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Usefulness of forward modelling in strong motion seismology: A few examples

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ABSTRACT: This paper is intended to illustrate, on a few examples, what are the present capabilities and limitations of forward modelling in engineering seismology. On one hand, simplified models may bring a qualitative understanding of the main physical phenomena, and help in identifying either the main parameters to be known, or the remaining knowledge gaps to be filled, regarding source as well as site effects. On the other hand, case studies on recent earthquakes, despite the sometimes surprisingly good agreement between theoretical results and actual observations, reveal the long way still to go before forward modelling may be reliably used, alone, for deterministic hazard estimates.

INTRODUCTION

Numerical simulation is one amongst the various tools that may be used in strong motion seismology. It is, however, much less used in routine engineering practice than purely empirical or semi empirical methods. The aim of the present paper is to illustrate, on a few examples, how a careful forward modelling may bring to earthquake engineers a useful information from both qualitative and quantitative points of view. The first series of examples are purely theoretical and show how forward modelling may either improve the understanding of (some of) the physical processes concerning source and/or propagation effects, or test the actual capabilities of some empirical or semi-empirical, widely used, techniques, or finally help in clarifying and identifying actual issues still unsolved. The second series consists of case studies, which simultaneously illustrate the potential capabilities of numerical modelling to explain the observed motion during past earthquakes, as well as their present limitations for the prediction of motion during future earthquakes. As a conclusion, the interests of numerical simulation for understanding the physics of the wave field incoming upon structures during an earthquake are outlined, while its practical usefulness for hazard assessment studies is discussed.

Almost all the following examples are based on the Discrete Wavenumber (DWN) method, which has proved to be very powerful in dealing with wave propagation in complex geological structures, in the linear viscoelasticity domain. Most of them have been already presented in detail in various publications mentionned in the reference list. The main purpose of this paper is therefore not to present once again these results, but to take advantage of them to illustrate the potential interest of forward modelling for a better, or at least better understood, estimation of ground motion during expected future earthquakes.

THEORETICAL INVESTIGATIONS OF ENGINEERING INTEREST

Ground motion is classically considered as the combination of three main contributions, *i.e.* source effects, path effects and site effects. The main, outstanding, interest of forward modelling, till now, has precisely been to allow careful investigations on each of these contributions, and to help in identifying the leading parameters that influence most the ground motion.

Reviewing all the contributions of numerical modelling is far beyond the scope of the present paper (and far beyond the capacities of the author). Even a simple listing of interesting results obtained with one single numerical technique (for instance the DWN method) would require a much larger space (see for instance Bard *et al.*, 1992). Three examples are thus selected to show the various possible theoretical outcomes of engineering interest: the first one addresses the empirical Green's function technique and examplifies the interest of using theoretical modelling to test some techniques more and more used in engineering practice; the second one, dealing with surface topography effects, presents one of the remaining knowledge gaps in site effects; the third one, finally, considers the theoretical seismic response of geological structures with strong lateral discontinuities, and suggests that the damage potential for common buildings may be related not only with translational ground motion, but also with differential motion.

Tests on the empirical Green's function technique

The use of the empirical Green's function technique is spreading in the engineering community, because of its (relative) simplicity, and of its automatic accounting for path and site effects (in the linear range). The extrapolation from a single, small size event to a larger size event is generally based on the simple formula given in Figure 1 where $R_c(\theta_{lm},\phi_{lm})$ and $R_c(\theta_{l_0m_0},\phi_{l_0m_0})$ are the radiation patterns associated with one single subevent characterized by indexes l and m , while $G_{e\,l_0m_0}$ is the *recorded* motion due a particular small event located on the same fault plane (see Figure 1 and Irikura, 1983 for more details).

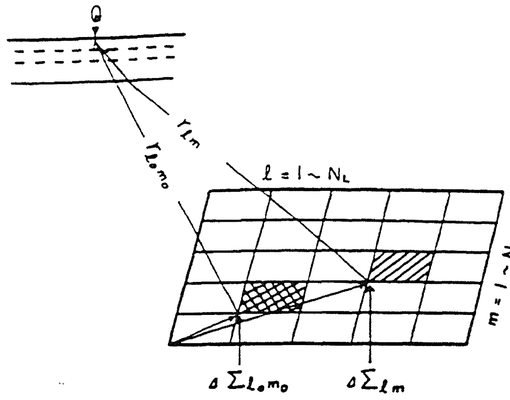
In this formulation, the $R_c(\theta_{lm},\phi_{lm})$ and $1./r_{lm}$ correction factors corresponds, theoretically, only to far-field body waves in a homogeneous space, and the Green's function signal shape is assumed to remain identical whatever the subevent location on the fault plane. These assumptions are very strong, and should be checked: numerical modelling may be used for that purpose.

Such a work was initiated recently (Riepl, 1992): the theoretical Green's functions for a double couple point-source in a stratified half-space were computed with the Discrete wavenumber technique (Bouchon, 1981), and synthetic accelerograms corresponding to step-sized source functions were obtained. The comparison of synthetics obtained for various source depths and focal mechanisms, receiver distances, and crustal structures then provided a useful information on the validity of the above mentionned assumptions. The first results, presented in more detail in Riepl (1992), allow to draw the following conclusions:

- the amplitude and shape of signals computed at distances ranging between 1 and 50 km, are highly sensitive on the depth of the source, and especially when the source is shallow (less than 4 km), and/or is located in the immediate vicinity of an interface between two different crustal layers. This induces variations on peak amplitudes which may reach one order of magnitude (Figure 2), and significant correlation and coherency drops when the source moves by less than 1 km.

- in the near source area, the radiation pattern corrections are extremely important, and must be accounted for. When this is done, the distance correction for the peak time domain amplitude appears to behave as $r^{-1.5}$ rather than as $r^{-1.0}$ (Figure 2). Such a high decay rate has to do with the large sentivity to source depth and the very efficient trapping of energy for shallow sources in stratified crusts.

These numerical results are preliminary, and should themselves be checked against



$$G(x,t) = \sum_{l=1}^{N_L}\sum_{m=1}^{N_W}\sum_{k=1}^{k=K}[R_c(\theta_{lm},\phi_{lm})/R_c(\theta_{l_0m_0},\phi_{l_0m_0})] \cdot (r_{l_0m_0}/r_{lm}) \cdot G_{e\,l_0m_0}(x,t-t_{i\,lm}-t_{d\,klm})$$

Figure 1: *Sketch of the empirical Green's function technique (after Irikura, 1983). Subevent with index* 0 *corresponds to the recorded weak earthquake,* $t_{i\,lm}$ *and* $t_{d\,klm}$ *are time lags corresponding to rupture propagation and hypocentral distance variations.*

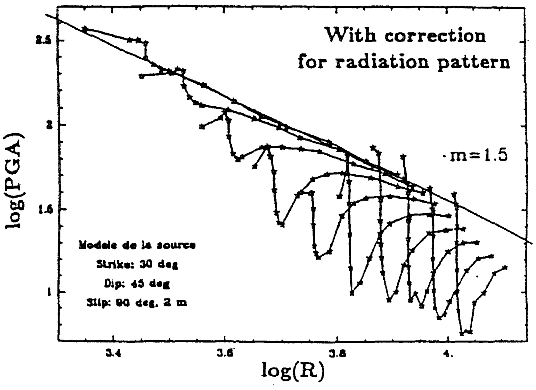


Figure 2: *An example of the computed dependence of horizontal peak surface acceleration on hypocentral distance R (measured in meters). Epicentral distances are in the range 1 - 10 km, while hypocenter depths are in the range 2 - 8 km; points linked with a solid line correspond to a fixed epicentral distance (depth variations only). P.g.a values are corrected for radiation pattern. (After Riepl, 1992)*

instrumental data . However, they indicate that one has to be cautious in basing the estimation of the strong motion expected from a future large earthquake, on the recordings of one single small event corresponding to a restricted part of the fault plane.

Predictability of surface topography effects

There have been in the two last decades repeated, consistent observations of large amplifications on the top of hills or mountains. They are not, however, numerous enough to ground a statistical analysis which would allow to derive meaningful results for incorporation in earthquake regulations, as was done two decades ago for soft site effects. As a consequence, several numerical investigations on surface topography effects have been also performed during the same period. A review of their results was given in Géli *et al.* (1988), whose main conclusions still agree with the most recent results (Gaffet and Bouchon, 1989; Sanchez-Sesma and Campillo, 1991), and are summarized hereafter.

Basically, plane waves incident on a smooth, isolated topography located at the surface of an otherwise homogeneous half-space are reflected at the surface with an angle which depends on the local slope, so that (i) the surface motion associated with the primary wave may change a lot along the topography because of the sensitivity of the reflection coefficients in the case of in-plane motion (SV waves), (ii) the energy is focused in the convex parts of the topography (mountain tops, valley edges), and defocused in concave parts (valley bottoms, mountain foot), and (iii) diffracted waves are generated which propagate horizontally away from the topographic feature (grazing SH or P waves, Rayleigh waves; the horizontal P waves are especially important in the case of incoming SV waves). For steeper topographies, the latest results obtained by Sanchez-Sesma and Campillo (1991), show that, in addition to the diffracted waves described above, there exist, at least for a certain class of topographies (including strong singularities such as wedges, rather rare in nature), *creeping* Rayleigh waves which propagate along the topography surface and may interfere together, which induce a significant duration increase and a late emission of outward propagating Rayleigh waves.

In any case, other similar computations with different incoming wavefields confirm that, even if sometimes the relative crest/base amplification reaches high values over some narrow frequency bandwidths, the absolute level of amplification is generally lower than about three to foir times the flat free surface motion, which is much smaller than what has been actually observed on real data. A few more complex models of topography have therefore been investigated. Besides some mechanical effects related with possible impedance variations inside a topography (in connection with weathering or decompaction processes), the main additional, purely geometrical effect that has been pointed out is the possibility of (leaking) lateral resonance between neighboring topographies, due to the interference between diffracted waves, a well known phenomenon in electromagnetism (Cutler modes, see Wirgin and Deleuil, 1969).

These theoretical results could have several important consequences in earthquake engineering:

- mountain tops experience amplification, and valley bottoms deamplification, the amplitude of which is the largest for incoming SV waves and the lowest for P waves. This amplification is frequency dependent, and is maximum for wavelengths comparable with the width of the topographic feature. Even in the SV case, no spectral amplification factors larger than a factor of 3 to 4 have been obtained (measured with respect to the amplitude on a flat free surface).
- the motion on the slopes and further outward exhibits rapid spatial and frequency changes due to destructive and constructive interferences between the primary and diffracted waves, which give rise to differential motions, and may play a significant role in the dynamic triggering of landslides or rockslides during earthquakes.

Nevertheless, the implementation of these theoretical results in quantitative hazard assessment studies, or in earthquake regulations is probably still premature, since, as outlined in Géli *et al.* (1988) and summarized in Figure 3, reported *in situ* amplification values are on average much higher than the theoretical predictions based on two-dimensional models, and even those few based on 3D models. The only agreement is a qualitative one, *i.e.* summits and ridges amplify over a broad frequency range. Reaching a satisfactory explanation of these quantitative differences, and hence opening the door to a efficient taking into account of surface topography effects in earthquake engineering practice, requires new, carefully prepared, experimental surveys, coupled with detailed geological and geotechnical investigations on the instrumented topographic feature, so as to constrain as much as possible the theoretical models which should be tried to interpret the recorded data. The design of such an experiment would certainly benefit a lot from being planned in close connection with theoreticians.

In such a case, despite its apparent failure, numerical simulation is nevertheless useful in identifying a knowledge gap, and presents the advantage of providing a guide for the design of further experimental surveys, aimed at filling these gaps.

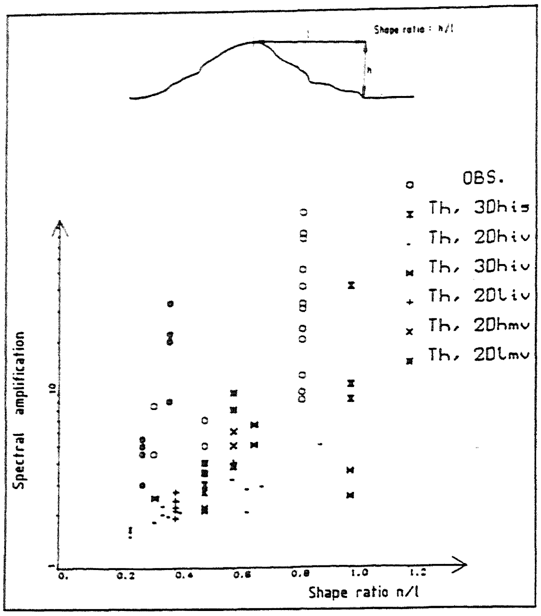


Figure 3: *Topographical effects: comparison between observations and theoretical results. The spectral crest/base amplification is displayed as a function of average shape ratio h/l, for the transverse (in-plane) component only. Open circles stand for observations (see Géli et al., 1988), while other symbols stand for various numerical models: 2D or 3D, for homogeneous ("h") or layered ("l") half-space models, for an isolated ("i") or multiple ("m") topography, impinged by body ("v") or surface ("s") waves.*

Effects of lateral discontinuities and differential motion

The numerous observations of significant damage increase due to local geological conditions have been almost systematically linked with the well known amplification of ground motion by surficial soft layers. Some intriguing observations, however, suggest that damage distribution on soft sites might *not* be linked only with the *translational* motion: Moczo and Bard (1993) mention numerous, consistent macroseismic observations showing a significant increase in damage intensity on stripes located along strong lateral discontinuities, *i.e.* areas such as faults, abnormal contacts or debris zones, where a soft material lies besides a more rigid one. As such observations cannot be explained using simple one-dimensional wave propagation methods, they investigate the SH seismic response of a simple example of such lateral discontinuities, *i.e.* a semi-infinite, horizontal soft layer embedded in a homogeneous half-space, as shown in Figure 4.

The response of such structures is shown to be characterized by two dominant effects, the "classical" 1D resonance and an efficient lateral diffraction from the discontinuity in the softer material, which has several consequences of engineering interest:

- first, as expected, an amplification of motion is observed on the soft side (Figures 4a and 4b). Of course, because of diffraction effects, the motion is larger in the immediate vicinity of the discontinuity than at some distance away, where it stabilizes around its 1D value. This

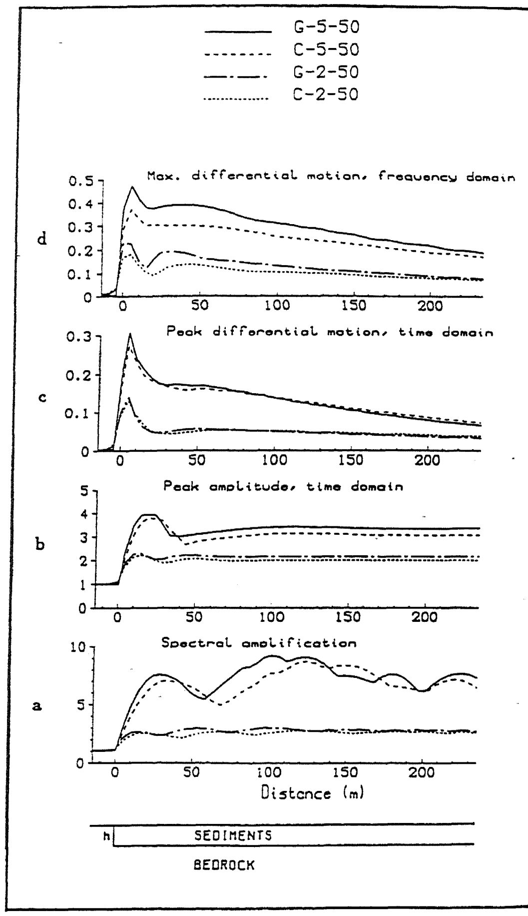


Figure 4: *Surface response across a sharp lateral discontinuity. The model geometry is depicted at the bottom of the figure, with h = 10 m.*

a) (Bottom) Spatial variation of the peak spectral amplification along a cross-section of the model.

b) (Second from bottom) Spatial variation of the peak time-domain amplitude for an incident normalized Gabor pulse.

c) (Third from bottom) Spatial variation of the peak time-domain surface strain for an incident normalized Gabor pulse.

d) (Top) Spatial variation of the peak spectral strain.

Solid, chain-dotted and dotted lines correspond to different velocity contrast and models. (After Moczo and Bard, 1993).

"overamplification", however, does not exceed 30 %, which does not explain the "stripe-like" distribution of intensity actually observed in reality.

- second, a slight increase in motion duration due to waves diffracted from the discontinuity. It cannot, however , explain the observations since this duration increase is significant only at sites away from the discontinuity (where local surface waves are late).
- finally, diffraction and subsequent laterally propagating waves induce a significant differential motion, which is, as shown in Figures 4c and 4d, sharply peaked close to the discontinuity, in a pattern similar to the macroseismic observations.

Based on these results, Moczo and Bard (1993) draw the - still qualitative - conclusion that damage in standard structures (including small size buildings) might be related not only with the amplitude of purely translational ground motion, but also with the level of differential motion, which they show, following several previous studies, to be very sensitive to local geology.

These potential effects of differential ground motions have been recognized for a long time for extended structures (large size buildings and, of course, lifeline facilities), because of the torsional and/or rocking effects they induce. Their true importance is, however, difficult to assess because of the sparcity of appropriate data. While a few instrumental arrays have been especially designed in that purpose (e.g. Katayama, 1991), numerical simulation provides an alternative approach to obtain some quantitative estimates of differential motion: Bard *et al.* (1992) for instance show that because of near surface irregularities, differential motion in soft ground may reach very large values (*i.e.* around $10^{-2} rad$), which is very noticeable and may induce damage, or even failure, in both surficial soils and man-made structures.

These three examples are simply a few amongst many others concerning source, path or site effects which, in our opinion, illustrate the capacity of numerical modelling to identify the main physical phenomena of engineering interest in the earthquake process, and to quantify their possible effects. A more complete - though non exhaustive - presentation of a wide variety of such theoretical investigations may be found in Bard *et al.* (1992).

CASE STUDIES

As numerical models may account for source, path and site effects, one may try and use them for predicting the expected strong motion for future earthquakes at specific sites, provided that the seismogenetic fault zones, the crustal structure and the site conditions are well known (which is already a very restrictive condition). While on one side "numericians" have a natural tendency to claim that their model may in effect perform such predictions without any major problems, earthquake engineers on the other have a rather skeptical attitude, and prefer, by far, empirically based approaches using, for instance, empirical attenuation laws. The two following examples are intended to illustrate the practical capabilities and limitations of models to reproduce the main qualitative and quantitative characteristics of observed ground motion. The first one concerns the Kalamata, Greece, event of September 13, 1986, while the second deals with the motion produced in Mexico City by the large subduction events occuring on the Mexico Pacific coast. The first one mixes source and site effects, while the second focuses on site effects only.

The Kalamata (Greece) event of September 13th, 1986

On September 13, 1986, a magnitude 6.2 (M_S) earthquake occurred in southwestern Peloponnese (Greece), causing heavy damage in the city of Kalamata, where it was recorded by two strong-motion instruments (SMA-1 type). Detailed post-earthquake investigations were conducted, leading to reliable estimates of source parameters, deep crustal structure and surficial geotechnical characteristics. Such a situation was therefore almost ideal for testing the actual capabilities of numerical modelling; a detailed presentation of such a comparison between numerical results and observed records may be found in Gariel *et al.* (1991), so that the present section will only - very briefly - outline the main conclusions of concern for engineering seismology.

The model

The epicenter was located about 10 km north of the city, with a hypocentral depth estimated between 5 and 7 km, and the seismic moment was estimated around $7.\,10^{17} N.m$ (Figure 5). Aftershock studies using a mobile network (LyonCaen *et al.*, 1988) showed that the ruptured area was located on a 45° dipping normal fault just beneath the city, covering an area of about $15 km$ (NS) $\times$ $10\ km$ (EW). In addition, the city of Kalamata, and more particularly the strong motion sites, are covered with young sediments, inducing low velocities over several tens of meters. Therefore, any realistic modelling should consider simultaneously three items: a) the finite dimensions of the source area, b) the rupture heterogeneities on the fault plane, and c) the actual crustal velocity structure at large depth as well as near the surface.

In order to account for items a) and b), a source model derived from Papageorgiou and Aki (1983) was considered, as illustrated on Figure 6. It consists of an aggregate of 12 identical, joint,

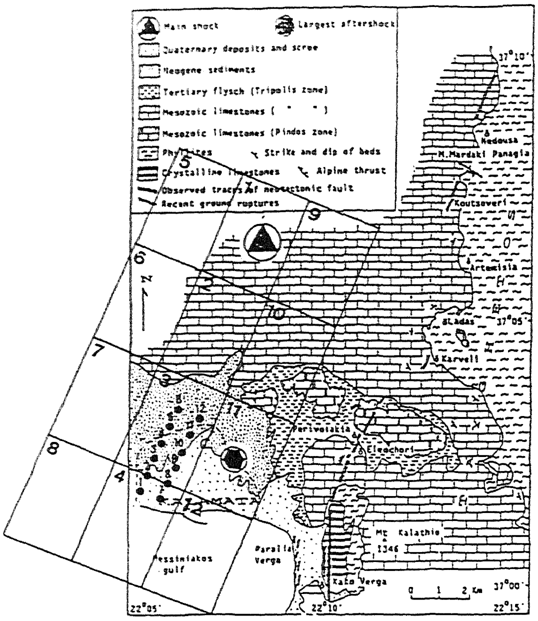


Figure 5: *Map of the epicentral area of the 1986 Kalamata earthquake. The large solid triangle corresponds to the mainshock epicenter, while the smaller octogon corresponds to the M=5 aftershock. The grid depicts the surface projection of the fault model (see Figure 6): numbers in each rectangle refer to each elementary crack. Small solid circles indicate locations where synthetics are computed. (After Gariel et al., 1991).*

circular cracks (radius: 1.8 km), included in a 14.4 $km \times 10.8$ km rectangle, lying on a dipping fault plane (dip: 45°) and having a pure normal faulting mechanism (rake: 90°). The axis of the central raw of cracks lies at a 6 km depth, which implies that the upper part of the broken area is at a 2.2 km depth. On each single circular crack, the slip history follows the dynamic solution given by Madariaga (1976). Unlike the conventional Papageorgiou-Aki model, the nucleation times of each crack are not random: the rupture initiates at the center of crack # 1, and every other crack i starts rupturing in its center at time $t_i = d_i/v_c$, where d_i is the distance between center of crack i and center of crack 1. The "sweeping" velocity v_c was taken equal to the rupture velocity on each single crack, estimated to be 75 % of the S wave velocity at hypocentral depth, *i.e.* 2.5 km/s. Each individual crack has the same local stress drop of 38 bars, which corresponds to an average slip of 16 cm (*i.e.*, a maximum slip of 24 cm at crack center). The total seismic moment is therefore 6. 10^{17} $N.m$, very close to the measured value. Although nothing in the Kalamata post-earthquake investigations can justify this particular choice of composite source, the number of "subevents" is in good agreement with the statistical results found by Kamiyama (1987).

Then, to account for item c), this composite source was embedded in a layered crust, including low velocity sedimentary layers at the surface (Figure 6, Model B). However, in order to measure the effect of these low velocity surface layers, similar computations were also performed for a two-layer crust model (Figure 6, Model A), corresponding to the model used for aftershock localisation.

The results

The resulting accelerations are computed at 12 surface sites, located above cracks # 3 and 4, over the 0 to 4 Hz frequency range, which was observed to contain the most energetic motion in the recorded accelerograms. Figure 7a displays the horizontal transverse accelerations (*i.e.*, parallel to the slip direction) resulting from the composite source for the 12 receivers. Several important comments can be drawn out of these synthetics:

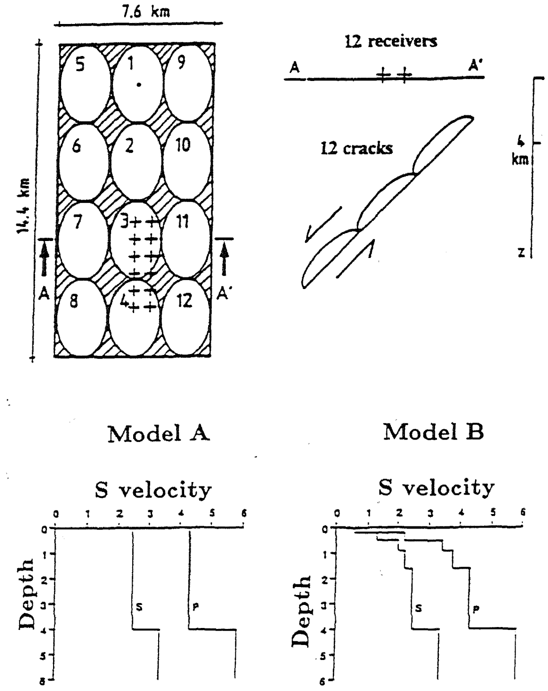


Figure 6: *Source and crust models used for modelling the Kalamata events (after Gariel et al., 1991). Top left: Multiple event source model for the Kalamata earthquake: the rupture starts at the center of crack # 1, and spreads circularly (see text). Top right: cross section of the fault plane, also displaying the final slip, and the location of the 12 receivers (surface crosses). Bottom: crustal velocity models used for computations.*

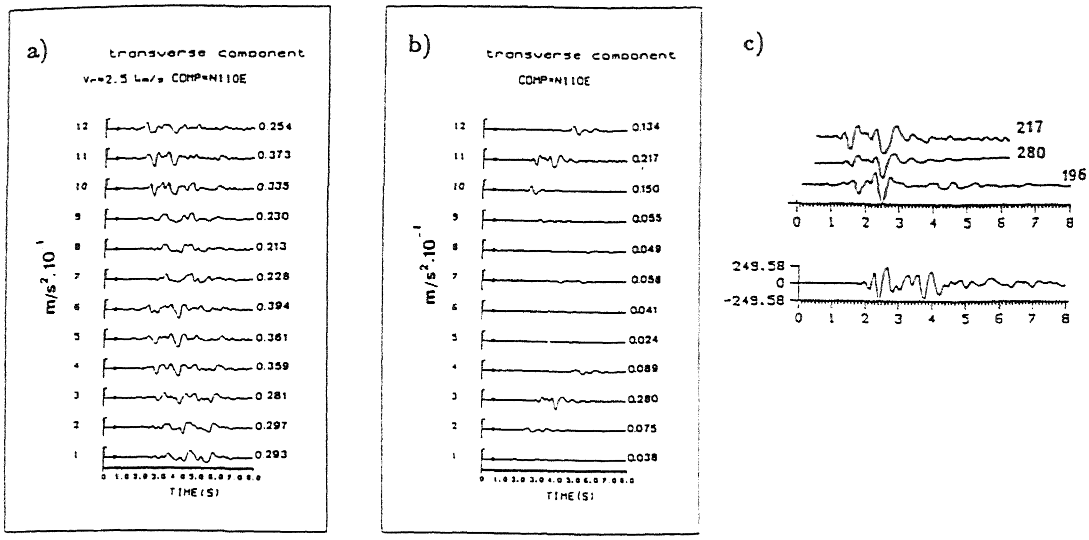


Figure 7: *Surface acceleration computed with model B.*

a) Transverse (i.e., EW) component of acceleration computed with the complete source model, at the 12 receivers. The numbers on the right of each trace represent the peak value.

b) Contribution of each crack to the acceleration time history at receiver # 4.

c) Comparisons between the accelerograms observed in Kalamata City for the main shock (bottom) and a M_S=5.2 aftershock, and the computed contributions of cracks # 3 and 11 at receiver # 4.

- traces in Figure 7a exhibit a significant spatial variation of both peak accelerations and waveforms. These variations are clearly associated with the heterogeneous faulting, and the constructive or destructive interferences between stopping phases of each elementary crack. Figure 7b displays the individual contribution of each crack at receiver # 4, and shows that the motion at this site is mostly influenced by the nearest subevents (# 3, 4, 10, 11 and 12), while remaining sub events have insignificant contributions. Thus, close to a fault, the peak acceleration values are governed by the radiation of the few nearest cracks - and especially the most surficial -, and are almost independent of the fault size: these results are of course in close agreement with the observed "saturation" of pga in the immediate vicinity of moderate to large earthquakes, and are moreover consistent with the similarity of pga values measured in Kalamata City for the main shock (M_s=6.2), and a M_s=5.2 aftershock located just underneath the accelerograph site: such a comparison is illustrated in Figure 7c, together with the computed contributions of cracks # 3 and 1.

- the acceleration level, as well as the duration of strong motion, are similar enough in synthetics and in the observations (Figure 7c), to consider this modelling as satisfactory from an engineering point of view. It must be noticed, however, that such a modelling is an *a posteriori* modelling, for which the macroscopic source parameters were

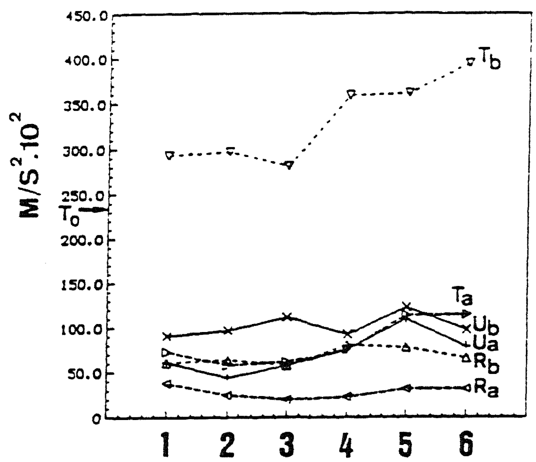


Figure 8: *Sensitivity of surface motion to near-surface crustal velocity structure. Peak accelerations computed at 12 receivers (1-6 and 7-12) with the two velocity models. Subscripts a and b refer to velocity models A and B. T is the transverse (N110E) component, R the radial component (N20E), and U the vertical component. T_0 is the recorded peak transverse acceleration.*

well constrained.

- from another point of view, the amplitude of surface ground motion is shown to depend a lot on the surficial crustal velocity structure (Figure 8): differences between the 2 models reach factors ranging from 3 to 5 in the time domain (for pga), and from 3 to 8 in the spectral domain. This is not a surprise; however, it has to be kept in mind, especially when considering other kinds of modelling (such as the so-called "spectral" models), where the transfer function of the crust is not always well accounted for.

Despite this satisfactory qualitative and quantitative *a posteriori* agreement, these results suggest that predicting pga values in the near source area wil remain uneasy, even with advanced numerical modelling, for some time: the rupture process of moderate to large earthquakes is too complex, including many sub events, rupture velocity variations, possible changes in the slip direction or fault plane orientation, etc., and is not, nowadays, predictable on a deterministic basis, while the near source acceleration is too sensitive to these rupture heterogeneities. Therefore, numerical models can only provide *a priori* estimations of the lower and upper bounds of near source ground motion, together with their physical interpretations. One must keep in mind, however, that all other estimation methods have similar limitations: estimates based on emirical attenuation laws have very large standard deviations for short distances, while estimates obtained with the empirical Green's function technique are also very sensitive to the assumed rupture process.

Local response of the Mexico City basin

It is well known that Mexico City basin exhibits a huge amplification of ground motion due to very soft lacustrine clay deposits in the ancient lake bed. Simple 1D models may satisfactorily explain the observed differences in the damped response spectra between soft and rock sites, provided an adequate choice of geotechnical parameters (Seed *et al.*, 1988). However, these models are *by no means*, even with especially adjusted soil parameters, able to explain the huge increase of strong motion duration consistently observed in the lake bed zone: the Mexico City records of the Michoacan earthquake and of three late events (February 1988, April 1989 and May 1990) last for more than 3' in the lake-bed zone (up to 3'), and less than 1' on nearby rock sites. As the duration issue is of primary importance for the resistance of structures (although not well accounted for in present regulations), explanations of this duration increase are needed.

A number of interpretations have therefore been proposed in recent years, some of them being very imaginative and exciting. Kawase and Aki (1989) interpret the last, long part of the accelerometric records in the lake bed zone as due to surface waves generated on the edges of the clay layer, whose low phase and group velocities result in a very long duration away from the edges. Seligman *et al.* (1989) propose a "membrane model", according which the long duration results from a (lateral) resonance of horizontally propagating P waves produced by a - postulated - energetic S to P wave conversion at clay surface. The pattern of such a resonance would then be controlled by the horizontal limits of the clay layer, and be similar - they say - to the irregular damage ditribution observed during the 1985 event. A few other authors suggest that local inhomogeneities in the surficial part of the ground (including clay layer and anthropic fill) could trap seismic waves and result locally in long wavetrains. Finally, Lomnitz (1990, 1991) advances an explanation in terms of gravity waves: according to him, non-linear effects would induce a drastic reduction in the shear modulus of the clay, which would then behave as a fluid with the associated, energetic gravity waves similar to sea or lake waves.

These explanations are mainly based on qualitative reasoning or incomplete quantitative computations, and are still controversial. However, as large magnitude earthquakes are still expected along the Mexican subduction zone, there exists an obvious need for a well-established, well-accepted explanation having an effective predictive power. In that purpose, numerical modelling may be used to test the underlying hypothesis and their postulated effects: such a testing work has been achieved recently by Chavez-Garcia (1991), through a thorough and careful numerical modelling, whose main findings are summarized hereafter.

The surface wave interpretation (Kawase and Aki, 1989) is shown to have serious limitations: indeed, when considering realistic damping values in the clay layer (*i.e.*, around 2%), surface waves generated along clay layer borders exhibit a very rapid spatial attenuation - because of their very large dispersion -, and cannot explain the high energy, late arrivals observed in lake bed zone accelerograms. Long enough seismograms (*i.e.*, comparable to the observed records) can only be obtained with very low damping values (as low as 0.25 %), which look unrealistic and unconsistent with the numerous laboratory measurements.

The membrane model is no more supported by the presently available numerical computations. On one hand, given the size and depth of the Mexico City basin, lateral resonance can occur only if the "deep sediments" underlying the clay layer are both very thick (more than 2 km) and relatively soft (S wave velocity not larger than 500 m/s, even at depth), and it would then correspond to frequencies much lower than 0.5 Hz (indeed, lower than 0.1 Hz). On the other hand, none of the various computations that were conducted up to now in the P-SV case, exhibit late, beating-like, waveforms similar to the observed records. It must be noticed, however, that no computations have yet been performed (because of numerical limitations) mixing very thick deep sediments and a thin clay layer, so that the door is not categorically closed for such an interpretation.

Local iregularities (geometric as well as mechanical inhomeneities) have indeed been shown to induce very significant effects; under realistic assumptions however, they affect mainly the amplitude of motion, and not so much its duration (by far not as much, in any case, as in the observations).

The gravity wave interpretation has been extensively examined, for very high Poisson's ratio elastic materials as well as for viscous fluids (Chavez-Garcia and Bard, 1992). Gravity perturbations in elastic solids are shown to affect significantly Rayleigh wave phase and group velocities only at low frequencies for very soft materials (Poisson's ratio higher than 0.4999...), but, even in that extreme case, can not result in very long durations. On the other hand, coupling between elastic and viscous fluid media through irregular interfaces may effectively induce gravity waves in the fluid, whose amplitude and propagation, however, are drastically sensitive to the viscosity value. As the "fluidization" of clay assumed by Lomnitz is hardly conceivable without a significant amount of viscosity, these computations lead to the conclusion that Lomnitz interpretation in terms of gravity waves is extremely improbable.

The conclusions of this testing work may seem very disappointing, since all the proposed interpretations finally appear as very unlikely, given our present knowledge of clay and sediments mechanical parameters. Such "negative" results are nevertheless meaningful, since they prompt the search for new ideas, and simultaneously provide precious indications on the "best" research axis for the coming years: for instance, *in situ* measurements of clay damping would bring a definite answer to the "surface wave" interpretation. Also, teaching modesty to numericians is not necessarily a bad thing, reminding them not only that a lot of numerical work is still ahead, but also, and above all, that in many cases instrumental data are irreplaceable.

CONCLUSIONS

The investigation field open to numerical simulation becomes wider and wider with the development and availability of either supercomputers, or very powerful workstations. Parameter studies with present-day models will probably turn into routine work in the current decade, while computations of the radiation of complex source models in a layered crust with 3D irregularities will be within reach very soon. Nevertheless, despite the satisfactory agreement that was obtained in a number of *a posteriori* comparisons with actual records, numerical simulation is still far from providing, on its own, reliable estimates of the seismic motion to be expected at a given site, for several important reasons:

- in many areas, the basic background geological or geophysical information is missing: precise location of fault planes, expected mechanism *and magnitude*, velocity and damping structure of the crust, velocity and damping structure of the soils surrounding the receiver site. Although this latter kind of information may seem the most easy to obtain, it still remains very expensive, and considerably increases the actual cost of numerical studies.
- even in the very few areas where such information is available, there still exist some unpredictable parameters that control at least some aspects of seismic motion: for instance the barrier or asperity distribution on the fault plane, or the precise location of the nucleation point and the subsequent rupture propagation direction (for directivity effects).
- finally, there still exist some observations that cannot be satisfactorily explained by presently available numerical models, or on the contrary the available data are not sufficient to discriminate between various theoretical explanations. This was in particular the case for Mexico City.

Earthquake engineers have long been aware of all these limitations of numerical modelling in strong motion seismology, which explains why the vast majority of current earthquake hazard studies basically rely on statistics of recorded intrumental data. We hope, however, that the investigations and the comparisons referred to in this paper (and in many other places by many other authors) will bring the earthquake engineering community to upgrade the credit it gives to results of numerical simulation in strong motion seismology, and simultaneously that the results of the sensitivity studies will bring it to pay more attention to the large standard deviations that exist in statistical studies of instrumental data. Thorough parameter studies and case studies, numerical investigations progressively improve the qualitative and quantitative understanding of seismic wavefields that shake man-made structures during earthquakes; they may also help a lot in checking the validity of some new hazard assessment techniques (such as the use of empirical Green's functions), or in performing some feasability studies for new prevention techniques (such as early warning systems presently under study in various parts of the world); last but not least, it may also provide some reliable upper and lower bounds for the major characteristics of ground motion to be expected at a given site in a given tectonic environment.

There is no doubt that all models are wrong; however, as soon as one recognizes his model to be very far from reality, he will look for the deep meaning of his results, and gain in the understanding of earthquake process, and then his model will start being very useful.

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