova, 1995).

The frequency dependence of τ_0 on magnitude M can be modeled by a quadratic function,

$$\tau_0 \approx a_1 + a_2 \cdot M + a_3 \cdot M^2 \tag{2}$$

where, a_i $(i = 1 \text{ to } 3)$ are regression coefficients. Although an exponential function would be more “natural”, this quadratic expression is preferred (Novikova and Trifunac, 1993a). The duration of the rupture process, estimated from the duration of high-frequency radiations from the source, can be approximated by $\tau_0 = \alpha \exp(\gamma M)$, where $\alpha \approx .01$ and $\gamma \approx 1$. This can be compared with other estimates of different durations of the earthquake source (Trifunac and Novikova, 1995; Trifunac and Todorovska, 2002).

The duration of faulting is

$$\tau_1 \approx L/v + 0.5W/\beta \tag{3}$$

where, W is the fault width, L is the fault length, v is average velocity of dislocation, and β is the shear wave velocity in the medium surrounding the source. τ_1 can be computed from estimates of L, W, v and β, but can also be evaluated from $f_1 = 1/\tau_1$, where f_1 is one of the two corner frequencies in the low-frequency part of the Fourier spectrum of strong motion acceleration (Trifunac, 1993).

Near long and narrow faults, it is useful to work with the time it takes the dislocation to reach its ultimate amplitude, $\bar{u}$. Designating this time by T_0, it can be shown that $T_0 \approx \frac{\bar{u}\mu}{\sigma\beta}$ (Trifunac, 1994b, 1998). Here, μ is the rigidity, and σ is the effective stress (Trifunac, 1993, 1994b, 1998). Another "local" estimate of duration of faulting corresponds to the time it takes the dislocation to spread over the entire fault width, $\tau_2 \approx W/v$. This is related to another corner frequency in the Fourier amplitude spectrum, $f_2 = 1/\tau_2$ (Trifunac, 1993, 1998).

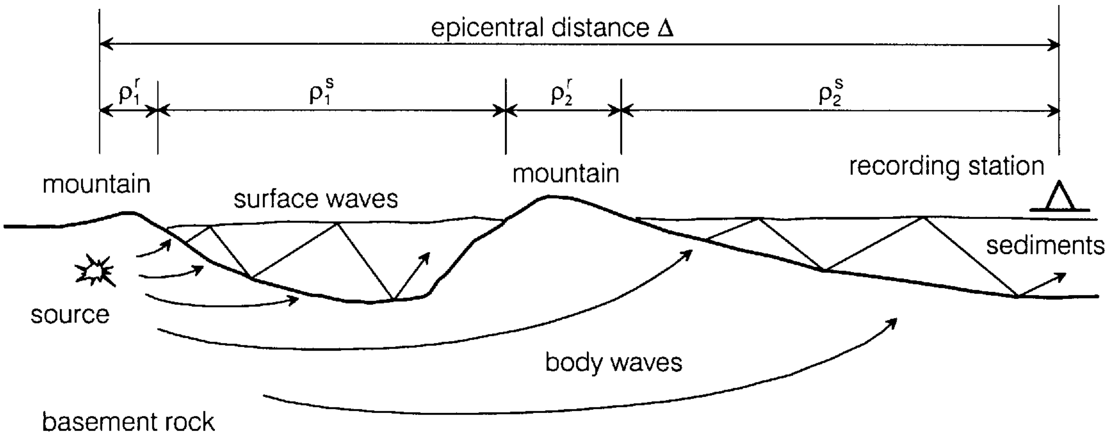

Fig. 2 The strong motion energy reaches the station in form of the surface waves and body wave

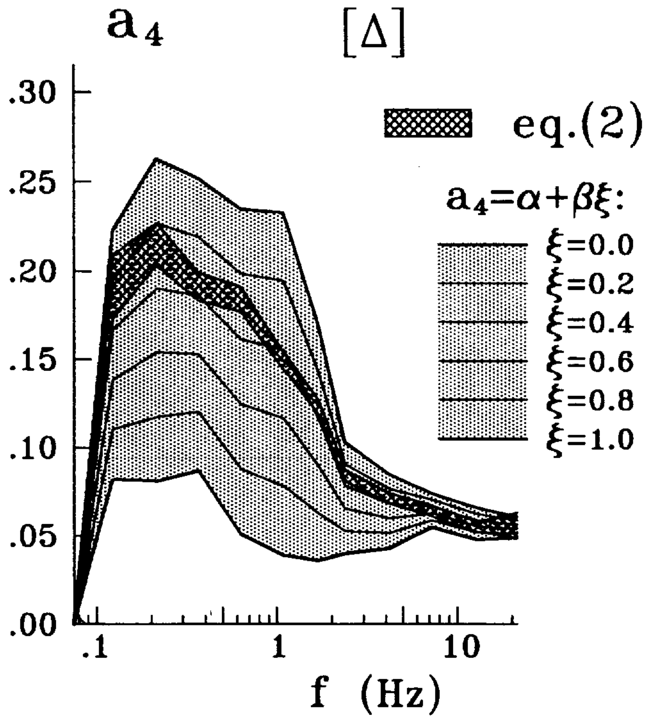

Fig. 3 "σ-interval" of the coefficient $a_4(f)$ in Equation (4) (cross-hatched area), and the coefficient a_4 with linear dependence of a_4 on the hardness of the propagation path, ξ (shaded zone)

2. Propagation Path Effects, τ_Δ

The second term in Equation (1), τ_Δ, describes the prolongations due to the distance traveled. It can be modeled by

$$\tau_\Delta(f) = a_4(f) \cdot \Delta \tag{4}$$

where, Δ is the epicentral distance, and f is the frequency.

This form has the interpretation that the duration of strong motion increases with the distance traveled due to dispersion of the strong motion waves (Figure 2). At its maximum (Figure 3, near frequency 0.2 Hz), the value of 0.2 corresponds to the increase of duration by 2 sec per 10 km of epicentral distance, and at $f \approx 15$ to 20 Hz, this value drops to 0.5 sec per 10 km. At low and intermediate frequencies, the main contribution to strong ground motion comes from surface waves, and the increase of the duration with distance can be explained by the dispersion of those waves, travelling through irregular, but generally "layered" structure of the upper crust (Trifunac, 1971). If $c_{\min}(f)$ and $c_{\max}(f)$ are the effective minimum and maximum phase velocities of the surface waves, and $v_{\min}$ and $v_{\max}$ are the lowest and the highest shear wave velocities in the layered half-space, then

$$a_4(f) \approx \frac{1}{c_{\min}(f)} - \frac{1}{c_{\max}(f)} < \frac{1}{v_{\min}} - \frac{1}{v_{\max}} \tag{5}$$

For long surface waves, only one mode of propagation is possible (at the local distances), and for a narrow frequency band, only minor dispersion occurs; so, $a_4(f) \to 0$ as $f \to 0$. Increasing f introduces additional modes; the variety of possible phase velocities increases, and $a_4(f)$ grows. Further increase in frequency causes concentration of the phase velocities of different modes at the smallest shear wave velocity of the region. Then, $c_{\min}(f)$ ($\approx v_{\min}$) and $c_{\max}(f)$ decrease, and so does $a_4(f)$. For high frequencies ($f > 5$ to 10 Hz), a_4 does not depend on frequency. The nature of the broadening of the strong motion pulses with distance differs here from the dispersive nature of the low-frequency wave propagation. For high frequencies, the strong motion appears to consist primarily of scattered body waves (Sato, 1989).

The propagation path effects depend on the percentage of the path traveled through rocks or through soft sediments. Using a map that shows the distribution of basement rocks on the earth's surface, we can characterize "hardness" of the transmission path for each pair source-station by the ratio of the portion of epicentral distance covered by rocks, as seen on the surface, to the total epicentral distance, $(\rho_1^r + \rho_2^r)/\Delta$ (Figure 2). We denote this ratio by ξ, and call paths with high ratio ξ as "hard" and with low ratio ξ as "soft". It may be assumed that $a_4 = \alpha + \beta\xi$, that is, the "prolongation due to propagation" coefficient is a linear function of the "hardness" of the path ξ (the regression analysis of this modified model gives practically the same results for all other scaling coefficients, as for the model in which ξ is ignored). The resulting dependence of a_4 on frequency is shown in Figure 3. The interval $a_4 \pm \sigma_4$ where $\beta = 0$, is shown for comparison. As expected, $a_4\,[\xi \approx 0] > a_4\,[\text{all cases}] > [\xi \approx 1]$.

The sum, $\tau_0 + \tau_\Delta$, represents the duration observed at the sites located on basement rock, and is called here as "basic duration". It serves as a basis for developing more "complete" accounting for prolongation of duration at the sites located on sediments.

3. Regional Effects, τ_{region}

Deviations from a uniform horizontally layered crust model occur everywhere along the path of the waves propagating from the fault to the recording site. For the San Fernando, 1971, California, earthquake, for example, substantial deviations from a multi-layered model dominate in the whole region shaken by this event. These deviations occur due to the topography of the basement rock, so that the several upper kilometers of the crust can be viewed as a collection of sedimentary basins separated by irregularly shaped basement rock "barriers". These barriers can be recognized on the surface as mountains, coupling geological and topographical irregularities (Figure 2). The influence of such structures on the propagating waves can be understood by studying idealized sedimentary basins (Todorovska and Lee, 1990, 1991a, 1991b), by generalization to arbitrarily shaped layers (Moeen-Vaziri and Trifunac, 1988a, 1988b), and by interpretation of recorded data and numerical modeling of the wave propagation.

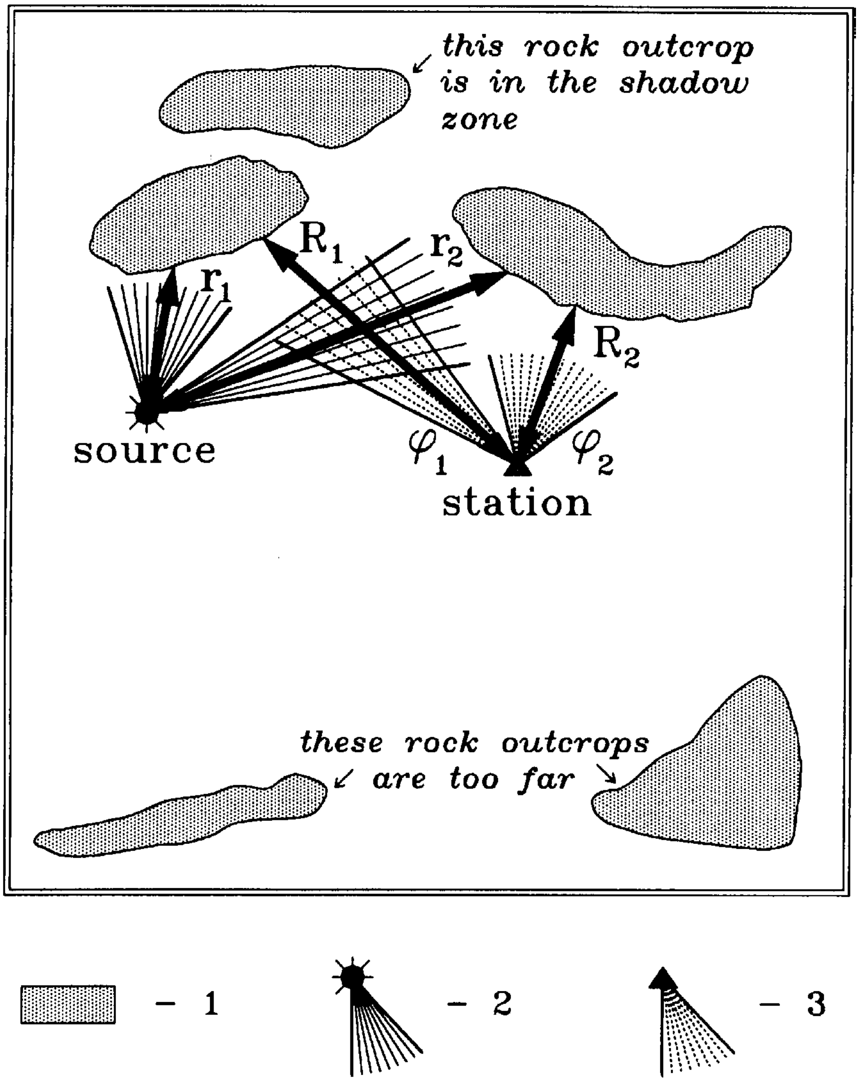

Fig. 4 A representation of R and φ ("1" shows the basements rocks on the earth's surface; "2" is the angle, subtended at the source by the rocks from which reflection occurs; the solid lines represent the direct waves; "3" is the angle, subtended at the recording station by the surface of the rocks from which reflection occurs; the dashed lines represent the waves reflected by the rock towards the station)

The reflection of waves back into the valley, and the conversion of body waves into "surface" waves at the boundaries of a sedimentary basin, suggest that the parameters describing some horizontal dimension of sedimentary basins should play a role in the description of the duration. The resulting prolongation of motion at stations, which are situated on sediments, has been studied by Novikova and Trifunac, while considering two parameters. One is the horizontal distance R (Figures 4 and 5) from the station to the basement rocks, appearing on the surface and producing reflections. The second parameter, φ, is the angle with which the reflecting surface of the rocks can be seen from the station. This parameter describes the "efficiency" of these reflections.

To scale the prolongation of motion in terms of R and φ, Novikova and Trifunac considered the energy of the waves reflected by individual "rocks" towards the station. The resulting equation ("energy equation"), however, is too complex to be considered in this review. Instead, we describe a simplification that ignores the geometrical spreading and attenuation. Then, the "energy equation" can be simplified to (Novikova and Trifunac, 1994a, 1994b)

$$R\varphi = \sum_i R_i \varphi_i \tag{6}$$

which states that the reflecting surface of the "fictitious" rock should be equal to the sum of the reflecting surfaces of the individual rocks.

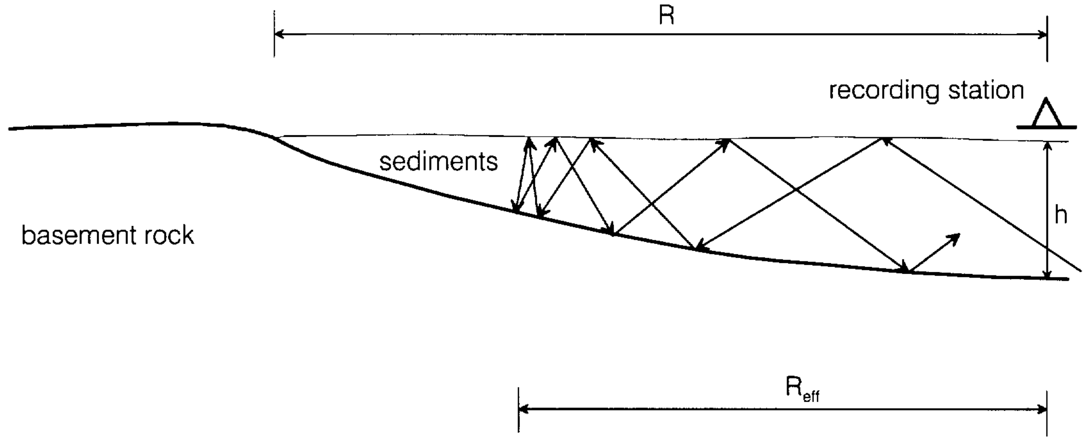

Fig. 5 Reflection of trapped waves from the edge of a sedimentary valley (h is the depth of sediments under the station, R is the distance to the reflecting rock, as it is seen on the earth's surface, and R_{eff} is the "effective" distance from the station to the region where the reflection actually occurs)

A pulse, reflected by a rock in the direction of the station, spends more time in the medium than a direct pulse; it travels distance $R_i + r_i$, and arrives at the station later. The "fictitious" rock has to be positioned at such a distance R from the station, that the delay of the pulses, reflected by it, represents some "proper" combination of the delays of pulses, coming from the individual rocks. The simplified "delay equation" is of the form (Novikova and Trifunac, 1994b)

$$R = \sum_i R_i \frac{R_i \varphi_i}{\sum_j R_j \varphi_j} \tag{7}$$

The increase of φ leads also to an increase of the duration of the reflected pulses, because a larger reflecting surface increases the azimuth and time windows of the sampled wave train. This suggests that τ_{region} is an increasing function of φ. The presently available data suggest that the dependence of τ_{region} on φ can be approximated by a linear function.

The dependence of τ_{region} on R is more complex. For small R, the time intervals, which correspond to the initial pulse (of duration dur_1) and to the reflected pulse (of duration dur_2), will be observed at the station almost simultaneously, without producing significant increase in the duration. For larger R, the time delay between the two pulses causes a complete separation of the corresponding intervals of strong motion, and the total duration is longer and equal to $dur_1 + dur_2$. Further increase of R causes an increase of the time, the reflected waves spend travelling through a dispersive medium. This causes increase of dur_2, and results in further prolongation of the total duration, $dur_1 + dur_2$. For large R, the second pulse, generated by reflection from a remote rock, experiences strong attenuation, and is so weak that it cannot be noticed relative to the background noise of the scattered waves. Therefore, two ranges of horizontal characteristic dimension exist: "small" R, where duration grows with increasing R, and "large" R, where the effect is opposite. A simple way to describe such dependence on R is to use,

$$\tau_{\text{region}}(R) = const_1 + const_2 R + const_3 R^2 \tag{8}$$

for some $const_i$, $i = 1,2,3$ (different at different frequency bands). We expect that $const_3 < 0$.

The depth of sediments, h, plays an important role in scaling various characteristics of strong earthquake ground motion (Trifunac and Lee, 1990; Lee, 1991). Studies of the influence of the depth of sediments on the duration of strong motion were also performed. In the more recent studies, a parabolic dependence has been assumed

$$\tau_{\text{region}}\left(h\right) = const_4 + const_5 h + const_6 h^2 \tag{9}$$

where, $const_i$, $i = 4,5,6$ depend on frequency. $\tau_{\text{region}}(h)$ should be similar to $\tau_{\text{region}}\left(R\right)$, with the difference in scale, since h describes the dimension perpendicular to the predominant direction of the strong motion wave propagation. The wave cannot penetrate to the edge of the shallowing basin, and refracts and reflects back into the valley (see Figure 5). The length of this "penetration" depends on the geometry of the basin, on the wavelength of the wave, and on the incident angle. The "effective" horizontal distance, R_{eff}, from the station to the region of reflection, is difficult to be determined, and so, Novikova and Trifunac employed R and h in all equations, and accounted for the reduction of distance via empirical regression coefficients. Fixing the horizontal dimension, R, and changing the depth of the sediments, h, changes R_{eff} also. The appropriate way to account for this coupling is to combine these effects by adding a coupling term. Recalling the contribution from the angle of reflection φ, the final form of τ_{region} (R, h, φ) becomes

$$\begin{aligned}\tau_{\text{region}}(R,h,\varphi) = a_5 \cdot h + a_6 \cdot R + a_7 \cdot hR \\ + a_8 \cdot R^2 + a_9 \cdot h^2 + a_{10} \cdot \varphi\end{aligned} \tag{10}$$

where, a_i, i = 5 to 10 are regression coefficients. The numbering of these and of all other coefficients in this paper has been chosen to maintain consistency with all previous works on this subject (Novikova and Trifunac, 1993a, 1993b, 1993c, 1994a, 1994b).

4. Effects of the Local Soil

At high frequencies (short wave lengths), τ_{region} should include additional terms representing the effects of local soil conditions. For consistency and continuity with previous work of Novikova and Trifunac, the soil conditions can be characterized by the parameter s_L only (s_L = 0 for “rock”, s_L = 1 for stiff soil sites, and s_L = 2 for deep soil sites (Trifunac, 1990)). In future, when more site-specific data becomes available, it will be possible to refine this classification and to consider more continuous soil site variables.

In the present database, any attempt to include s_L in the equations dealing with parameters h, R and φ, fails because of instability of the solution of the regression equations (due to the lack of data on s_L). At the present time, the effects of the local soil conditions can be studied, if the geology of the recording site is modeled in a simplified manner, using qualitative variable s (s = 0 for sediments, s = 2 for basement rock, and s = 1 for intermediate sites, see Trifunac (1990)). In this paper, we illustrate a duration model, where both "geological" parameter s (characterizing a site on the scale of kilometers) and local soil conditions parameter s_L (geotechnical description on a scale of tens of meters) are used simultaneously.

REGRESSION MODELS

1. Modeling in Terms of Magnitude, Epicentral Distance and Geometry of a Sedimentary Valley

A variety of regression models of the duration of strong ground motion have been presented by Novikova and Trifunac (1993a, 1993b, 1993c, 1994a, 1994b, 1994c, 1995a, 1995b, 1995c). Most of these are constructed starting with the "basic" model, which scales the duration in terms of the earthquake magnitude and the epicentral distance. The "basic" duration is a sum of τ_0 and τ_Δ, the duration of the rupture process and the prolongation due to propagation path effects, as discussed in the previous section. A more complete description also considers regional effects, as included in the function τ_{region}. In what follows, we illustrate a model, which uses the most "complete" representation, given by Equation (10). In this model, the duration at frequency f is represented by

$$\begin{Bmatrix} dur^{(h)}(f) \\ dur^{(v)}(f) \end{Bmatrix} = \begin{Bmatrix} a_1^{(h)}(f) \\ a_1^{(v)}(f) \end{Bmatrix} +$$

$$a_2(f) \cdot \bar{M} + a_3(f) \cdot \bar{M}^2 + a_4(f) \cdot \Delta +$$

$$\begin{Bmatrix} \left[\begin{matrix} a_5^{(h)}(f) \cdot h + a_6^{(h)}(f) \cdot R + a_7^{(h)}(f) \cdot hR \\ + a_8^{(h)}(f) \cdot R^2 + a_9^{(h)}(f) \cdot h^2 + a_{10}^{(h)}(f) \cdot \varphi \end{matrix} \right] \\ \left[\begin{matrix} a_5^{(v)}(f) \cdot h + a_6^{(v)}(f) \cdot R + a_7^{(v)}(f) \cdot hR \\ + a_8^{(v)}(f) \cdot R^2 + a_9^{(v)}(f) \cdot h^2 + a_{10}^{(v)}(f) \cdot \varphi \end{matrix} \right] \end{Bmatrix} \tag{11}$$

where, the epicentral distance, Δ, the depth of sediments, h, and the distance to the reflecting rocks, R, are measured in kilometers. The angle φ is measured in degrees, and $\bar{M} = \max\{M, M_{\min}(f)\}$, where $M_{\min}(f) = \dfrac{-a_2(f)}{2a_3(f)}$. $\bar{M}$ is introduced to keep the duration of strong motion a non-decreasing function of magnitude. The expression in the square bracket in Equation (11) represents the frequency-dependent term τ_{region}, when it is given by Equation (10), and $\left[\tau_{\text{region}}(f)\right]$ = max $\{0, \tau_{\text{region}}(f)\}$. The values of R, h and φ are assumed to be zero, if the station is located on rock. Two different sets of coefficients, $\{a_i^{(h)}(f),\ i = 1, 5 \text{ to } 10\}$ for the horizontal components, and $\{a_i^{(v)}(f),\ i = 1, 5 \text{ to } 10\}$ for the vertical ones, are considered.

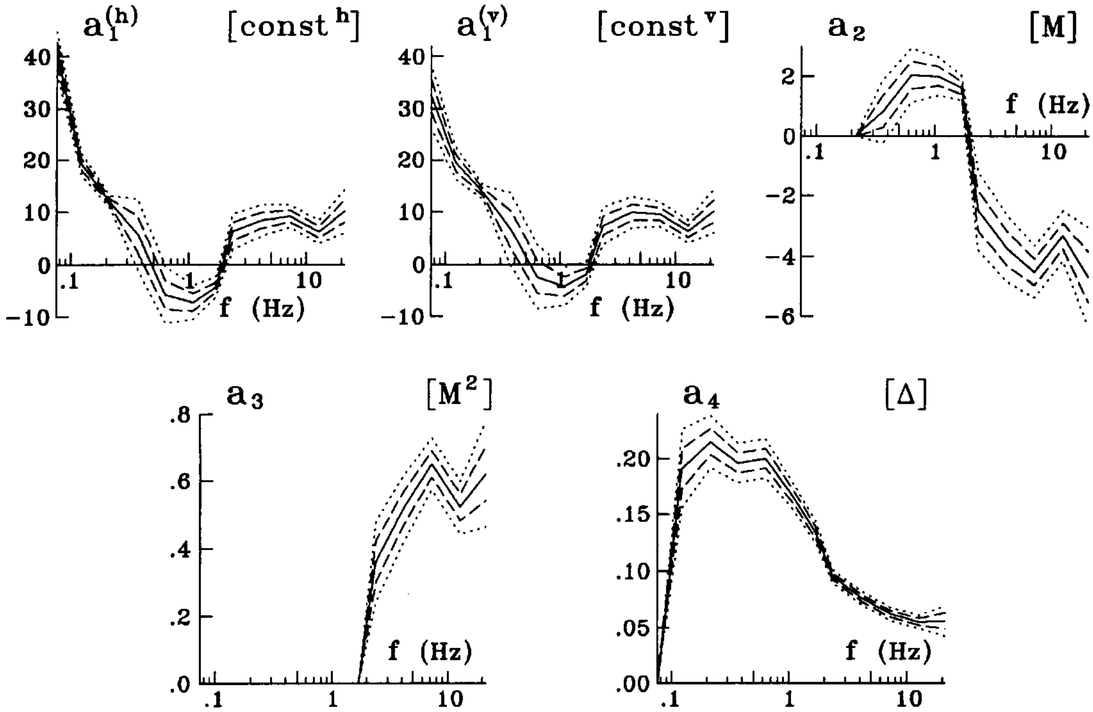

Fig. 6(a) Coefficients a_1 through a_4 in Equation (11), versus central frequency of the channels (solid lines) (the dashed lines indicate the "σ-intervals" and the dotted lines indicate the 95% confidence intervals)

Figures 6(a) and 6(b) show the dependence of the coefficients of the regression model on frequency (solid lines), and their reliability in terms of the "σ-interval" and the 95% confidence interval. The coefficients $\{a_i(f),\ i = 1 \text{ to } 4\}$ show how the duration depends on the magnitude and on the epicentral distance. No dependence of the duration on magnitude M can be detected at low frequencies, because the dimension of the source is smaller than the wavelength of the seismic waves at these frequencies. As the frequency increases, first linear, and then quadratic dependence on M can be observed.

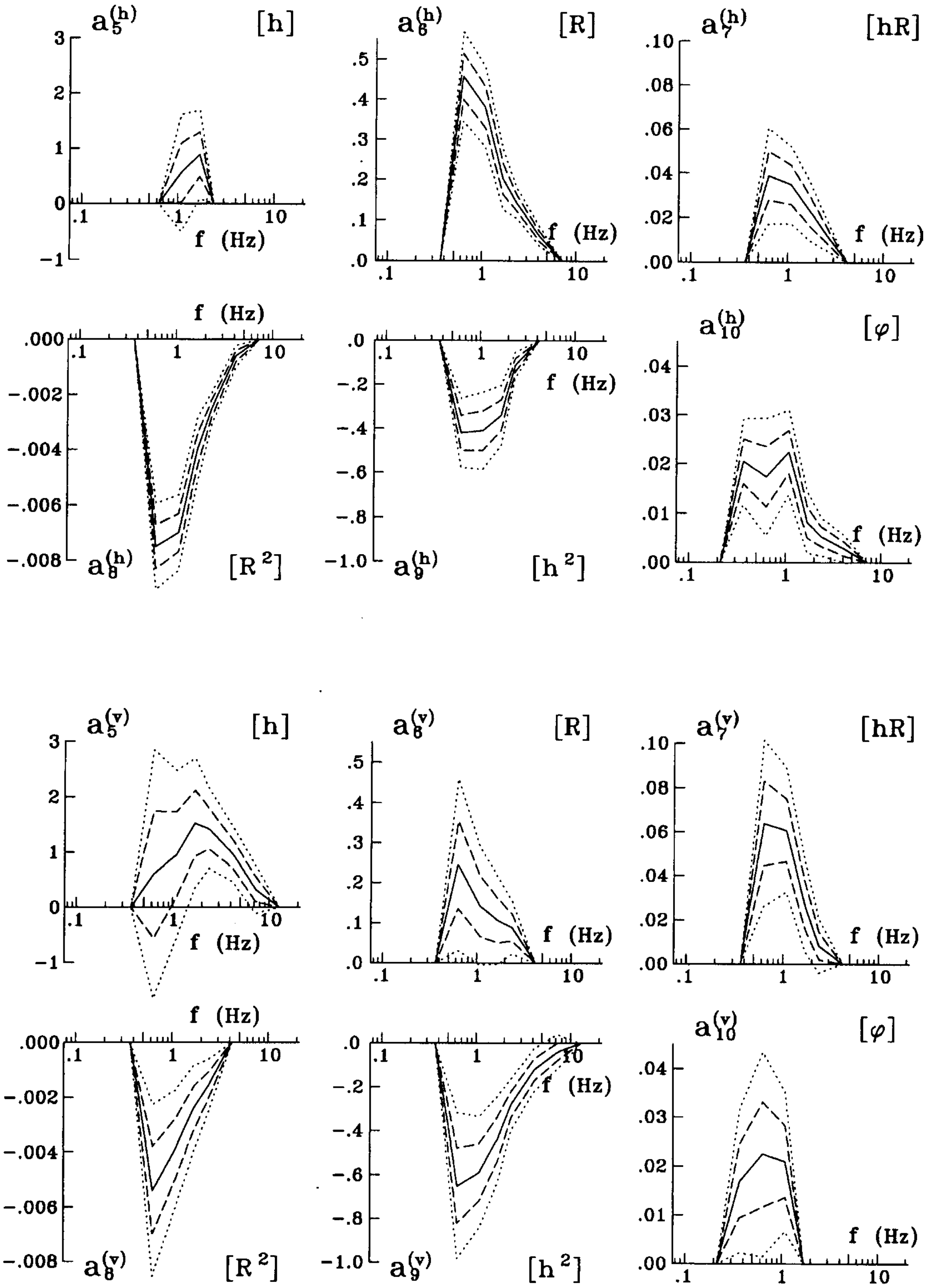

Fig. 6(b) Coefficients a_5 through a_{10} in Equation (11), versus central frequency of the channels (solid lines) (the dashed lines indicate the "σ-intervals" and the dotted lines indicate the 95% confidence intervals)

The coefficients $\{a_i(f), i = 5 \text{ to } 10\}$ describe the influence of the geometry of the sedimentary basin. The coefficient $a_{10}(f)$ represents the "strength" of the horizontal reflections, measured by the angle φ. The coefficients scaling the contributions of the quadratic terms R^2 and h^2 are negative. Thus, the duration increases for intermediate values of R and h, and there is no increase for small or large R and

h. Figure 7 shows isolines (in seconds) of the positive contribution to the overall duration, predicted by Equation (11), made by the following sum of terms involving R and h:

$$\begin{Bmatrix} \tau_{\text{region}}^{(h)}(R,h) \\ \tau_{\text{region}}^{(v)}(R,h) \end{Bmatrix} = \begin{Bmatrix} \begin{bmatrix} a_5^{(h)} \cdot h + a_6^{(h)} \cdot R + a_7^{(h)} \cdot hR \\ + a_8^{(h)} \cdot R^2 + a_9^{(h)} \cdot h^2 \end{bmatrix} \\ \begin{bmatrix} a_5^{(v)} \cdot h + a_6^{(v)} \cdot R + a_7^{(v)} \cdot hR \\ + a_8^{(v)} \cdot R^2 + a_9^{(v)} \cdot h^2 \end{bmatrix} \end{Bmatrix} \tag{12}$$

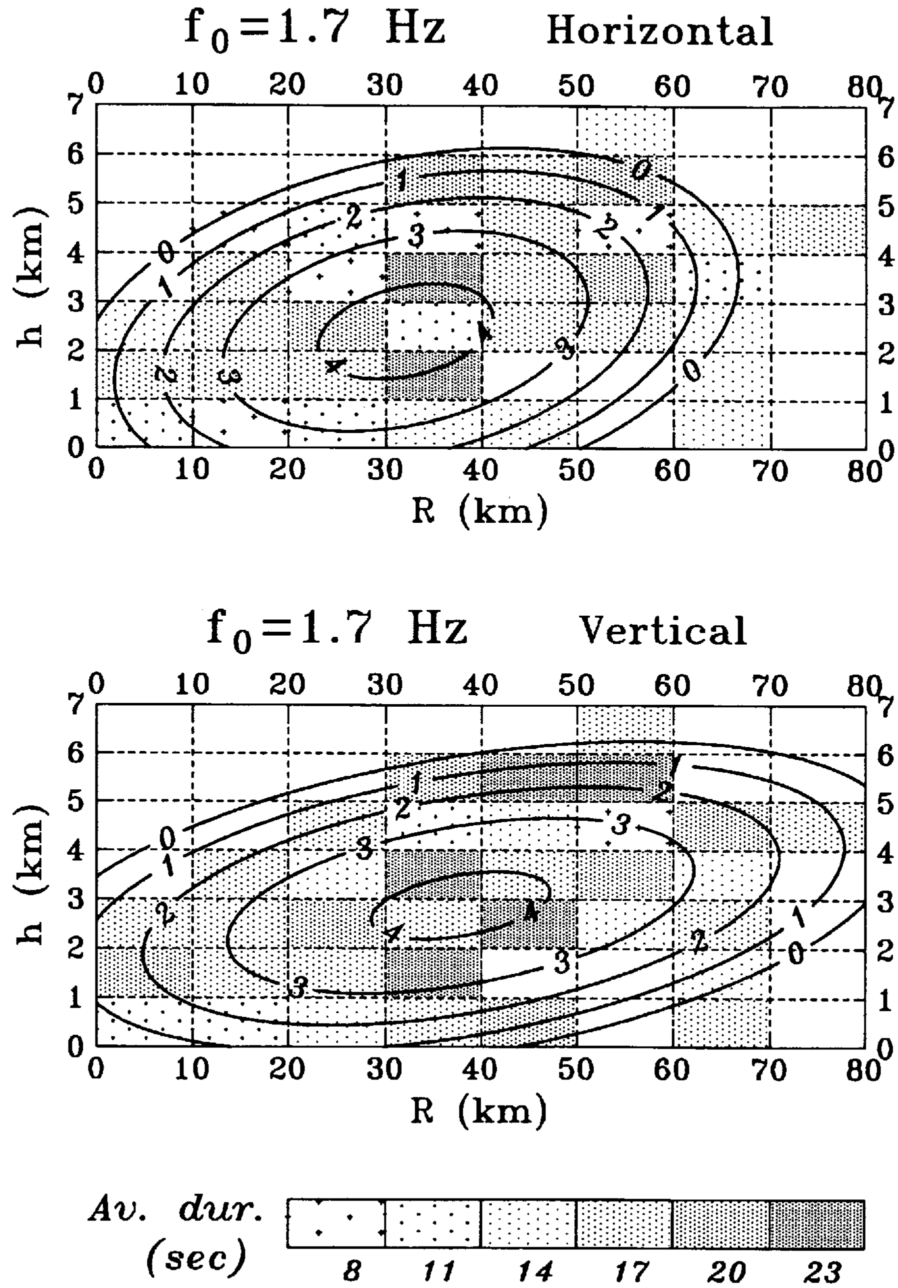

Fig. 7 Isolines of prolongation of motion (in seconds) defined by Equation (12) (the observed duration is shown averaged over ranges of R and h, as specified by the dashed mesh)

Figure 7 also shows the observed duration, averaged over the ranges of R and h. The darker shade corresponds to longer observed durations. The coefficients representing the prolongation of motion by the specific shape of the valley, are distinct from zero in the intermediate frequency range only. At low frequencies ($f \leq 0.3$ Hz), all the coefficients $\{a_i(f),\ i = 5 \text{ to } 10\}$ are equal to zero, and no influence of the sedimentary basin can be noticed. At channel # 4 ($f_0 = 0.37$ Hz), the prolongation is expressed by a term involving φ only. The angle of effective reflection, φ, appears to be more sensitive to long waves. At $f = 0.63$ to 2.5 Hz, the geometrical properties of the sedimentary basin "work" in full strength, and all the terms in τ_{region} have non-zero values. It is interesting to notice that the range of parameters R and h where τ_{region} $(R, h) > 0$, preserves itself for both components in the frequency range $f = 0.63$-2.5 Hz. This effect may be similar in nature to the independence of the amplification factor from the frequency of motion inside the interval $f_2 < f < f_1$ (Trifunac, 1990). Inside the frequency range where τ_{region} $(R, h) > 0$, the value of τ_{region} (R, h), however, does depend on frequency. The maximum possible contribution of τ_{region} (R, h) changes from 7.5 s to 2.5 s for the horizontal components, and from 5 s to 3.5 s for the vertical component, when the frequency changes from 0.63 Hz to 2.5 Hz. After the transition range (4.2 to 7.2 Hz), where some effects of the geometry of the sedimentary basin can still be noticed, the short wave range ($f > 5.0$ to 8.5 Hz) sets in. For these frequencies, no influence on the duration by the presence of a sedimentary basin can be observed.

We consider next the differences between the coefficients for the horizontal $\{a_i^{(h)}(f),\ i = 5 \text{ to } 10\}$ and vertical $\{a_i^{(v)}(f),\ i = 5 \text{ to } 10\}$ components. The coefficients, $a_6^{(h)}(f)$ and $a_8^{(h)}(f)$, which scale the influence of the parameter R on the horizontal component of motion, are better defined and can be followed in a wider frequency range than their vertical counterparts, $a_6^{(v)}(f)$ and $a_8^{(v)}(f)$. Conversely, the coefficients that describe the contribution of the depth of sediments h to the duration of the horizontal component, i.e. $a_5^{(h)}(f)$ and $a_7^{(h)}(f)$, have larger variances, and are distinct from zero in a narrower frequency range, compared to $a_5^{(v)}(f)$ and $a_7^{(v)}(f)$. Thus, the strong motion duration of the horizontal components appears to be more sensitive to the horizontal characteristic dimension R, while the duration of the vertical component "feels" the depth of sediments under the station, h, better, than it "feels" R. The parameter R describes the geometry of the basin on a large scale, while h gives a more local description in terms of the depth of sediments right under the recording station.

The angle φ with which the reflecting rocks are "seen" from the station, is a measure of the contributions to the duration from the horizontal reflections. The characteristic "dimension" of these reflections is described by R. Both φ– and R–related coefficients are better defined for the horizontal components. The typical values obtained for $a_0(f)$ give an increase in duration by about 2 s for $f \approx 0.37$ to 1.1 Hz, and by about 0.5 s and less for $f \approx 2.5$ to 1.1 Hz, per each 100° of φ.

2. Modeling in Terms of Magnitude, Epicentral Distance and Geological and Local Soil Conditions

To examine the influence of the local soil conditions on the duration of strong ground motion, the sites are divided into three groups according to the value of the soil parameter s_L. Deep soil sites have $s_L = 2$, stiff soil is designated as $s_L = 1$, and $s_L = 0$ stands for a "rock" site. Following Novikova and Trifunac (1994a), the qualitative indicator variables ("geological" parameter s, and "local" soil parameter s_L) are included in the model equation in the form, $a_{11}(f)\cdot S_L^{(1)} + a_{12}(f)\cdot S_L^{(2)} + a_{13}(f)\cdot S^{(1)} + a_{14}(f)\cdot S^{(0)}$, where $S^{(1)}$ and $S^{(0)}$ are indicator variables for $s = 1$ and $s = 0$ ($s = 0$ for sediments, $s = 2$ for basement rock, and $s = 1$ for intermediate sites), and $S_L^{(1)}$ and $S_L^{(2)}$ are the indicator variables for $s_L = 1$ and $s_L = 2$. However, a substantial reduction in the number of available data points (because of the lack of information about S_L for many sites), causes instability in the regression analysis. Therefore, Novikova and Trifunac reduced the number of unknown coefficients (to "improve" numerical stability) by treating the parameter s as a regular quantitative "continuous" variable.

The model that follows is then

$$\begin{Bmatrix} dur^{(h)}(f) \\ dur^{(v)}(f) \end{Bmatrix} = \max\left[\left(\begin{Bmatrix} a_1^h(f) \\ a_1^v(f) \end{Bmatrix} + a_2(f)\cdot M\right), 1\right] + a_4(f)\cdot\Delta + a_{15}(f)\cdot(2-s) + a_{11}(f)\cdot S_L^{(1)} + a_{12}(f)\cdot S_L^{(2)} \tag{13}$$

To avoid negative values of the duration, at locations close to a source, for small magnitude events, it is assumed that the duration at the source should not be less than 1 s. The reason to consider the term $a_{15}(f)\cdot(2 - s)$ instead of $a_{15}(f)\cdot s$ is that it is convenient to have basement rock as a reference, and to deal with positive a_{15}, if the duration on sediments is longer than that on the rock sites.

The results of the regression analysis with Equation (13) are shown in Figures 8(a) and 8(b). The coefficient $a_{15}(f)$, responsible for scaling the influence of the epicentral distance Δ, does not change, when compared to other related models. The coefficients $a_1(f)$ and $a_2(f)$ have different meanings now due to a linear approximation of τ_0. For low frequencies ($f_0 = 0.37$ Hz), the duration of motion on sediments ($s = 0$) is about 4 s longer, than that on rock sites. At high frequencies, $a_{15}(f)$ is not well defined (the condition $|\sigma_{15}(f)/a_{12}(f)| < 1$ is hardly satisfied). The inequality shows that the duration of strong motion is longer on deep soil ($s_L = 2$) and shorter on "rock" sites ($s_L = 0$), with stiff sites being in the middle. Also, the influence of the local soil conditions on the duration can be noticed at higher frequencies, compared to the influence of the geological conditions.

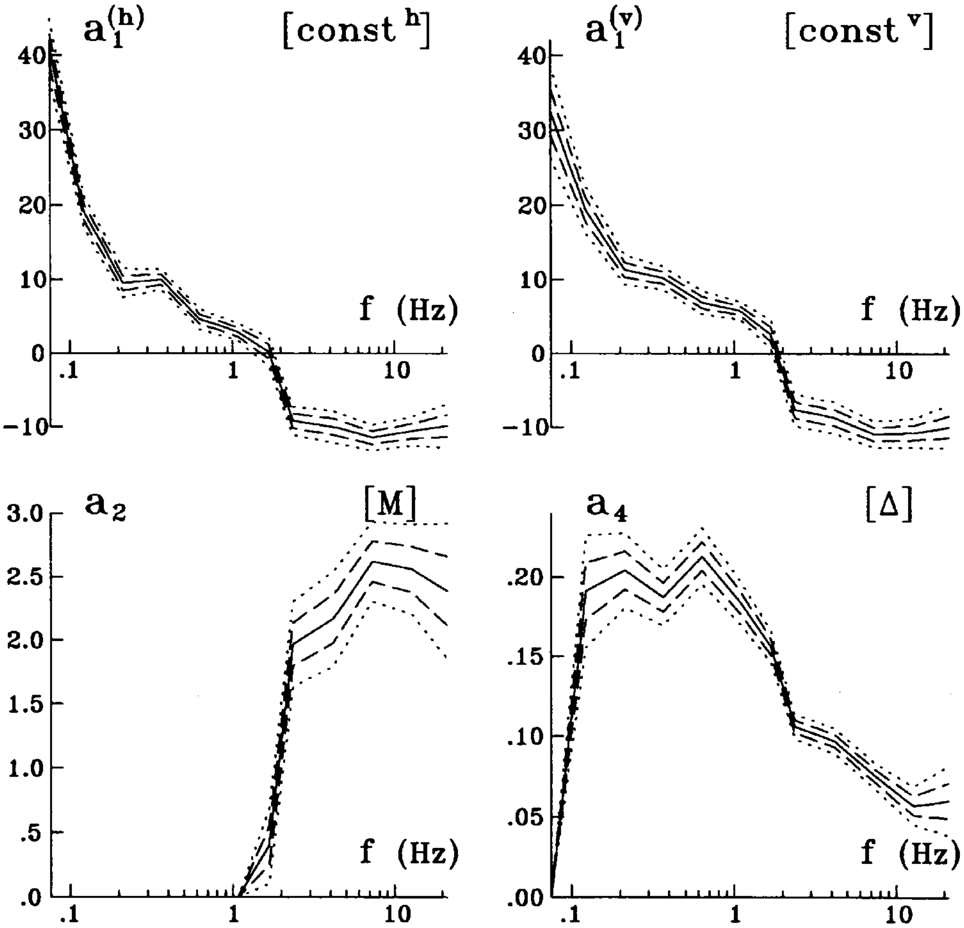

Fig. 8(a) The coefficients $a_1(f)$, $a_2(f)$ and $a_4(f)$, in the model in Equation (13) plotted versus central frequency of the channel (solid lines) (the dashed lines indicate the "σ – intervals", and the dotted lines indicate the 95% confidence intervals)

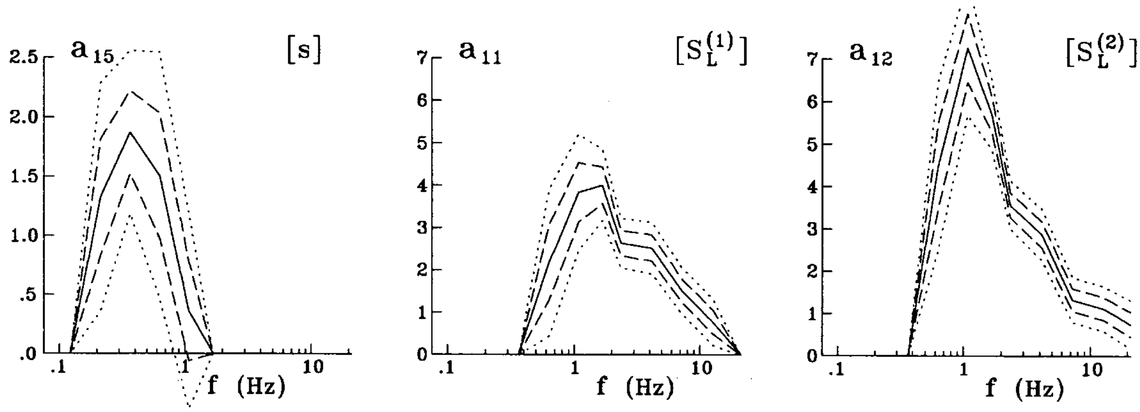

Fig. 8(b) The coefficients $a_{15}(f)$, $a_{11}(f)$ and $a_{12}(f)$ in the model in Equation (13) plotted versus central frequency of the channel (solid lines) (the dashed lines indicate the "σ - intervals", and the dotted lines indicate the 95% confidence intervals)

DISCUSSION AND CONCLUSIONS

The models reviewed in this paper suggest the physical mechanisms, which may be responsible for the prolongation of strong ground motion at sites located on sediments. However, the prolongation mechanisms discussed here (e.g., reflection of the waves at the boundaries of a valley back inside the valley) represent only one possible way to interpret these observations.

Consider, for example, the case shown in Figure 2. The earthquake source generates body (and surface) seismic waves. Some body waves penetrate deep into the crust, and reach the recording station from below. Some body waves are converted into surface waves at the first boundary rock-sediments, at a distance ρ_1^r from the source. Together with the surface waves generated in the epicentral region, these waves propagate through the sedimentary valley, and reach the boundary at epicentral distance $\rho_1^r + \rho_1^s$. There, they are partially reflected back into the valley (this surface wave energy is "trapped" in the valley), and partially continue to propagate away from the source in the form of body and surface waves. Similar processes of surface-body wave conversions and reflections from the edges of the valley repeat in each sufficiently deep alluvial valley. As a result, a complex picture of overlapping strong motion pulses consisting of body and surface waves is recorded at the station.

When the three-dimensional geometry of the region is irregular, and when many factors contribute to the formation of the signal at the location of the station, it is difficult to decide how to describe these factors by using just a few parameters which can be included in the regression analysis. For example, Novikova and Trifunac considered the percentage of epicentral distance covered by rocks on the earth surface $\xi = \left(\rho_1^r + \rho_2^r\right)/\Delta$, and looked at $\tau_\Delta = \left(a_4^{(1)} + a_4^{(2)}\xi\right)\cdot\Delta$ instead of $\tau_\Delta = a_4 \cdot \Delta$. It appears that, per each 10 km of epicentral distance at frequencies near 0.3 Hz, the motion is prolonged by 2.5 s if $\xi = 0$ (the direct surface path from the source to the station does not cross any rocks), and by only 0.8 s if $\xi = 1$ (the epicenter and the station are located in the same rock outcrop). We explain this by dispersion along the path through alluvium. Choosing ξ as one of the parameters describing the geometry of the region is only an approximation. If the deep source and the recording station are separated by several valleys, the strong motion energy observed at the site could be coming from direct body waves and from surface waves generated at the edges of the valley, where the station is located. In this case, one might consider the percentage of path covered by alluvium valleys located far from the station. If the source in Figure 2 is shifted to a greater depth, this parameter can be expressed as the ratio ρ_2^s/Δ. Another parameter, useful in the consideration of shallow sources, is the number of boundaries rock-sediments and sediments-rock, crossed by the waves on their way to the recording station. It may influence, not only the duration of strong motion, but also the amplitude of Fourier spectrum observed at the site. Large number of boundaries would weaken the signal. Also, the presence of several valleys reduces the amplitudes of strong motion, because some energy is "trapped" in these valleys and cannot reach the recording station.

When a detailed description of the geological conditions at the site is not available, the parameter s can be used instead. The importance of considering the local soil together with the geological site conditions was also demonstrated. The influence of the geological and soil site conditions on the duration of strong ground motion prevails at different frequencies. The duration can be prolonged by 3.5 sec at a site located on a deep sedimentary layer at frequencies about 0.5 Hz, and by as much as 5 to 6 sec due to the presence of soft soil underneath the station at frequency of about 1 Hz.

The results of this and of all cited works can be used for the prediction of the duration of shaking expected during future earthquakes, when the parameters of the shock (M and Δ, or I_{MM}) and the site (R, φ and h, or s and s_L) can be specified. However, we note that the equations and the regression coefficients can be employed for such a prediction, only in the region where the data were recorded (western U.S., and primarily southern California). A different geological environment may be associated with different earthquake mechanism, distribution of hypocentral depths of the sources, velocities and attenuation factors, thus changing the values of the regression coefficients (Novikova et al., 1993).

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