Long period microtremors, microseisms and earthquake damage: Northridge, CA, earthquake of 17 January 1994

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Abstract

Contour maps of site amplification at long periods (3–10 s), estimated from recorded ambient noise in San Fernando Valley, are compared with simple indicators of the strong motion amplitudes and damage from the 1994 Northridge, CA, earthquake. The results show that there is no simple correlation between these two sets of observations. Ambient noise measurements have been successful in identifying the model parameters of full-scale structures, and may become successful in identifying parameters of soil and sedimentary layers that are related to strong motion site amplification in the near field, but new methods of measurement and analysis are needed to achieve that.

Keywords: Northridge earthquake; Microtremors; Microseisms; Ambient noise; Site amplification; Site response; Microzonation; Seismic hazard; Seismic hazard mapping

1. Introduction

A popular, albeit simplistic, hypothesis is that the degree of damage caused by earthquake shaking is larger when the amplitudes of ground motion are larger, and when the ground motion has significant energy near the period of the structure. Another popular view is that the site has its own "predominant period" $T = 4H/v_s$, where $H$ and $v_s$ are the thickness and average shear-wave velocity in the top soil layer. This hypothesis is based on a model of a "soft" elastic surface layer over elastic half-space, excited by vertically incident shear-waves. Constructive interference of the incident and reflected wave from the "free" surface causes amplification of the motion at the surface, at periods $T_n = (4H/v_s)/(2n - 1)$, $n = 1, 2, 3, ...$. The “predominant period” is the longest of these periods. Starting with such simple models, numerous investigators tried to interpret, map, and find practical use of the spatial distribution of “predominant period of strong ground motion” [1–4] and of the inferred relative amplification of the incident waves by the local soil and geologic conditions.

As destructive earthquakes occur infrequently, to interpret patterns of damage from past events and to “forecast” the effects from future events, alternative sources of excitations have to be sought, such as large distant earthquakes, small near earthquakes, explosions, aftershocks, microseisms and microtremors. The motions at a site from such sources have different nature from damaging earthquake motions, e.g. different amplitudes, duration and frequency content (see Fig. 1 for differences in amplitudes of Fourier spectra of acceleration). Besides the differences in the nature of the sources, the radiated waves reach the site via different propagation paths, and therefore carry information on different parts of the geologic structure, between the source and the site, and near the site. In the studies using such alternative sources it has been assumed that these differences may be ignored or accounted for.

The Northridge earthquake of 17 January 1994 ($M_L = 6.4$, epicenter at 34.21°N 118.55°W, focal depth 18 km) offered an excellent opportunity to test different hypotheses of predicting site amplification and earthquake damage. Though of moderate size, it occurred beneath a densely populated metropolitan area (San Fernando Valley of metropolitan Los Angeles) and caused extensive damage. Ground motion was recorded by more than 200 strong motion stations [5], many of which were in the near-field. The damage was documented by the City of Los Angeles, Department of Water and Power [6], and by Office of Emergency Services [7], and it was reviewed and summarized in Ref. [8]. Following the main event, weak motions from many small aftershocks were recorded by temporary arrays. These were used to estimate site amplification [9–11], and
to study the correlation of those with the observed damage from the main event. Site amplification factors were also estimated from coda waves of Northridge aftershocks in San Fernando Valley [9] and from other earthquakes in Southern California [12]. Finally, long period microseisms (caused by ocean waves), and short period microtremors (caused by local traffic and cultural noise), were recorded in the same area before and after the Northridge earthquake [13–15] and were used to map site amplification.

In another paper [16], we reviewed the site amplification studies for San Fernando Valley, Los Angeles and Santa Monica estimated from weak Northridge aftershock motions [9–11] and from strong motion [17], and studied the agreement of the results of these studies with each other and with the observed patterns of damage from the Northridge main event. In this paper, we do a similar comparison, but for site amplification estimated from ambient noise. Kagami et al. [14] evaluated the mean Fourier spectral ratios of 3–5 s microseisms recorded at a grid of 50 sites, 3 km apart. Dravinski et al. [15] used a grid of 74 sites, 2 km apart, and computed the average Fourier spectrum amplitudes (for periods 4–10 s) of horizontal and vertical microseism motions. Then, they computed (a) the ratio between the average spectra for horizontal motions recorded on sediments and the average of spectra for horizontal motion recorded at a reference (rock) site, and (b) the ratio of the average spectral amplitudes for horizontal motion (between 4 and 10 s) and spectra of vertical motions recorded at the same site. Kagami et al. [14] suggest that their results can be “regarded as an empirical qualitative estimate of the amplification effects of the sedimentary layers present in the valley”. In this paper, we examine the validity of this claim, by comparing their results (i) with strong motion estimated from recorded data and (ii) with its consequences in San Fernando valley. Also, we try to explore in general the usefulness of measuring ambient noise.

2. Microtremors and site effects

In this section, we review some of the basic features of microtremors, as to how they compare with strong earthquake shaking, and we re-examine the validity of the assumptions in procedures that use microtremors to estimate site amplification.
2.1. Amplitudes and spectral characteristics of microtremors

Fig. 1 illustrates Fourier spectra of acceleration of microtremor motion, and the spectra of earthquake strong motion for magnitudes $M = 1$ to $7$, at 10 km distance from the source. These were calculated using empirical regression models, and their extensions for long periods (where the signal to noise ratio for recorded strong motion is $<1$ [18]). Spectra corresponding to destructive strong motion are highlighted. The heavy line (for $M = 6.6$) illustrates motions in San Fernando Valley from the Northridge earthquake. These motions had most energy in the frequency range 0.2–10 Hz. Spectra corresponding to motions of Northridge aftershocks, used by Gao et al. [9], Hartzell et al. [10] and Field and Hough [11], are also highlighted. The Fourier amplitudes of these motions are 10–100 times smaller than those for the main event, and the frequencies used in these studies are mainly in the range 1–10 Hz. The Fourier amplitudes for the ambient noise are even smaller, by 5 orders of magnitude in very noisy areas and by 7 orders of magnitude in quiet areas.

The Fourier spectra of ambient noise usually have two characteristic peaks, one at 0.07 Hz and the other one at 0.14 Hz. The energy near 0.07 Hz is believed to result from the action of ocean waves along the coastline, while the source of the larger peak near 0.14 Hz has been ascribed to pressure caused by standing ocean waves [19]. These waves are also called microseisms. The higher frequency waves in ambient noise are caused by wind and by random sources of cultural noise (traffic, factories, ...) and are believed to consist mainly of high frequency Rayleigh waves propagating through the shallow soil layers. These waves are also called microtremors.

2.2. Procedures and assumptions in site effects prediction from ambient noise

2.2.1. Common procedures

There are two frequently used procedures to estimate the site effects (predominant period and site amplification). The first procedure uses the ratio of spectra of motions recorded at a sediment site and at a reference “rock” site, recorded simultaneously. The second procedure does not rely on measurements at a reference site, but uses the ratio of spectra of horizontal motion and vertical motion recorded at the same site, simultaneously [20]. It is assumed that this ratio is capable of identifying the site amplification and the predominant period of the site [21,22]. In the notation of Dravinski et al. [15], we define those two
spectral ratios as

\[ K_{SR}(\omega) = H_S(\omega)/H_R(\omega) \]  

(1)

\[ N(\omega) = H_S(\omega)/V_S(\omega) \]  

(2)

where \(H_S(\omega)\) and \(V_S(\omega)\) are Fourier amplitude spectra of horizontal and of vertical ground noise at the sediment site, and \(H_R(\omega)\) the Fourier amplitude spectrum of horizontal ground noise at the reference (rock) site.

2.2.2. Transfer-function representation

The first assumption in the use of microtremors for prediction of site effects for use in seismic hazard mapping is that the wave paths are “simple” and “essentially the same” (i.e. there are no complex three-dimensional scattering and focusing environments and the path effects can be approximated by the elementary form of geometric spreading and anelastic attenuation). The second assumption is that the response of the ground during both ambient noise and strong earthquake shaking is linear. If these two assumptions are satisfied, then the modification of ground motions at the site and at the source, \(P(f)\) and \(S(f)\) are the transfer functions of the propagation path and of the local site effects, and \(f\) is frequency [22,23].

2.2.3. Equal paths assumption

The second assumption was that the effects of different and complex three-dimensional scattering and focusing along the propagation paths can be neglected or accounted for both in case of strong earthquakes and ambient noise excitation. This condition is often difficult to achieve even in ambient noise experiments. An example of minimum requirement would be to have a reference (rock) site, but even this simple requirement is not adequate in the near-field of the earthquake.

Fig. 2 shows the horizontal projection of the fault surfaces of the two largest earthquakes that shook San Fernando Valley during the past 30 years: the 1971 San Fernando [24] and the 1994 Northridge earthquakes [25]. During the latter, the faulting started at depth (17.5 km the epicenter at 32.21°N and 118.55°W, north of Reseda in Fig. 2), and then propagated to the north and north–west. During the first 6–7 s of strong shaking, a site in San Fernando Valley received waves from many different directions. Consequently, in the near-field, during the most intense shaking, the waves arrived along many different paths, with different azimuths of approach, and with different incident angles relative to the horizontal. This is illustrated in Fig. 2 for a site in the central part of the valley, near the intersection of Roscoe and San Diego Freeway.

Dravinski et al. [15] used a reference site about 25 km away from the center of San Fernando Valley. They assumed that the long period ambient noise was caused mostly from excitation by sea waves along the Pacific coast, and that the long period waves in San Fernando Valley, and at the reference sites, were caused by the “same” train of waves. For periods of motion 3–10 s, the corresponding wavelengths of ground noise are in the range from ~3 to ~20 km. These wavelengths are “short” relative to the horizontal dimensions of San Fernando Valley and the size of Santa Monica and Verdugo Mountains. Therefore significant differences (direction of arrival, lateral reflections, lateral shadow zones) are to be expected between the motion in the center of the valley and at the reference site R2. (This reference site is on basement rock, 9.2 km east of 118°18′W longitude, see Fig. 3b).

2.2.4. Linearity assumption

The first assumption was linearity of response. As Fig. 1 shows, the amplitudes of strong earthquake motion can be orders of magnitude lager than those of ambient noise. During the Northridge earthquake, there was a wide spread evidence of nonlinear response of the near-surface soils in San Fernando Valley and in the northern part of Los Angeles–Santa Monica area (such as numerous breaks in the water pipes, and various signs of surface distress like fissures, slides, grabens, dynamic settlement, …; [26,27]). There is some evidence that, in the areas mapped, the recorded peak accelerations were noticeably reduced by nonlinear response (at Jensen F.P. Administration Building, and at USC stations 53, 3, 9 and 13; [26]). However, this may have been associated with the “high” frequencies only, and the spectra of intermediate and long period amplitudes may have changed only little. Simple foreword theoretical models of nonlinear soil response can produce a variety of trends reflecting the model assumptions. Unfortunately, recorded strong motion data on large (e.g. peak ground velocity \(v_m > 100\) cm/s) nonlinear soil response in San Fernando Valley does not exist. In spite of the evidence of nonlinear response during the Northridge earthquake, the details of its effects on the actual amplitudes of recorded motion are not understood. To learn about these effects, it will be necessary to install and to maintain dense strong motion networks, with density of strong motion stations per unit area, at least 100 times greater than what is available at present (e.g. the density of strong motion stations as in Fig. 3a).

We conclude that the transfer-function representation, given by Eq. (3), might give useful results for long waves and for epicentral distances large relative to the source dimensions (i.e. when point source representation is acceptable), and for linear motions. We saw that, in San Fernando Valley, however, this representation is not valid for the San Fernando and Northridge earthquakes (Fig. 2). It might be
valid approximately for motions arriving from individual asperities, and it may become valid if it is converted to an integral of the contributions from all asperities, with $E(f)$, $P(f)$ and $S(f)$ becoming (nonlinear) functions of, e.g. the position and size of the contributing asperities, orientation of the slip vectors, geometry of the wave paths, azimuth and incident angle of waves arriving to the station, and three-dimensional geometry of the geologic environment surrounding the source wave path and the site.

2.3. Can this work?

It should be obvious that, $S(f)$ for excitation by strong motion waves (for the Northridge earthquake, arriving from depth, almost vertically), must be very different from $S(f)$ for long period surface waves (3–10 s microseisms). In order to determine $S(f)$ for strong motion from $S(f)$ measured from long period microseisms, one should know how these two are related and whether the latter can provide sufficient information to determine the former. Finding this answer is a difficult task and a challenge for future research.

The transfer-function representation may work in some instances, e.g. when local alluvium and soil have very different impedances relative to the surrounding rocks (e.g. San Francisco bay mud, or former lake bed in Mexico City), when the soil material is linear up to high levels of strain (e.g. as suggested by some authors for the clays of the former lake bed in Mexico City), and when for other site specific reasons the site effects ($S(f)$) are “very strong” (compared to effects due to differences in wave paths). In such cases, the linearized representation in Eq. (3) and simplified measurements of ambient noise based on Eq. (2), perhaps may lead to useful results [21].

For typical soils and sediments in southern California, Udwadia and Trifunac [28] compared directly spectral characteristics of strong motion and of microtremors. They concluded that no simple features, e.g. local site conditions, govern the details of strong ground shaking. In Imperial Valley, in El Centro, the effects of the local site conditions appeared to be overshadowed by the source mechanism and the propagation path.

Chávez-García et al. [22] compared spectral ratios from ambient noise motions and from $S$-waves of small earthquakes and explosions, recorded by a temporary array in Parkway Valley in New Zealand. Their results showed good agreement, in spite of the obvious differences in the nature of the waves from which the ratios $K_{SR}$ and $N$ were calculated (body waves for the earthquake data and surface waves for the ambient noise data). They performed a detailed dispersion analysis to find out the reason for the agreement. They concluded that it was due to the fact that the $S$-waves were contaminated with locally generated Rayleigh waves, and due to the coincidence that, for the geology of that site, 1D and 2D effects produced similar results.

For engineering applications, $S(f)$ is needed in a broad frequency range (e.g. $0.1 < f < 30$ Hz). The long periods are required for analyses of tall buildings and very short periods are required for analyses of brittle and stiff equipment. The strong motion records provide data in this entire frequency range and for amplitudes in the range from $M \sim 4$ to $M = 7$ (see Fig. 1). The aftershock studies can contribute information on linear forms of $S(f)$, for $f > 1$ Hz (e.g. Ref. [9]), while analyses of ambient noise can address all the frequencies of interest for earthquake engineering [28], or selected frequencies (e.g. Refs. [14,15]). For linear response amplitudes, $S(f)$ must incorporate amplification of long period motion ($f \leq 2$ Hz) and attenuation ($f \geq 10$ Hz) on “soft” sediment and soil sites [29]. Direct comparison of average estimates of $S(f)$ in different frequency bands is not possible and must be done with appropriate corrections and transformations.

Gao et al. [9] argued that the dominating factor in producing the observed large strong motion amplitudes in Santa Monica during the Northridge earthquake was associated with focusing of waves by the three-dimensional geometry of the geologic basement. Todorovska and Trifunac [30,31] suggested that lateral scattering and diffraction, along with “flow” of strong motion wave energy through the deep sediments of Los Angeles basin (by reflections from the valley floor and from the nearly vertical edges of the valley formed by local faults) influenced the attenuation of strong motion amplitudes significantly. Since focusing, scattering and diffraction are highly dependent on the 3D geometry of the geologic environment, in relation to the orientation of the fault plane and the geometry of the earthquake-station pairs, it is not likely that the observed patterns of amplification during the Northridge earthquake will be repeated in future, and so it is naive to search for some $S(f)$ with linear transfer function properties.

There is no doubt that focusing and local site amplification do play some role in every recorded strong ground motion. But, in the area of strong shaking, where ground motion begins to damage engineering structures (e.g. for peak velocities $v_{\text{max}}$ greater than $\sim 20$ cm/s), the distribution of soil properties begins to modify the effects of linear wave propagation, and for very strong motion ($v_{\text{max}} > 40$ cm/s) may overshadow the final outcome via nonlinear responses [27,32,33]. Beyond certain threshold levels of strong motion amplitudes, the spatial distribution of local susceptibility to nonlinear response may have little to do with the type of incident waves, with direction of their approach, or with initial amplification. Then, mapping susceptibility and threshold amplitudes for nonlinear response of local soil may turn out to be the simplest and the most practical engineering tool for seismic hazard zonation so far. If analyses of microtremors and microseisms can be so developed to help outline this local susceptibility to nonlinear response, simple and expedient measurements of ambient noise will become a valuable engineering tool.

In this paper we compare only the simple and direct results of published work on long period microseisms,
Fig. 3. (a) Comparison of contours of Mean Spectral Ratio (MSR) of microtremor and microseism noise after Kagami et al. [14] (solid lines), averaged over periods 3–5 s and relative to reference site R1, with smoothed contours of PSV in cm/s at $T = 4.4$ s (dashed lines), (b) comparison of contours of $K_{SR}$ ratio after Dravinski et al. [15] (solid line), averaged over periods 4–10 s and relative to reference site R2 (9.2 km east of 118°18'), with smoothed contours of PSV in cm/s at $T = 7.5$ s (dashed lines), (c) comparison of contours of N ratio after Dravinski et al. [15] (solid lines), averaged over periods 4–10 s and relative to smoothed contours of PSV in cm/s at $T = 7.5$ s (dashed lines) and with the “gray” zones where buildings were damaged (red-tagged) during the 1994 Northridge earthquake, after Trifunac and Todorovska [27].
with equally simple (and not always related) analyses of strong motion amplitudes, and of their consequences. Our aim is not to show that simplistic interpretations do not work (this should be obvious without any analysis), but to explore whether some subtle similarities, consistencies and inconsistencies can suggest a useful direction for future research.

3. Observations in San Fernando Valley

3.1. Review of published results for San Fernando Valley

3.1.1. Study I
Kagami et al. [14] measured long period ambient noise between 11 and 15 July 1984, at a grid of 50 sites 3 km apart, in San Fernando Valley, and at a reference point, R1, in Santa Monica Mountains (near the intersection of Mulholland Dr. and Beverly Glen Blvd.). They took all the measurements during the day, for 15 minutes at each site. They used seismometers with $T = 10$ s natural period, and with low-pass filters and amplifiers with approximately constant gain in the operating range from about 1–0.1 Hz [13]. Each record was sampled at a rate of 4 pts/s, and segments 256 s long (1024 samples) were then Fourier transformed. The spectra of motions recorded in the valley were first normalized by the spectra at the reference site. Mean spectral ratios (MSR) were then calculated for periods 3–5 s.

Based on these ratios, Kagami et al. [14] presented contour maps for San Fernando Valley. We redrew their contours in Fig. 3a (the solid lines) against the projection of the fault plane of the Northridge earthquake (the heavier dashed line), locations of the strong motion stations (full and open circles, with the station name or number next to it) our contours of Pseudo Spectral velocity (the dashed lines) based on strong motion recordings, and the distribution of “gray” and “white” zones [27]. Their reference site, R1, is shown by a gray circle, in the bottom of the figure, near USC station 13.

The contours of Kagami et al. [14] indicate 2–3 times larger spectral amplitudes (1) along Coldwater Canyon Ave., between Magnolia Blvd. and Roscoe Blvd. (near USC stations 6 and 9 in Fig. 3a) and (2) in Winetka–Northridge area centered along Wilbur Ave., between Saticoy St. and Lassen St. (north of Reseda, in Fig. 3a). In their paper, they suggest that this contour map “could be regarded as an empirical qualitative estimate of the amplification effects of the sedimentary layers present in the valley”.

3.1.2. Study II
Dravinski et al. [15] measured long period (0.1–2 Hz) ambient ground noise after the Northridge earthquake, at 74 sites 2 km apart, in San Fernando Valley, and at one reference bedrock site, R2, in La Canada (at 34.21 N and 118.20 W, about 25 km east of the valley center). Most of their measurements were completed between 17 and 22 April 1994, and few additional measurements were taken...
Fig. 4. (a–c) Same as Fig. 3 (a–c) but with the gray zones replaced by triangles showing the locations of damaged (red-tagged) buildings.
3.2. Comparison with observations during the Northridge earthquake

In Figs. 3–6, we show overlays of the results of Kagami et al. [14] and of Dravinski et al. [15] with different indicators of ground motion and of damage from the Northridge earthquake. Parts (a), (b) and (c) of these figures correspond to different results of the microtremor studies. In parts (a), we show Kagami et al. [14] contours of MSR, the mean spectral ratios of long period microseisms, averaged over periods 4–10 s and normalized with respect to reference site R1. In parts (b) and (c), we show, respectively, Dravinski et al. [15] contours of $K_{SR}$ and $N$ defined by Eqs. (1) and (2), with R2 as the reference (rock) site.

Computation of $K_{SR}$ ratio requires simultaneous recording of noise at a site on sediments and at a reference site. These sites were separated by about 25 km. They assumed that the recorded noise consists of microseisms generated along the Pacific Coast, propagating inland along the “same” path both to the sediment sites in the valley and to the reference site, R2, in La Canada.

Dravinski et al. [15] present their results in form of contour maps showing spectral ratios $K_{SR}$ and $N$ averaged over periods 4–10 s. We redrew their contours in Fig. 3b and c, against the same background as in Fig. 3a. Their reference site R2 is about 9.2 km off the right margin of Fig. 3b, as indicated by the arrow. Their results show $K_{SR}$ ratio as large as 5.5, near the intersection of Victory Blvd. and San Diego Freeway (I-405), north-west of Sherman Oaks.

The largest average value of the $N$ ratios reach 7.5 in the same general area, along Sherman Way and between White Oak Avenue and Van Nuys Avenue (i.e. between USC stations 3 and 9, see Fig. 3c).

### 3.2.1. PSV spectra at long periods

In Fig. 3, the indicator of ground motion amplitudes are contours of $\log_{10} PSV$ (in cm/s) for horizontal motion, for oscillator period $T = 4.4$ s (part a) and $T = 7.5$ s (parts b and c), and for fraction of critical damping $\zeta = 0.05$. These contours were drawn based on recorded strong ground motion, after standard data processing and application of the SRSS (square root of the sum of squares) combination rule to the two recorded horizontal components of motion. Todorovska and Trifunac [30] presented such contour maps for periods $T = 0.04$, 0.11, 0.34, 0.9 and 2.6 s. For the purpose of this paper, we followed the same procedure, and extended their set to $T = 4.4$ and 7.5 s, values chosen from the set of 91 periods for which response spectra are calculated in routine accelerogram data processing [34]. We chose $T = 4.4$ s for comparison of $\log_{10} PSV$ with MSR for...
period range 3–5 s, and \( T = 7.5 \) s for comparison with average values of \( K_{SR} \) and \( N \), averaged for periods 4–10 s.

We used accelerograms at 24 stations to construct the contours in the area shown in Fig. 3, and 13 of those fell within this area shown in Fig. 3 (black and white circles). A small full circle indicates that the peak accelerations were within the expected range, a large full circle indicates peak acceleration smaller than expected, and an open circle

Fig. 5. (a) Comparison of contours of Mean Spectral Ratio after Kagami et al. [14] (solid lines) and as in Fig. 3a, (b) comparison of contours of \( K_{SR} \) ratio after Dravinski et al. [15] (solid lines) and as in Fig. 3b, (c) comparison of contours of \( N \) ratio after Dravinski et al. [15] (solid lines) and as in Fig. 3b, with estimated Modified Mercalli site intensities during the 1994 Northridge earthquake, (after Trifunac and Todorovska [32]).
indicates peak acceleration larger than expected [26]. This density of strong motion stations was not adequate to capture details in the variations of \( \log_{10} \) PSV similar to those in the MSR, \( K_{SR} \) and \( N \) contours. Except for systematically smaller \( \log_{10} \) PSV at CDMG station Arleta and at USC station #6, the comparison of the trends in the other areas is not conclusive. CDMG station Tarzana, where large motions were recorded, is at the border of the maps of Kagami et al. [14] and Dravinski et al. [15], so comparison is not possible.

3.2.2. Patterns of “gray” and “white” zones

Trifunac and Todorovska [27] studied the spatial distribution of heavily damaged (red-tagged) buildings [7] and of breaks in the water pipes gathered by the City of Los Angeles [6] and their correlation. They noticed that in the areas with many pipe breaks there were no red-tagged buildings, except in few localized areas where the shaking was most severe (e.g. Sherman Oaks area). They outlined the areas with red-tagged building, and colored them gray. Their “gray” zones are reproduced in Fig. 3a–c. Most (~90%) of the reported pipe breaks occurred outside the “gray” zones. We refer to these areas as “white” zones. Their interpretation was that, within the “gray” zones, the seismic waves propagated to the surface in an essentially linear manner, i.e. without dissipating much energy, and thus damaging many structures. In the “white” zones, the soil response was nonlinear, to a degree shown by the density of breaks in the water pipes. Via nonlinear response, the incoming wave energy was dissipated, and the structures were damaged less or were not damaged.

It is reasonable to assume that the distribution of “gray” and “white” zones is related to some features of the soil and geological structure of San Fernando Valley. For example, the geology underneath the “white” zones is such that these areas would respond in a nonlinear manner at a lower level of shaking than the “gray” zones. In this sense, it would be valuable to develop a procedure for mapping of “gray” and of “white” zones prior to a devastating earthquake. Developing a method to accomplish this via measurement of ground noise would be very valuable, because such measurements are simple and fast.

Fig. 3a–c shows that there is no apparent correspondence between the amplitudes of MSR, \( K_{SR} \), and \( N \) and the pattern of “gray” and “white” zones. This is not surprising since the “gray” zones reflect mostly location of damaged woodframe single-family dwellings (SFD) which are more sensitive to high-frequency strong motion (say 5–10 Hz and higher). Many intermediate and tall buildings were also red-tagged, but their number and geographical distribution are not dense enough to examine the correlation of their levels of damage with the amplitudes of MSR, \( M_{SR} \) and \( N \).

3.2.3. Spatial distribution of red-tagged building

In Fig. 4a–c we show individual locations of all red-tagged buildings. These figures show that there is no simple and direct correlation between the location and the density of the red-tagged buildings with the average amplitudes of MSR, \( K_{SR} \) or \( N \).
3.2.4. Refined estimates of site intensities

Trifunac and Todorovska [32] used the data on red-tagged buildings and breaks in the water pipes to come up with a more detailed map of site intensities than based on felt reports at selected posts [35], and taking into account that nonlinear soil response may reduce the degree of damage. They first developed empirical scaling equations of the form

\[ N_b = f_1(n, I_{MM}) \]

where \( N_b \) is the number of red-tagged buildings per km\(^2\), \( n \) is the number of breaks in water pipes per km\(^2\) and \( I_{MM} \) is the reported site intensity on the

![Fig. 6. (a–c) Same as Fig. 5 (a–c) but with estimated peak site velocity (cm/s) (after Trifunac and Todorovska [32]) instead of Modified Mercalli intensities.](image)
Modified Mescalli scale. Then they inverted these equations to obtain $I_{MM}$ as function of $N_b$ and $n$, and produced a map of $I_{MM}$ with the improved estimates for the areas with $N_b \geq 1$ and $n \approx 0$, and with the original estimates for sites with $N_b = 0$. This map agrees with the overall average trends of the map of Dewey et al. [35], but reflects better the fluctuations of observed damage.

In Fig. 5a–c, we compare MSR, $K_{SR}$ and $N$ with the distribution of $I_{MM}$ from the map of Trifunac and Todorovska [32]. We recall that $I_{MM}$ represents subjective and weighted indicator of the severity of shaking and of its consequences, and it reflects the amplitudes of strong motion in a broad frequency range. It can be seen that the correlation of MSR, $K_{SR}$ and $N$ with the refined $I_{MM}$ is not good. In the western half of San Fernando Valley, the MSR contours appear to follow the trends of $I_{MM}$, but in the eastern part the largest MSR amplitudes fall in the area with low intensity VII. For $K_{SR}$ and $N$, the correlation is also poor.

### 3.2.5. Refined contours of peak velocity $v_{max}$

Following a similar procedure as for the site intensity, Trifunac and Todorovska [32] drew a more detailed map of peak ground velocity $v_{max}$, than the map based on strong motion recordings [36]. They first developed empirical scaling law for $N_b = f_2(v_{max}, n)$ and then inverted it to estimate $v_{max}$ from the distribution of $N_b$ and $n$. Their map is redrawn in Fig. 6a–c against the contours of MSR, $K_{SR}$ and $N$. It is seen that the two sets of data are not correlated.

### 4. Discussion and conclusions

In the above comparison of contours of MSR, $K_{SR}$ and $N$, for long period (3–10 s) ambient noise with several simple and direct indicators of strong motion in the near-field of the Northridge earthquake, we found no obvious correlation. This is not surprising, because of obvious differences in frequencies, amplitudes, or meaning of the mapped parameters, and because of the strong influence of the source radiation pattern in the near-field.

The comparison of log $10$ PSV at $T = 4.4$ and 7.5 s with contours of MSR (Fig. 3a) and the contours of $K_{SR}$ and $N$ (in Fig. 3b and c) is only qualitative, because the density of strong motion accelerographs stations did not provide adequate resolution for this comparison. It could be argued also that comparisons with spatial distribution of “gray” zones and with actual distribution of red-tagged buildings are not appropriate because the woodframe SFD (representing ~90% of all red-tagged buildings) are “sensitive” to “high” frequency ground motions, which may have different amplification than the long period strong motion. Similar arguments could be made against the comparisons with $I_{MM}$ or $v_{max}$ in Figs. 5 and 6.

This lack of obvious correlation does not necessarily mean that measurement of long period microtremors and microseisms is useless. It only means that the simple methods of analysis, based on transfer-function representation, are not adequate, and that other physically sound analyses and indicators must be tried, tested and verified against actual earthquake effects.
Aftershock studies also could not explain the spatial distributions of observed damage in San Fernando valley [9,32], because of nonlinear site response during strong shaking [16].

The lack of simple correlations of site amplification derived from analyses of ambient noise also calls for more conservative presentation of the results, and for more rigorous tests of the methods of analysis. Statements that the analyses of long period ambient noise “appear to be highly practical and attractive as an engineering tool” for determination of the site amplification and of the predominant site periods are at best optimistic, and at present are not justified. The common conclusions of many studies of ambient noise, that those “can reveal the fundamental resonant frequency of surface sediments” and the associated site amplification, might be useful only at some distance from the earthquake fault (where the motions tend to be linear), but this must be verified by comparison with other independently derived indicators of site amplification.

It is clear from the above comparisons that measurements of both strong ground motion and of ambient ground noise must be carried out with much more dense arrays than what we have at present. Perhaps, a selected group of dense arrays (e.g. grid size 100 × 100 m²) will teach us how to analyze ambient noise and discover useful characteristics of a site during future strong earthquakes. Ambient noise measurements have been very successful in providing full-scale experimental data for analyses of buildings and bridges [37–40], and there is no reason why this should not be possible also for soil and geological deposits in general. But, to achieve this capability, we must first recognize that the linear “transfer-function” methods usually do not work in the near-field. Then we can embark on search for physically sound new methods. Finally, we must verify those new methods against actual, full scale “experiments”, involving recorded strong motion, and observed spatial distributions of damage to structures and soil, during real earthquakes.

References


