UNIVERSIDAD NACIONAL AUTÓNOMA
DE MÉXICO

INSTITUTO DE GEOFÍSICA
POSGRADO EN CIENCIAS DE LA TIERRA

LA RESPUESTA SÍSMICA DEL TERRENO FIRME EN LA CIUDAD DE MÉXICO. OBSERVACIONES Y MODELOS

TESIS

QUE PARA OBTENER EL GRADO DE
DOCTOR EN CIENCIAS (SISMOLOGÍA)

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MÉXICO, D.F. 2002

Agradecimientos

Al Dr. Francisco José Sánchez Sesma por haberme permitido ser actor estelar en esta aventura.

Al Dr. Eduardo Reinoso Ángulo por contar con su apoyo, sus comentarios y su guía en gran parte de este trabajo.

Al Dr. Shri Rrishna Singh por confiar en mi y por sus enriquecedoras enseñanzas.

Al Dr. Javier Pacheco por darme la oportunidad de colaborar con El. Las discusiones y sus consejos son muy valiosos para mi.

Al Dr. Cinna Lomintz por interesarse en este proyecto de tesis.

Al Dr. Mario Ordaz por sus comentarios y sugerencias en algunas partes de este trabajo.

Al Dr. Sergio Rodríguez Elizarraras por ser miembro del jurado.

A la Universidad Nacional Autónoma de México (UNAM), a través de los Institutos de Geofísica (por haberme recibido y formado académicamente dentro del Programa de Doctorado de Ciencias de la Tierra) e Ingeniería (en donde lleve acabo la mayor parte de este trabajo).

A mis grandes amigos: Fernando Velasco, Mario Rebolledo, Hortencia Flores, Leonardo Ramírez, Moisés Contreras, Jorge Andaverde, Jorge Aguirre, Sergio y Eloísa, Ramón Mendoza, Mario González, Efraín Alva, Ulises Rodríguez y Gabriel Chávez. Por compartir con ustedes mis aventuras y desventuras, por formar parte de mi vida y por demostrarme que un mundo mejor es posible.

Al Consejo Nacional de Ciencia y Tecnología (CONACYT), al Sistema Nacional de Investigadores (SNI), a la Dirección General de Estudios de Posgrado (DGEP), a la Dirección General de Asuntos del Persona Académico (DGAPA) y al Fondo de Becas del Instituto de Ingeniería, UNAM. Por el apoyo económico brindado durante algunas etapas de mi trabajo doctoral.

A la memoria de mi padre, Alfonso

Para Claudia,
porque eres mi amor mi cómplice y todo

A mis hijos: Carlos Emilio, Alejandro y Mariana

A mi madre, Alicia
A mis hermanos: Teresa, Martha Alicia y Alfonso
y mi sobrina Cecilia del Carmen

IV. THE SEISMTC RESPONSE OF THE HILL ZONE IN MÉXICO CITY: A REVIEW AND NEW FINDINGS (*BuBetín of the Seismological Society of America, enviado)*

V. HYBRID INDIRECT BOUNDARY ELEMENT - DISCRETE WAVE NUMBER METHOD APPLIED TO SIMÚLATE THE SEISMIC RESPONSE OF STRATIFIED ALLUVIAL VALLEYS (*Soü Dynamics and Earthquake Engineering, aceptado)*

RESUMEN

En el presente estudio se hizó un análisis de la respuesta sísmica de las estaciones de la Red Acelerométrica de la Ciudad de México (RACM) localizadas en la *zona de lomas.* En la primera parte de este trabajo se estudió el comportamiento de las estaciones del terreno firme y se propusieron modelos para explicar las diferencias observadas entre ellas. Se estudió la relación entre la estructura profunda del valle y los depósitos someros y los efectos del campo incidente. Además, se obtuvieron nuevos coeficientes en las leyes de atenuación para predecir el movimiento sísmico de futuros temblores en la estación CU. Finalmente, se presenta una metodología para estudiar la respuesta sísmica de valles aluviales estratificados en 2D bajo incidencia de ondas *SH,* con algunas aplicaciones al Valle de México.

ABSTRACT

In this study, we analyze the seismic response of the *hill zone* in México City. We studied the behavior of the stations located at this zone and proposed models to try to explain the observed differences. Also, we studied tbe relationships between the shallow and deep stmcture in the Valley of México and the effect of the incident wave-field. We obtained new coefñcients for attenuation laws to predict the strong ground motion and response spectra at CU station for two scenarios. Finally, we introduce a hybrid indirect boundary element-discrete wave mimber method applied to simúlate the seismic response of stratified alluvial valleys in México City.

1. INTRODUCCIÓN

Desde el punto de vista sísmico la *zona de lomas* de la Ciudad de México es considerada como el sitio de referencia. Esta región representa al terreno firme y está compuesta principalmente por flujos de lava de composición basáltica y tobas de diferentes edades. En las últimas 2 décadas, el crecimiento de la ciudad ha originado un incremento en el desarrollo urbano en esta región, en donde se han construido modernos centros comerciales, de negocios, educativos y habitacionales. Una característica importante es que es considerada como una zona de bajo riesgo para sismos provenientes de la zona de subducción, no así la llamada *zona del lago* en donde este tipo de sismos han provocado severos daños. Sin embargo, el riesgo sísmico en el Valle de México no está relacionado sólo a este tipo de sismos. Existen otros tres tipos de fuentes sísmicas que pueden causar daños a la ciudad, estos son: intraplaca, continentales y locales. Como se analizará más adelante, estos tres tipos de fuentes son los más peligrosos para *la zona de lomas.*

Una de las características de los sismos provenientes de la trinchera meso americana, es que sus periodos de recurrencia son muy cortos comparados con los otros tipos de fuentes. En general, el daño en la ciudad (principalmente en la *zona del lago)* está asociado a sismos costeros como el terremoto del 19 de septiembre de 1985 (Mw = 8.0). Sin embargo, el 15 de junio de 1999 y 21 de julio de 2001 se produjeron dos sismos intraplaca: Tehuacán (Mw = 7.0) y Copalillo (Mw = 5.9). El primero de ellos, causó daños en algunas ciudades de los estados de Puebla y Morelos, debido a su proximidad a la fuente. En la Ciudad de México localizada a ~200 km los perjucios fueron menores. Para el segundo sismo (Copalillo), no se reportaron daños en la ciudad, pero debido a su gran proximidad, los registros de aceleración fueron de muy buena calidad, esto condujo a que en algunos trabajos se tratara de evaluar el efecto que causaría este sismo en la ciudad, si hubiera sido de una magnitud mayor (Mw = 7.0), equiparable al temblor de Tehuacán. Los resultados muestran que también *la zona de lomas* es de alto riesgo. Para los otros tipos de fuente (continental y local) la sismicidad es baja y dispersa comparada con los eventos de la trinchera e intraplaca. Sin embargo, si un terremoto con M > 5.0 llega a ocurrir, grandes daños se producirían en la Ciudad de México, principalmente en la *zona de lomas.*

En la Ciudad de México, el sitio tradicional de referencia ha sido la estación CU. La razón es muy simple: desde 1964 más de 60 sismos moderados y grandes (Mw > 5.0) han sido registrados. Durante el terremoto del 19 de septiembre de 1985 (Mw = 8.0) tres acelerómetros funcionaban en CU: MV, CUO1 y CUIP los cuales se localizaron en un área de aproximadamente 2 x 2 km. En la actualidad, hay dos instrumentos digitales de tres componentes operados uno por el Instituto de Ingeniería (Kinemetrics K2) y el otro por el Instituto de Geofísica (broadband FBA23 + Quanterra de 24 bits), ambos en la UNAM.

Después del terremoto del 19 de septiembre, la red acelerométrica en la Ciudad de México creció considerablemente. En la actualidad esta formada por más de 90 estaciones digitales ubicadas en superficie, 18 de ellas localizadas en \&*zona de lomas* (estaciones 07, 13, 18, 21, 28, 34, 40, 50, 64, 74, 78, CH, CU, CN, ES MD, TX, y TY; los sitios 18 y 28 se localizan en los Cerros de La Estrella y El Peñón, respectivamente). Con el aumento de las estaciones en esta región, es obvio que el conocimiento de la respuesta sísmica será mejor.

Uno de los usos que se le ha dado al sitio de referencia, es que se asume que el registro de superficie libre es equivalente a la señal de entrada en la base de las capas del suelo. A partir de esto, se han llevado a cabo muchos análisis para cuantificar la amplificación espectral en *\azona del lago* respecto al sitio de referencia. Singh *et al.* (1988a,b) utilizando a CU como estación de referencia, encontraron que los cocientes espectrales entre estaciones de la *zona del lago* y la *zona de lomas* presentan comportamientos muy similares durante diferentes temblores. Estos autores obtuvieron amplificaciones espectrales de 50 veces en 0.5 Hz en la zona del lago durante el sismo del 19 de septiembre de 1985. Reinoso y Ordaz (1999) mencionan que el promedio espectral de las estaciones localizadas en la porción sur del terreno firme es una medida más estable que cuando se usa sólo una estación como sitio de referencia. Además, estos autores mencionan que este promedio, representa la forma espectral general del campo incidente en al base de los depósitos de la *zona del lago,* a una profundidad aproximada de 100 m. También, comparan los cocientes espectrales de algunos sitios en la zona lacustre con la obtenida mediante el modelo unidimensional, ellos encontraron que las respuestas son muy similares.

Singh *et al.* (1988a), Ordaz y Singh (1992) y Sánchez-Sesma *et al.* (1993) señalan que la *zona de lomas,* presenta amplificación con respecto a la predicha por leyes de atenuación para sismos provenientes de la zona de subducción. Esta amplificación llega a ser hasta 10 veces mayor en el rango de frecuencias de 0.2 a 0.7 Hz con respecto a sitios de terreno firme fuera del Valle de México. Pacheco y Singh (1995) a partir de analizar varios sismos de fallamiento normal de profundidad intermedia, grabados en la estación CU, encontraron que las ondas *S* presentan una amplificación de 2.5 veces entre 0.2 a 3.0 Hz. Pérez-Rocha (1998) encontró que existen claras diferencias en el espectro de amplitud de Fourier, para cada zona sismogénica en la trinchera Meso americana, de donde la región Guerrero - San Marcos representa la zona con mayor riesgo para la Ciudad de México, principalmente la *zona del lago.*

Estudios recientes, muestran diferencias importantes entre estaciones localizadas en la *zona de lomas* (Reinoso y Ordaz, 1999). Ordaz y Singh (1992) encontraron que los movimientos fuertes del terreno en la estación MD localizada al norte de la Ciudad, son menores comparados con estaciones localizadas en el sur de la ciudad (CU). Después de analizar los trabajos realizados en esta región se plantea como objetivo general de esta tesis estudiar los posibles efectos de cámaras magmáticas en el movimiento sísmico observado en la Cuenca de México y el modelado de la respuesta sísmica de valles aluviales, los cuales se dividieron en los siguientes objetivos específicos:

- Analizar la respuesta sísmica en las estaciones localizadas en la *zona de lomas,* con la finalidad de evaluar los efectos de fuente, magnitud, distancia epicentral y acimut de diferentes sismos.
- Determinar la estación o estaciones de referencia.
- Proponer un modelo para explicar las diferencias observadas.
- Obtener nuevos coeficientes en las ecuaciones para predecir el movimiento del terreno, en la estación de referencia.
- Proponer una metodología para modelar la respuesta sísmica en 2D del Valle de México e impl ementarla.

El contenido de este trabajo, está centrado principalmente en los apéndices, en ellos se muestran estudios detallados de los tópicos que se mencionarán en los subsecuentes capítulos. Los textos presentados en los cinco apéndices, representan artículos de investigación arbitrados, los cuales se encuentran en etapa de edición y de revisión en algunas revistas de circulación internacional.

Para llevar a cabo este trabajo se ha revisado toda la información existente en la zona (datos y modelos), además se analizaron todos los registros acelerométricos en la estación CU (localizada en el campus de la Univesidad Nacional Autónoma de México, UNAM), desde 1964 con Mw > 5.0, con la finalidad de obtener nuevos coeficientes en las expresiones para predecir el movimiento del suelo *(amax₃ ymax,* amplitud espectral, espectros de repuesta de aceleración, velocidad y desplazamiento).

2. PRINCIPALES FUENTES SÍSMICAS QUE AFECTAN A LA CIUDAD DE MÉXICO

El riesgo sísmico en la Ciudad de México está relacionado a cuarto tipos de fuentes: subducción, intraplaca, continentales y locales (Figura 1); las últimas tres regiones sismogénicas son las más peligrosas para las estructuras localizadas en la *zona de lomas:*

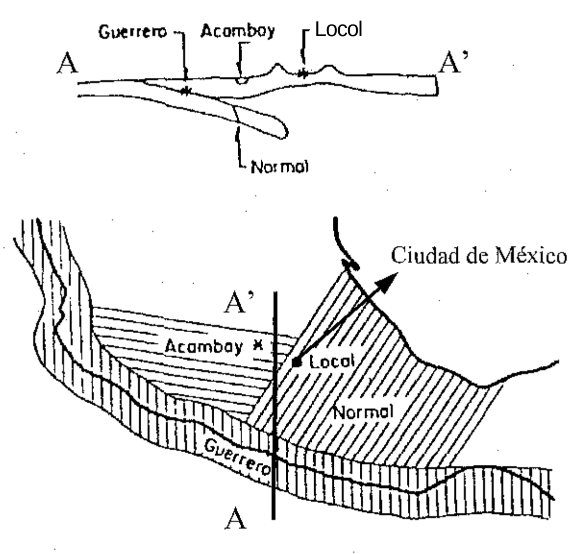

Figura 1. Ubicación de las principales fuentes sismogénicas que afectan a la Ciudad de México (modificado de Rosenblueth *eí al.,* 1989).

(a) Terremotos de subducción: Los sismos originados en esta región, históricamente son los que más daños han generado en la Ciudad de México, principalmente en la *zona del lago,* mientras que en la *zona de lomas* no se han reportado daños. La magnitud máxima registrada desde 1800 es Ms = 8.4 para el sismo de Jalisco de 1932. Sin embargo, los sismos originados cerca de las costas de Guerrero y Michoacán producen los movimientos más violentos en el valle. Estos sismos presentan magnitudes menores que Ms = 8.4. Existe

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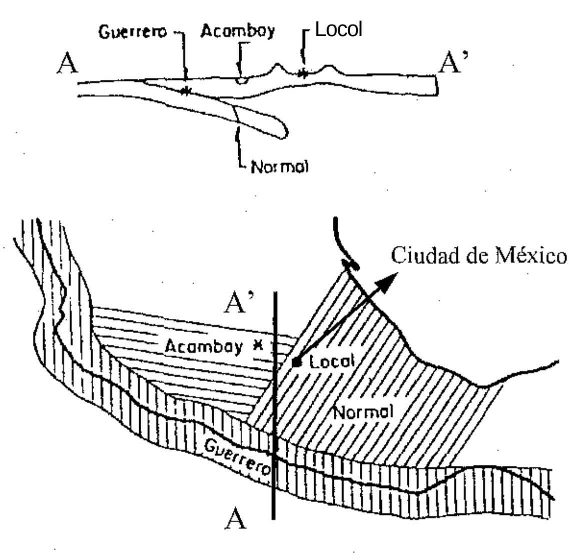

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(b) Terremotos intraplaca: Estos terremotos se originan en la placa subducida de Cocos en la parte centro-sur de México a profundidades menores de 80 km y distancias menores de 200 km de la costa (Iglesias *et al.,* 2002). Rosenblueth *et al.* (1989) consideran que el terremoto más peligroso para el valle tiene una Mw = 6.5 y H = 80 km. Sin embargo, el sismo de Tehuacán de 1999 tuvo Mw = 7.0 y H = 60 km. Este terremoto causó gran daño en el estado de Puebla y Morelos. En la Ciudad de México a una distancia aproximada de 200 km, el daño estuvo asociado con elementos no-estructurales (Singh *et al.,* 1999 y Shapiro *et al,* 2000). Iglesias *et al.* (2002) y Montalvo-Arrieta *et al.* (2002a) mencionan que si ocurre un sismo con Mw = 7.0 a una distancia epicentral entre 120 a 150 km, se presentarían grandes daños en la ciudad, principalmente en la *zona de lomas.* García y Suárez (1996) recopilaron información de los efectos sentidos por terremotos históricos en la República Mexicana, incluida la Ciudad de México, Ellos encontraron que el temblor de 1858 causó muchos perjuicios en el estado de Michoacán y en la Ciudad de México. Periódicos de esa época mencionan que en Coyoacán muchas casas fueron dañadas. Esta localidad, ahora la Delegación Coyoacán, se ubica entre *la zona de lomas y transición.* Singh *et al.* (1996) obtuvieron una magnitud de 7.7 para este sismo y sugieren como fuente un sismo intraplaca.

(c) Terremotos continentales: La actividad en esta región está caracterizada por sismicidad somera y de tipo extensional. Los grandes temblores ocurridos en esta región son el terremoto de Acambay de 1912 (Ms = 6.7) y el sismo de Jalapa de 1920 (Ms = 6.2) (Suter *et al.,* 1996). Estos autores estimaron valores de intensidad VIII en la Ciudad de México, para el sismo de Acambay. El sismo que más daños ocasionaría en el valle es uno de Ms = 6.7 (tipo Acambay) a una distancia de 100 km. Los valores de deslizamiento en la región de Acambay son de aproximadamente 0.4 mm/año (Rosenblueth *et al.,* 1989) y Suter *et al.* (2001) encuentran que estos valores varían entre 0.02 a 0.18 mm/año para la parte central de México. Rosenblueth *et al.* (1989) mencionan que los periodos de retorno para estos sismos son de aproximadamente 1500 años.

(d) Terremotos locales: Este tipo de eventos se localizan bajo la Cuenca de México. Una de las características de esta zona, es que los sismos son someros, de tipo extensional y de magnitudes pequeñas. La máxima magnitud registrada instrumentalmente es Me = 3.9, en la región de Milpa Alta (UNAM and CENAPRED Seismology Group, 1995). Mooser (1987) menciona que el máximo sismo registrado en el siglo XX fue de M = 5.1. Rosenblueth *et al.* (1989) mencionan que el escenario más peligroso para la ciudad corresponde a un evento de M = 4.7 con una distancia focal de 11 km.

La figura 2 muestra el mapa del sur de México con las localizaciones epicentrales y áreas de ruptura de algunos de los terremotos usados en este trabajo, que corresponden a sismos de subducción e intraplaca con Mw > 5.0.

Aunque los terremotos locales y continentales representan el primordial peligro para la Ciudad de México, principalmente para *la zona de lomas,* el riesgo sísmico es bajo debido a los grandes periodos de recurrencia para este tipo de fuentes (> 1000 años). En este caso los sismos intraplaca representan el máximo riesgo para *lo, zona de lomas.*

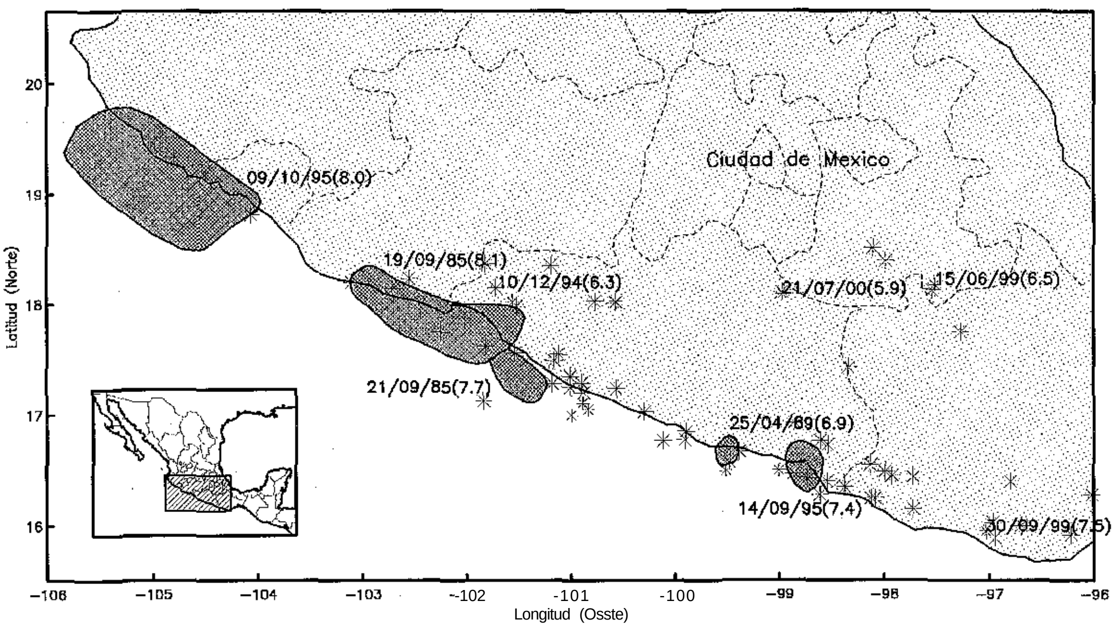

Figura 2. Ubicación de los sismos utilizados en este trabajo con Mw > 5.0, y de las principales áreas de ruptura, se señala además, para algunos eventos importantes fecha de ocurrencia y magnitud (modificado de Reinoso y Ordaz, 2001).

3. EFECTOS DE LA TRAYECTORIA DE LAS ONDAS SÍSMICAS EN LA CIUDAD DE MÉXICO

El estudio de los efectos del campo de ondas incidente en la Ciudad de México, está relacionado a tratar de comprender y explicar, en gran medida, el origen de las largas duraciones observadas en los registros de aceleración generadas a partir de sismos ubicados en la zona de subducción (Campillo *et al.,* 1988; Sánchez-Sesma *et al,* 1988; Kawase and Aki, 1989; Sánchez-Sesma *et al,* 1993; Singh y Ordaz, 1993; Chávez-García, *et al,* 1995; Sánchez-Sesma *et al,* 1995; Sánchez-Sesma, 1996; Shapiro *et al.,* 1997; Furumura and Kennett, 1998 and Shapiro *et al,* 2001). El interés se ha centrado en tratar de explicar estas largas duraciones como una combinación entre las estructuras regional y local en la respuesta sísmica (Cuenca de México, Cinturón Volcánico Mexicano (CVM) y la Trinchera Mesoamericana). La presencia de una zona de baja velocidad debajo de la parte centro-sur del CVM produce que la señal sísmica sea amplificada en el centro de México (Shapiro *et al,* 1997). Esta estructura fue inferida a partir de las observaciones de amplificación

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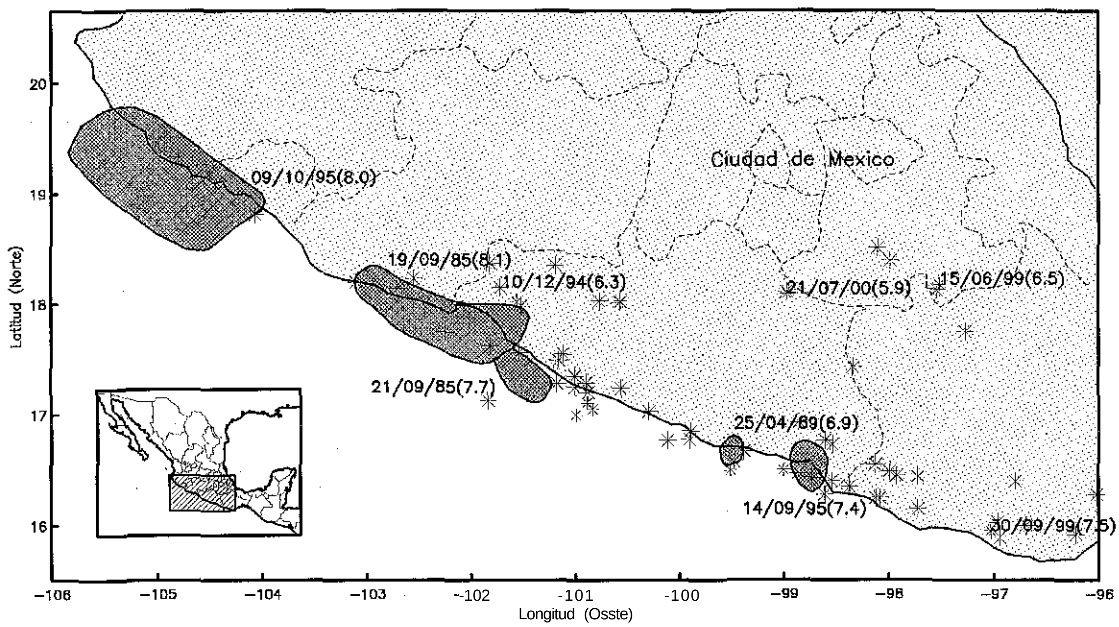

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3. EFECTOS DE LA TRAYECTORIA DE LAS ONDAS SÍSMICAS EN LA CIUDAD DE MÉXICO

El estudio de los efectos del campo de ondas incidente en la Ciudad de México, está relacionado a tratar de comprender y explicar, en gran medida, el origen de las largas duraciones observadas en los registros de aceleración generadas a partir de sismos ubicados en la zona de subducción (Campillo *et al.,* 1988; Sánchez-Sesma *et al,* 1988; Kawase and Aki, 1989; Sánchez-Sesma *et al,* 1993; Singh y Ordaz, 1993; Chávez-García, *et al,* 1995; Sánchez-Sesma *et al,* 1995; Sánchez-Sesma, 1996; Shapiro *et al.,* 1997; Furumura and Kennett, 1998 and Shapiro *et al,* 2001). El interés se ha centrado en tratar de explicar estas largas duraciones como una combinación entre las estructuras regional y local en la respuesta sísmica (Cuenca de México, Cinturón Volcánico Mexicano (CVM) y la Trinchera Mesoamericana). La presencia de una zona de baja velocidad debajo de la parte centro-sur del CVM produce que la señal sísmica sea amplificada en el centro de México (Shapiro *et al,* 1997). Esta estructura fue inferida a partir de las observaciones de amplificación regional hechas por Ordaz y Singh (1992); Sánchez-Sesma *et al.* (1993); Síngh *et al.* (1995). Los resultados de Furumura y Kennett (1998) muestran que las largas duraciones en la Ciudad de México se pueden reproducir a partir de modelos regionales 2D. Por otro lado, Shapiro *et al.* (2000); Montalvo-Arrieta *et al.* (2002a); Iglesias *et al.* (2002) encontraron que los registros de sismogramas en la estación CU muestran que las amplitudes de las ondas sísmicas son fuertemente disminuidas cuando sus trayectorias pasan a través del volcán Popocatépetl antes de alcanzar la ciudad. Estos autores mencionan que la fuerte atenuación es atribuida a la presencia de magma y fusión parcial de rocas debajo del volcán. Shapiro *et al.* (2000) mencionan que las ondas sísmicas son disminuidas por un factor de alrededor de 1/3 en frecuencias mayores a 1 Hz, comparadas con respecto a las ondas que no pasan bajo la estructura volcánica. Estas observaciones se hicieron durante el sismo del 15 de junio de 1999 (Mw = 7.0), cuyo epicentro fue ubicado al sureste de Tehuacán, Puebla (figura 2). Como se mencionó anteriormente, este evento causó severos daños en el estado de Puebla y Morelos. En la Ciudad de México localizada a 200 km del epicentro, el daño estuvo asociado solamente a elementos no-estructurales. El movimiento del suelo observado durante este sismo fue mucho más pequeño de lo predicho (Singh *et al.,* 1999; Shapiro *et al.*, 2000). El riesgo sísmico en la *zona de lomas* por sismos intraplaca es alto, sin embargo, con este sismo no se reportaron graves daños, ya que la ciudad queda incluida en la zona de sombra producida por el volcán. Debido a la alta atenuación observada en el temblor de Tehuacán, Montalvo-Arrieta *et al.* (2002a) e Tglesias *et al.* (2002), a partir del terremoto de Copalillo del 21 de julio de 2001 (Mw = 5.9) escalaron la respuesta sísmica en CU para un sismo postulado con Mw = 7.0 igual al de Tehuacán. Estos autores encontraron que las máximas aceleraciones en el dominio del tiempo y de la frecuencia para el sismo escalado, son mayores que para el terremoto del 15 de junio de 1999. El sismo de Copalillo se localizó al suroeste del volcán Popocatépetl. En el apéndice 1 (Montalvo-Arrieta *et al,* 2002a) se presentan los resultados del escalamiento realizado al evento del 21 de julio de 2001 (Mw = 5.9).

En el apéndice IV (Montalvo-Arrieta *et al.,* 2002d) se presentan resultados de una red temporal de banda ancha, localizada alrededor de los volcanes Popocatépetl e Iztaccíhuatl, en donde se hace un estudio de los efectos de atenuación provocados por el paso de ondas sísmicas a través de volcanes activos.

4. GEOLOGÍA

La Cuenca de México se encuentra localizada en la parte central del CVM. Las rocas volcánicas forman el contenedor de la cuenca, la cual está formada por depósitos lacustres, volcánicos, arenas y gravas. A partir de datos geotécnicos la Ciudad de México ha sido dividida en tres regiones (1) la *zona de lomas,* formada por terrenos compactos, arenolimosos con alto contenido de grava unas veces, y otras por tobas pumiciticas bien cementadas y derrames basálticos; (2) la *zona del lago,* la cual está formada por arcillas con espesores que varían de 10 a 100 m y que se encuentran sobreyaciendo arenas; y (3) la *zona de transición,* compuesta por arenas aluviales, limos y capas esporádicas de arcillas (Marsal y Mazarí, 1962). La figura 3, muestra la zonación geotécnica, la localización de las estaciones en la *zona de lomas* y algunas calles y sitios de referencia.

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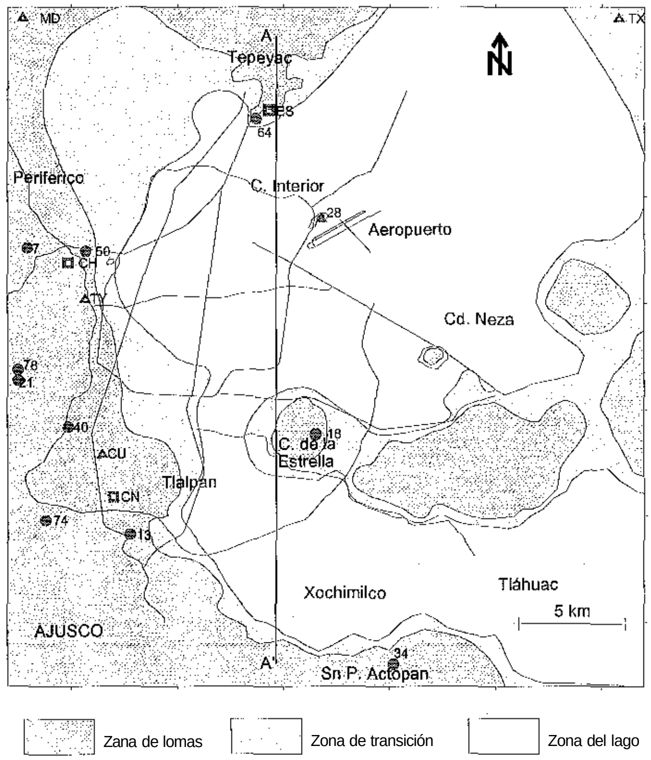


Figura 3. Ciudad de México: ubicación de las zonas geotécnicas, localización de las estaciones acelerométricas en la *zona de lomas* y algunos lugares de referencia en la ciudad (modificado de Reinoso y Ordaz, 1999).

La porción sur de la *zona de lomas* esta formada por depósitos recientes de flujos de lavas provenientes de la Sierra de Chichinautzin (sitios CU, 13, 34, 40 y 74, figura 3). La parte oeste esta dominada por tobas (sitios CH, TY, 07 y 78), mientras que la zona norte de la ciudad esta compuesta por lavas de edad Mioceno (sitios MD, TX, ES y 64), que sobreyacen rocas volcánicas de edad Oligoceno. Las lavas Miocénicas subyacen las capas lacustres de la zona del lago, los depósitos de la Sierra de Chichinautzin y las tobas de la porción oeste. La figura 4 muestra una sección geológica en la dirección Norte - Sur (Perñl A-A', figura 3) que atraviesa la Cuenca de México desde la Sierra de Guadalupe hasta la Sierra de Chichinautzin (Mooser *et al.* 1996), en esta sección es claro como la capa de edad Mioceno esta inclinada hacia el sur. En los apéndices I y IV (Montalvo-Arrieta *et al.*, 2002a; Montalvo-Arrieta *et al.*, 2002d) se presentan los perfiles geológicos utilizados para hacer modelado ID en la *zona de lomas.* En el apéndice II (Montalvo-Arrieta *et al,* 2002b) se presenta un modelo en 2D para tratar de explicar el efecto sísmico de una capa inclinada, tomando en cuenta el perfil geológico de la figura 4.

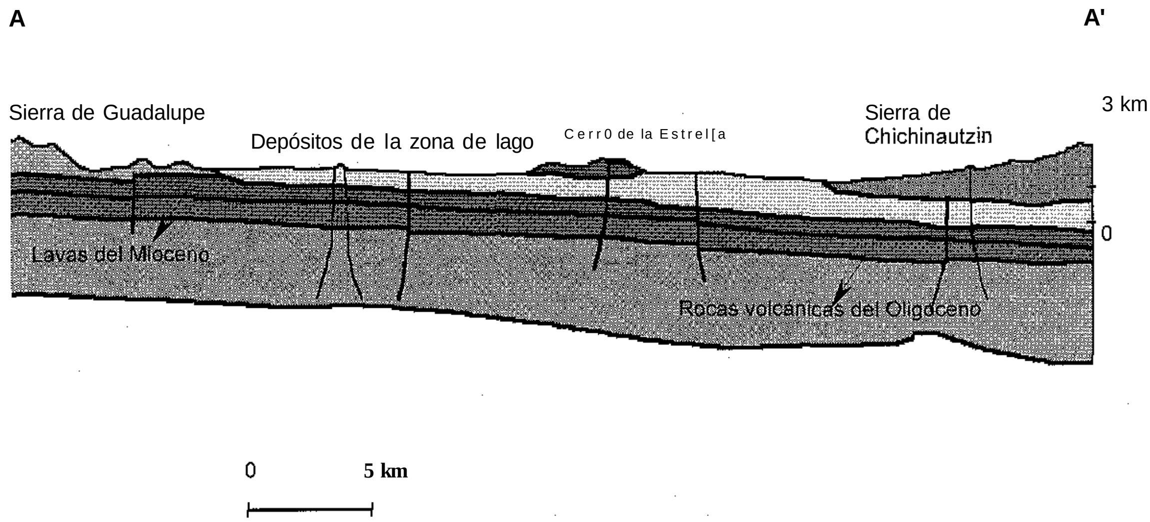
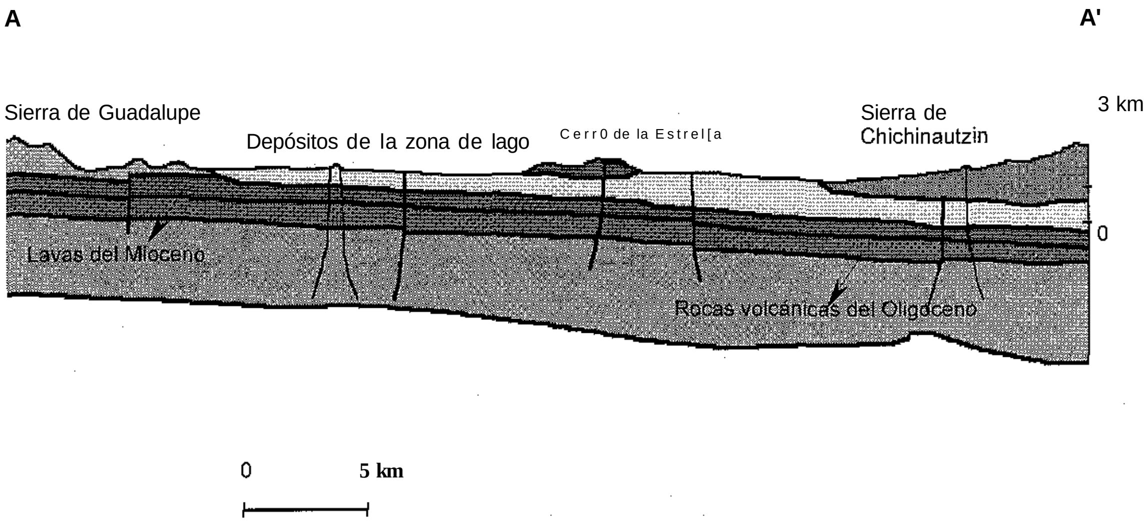


Figura 4. Sección geológica en dirección Norte - Sur, desde la Sierra de Guadalupe a la Sierra de Chichinautzin (modificada de Mooser *et al.*, 1996)

5. DIFERENCIAS EN LA RESPUESTA SÍSMICA ENTRE LAS ESTACIONES DE LA ZONA DE LOMAS

La observación instrumental de los movimientos del terreno ocurridos en los últimos años en el Valle de México, ha permitido identificar grandes variaciones espaciales en los registros del movimiento del terreno, sobre todo en sitios de la zona del lago. Respecto a esta variación en las observaciones, en este trabajo se hace un estudio de la respuesta sísmica de la *zona de lomas*. La estación CU tradicionalmente ha sido utilizada como sitio de referencia en la ciudad, este sitio cuenta con registros desde 1964 hasta la fecha. Durante el sismo del 19 de septiembre de 1985 (Mw = 8.0), sólo dos estaciones se localizaban en esta zona; Tacubaya (TY) y CU. Antes de contar con varios registros de terreno firme para un mismo temblor, se consideraba que las diferencias del movimiento entre los sitios de esta zona eran despreciables. Durante el temblor de 25 de abril de 1989 estas diferencias se hicieron evidentes, siendo hasta el sismo del 14 de septiembre de 1995 cuando se miden con mayor calidad (Reinoso y Ordaz, 1999).

Tanto los registros de aceleración, como los espectros de respuesta de seudoaceleración y desplazamiento correspondientes, presentan diferencian notables. Por ejemplo, en aceleración máxima: 0.05 m/s^2 para el sitio 64 y 0.25 m/s^2 para el 21, un factor mayor que cuatro (Reinoso y Ordaz, 1999). Analizando estos registros en el dominio de la frecuencia se observan dos grupos que, atendiendo a la amplitud de sus espectros de Fourier, presentan un comportamiento bien definido. El grupo sureste (estaciones 07, 13, 21, 34, 40, 50, 74, 78, CU, CH y TY) tiene mayor amplitud que el grupo noreste (64, ES, TX y MR), mientras



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6. RESPUESTA SÍSMICA DE LOS DEPÓSITOS PROFUNDOS EN EL VALLE DE MÉXICO

A partir de estudios recientes se ha demostrado la importancia de la interacción entre la propagación de ondas guiadas por las capas profundas y los depósitos someros de las capas de arcilla en la respuesta sísmica del Valle de México (Chávez-García, *et al.* 1995; Furumura y Kennett (1998) y Shapiro *et al.* 2001). Esta relación podría explicar la larga duración del movimiento y algunas amplificaciones observadas en la *zona del lago* de la ciudad. En el apéndice III (Montalvo-Arrieta *et al*, 2002c), se presenta una nueva hipótesis para estudiar la respuesta sísmica de la interacción entre los depósitos profundos y las capas someras de la zona del lago de la Ciudad de México.

7. LEYES DE ATENUACIÓN

En esta parte del trabajo, se utilizaron todos los sismos ($Mw > 5.0$) registrados en CU desde 1964 al 2001, con la finalidad de obtener nuevos coeficientes en las leyes de atenuación para valores pico o máximos de aceleración y velocidad del terreno, espectros de amplitud de Fourier. También se obtuvieron ecuaciones predictivas para espectros de respuesta de aceleración, velocidad y desplazamiento para el 5% de amortiguamiento. Estas leyes pretenden hacer estimaciones más reales del movimiento del suelo y de los espectros de respuesta para futuros terremotos. El estudio estará dirijido a predecir la respuesta sísmica para un terremoto hipotético de $Mw = 8.2$ y distancia $R = 280$ km, que se origine en el gap de Guerrero y un temblor intraplaca de $Mw = 7.0$ con distancia epicentral de $R = 137$ km.

El análisis para eventos de subducción se llevó acabo a partir de 44 sismos, estos eventos se localizaron a lo largo de la Trinchera Mesoamericana y representan la ocurrencia de todos los grandes sismos registrados en CU, a partir de 1965, con $5.0 < Mw < 8.0$ y distancias a la fuente o zona de ruptura de $270 < R < 545$ km. Para los sismos intraplaca se utilizaron 14 eventos con $5.2 < Mw < 7.4$ y distancias a la fuente o zona de ruptura de $136 < R < 435$ km.

En el apéndice IV (Montalvo-Arrieta *et al*, 2002d) se presenta el estudio detallado de este análisis. En los resultados de las predicciones se pueden observar los efectos de la magnitud y distancia en los espectros de amplitud de Fourier y espectros de respuesta. Además, se que las estaciones que no pertenecen a estos grupos (18 y 28) tienen una respuesta intermedia. Esta observación puede explicarse por la presencia de formaciones recientes en la zona suroeste de la ciudad, lo que provoca una considerable amplificación de las ondas que llegan a la superficie. Lo contrario sucede con el grupo noreste cuyas estaciones se encuentran localizadas en depósitos más antiguos. En resumen, a partir de la instalación de más estaciones en la *zona de lomas*, se han podido observar variaciones considerables en el movimiento sísmico en el terreno firme de la Ciudad de México. Esto nos ha llevado a hacer un estudio del comportamiento de la respuesta sísmica de la *zona de lomas*. En los apéndices I, II y IV (Montalvo-Arrieta *et al*, 2002a; Montalvo-Arrieta *et al*, 2002b; Montalvo-Arrieta *et al*, 2002d) se presenta con detalle estos trabajos.

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muestra que para la predicción del sismo proveniente de la Brecha de Guerrero (Mw = 8.2) las amplitudes espectrales y de los espectros de respuesta son mucho más grandes que para el sismo de Michoacán de 1985, éste no es un resultado inesperado, sin embargo, implica que el movimiento experimentado durante el sismo de 1985 no representa el rango límite. Semejantes resultados fueron obtenidos para la predicción del sismo intraplaca.

8. RESPUESTA SÍSMICA DE VALLES ALUVIALES ESTRATIFICADOS EN DOS DIMENSIONES

Finalmente, en el apéndice V (Gil-Zepeda *et al*, 2002), se presenta un método Híbrido de Elementos de Frontera Indirecto y Número de Onda Discreto, para calcular la repuesta sísmica de valles aluviales irregulares estratificados en 2D bajo incidencia de ondas *SH*. En este trabajo se presentan comparaciones con otros métodos utilizados en la literatura, principalrnente con el modelo de Kawase y Aki (1988), estos autores analizaron la respuesta sísmica de un modelo simple del Valle de México (un estrato), los resultados de la comparación muestran acuerdos muy buenos, lo que da pie, para realizar modelos más realistas de la Cuenca de México, que consideran una estratificación múltiple, en este caso de sedimentos lacustres, aluviales y flujos de lava entre otros.

9. DISCUSIÓN Y CONCLUSIONES

En este trabajo, se ha estudiado con detalle la respuesta sísmica de las estaciones localizadas en el terreno firme de la Ciudad de México. El interés en el estudio de *la zona de lomas* se ha acentuado por las siguientes razones: (a) La *zona de lomas* representa sísmicamente el sitio de referencia de la Ciudad de México, (b) En los últimos años esta porción de la ciudad, se ha desarrollado enormemente, ubicándose en ella grandes sitios habitacionales, modernos centros comerciales y de negocios.

Uno de los motivos por la ausencia de estudios de la respuesta sísmica a detalle en esta región, es debido a que se ha considerado como de bajo riesgo para sismos provenientes de la zona de subducción, los cuales, históricamente, son los que mayor daño han causado principalmente en la *zona del lago* de la Ciudad de México. Este tipo de eventos no son los únicos que pueden perjudicar a la ciudad. Las principales fuentes sismogénicas son: (a) subducción, (b) intraplaca, (c) continentales y (d) locales; siendo los últimos tres, los más peligrosos para el terreno firme; los sismos locales y continentales si bien son los más dañinos, debido a sus grandes periodos de recurrencia (> 1000 años, continentales) representan menor riesgo. No así los sismos intraplaca, aunque sus periodos de recurrencia no están muy bien cuantificados, en los últimos cuatro años se han generado dos grandes sismos (Mw > 7.0) que han causado severos daños en algunas ciudades del centro-sur de la república. Un temblor intraplaca que no causó perjuicios debido a su magnitud (Mw = 5.9), pero que es importante tener en cuenta por la cercanía de su origen a la Ciudad de México (alrededor de 137 km) es el sismo ocurrido en la región de Copalillo, Guerrero. Si un terremoto se origina en esa región con una Mw > 7.0, el daño a la ciudad sería importante principalmente en \a*zona de lomas.*

muestra que para la predicción del sismo proveniente de la Brecha de Guerrero (Mw = 8.2) las amplitudes espectrales y de los espectros de respuesta son mucho más grandes que para el sismo de Michoacán de 1985, éste no es un resultado inesperado, sin embargo, implica que el movimiento experimentado durante el sismo de 1985 no representa el rango límite. Semejantes resultados fueron obtenidos para la predicción del sismo intraplaca.

8. RESPUESTA SÍSMICA DE VALLES ALUVIALES ESTRATIFICADOS EN DOS DIMENSIONES

Finalmente, en el apéndice V (Gil-Zepeda *et al*, 2002), se presenta un método Híbrido de Elementos de Frontera Indirecto y Número de Onda Discreto, para calcular la repuesta sísmica de valles aluviales irregulares estratificados en 2D bajo incidencia de ondas *SH.* En este trabajo se presentan comparaciones con otros métodos utilizados en la literatura, principalrnente con el modelo de Kawase y Aki (1988), estos autores analizaron la respuesta sísmica de un modelo simple del Valle de México (un estrato), los resultados de la comparación muestran acuerdos muy buenos, lo que da pie, para realizar modelos más realistas de la Cuenca de México, que consideran una estratificación múltiple, en este caso de sedimentos lacustres, aluviales y flujos de lava entre otros.

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Como se ha mencionado, el riesgo sísmico en la *zona de lomas* es bajo, sin embargo, el hecho que en pocos años hallan ocurrido tres sismos intraplaca importantes en e] centro-sur de la república, crea el interés y/o necesidad de conocer y estudiar la respuesta de esta zona considerada como la más segura de la ciudad.

Desde el punto de vista sísmico, diversos autores han encontrado diferencias importantes en la respuesta entre estaciones localizadas en esta región. Se analizaron con detalle los dos tipos de diferencias existentes: a) local y b) regional. En la primera de ellas se observó, que existe un efecto de sitio importante en distancias de alrededor de 3 km (sitios CU, 74 y 40). Las diferencias encontradas se explican por la presencia de capas recientes y poco consolidadas que varían en espesores debajo de los depósitos de lava. El segundo efecto, el que se denominó como regional (diferencias entre el sur y el norte), se puede explicar por la presencia de formaciones recientes en la zona suroeste de la ciudad, mientras que en la porción norte, los depósitos aflorantes son mucho más antiguos y compactos. Se obtuvo un valor de amplificación espectral promedio entre las estaciones del sur y las del norte de 3 veces y, además se muestra que con un modelo simple de una capa inclinada bajo incidencia de ondas SH se puede explicar esa amplificación.

Por otra parte, se generaron leyes de predicción para valores pico o máximos de aceleración y velocidad del terreno, espectros de amplitud de Fourier. También se obtuvieron ecuaciones predictivas para espectros de respuesta de aceleración, velocidad y desplazamiento para el 5% de amortiguamiento. Estas leyes pretenden hacer estimaciones más reales del movimiento del suelo y de los espectros de respuesta para futuros terremotos. El estudio se encaminó a predecir la respuesta sísmica para dos de los peores escenarios en la Ciudad de México: un terremoto hipotético de Mw = 8.2 y distancia R = 280 km, que se origine en el Gap de Guerrero y un temblor intraplaca de Mw = 7.0 y una distancia epicentral d e R ^ 137 Ion. Los resultados muestran que los valores predichos para cada uno de los parámetros en el caso del sismo de subducción son mucho más grandes que para el sismo de Michoacán de 1985, este no es un resultado inesperado, sin embargo, implica que el movimiento experimentado durante el sismo de 1985 no representa el rango límite, semejantes resultados fueron obtenidos para la predicción del sismo intraplaca, las diferencias encontradas en ambas predicciones es que el daño ocasionado en la ciudad variará de acuerdo al tipo de estructuras y el sitio donde se encuentren. Siendo los sismos de subducción los más peligros para la *zona del lago* en frecuencias menores a 1 Hz, mientras que los eventos intraplaca ocasionarán mayores perjuicios en la *zona de lomas* en el rango de frecuencias mayores a 1 Hz.

Otro aspecto de interés es que se confirmo que el efecto de sitio en la *zona del lago* está controlado por el efecto de capas someras y profundas, siendo el promedio de las estaciones del sur, las que representarán a la señal de entrada en la base de las capas del suelo a una profundidad de aproximadamente 100 m, mientras que el promedio espectral de las estaciones de la zona norte representan la respuesta de una estructura más profunda a una distancia mayor a 1000 m.

Una observación importante, es que el riesgo sísmico en la ciudad es menor cuando los sismos se originan al sureste del volcán Popocatépetl, este análisis se llevo acabo a partir de los resultados obtenidos por una red temporal de banda ancha localizada alrededor de los volcanes Popocatépetl e íztaccíhuatl. Se pudo confirmar la hipótesis hecha por Shapiro *et al.* (2000), a partir del sismo de Tehuacán del 15 de junio de 1999, en donde mencionan una fuerte atenuación en las ondas sísmicas (en frecuencias mayores a 1 Hz) cuando estas atraviesan el volcán y llegan a la ciudad. Estas observaciones indican que si bien los sismos intraplaca pueden ocasionar graves daños en la Ciudad de México principalmente en la *zona de lomas,* si estos se llegan a generar al este o sureste de la estructura volcánica, los perjuicios serán menores que si el sismo ocurre en la región de Copalillo. Por consiguiente, el volcán activo sirve como una barrera natural para la ciudad.

Finalmente, se presentó un método híbrido de Elementos de Frontera Indirecto y Número de Onda Discreto, para calcular la repuesta sísmica de valles aluviales de forma irregular estratificados en 2D, bajo incidencia de ondas SH, los resultados obtenidos muestran que la variación en las propiedades de las capas en 2D puede producir importantes efectos en la respuesta sísmica, los cuales no pueden ser observados en modelos ID. Además, una ventaja de este método, es que la solución numérica es muy rápida y económica, lo que permitió hacer una aplicación al Valle de México.

De todo lo anterior y en concordancia con los objetivos se obtienen las siguientes conclusiones:

- Para sismos de subducción las mayores amplitudes se encuentran en frecuencias menores a 1 Hz, mientras que para sismos intraplaca, éstas se presentan a frecuencias mayores a lHz. En el dominio del tiempo las mayores amplitudes se observan en las estaciones del sur respecto a las del norte, y en el dominio de la frecuencia la amplificación promedio entre estas zonas es de 3 veces en el rago de frecuencias de 0.7 a 10.0 Hz. En la porción sur de *la zona de lomas,* se presentan marcados efectos de sitio en frecuencias mayores a 1 Hz, entre las estaciones 34, 40 y 74 respecto a CU. El riesgo sísmico en la Ciudad de México es menor cuando los terremotos ocurren al este y sureste del volcán Popocatépetl,
- El promedio espectral de las estaciones localizadas al sur del terreno firme representa el sitio de referencia y equivale a la señal de entrada en la base de las capas del suelo, a una profundidad no mayor a 100 m mientras que el promedio espectral de las estaciones localizadas al norte de la *zona de lomas* representan la respuesta más profunda de la cuenca (> lOOOm).
- La diferencia entre las estaciones del norte y sur se puede explicar a partir de un modelo de una capa (lavas de edad Mioceno) inclinada (3^{o}) bajo incidencia de ondas *SH.*
- Se obtuvieron nuevos coeficientes en las ecuaciones utilizadas para predecir el movimiento del terreno. Se presentaron los escenarios más peligrosos para la ciudad: los sismos hipotéticos con Mw = 8.2 y 7.0 generados en la brecha de Guerrero (subducción) y región de Copalillo (intraplaca), respectivamente.
- Se presentó una metodología para evaluar la respuesta sísmica de valles aluviales estratificados bajo incidencia de ondas *SH* en 2D y se aplicó en el Valle de México.

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APÉNDICE I

OBSERVATIONS OF STRONG GROUND MOTION AT HILL STTES IN MÉXICO CITY FROM RECENT EARTHQUAKES

(Geofísica Internacional en Prensa)

Geofísica internacional (2002), Yol 41, Num. 1, pp.

Observations of strong ground motíon at hill sites in México City from recent earthquakes

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Received: December 12, 2000; accepted: October 2, 2001.

RESUMEN

Se presentan resultados del estudio de la respuesta sísmica de las estaciones de ía Red Acelerométriea de ¡a ciudad, de México localizadas en la zona de lomas. Se analizaron siete sismos provenientes de diferentes fuenies sismogéiúcas, con magnitudes entre 5.9 y 7.3. Existe una sensible dependencia de la impuesta sísmica en las estaciones con respecto a las características del terremoto (magnitud, acimut, distancia epicentra] y profundidad). Los sismos provenientes de ía costa concentran su mayor energía en frecuencias menores de 1 Hz, Mientras que sismos intraptaca en frecuencias mayores de 1 Hz,. El análisis de ios acelerogramas en el dominio del tiempo muestra dos tipos significantes de amplificación: Regional (estaciones localizadas al suroeste de ia zona de íornas presentan mayores amplificaciones que !as localizadas ai norte de la ciudad) y local (ías estaciones localizadas en ¡a zona suroeste presentan una mayor amplificación respecto a CU). Se realizaron además cocientes espectrales de las estaciones localizadas en la parte central y suroeste respecto a la estación Eslanzuela (ES), localizada en la zona norte de la ciudad. La comparación de estos cocientes espectrales muestra que existen amplificaciones relativas hasta de cuatro veces entre algunas de ias estaciones localizadas en la zona suroeste de ía ciudad en el rango de frecuencias de i a 3 ÍIz. Estos efectos observados se pueden deber a la presencia de material mucho más suave debajo de ios (lujos de iava donde se localizan las estaciones,

PALABRAS CLAVES: Ciudad de México, zona de lomas, cocientes espectrales, electos de sitio, amplificación y trayectoria.

ABSTRACT

Resnlís of seismic response of the siations of ¡he Acclerometric Network of México City located in tíiehiíl 20ne are prcseiited. Seven eanhquakes from different seismic sonrees were analyzed, wifti magnitudes beíween 5.9 and 7.3. There is a dependence of the seismic respe-use at hüi sites on earfhqiiake magnitude, azimtirh, epicentral distance aiid depth. Subduction earthquakes concéntrate their encrgy at iow frequencies (<1 1-íz), while intraplaie earthquakes have Crcquencies liigher than 1 i-iz. We find two significaní íypes of amplification: regionaJ, for sites located to the sonthwest hill zone over sites located tn the north of the city, and local. Stations located ín Ule srmthwcst área show higher amplification wilíi respect to tile referen ce station. The spcctai! futios of stations located in the central and souüiwesí part of the hiíl zone showreíaiive amplifications up to four times higher than stations located in the north área of the city for frequencies between 1 and 3 Hz. These iocai amplifications may be due to the piesence of sofi material under ihe lava fíows where tiie stations are located.

KEY WORDS: México City, iiilí zone, spectral raüo, site eifecta, ampliücation and wave path.

INTRODUCCIÓN

Atnrilificattón of ground motion during eanhquakes in sedimenlary alluvial valleys can cause considerable damage, as in the earthquake of Míchoacán 1985 in México City. Nitmerotis autliors have been working on Ihe lask of explaitv ing the hehavior of the seismic response of this basin. The observed effeets are euormous spectral amplifications and targe strong motion dwatíon and spatíat variability. However, the seismic response oí firm soil in ihe hill zone of México City has not been much studíed.

Singh *et al* (1988) and Ordaz and Singh (1992) suggest that Ihe hill zone in the Valley of México features important ampliflcalion wifh respect to Ihe attenuation laws. This amplification eould reach í 0 times for frequencies between 0,2 and 0.7 Hz. Singh *ei al* (1995) conclude that lwo norlhern stations of the hill zone (MADI and TEXC), where the motion is less amplified ihan at other sites, siill present a signíHcant amplificalion. Thcy sugge^i Iow S wave vclocities and complex síructure of tiie upper (ayers of volcanic rocks as a possible explanation.

Reinoso *and* Ordaz (1999) fbund imporiani differences in ihe Fourier amplitude spectra in the hiil zone between áreas inlhe southwest and the north for the earthquake of September 14, 1995. The spectral amplitude of the stíitions of Ihe southwest is cight times larger tlian to ¡he north for all frequencies.

Pérez-Rocha (1.998) carried oni the scalirtg of amplitudes of groiind motion in the liül zone for severa] earmquakes using simple theorelical raodels of the seismic source. He found that the strongest earth.qua.kes for structures in the Valley of México are those from the Guerrero coast. íf an earthquake with M = 8.1 originales at this coast, the design spectra specil'ied inthebuilding code may be underestimatedfor short-period structures. This is particulaiiy dangerous for structures located in the hill zone. He concluded that a detai led study of the seismie respon.se of Ihe hill zone is justified.

As for frequencies larger than í Hz, Pérez-Rocha (1998) suggested Ihaí earthquak.es fromthePetatlán gapare veryenergelic for tlíese frequencies. This is very important for the hill zone since structures there are vulnerable to motions with this freqiiency contení. An earíhquake origínated in this subduction zone caused the collapse of ihe Ibero-American Univereity in 1979.

García and Suárez (1996) compíled the effeets of historícal earthquak.es felt in México, including México City and its liill zone. The earthquake of 1858 caused extensive damage in the state *of* Michoacán and i ti México Cily. Newspapers of that time describe thal in Coyoacán "many houses suffer daniages". This town is located al the border between the transition and hill zone. Singh *et al.* (1996) estímate a magnilude of 7.7 for Hits earthquake and suggesl a normal fault oriein.

GEOLOGY

The basírt of Mex ico is located ín the central part of the Mexican Volcanic Bell (MVB). Vbloanic rocks form the geology of thebasín. It ís fillcd by lacustrinc sediments, volcanic tuff, sands and gravéis. Based on Ihe geoteclmical chajacteristíes of its shallow layers, the Valley of México has been dividedmto three regions: (1) the hill zone, íbrmedby volcante tuffs and lava flows. (2) the lake bed zone, l'orrned by clays with thickness vaiying from 10 lo 100 m; and (3) the transilion zone, composed by alluviál sandy and silty layers. with scattered clay layers (Marsal and Mazai'i, 1959).

The Southwest paitof ihe hill zone is formed by recent Qnaternary deposits (lava flows from Xille volcano) that overfie sol't material (Delgado *et al.,* \999). The north part is composed of hard Pleistocene lavaofafcw meters in ihickness, that overlies volcante rocks of Oligocene age (Singh *et al,* 1.995).

DATA AND ANALYSIS

We examine the effeets of azimuth, magnitude, ep¡central distante and depth ofearthqnak.es on ihe seísmic response of the stations located in fheiuli zone of México Cily.

Seven earthquakes iram differenl seismíc regions were used (Figure í). Most of the data, come from subduction eaithquakes with epiceniral dislanccs larger than i 30 km from

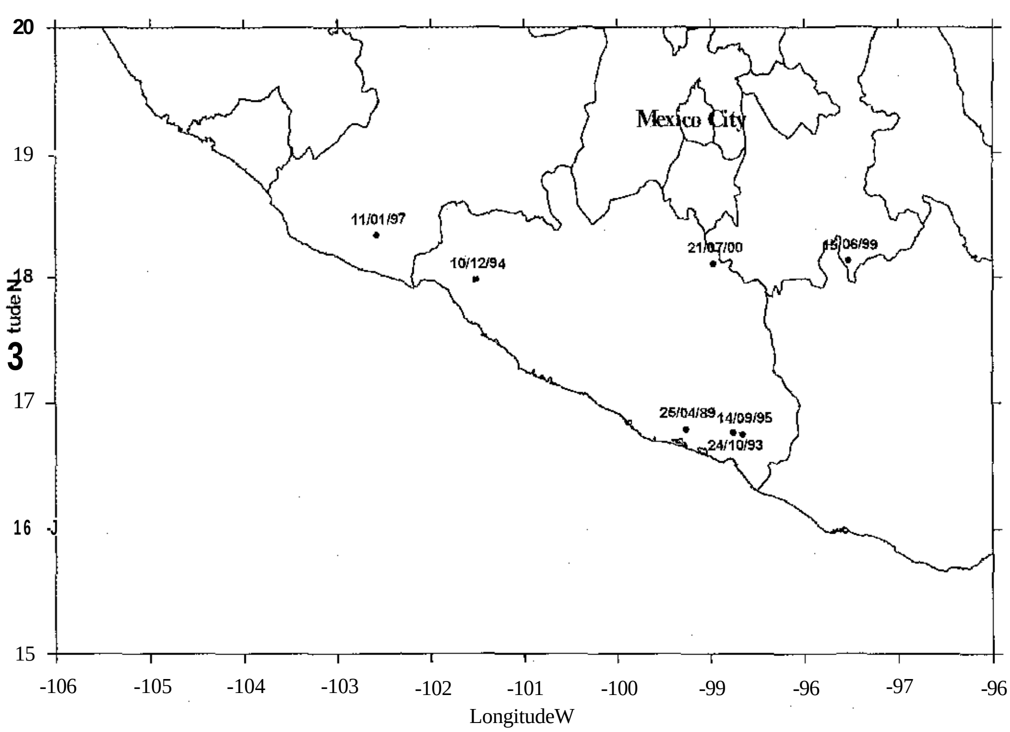


Fig. 1. Location of epiceníers and México City.

t.he aecelerometer station at Ciudad Universitaria. UNAM (CU). Tabíe 1 shows the parame ters of thes-e earthquakes. We shall discusa events 1, 2 and 7 ín particular, dueto the ir combination of relatívely near epicentral distance and large magnitude.

Ar'ler íhe extensíve damage cau.sed by íhe 1985 Míchoacán eailhquake in México City, the accelerometer array has grown consicierably. It now consists ol' more than 90 frtie-field digital aecelerometric stations, 18 of them in the hil! zone (stations 07. 13, 18, 21, 28, 34, 40, 50, 64, 74, 78, ES, CU, CH, CN, TX, MD and TY). Figure 2 shows the locaüon of these stalions.

The earthqn&ke of June 15,1999, M 7.0, H ^ 60 km

The cpícenter of this earthquake (see Tablc 1 and Figure I) was located soulliwest of Tehuacán, Puebla, It caused severe damage in the state of Puebla and Morel os. In México City, about 200 km from ihe epieenter, damage was associated otily to non-strucHiral elemcnts (Singh *et al..*. 1999). The ground motion observed in México City during this earthquafce was much smaller than the predicted one (Shapiro *etal,* 2000; *Sihgh etal...* 1999).

Figure 3 shows the accelerograms from this earthquake recorded ai hill zone sites and at station CUER in Cuemavaca, Mor, on fírm soil, but outside the valley. We also show Üie response spectra of absolute acceleration and dísplacements (with damping ratio of 5%) for (he norih-south component. The accclerograms show unusually large amplitudes and high frequency content. Stations 40, 34,74 and CU tire located in the southwest part of the ctty, very cióse lo each other (-3.0 km). The motion was expecled to be very similar, because four siles are on lava flows. However. station. CU exhibíts a smaller amplitude than sites 40, 34 and 74. Response spectra show similar hehavior. Figure 4 shows Fourier amplitude spectra for the horizontal components of motion. The largesi spectral amplitude is at stalions CUER, 34 and 74. The average spectra for the southwest stations, AS W, and for the north. station, AN, are shown.

The northern stations 07, 64 and ES show the smallest amplitude, bothin the frequency and time domains. This may be due to a strong change in the geological condilions between the southwest and north parts of the hill zone. The older rocks are norlh of the cíty. Attenuation ís not an important factor since Ihe dislance between the southwest stations and the north stations is small. The epicentral distance for sites CU and ES is 222 and 227 km respecttvely.

The earthquaiíe of September 14,1995, M 7.3

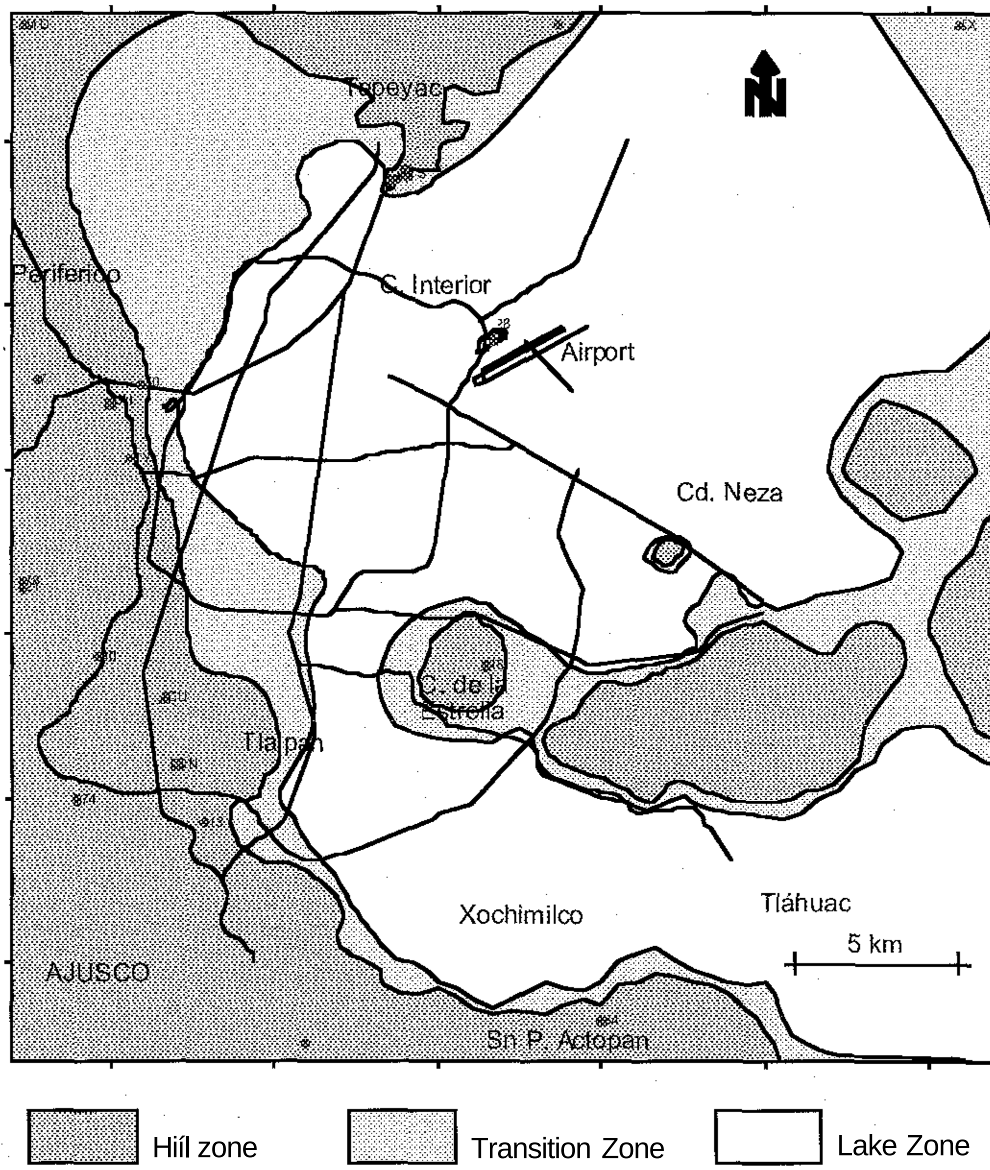
Figure 5 shows the accelerograms and the acceleration response spectra and dìsplacement (with dumping ratio of *5%)* corresponding lo the north-soulh componenL Inthe time domain, the accelerogrums diífer in amplitude and duratfon of motion. Figure 6 displays the smoothed Fourier amplitude spectra for all stations, using a one-sixth-octave band **ftlter.** Note the impoitant diñ'erences of Irequency conietu (Reinoso-Angulo, 1996; Reinoso and Orclaz, 1999).

From Figure 5, two groups of stations follow the same behavior: tlie southwest station and the north stations (Figure 2). North stations have smaller amplitudes than soutlv wesi stations. From Figures 3,4,5 and 6, the largcst spectral amplitudes arefound in the southwest part. of the city, while the smallest are observed in the northern part. The máximum spectral amplitude varíes for bolh earihquakes. For the 1995 eveni the máximum peaks are in Ihe range of 0.2 to 0.7 Hz, typical of subduciion eaithquakes. while for (he 1999 event the máximum peaks are between 1.0 and 3.0 Hz.

Table 1

fíarthquakes used in this work

Date dd/mm/yy	Origin	Magnitu.de	Depth (km)	Latitud*; (N)	Longitude (W)
(1) 15/06/99	Tntraplate	7.0	63	18.133	97.539
(2) 14/09/95	Subduction	7.3	21	16.752	98.667
(3) 10/12/94	Litraplate	6.3	53	17,980	101.520
(4) 24/10/93	Subduction	6.6	30	16.767	98.767
(5) 11/01/97	Intraplate	7.3	40	18.340	102.580
(6) 25/04/89	Subduction	6.9	23	16.795	99.275
(7)21/07/00	Intraplate	5.9	58	18.1)0	98.974



pig. 2. México City: acceteromctric stations, geoieelmícaí zones and rer̃crenee siles. The ai-ceieromclric amiy consists of inore than 90 free-field digitai acceJeronielric stations, 18 in üie Mil zone (07,13,18, 21, 28, 34, 40, 50 64, 74, 78, ES, CU, CH, CN, TX, MD and TY)

Figure 7 shows tlie average Fourier spectraof the noith-south component for stations in the southwest and in the north, for bolh events. Similar Lo Figures 4 and 6, Figuré 7 shows rhat foreach earthquake spectral shapes are very similar, aod tliat strong variations exist in ihe amplitudes for bolh groups of stations. The difference *h* up to 4 limes for both carth-quakes. Notice the diJíerence between tlie two lypes oí'earth-quakes; for the 1999 intraplale eartliquake the masiiman speciral amplitudes occur at í'requencies liigher than 1 Hz, while in the 1995 subduction earthquake Ihe máximum am-plitude occur at frequencies below 1 Hz.

The earthquake of July 21, 2000, M 5.9

This intraplate earthquake was t'elt in México City but no damage was reported. Due lo the umisual proximity to the cily (~ 137 km), acceleration records of very good qual-ity were obtaíned, and P and S wave arrivals are clearly seen (Figure 8). Figure 9 shows Ihe Fourier amplitude spectra. As expected, the accelerograms contain high energy at high fre-quencies. Southwest stations show ihe largest. amplitudes wíth respect to ES statíon located in the north of the city. This is also observed in the Fourier amplitude speclra shown in Fig-

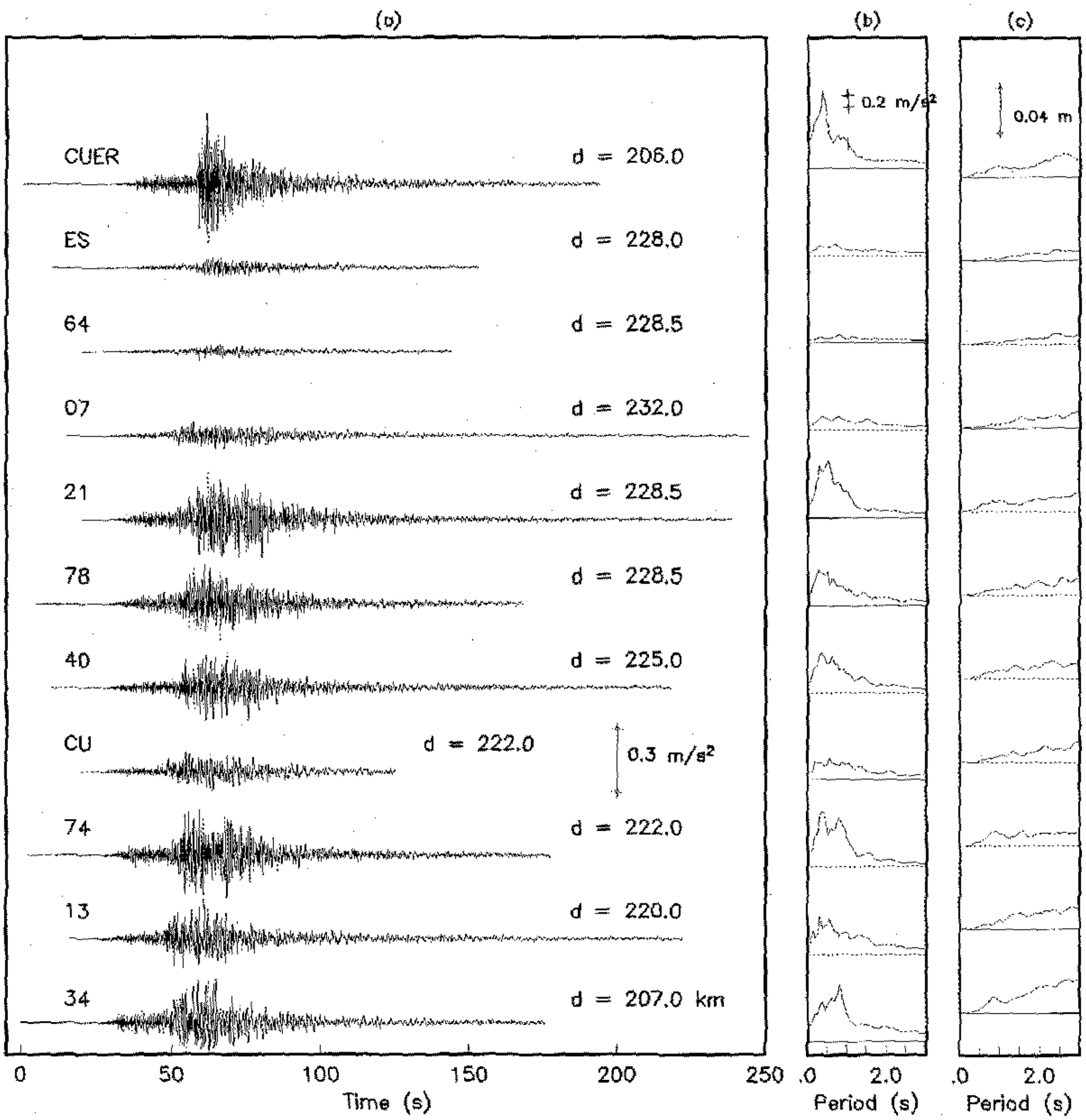
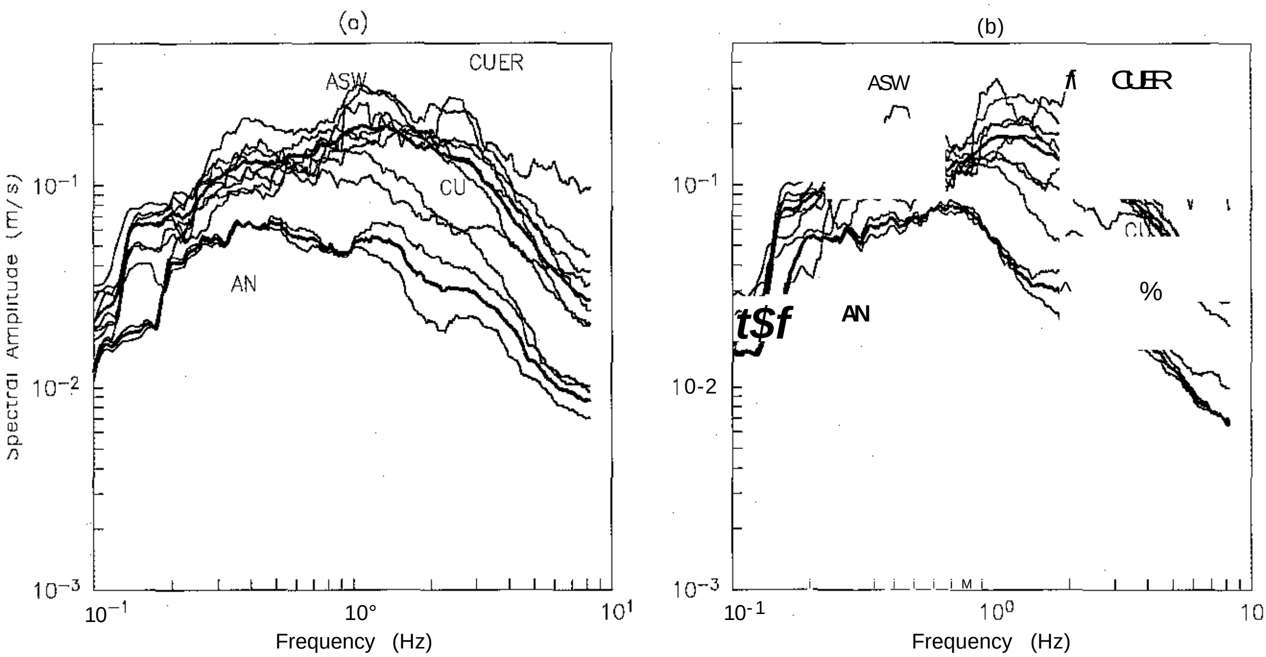


Fig. 3. Strong ground motion at hill-zone sites during the June 15, 1999 earthquake. (a) Accelerograms, (b) response acceleration spectra, (c) response displacement spectra.

ure 9, where the maximum spectral amplitudes are found around 2 Hz. For low frequencies, the amplitude is low. Station CU has smaller amplitudes (in the north-south component) that the other stations located in the southwest; but at low frequencies in the east-west component, the amplitudes with respect to the other stations are large. Figure 9 also shows the average of the Fourier amplitude spectra at the southwest stations. Station ES is located in the north of the city and presents the smallest spectral amplitude. The average spectra for the southwest stations, ASW, and for the north station, AN, are shown.

We compare the effects of path and magnitude of this earthquake (Figure 1) with those of the 1999 event. These earthquakes had similar source mechanics and depth, but they differ in azimuth, magnitude and epicentral distance. The wave path for the 2000 event does not pass through the volcano. We use a theoretical ω^2 scaling (Pacheco and Singh, 1995; Pérez-Rocha, 1998) for the 2000 event to estimate the Fourier amplitude spectra adjusted to M = 7.0. These spectra were also corrected with distance using a regional spectral attenuation relation for south-central Mexico (Ordaz and Singh, 1992).



Pig. 4. Fourícr ampiituclespetiraof'ttie June 15, J999 earihquaice. (a) Nortli-souih and (b) easi-west (ASW, average of southwcsl.siations. AN, avera^e ofnortfi stauons).

Figure !.() shows (he Fourier atnp].ítude spcctra of the grouncl molion al CU stalion of tlie earlhquake of 1999 and the scaled event of 2000. The «caled spectrum has higher spectral amplitudes íbr the wholerangeoffreq uencies. This muy be explaincd by (he fací that the July 21 carthquake *has* a clifferent azimuíh and its palh (o CU does not cross Popocatépetl volcano.

On October 24, 1980 an eartliquake of M = 7.0 occurred in the región of Huajuapan de León near tlie epicenters of events (1) and (7). This event had sitnilai" source mechanics and dcep of the 1999 earthquake; yet, the 1980 eartliquake showed larger spectral amplitudes in the time doniain and in tlie whole range of frequencíes (Singli *et al,* 1999). The differences at site CU for the se two eanhquakes may be due Co the fací. Ihai the palh for the earlhquake of Octobcr 24 *áoes* not pass Ihrough Popocatépetl volcano. Sliapiro *et al.* (2000) proposed that the seismic waves Ihal. traveled toward México City were slrongly attenuated by Popocatépeti volcano.

A similar analysis was carried out for the other earthquakes in Tablü 1. Similar behaviors are observed between southwesl.ern *and* northem siations.

Figure 11 shows the average spectra for the soulhwest stütions oí" all sevcn eatthquakes. The earthquakes of June 15, 1999 and July 21, 2000 show larger amplitudes for frequeiicíes higher than 1 Hz, whíle the other events present greater amplitudes for lowfrequencies. The spectra ofevenls (3), (4) and (5) are similar to (he coastal events; however, due to their larger epicentral distances and íheir smaller magnitude, the amplitudes are shorter.

We conclude ihal. (1) there is a dependence of the seismic motion in the hili zone due lo azimuth, magnitude, epicentral distances and deplh; (2) the aUtions in the southwest of the hill zone have larger spectral amplitudes ihan the statioDS in the noith área of the city, and (3) Ihere are impoitant Jocal differences of the seismic motion for the statíons in tlie southwest región even when they are very elose lo each oiher.

RELATÍVE.\MPLIFiCATION BETWEEN HILL ZONE SÍTES

To compute amplification factors we decided to elimínale the effeets of azimuth, path, magnitude and depth using the speciral ratio technique. Ampiitude rallos were obiained for stations CU and 74 with respect to station Estanzuela (ES), localed on the oldesi rocks (Siiigh *et al,* 1995; Reinoso and Ordaz, 1999). Stations 74 and CU are on lava flows from the Xitle volcano {Delgado *el al,* 1999); they are only 3.5 krn away. The spectral ratios ofthese two stations were compared with the 1-D response using the Haskell method. The

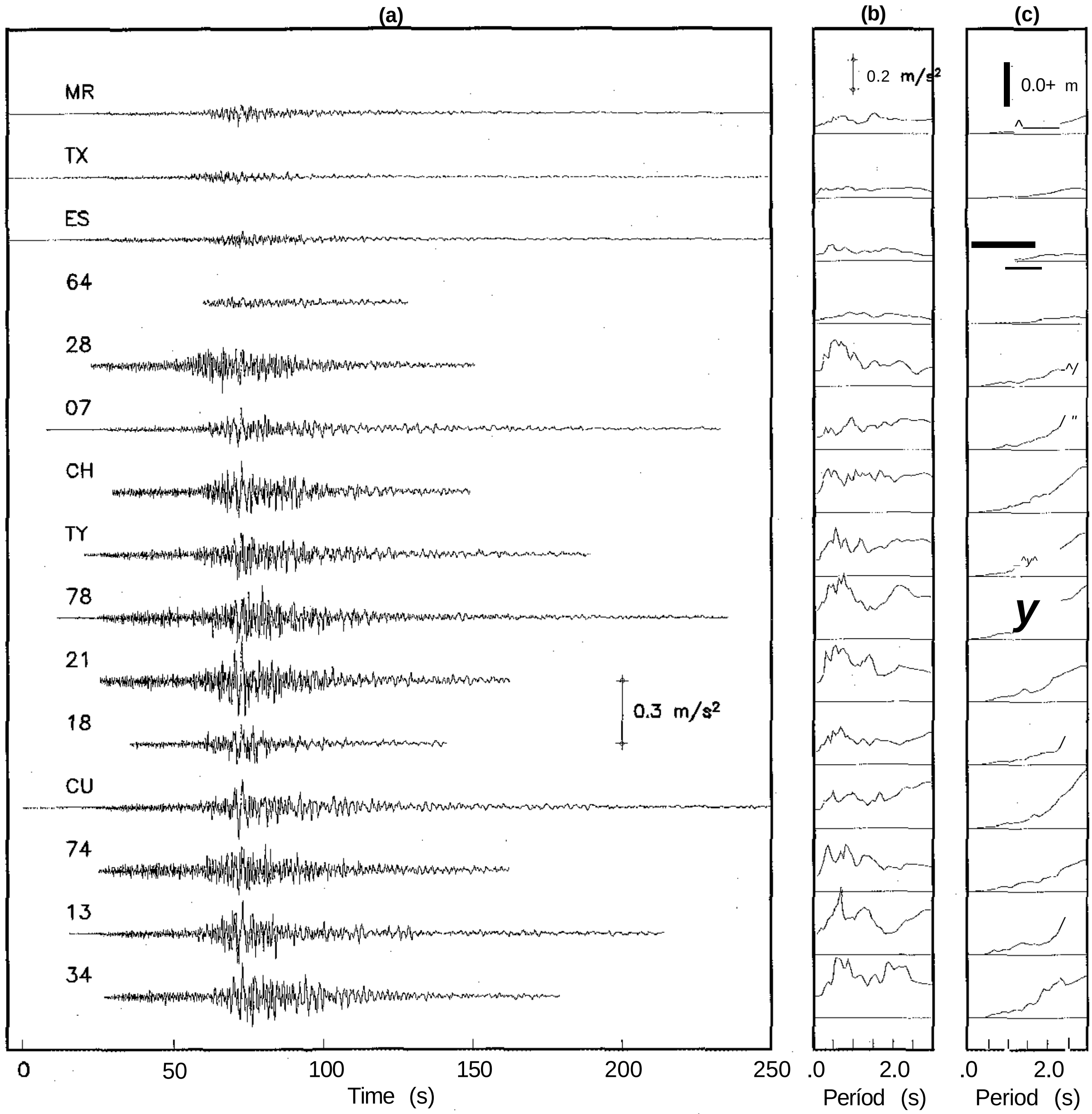


Fig. 5. Strong groimd raotion at hiil-zone siles during the Sepiembe-r 14,1995 earthquake. (a) Accelerometric data, (b) response specíra of acceleraíion, *(c)* response Kpectia. of displacement

geologic model corresponda to site CU (Table 2) after Gutiérrez *et al.* (1994).

Figure 12a shows the Fourier amplitude spectra of thc three stations, Figure i2b shows the transfer functíons CU/ES,74/ES and the 1-D response, and Figure 12c shows the accelerograms. It appears that the i-D model reproduces reasonably well the response of Ihe transfer funetion of site *CV* for Ircquencies hel^veen 1 and 4 Hz. Thc ratio 74/E,S shows impoitant differences wit.li the 1-D transfer funetion; amplitudes are up to 4 times larger. This effecl. may be due ío geológical conclitioiis at site 74. Delgado-Granados and Mooser

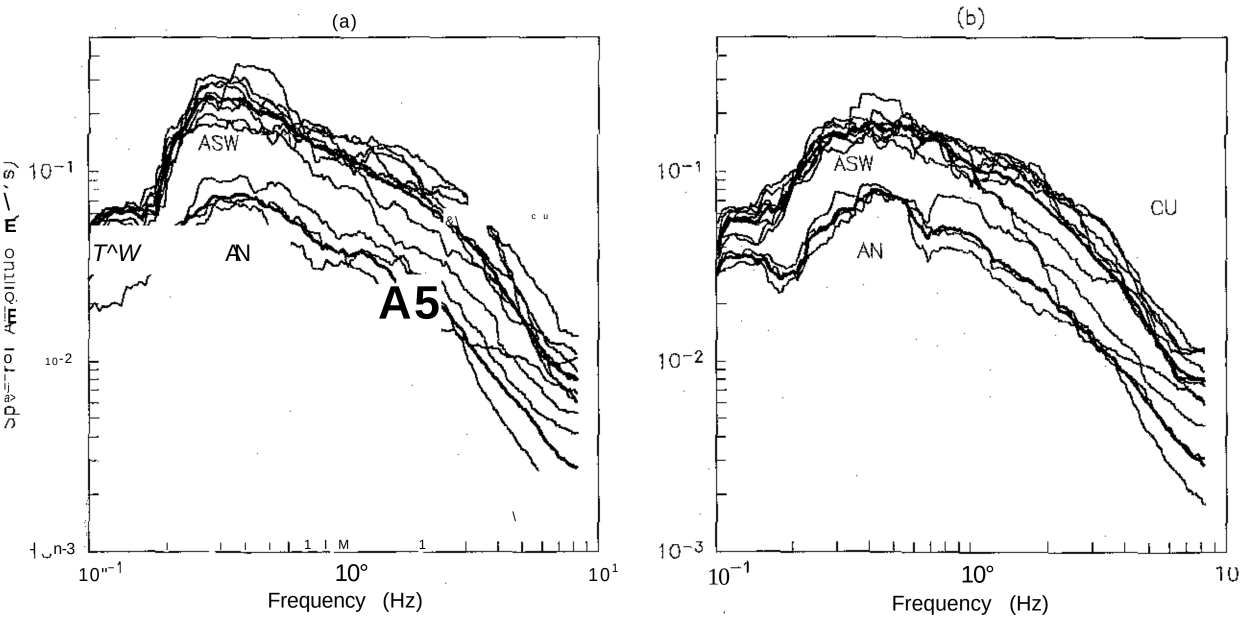


Fig. 6. Fourier amplhude spectra of the September 14.1995 eanhqiiake. (a) Nortii-south. and (bj east-west (ASW, average of soirthwest stations, AN, average of'norfh síatiojis).

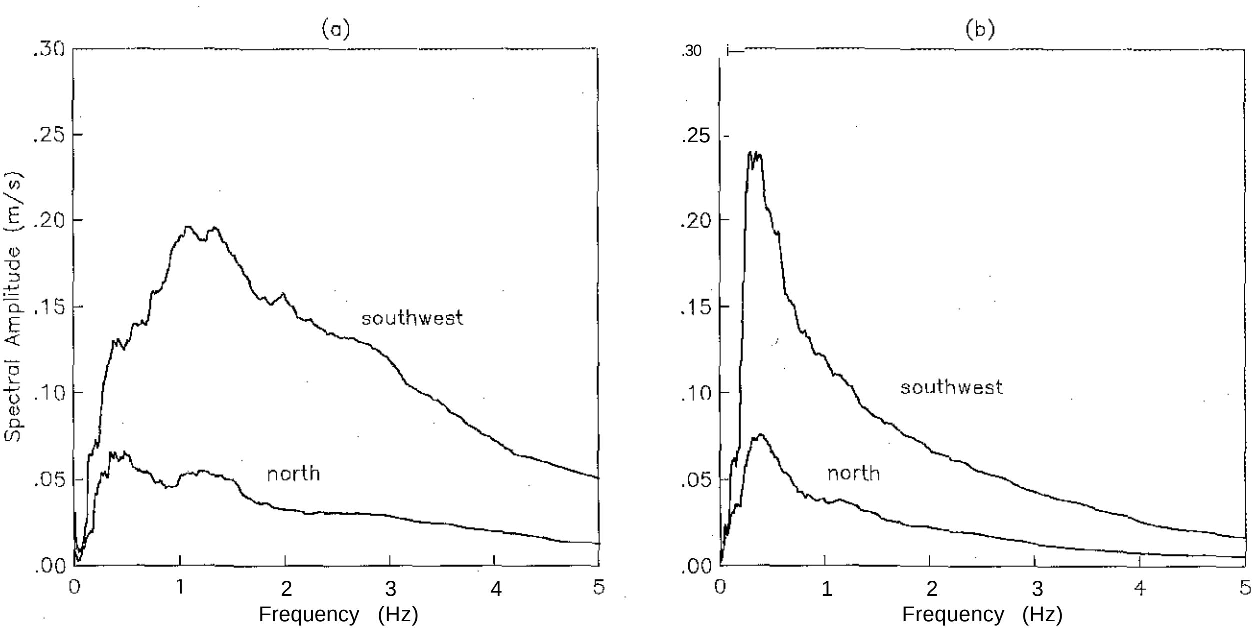


Fig. 7. Average oí Foorier ampltiiide spectra fot the north-souíh component of motion for sonthwest aad tiorth stations. (a) Jtme i5,1999 and (b) September 14,1995.

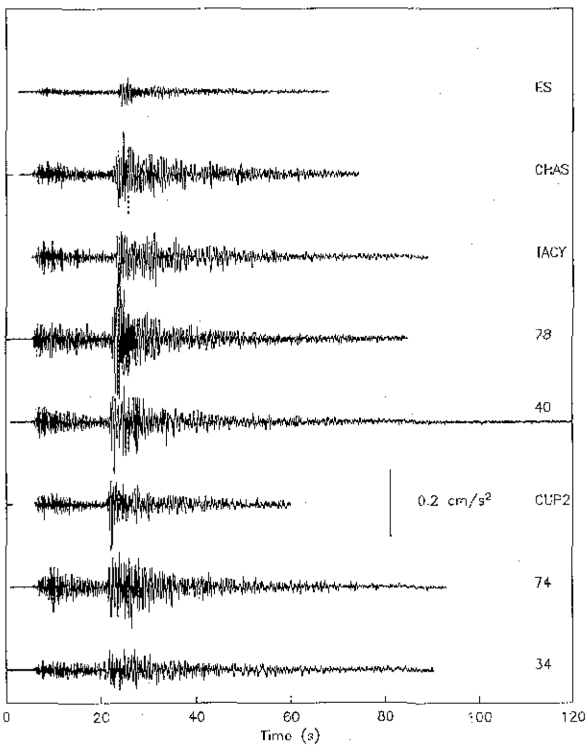


Fig. 8. Strong ground motioii at hiil-zoiie sites daring the July 21, 2000 eartfiquake.

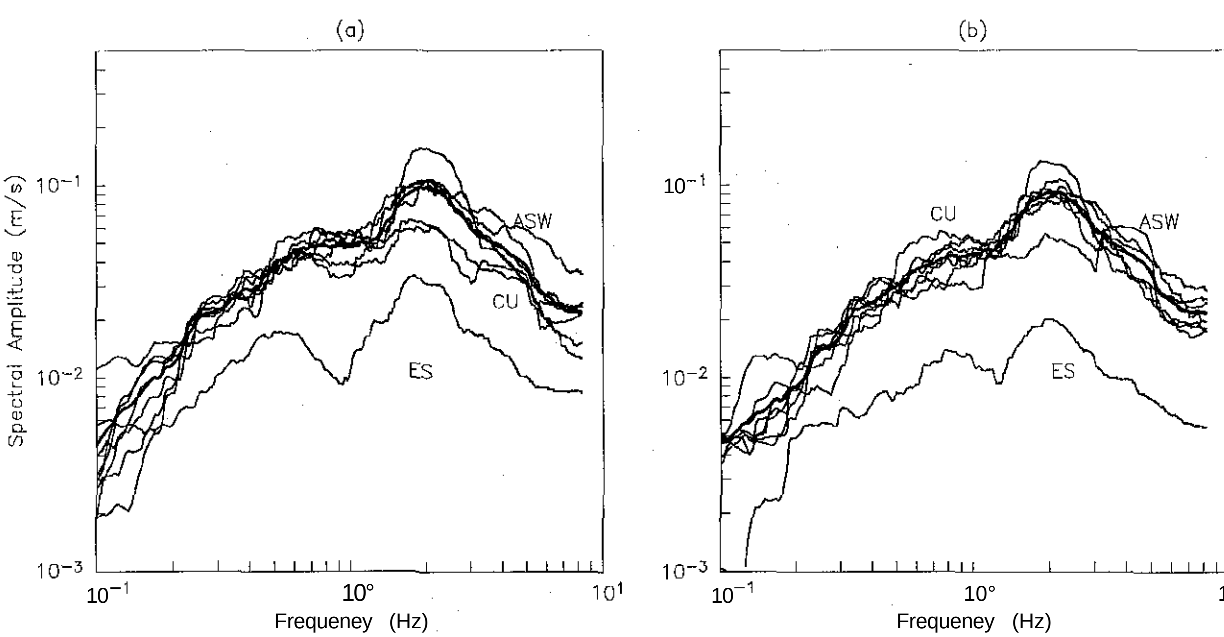
(personal communication) belíeve that the material under ihe lavas beneath station 74 is much KO.11 er. Thi« causes the higher amplíñcations observed in Figure 12 compared lo station CU.

Figure 13, shows tlie average spectral ratios of all souih-wesl statíons with respect lo station ES for two earlhquakes (1999 and 1995), with magnitudes greater than 7.0. Tfaey are plotted against period. From this ligure, the transíer iimc-tions are not as similar as might be expected. For pcriods between 1 and 2.5 s, l.he íargesl amplitudes are for tliecoasíal. earthqoakes, while for periods smaller than 0.6 s (Figure 13b), the inlraplate eavthquake of 1999 has larger ampli-tudes. ThtiR ihe seísraic beliavior ofthe structure of ihe Mexí-can Volcanic Belt, where México City is located, depcnds on the type of earlhquake, and the input motion will shake the structare difte-rently depending *on* the azimuth, distances and magnitude of the earthquake.

CONCLUSIONS

1) In the time domain, higher amplitudes are observed in the stations located in the southwest compared lo ihose in the noith. In the frequeney domain. the average spec-tral amplitudes are up to 4 times larger.

2) Forsubductionearlhquakes, the higher amplitudc is found at frequencies below 1 Hz, while for intraplale eaith-



Fíg. 9. Foiirier ampütude,spectra ofthe .fuly 21, 20(X) e;irthqüake. (a) North-south and(b) east-west. The ES síation is stown atibe bottom of ihe figure.

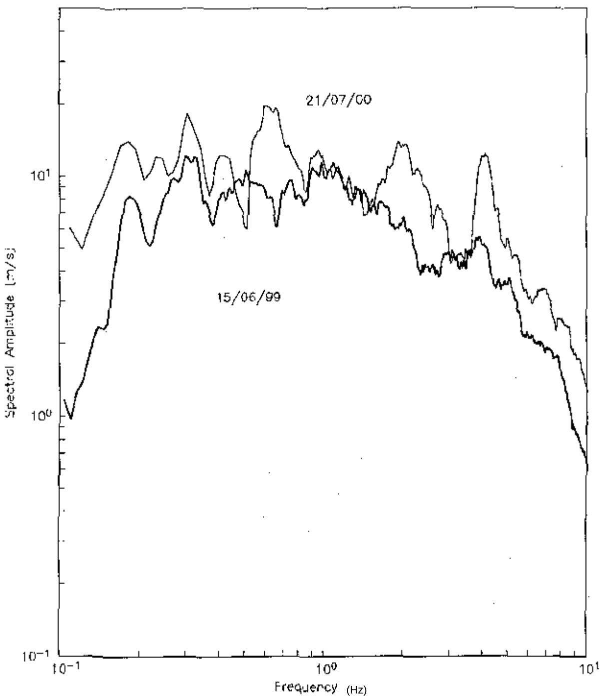


Fig. 10. Fourier ampJitude spectra of Uie June 15,1999 earthquake and scaied (M = 7) Fourier ampiitude spectra of the *íuYy* 21, 2000 eurthqnake.

quakes the higher amplitudes correspemd to írequencies above 1 Hz.

3) Regional site efí'ects are observed at southwest stations at íxequencies over 1 Hz, for stalions 34, 40, and 74 wilh respect lo CU.

4) Important differences ia the transfer functions at tlie southwesl slations are fonnd. This suggests a dependence oí the amplification palterns on Ihe origin and localizatioti of tlie earlliquake.

5) The seismic risk in México City is smalier when the earthquake oceurs easl. of Popocatépet! volcano.

ACKNOWIvEDGMENTS

We are gratefi.il to Cinna Lomnitz and Raúl R. Castro-Escamilla for their comments on Ihe mamiscript. This work. was partially suppoited by DGAPA-UNAM, under granl IN 104998 and by the National Council for Science and Technology (CON ACyT).

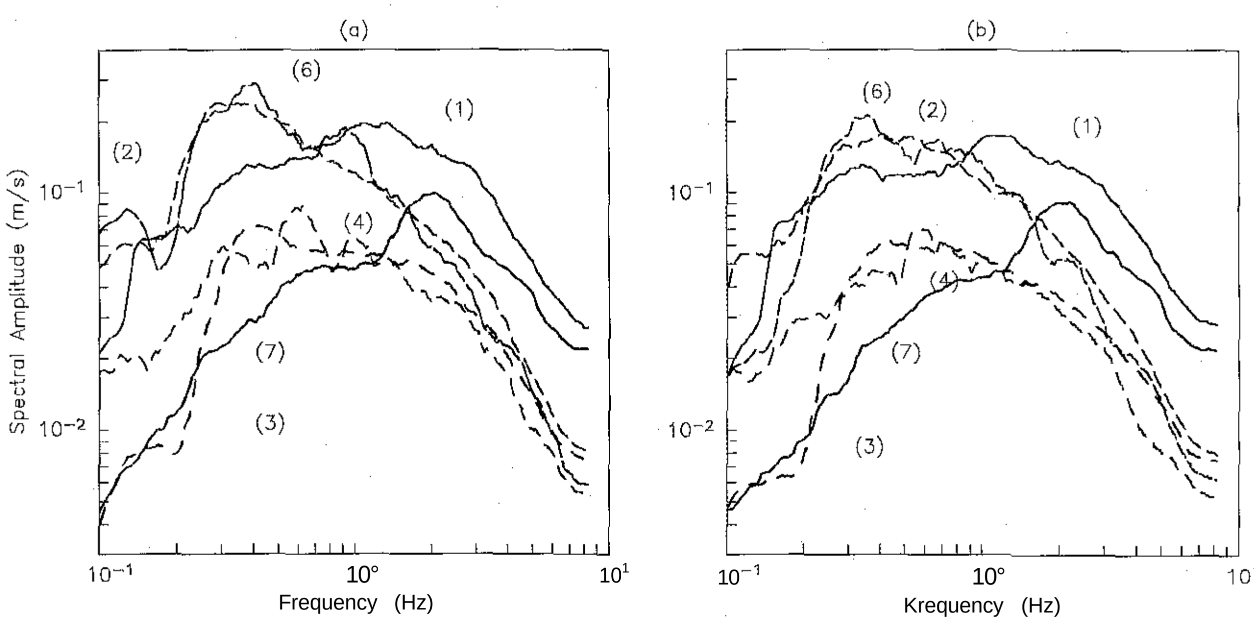


Fig. í í - Average of Fourier ampiitude .spectra of seven eanhquakes for southwest stations. (a) North-south and (h) east-west.

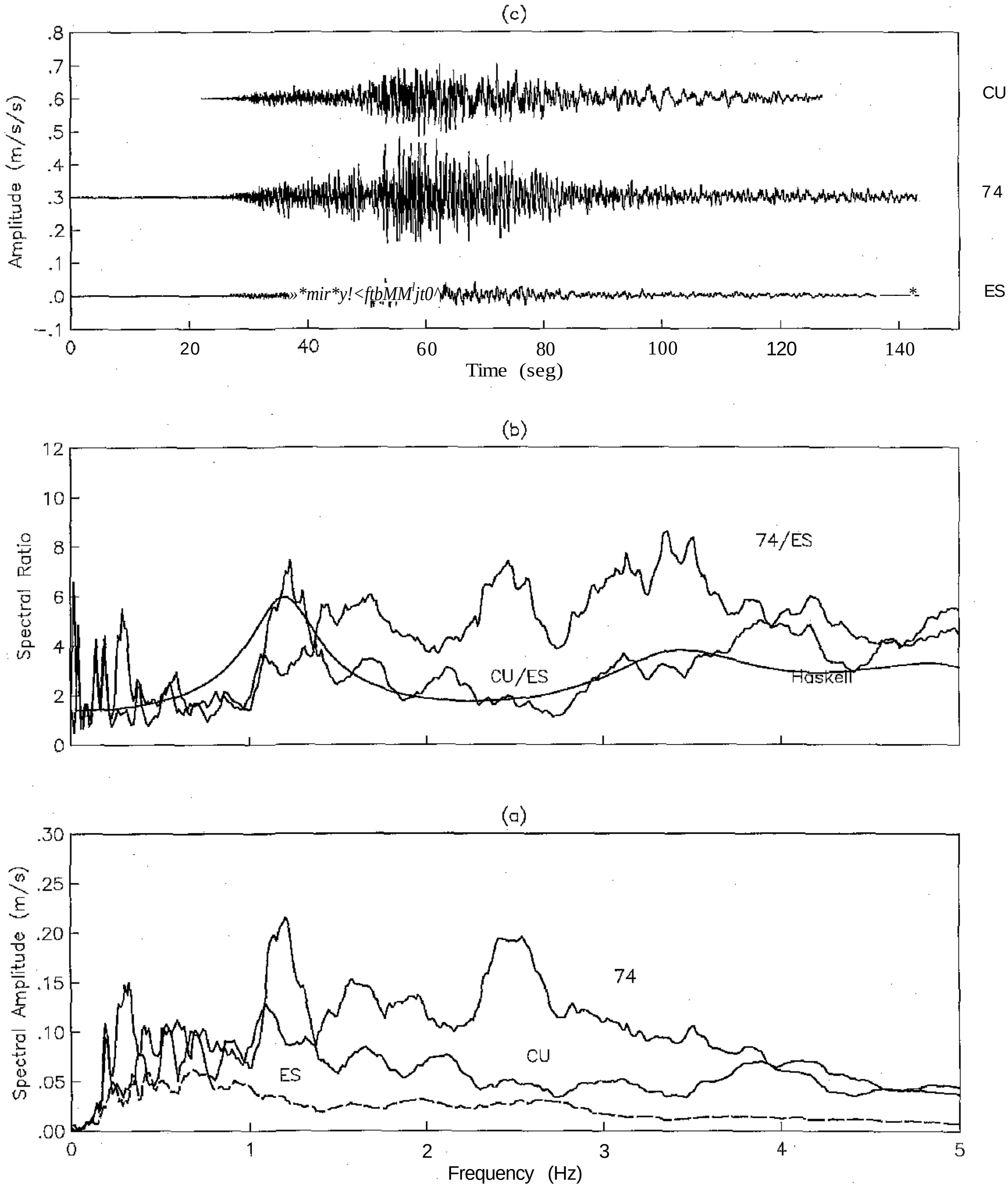


Fig. Í2. Strong groimd moñón ai sialicms74, CU and ES, ofthe June 15,1999 eanhquake easí-west component. (a) Fourier amplitude spectra. (b) spectral ratios 74/Es$_f$ CU/ES, and ÍDresponse, *(c)* accelerometvic data.

Table 2

Sile proíile tised to compute orie-dimensional transfer funclion at sile CU

Thickness fm)	Velocity P waves (m/s)	Veiocity S waves (m/s)	Density T/nr[1]	QP	QS
80.0	1000.0	430.0	1.9	280	140
50.0	2062.0	875.0	2.0	600	300
∞	3000.0	1500.0	2.5	5000	5000

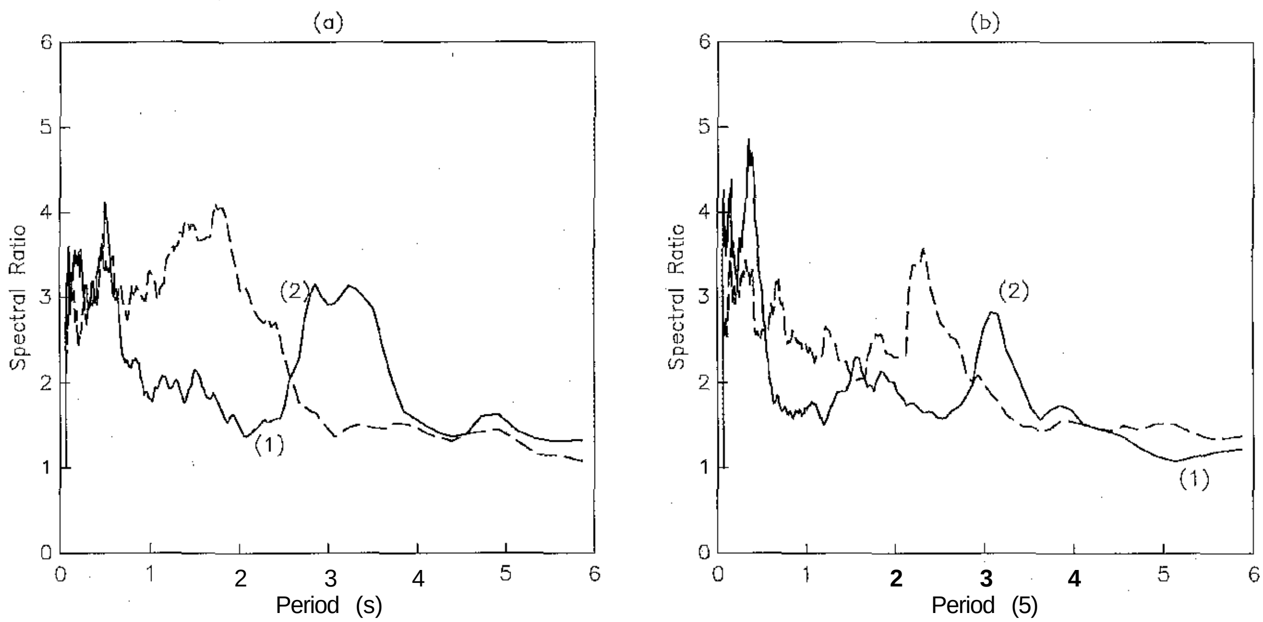


Fig. 13. Spetfral ratios of earthquakes with magnitudes ≥ 6.9 (1999 and 1995 events), (a) Noilh-KOiiíli and (b) easf-west.

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APÉNDICE II

A VIRTUAL REFERENCE SITE FOR THE VALLEY OF MÉXICO

(*Bulletin ofthe Seismological Society of America**en Piensa)

BSSA

Bulletin of the Seismological Society of America

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MS Number 2001257

Lead Author Montalvo-Arr

Title A Virtual Reference Site for the Valley of Mexico

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A VIRTUAL REFERENCE SITE FOR THE VALLEY OF MÉXICO

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Submitted to
Bulletin of the Seismological Society of America
Revised Versión
January 2002

16

ABSTRACT

We propose the use of the average spectmm of northern strong motion stations located at the hill zone *(e. g.* MD, TX, ES, 64) as the reference for México City's ground motion. This virtual site proposal is based upon the analysis of recent data from the México City Acelerometric Network. The northern stations show amplitudes, both in time and frequency, that consistently are smaller than hill zone stations located south and west of the city *(e. g.* CU). It is well known that CU, the histórica! reference site in México City, and other sites at the southwest present ampliñcations, while the northern ones appear to be free of such effects. The spectral ratio of the averages of the stations located in the south and west with respect to the northern stations show a relatively constant amplificaron of up to 3 times in the 0.7 to 10.0 Hz frequency band. This amplification is a very unusual feature that should be explained. The geological conditions at the hill zone show that older, Miocene age deposits are located north of the City. Considering that northern sites represent the basement, we assume that the configuration along the hill zone in the N-S direction can be approximated by a simple dipping homogeneous layer. We computed the antiplane seismic response for this model, and averaged and compared it with the spectral ratio obtained from strong ground motion data. The agreement is good and suggests how a smooth, large scale feature could amplify seismic ground motion in a broad frequency band.

INTRODUCTION

It has long been established that soil types respond differently when subjected to ground motion from earthquakes. Usually, the younger, softer soils amplify ground motion relative to older, more competent soils or bedrock (Aid, 1988). One of the goals of engineering seismology is to obtain reliable estimates of ground motion amplification throughout metropolitan regions in earthquake-prone áreas. The measurements of ground motion are useful to distinguish regions where the seismic hazard is large due to local amplification from the surface geology and subsurface structure. To define seismic ground motion amplification it is mandatory to establish with respect to what motion the amplification takes place. Such reference motion is thus characteristic of a reference site. In fact, an ideal reference site allows to get trustworthy estimates of the input motion that affects the soil at nearby locations.

Regarding México City, Singh *et al* (1988) computed spectral ratios for some events using the station CU as reference motion. The use of CU as reference was obvious: since 1964 more than 20 modérate and strong earthquakes were recorded there. The ratios computed at each site by Singh *et al.* showed roughly the same behavior during different earthquakes. However, such ratios exhibited significant variations if used to predict site amplification effects. Reinoso (1996) and Reinoso and Ordaz (1999) proved that these differences were smaller, and sometimes much smaller, when the average Fourier amplitude spectmm of south and west hill-zone stations is used as the reference motion.

Singh *et al.* (1988, 1995) and Ordaz and Singh (1992) pointed out that the hi1l zone in the Valley of México suffers important amplification with respect to that predicted by the attenuation laws. This amplification could be as large as 10 for frequencies between 0.2 and 0.7 Hz. On the other hand, Pacheco and Singh (1995) studied several intermediate-depth normal faulting events recorded in México City. They found that *S* waves are amplified 2.5 times between 0.2 and 3.0 Hz. To correctly estímate ground motion at various sites using CU as reference, the station site effects has to be removed.

Recent studies of México City Valley report significant differences between southwestern and northern hill zone stations (Reinoso and Ordaz, 1999; Montalvo-Arrieta *et al.*, 2001). Putting aside the spectacular amplifications of ground motion in the lake zone, the differences discovered in the hill zone are striking. Ordaz and Singh (1992) found that strong ground motion at station MD, located in the northern part of the hill zone, was smaller when compared to the southern and western hill zone motion. They suggested that the cause of the observed increase in the amplitudes of seismic waves at the latter hill sites is related to a relatively shallow (<1 km) clay deposit which was emplaced in a basin that existed in late Oligocene to Pliocene times.

Average spectral ratios of southern and western sites relative to northern stations displays an approximately constant amplification factor of about three in the frequency band between 0.7 and 10.0 Hz. This constant average amplification in such a wide frequency range needs to be explained. This behavior is unusual for the majority of models employed in seismology. A possible exception is the wedge. In fact, considering that (1) the northern sites reasonably represent the basement rock and (2) the Quaternary deposits pinch out smoothly towards the north, we assumed a simple homogeneous dipping íayer and computed the average amplification for incident *SH* waves.

In this work we have found new evidence that the northern sites in the hill zone appear to be free of site effects. We propose to use northern strong ground motion as a virtual reference site free of local an regional amplification effects, since a regional amplification factor of 3 times for CU and other sites located at the southern hill zone have been observed. This virtual site proposal is based upon the analysis of recent data from the México City Accelerometric Network.

GEOLOGICAL SETT1NG

The basin of México is located in the central part of the Mexican Volcanic Belt (MVB). Volcanic rocks form the geology of the basin. It is filled by lacustrine sediments, volcanic tuff, sands and gravéis. The Valley of México has been divided into three regions based on the geotechnical characteristics of its shallow layers: (1) the hill zone, formed by volcanic tuffs and lava flows, (2) the lake bed zone, formed by clays with thickness varying from 10 to 130 m; and (3) the transition zone, composed by alluvial sandy and silty layers, with scattered clay layers (Marsal and Mazari, 1959). Figure 1 shows these geotechnical zones together with the accelerometric stations and some reference sites.

The southern part of the hill zone is formed by recent Quaternary lava flows from the Sierra de Chichinautzin (sites CU, 13, 34, 40 and 74, Figure 1). The western portion is dominated by tuffs (sites CH, TY, 07 and 74) while the northern part is composed of Miocene lava (sites MD, TX, ES and 64) that overlies voícanic rocks from the Oligocene age. This Miocene lava underlies the deposits from Sierra de Chichinautzin and the tuffs of the western portion. Figure *2* shows a geological cross section across the Valley in the north-south direction (Mooser *et al.* 1996), from Sierra de Guadalupe to Sierra de Chichinautzin (Figure 1). In this cross section, it is clear that the horizon of Miocene lava is dipping to the south.

DATA AND ANALYSIS

After the extensive damage caused by the Ms = 8.1, 1985 Michoacán earthquake in México City, the accelerometer array has grown considerably. It now consists of more than 90 free-field digital accelerometric stations, 18 of them in the hill zone (stations 07, 13, 18, 21, 28, 34, 40, 50, 64, 74, 78, ES, CH, CU, CN, TX, MD and TY). The stations 18 and 28 at the Cerro de la Estrella and Peñón outcrop are not included in this work since they probably present signiñcant topographic effects (Reinoso and Ordaz, 1999).

Data from two recent earthquakes with M $\geq$ 7 are presented. Records correspond to hill zone stations and emphasis is laid upon the northern ones.

The earthquake of September 14,1995, Ms 7.3

The epicenter of this subduction zone earthquake is located 300 km south of México City. The hypocenter was located at 17 km depth, and the focal mechanism was found to be shallow angle thrust. Reinoso and Ordaz (1999) pointed out that the northern stations MD, TX, ES and 64 show the smallest amplitudes, botb in frequency and time domains. This may be due to a strong change in the geological conditions between the southwestern and northern parís of the hill zone. The older rocks are located to the north of the city. Attenuation is not an important factor since the distance between the southern and western stations and the northern stations is small compared to the epicentral distance. Thus, the distance from CU to ES is approximately 20 km, less than 10 per cent the epicentral distance.

Figure 3 shows the accelerograms corresponding to the north-south component of motion. The accelerograms in figure 3 differ in amplitude and duration of motion (Reinoso and Ordaz, 1999). Figure 4 displays the smoothed Fourier amplitude spectra of both horizontal componente of motion (Reinoso and Ordaz, 1999). The largest spectral amplitude is presented for the stations of the southwest. Also shown in figure 4 are the average spectra of the southwestern average station, ASW, and northern average station AN (thick lines).

The earthquake of June 15,1999, Mw 7.0

The epicenter of this normal fault earthquake (60 km deep) was located southwest of Tehuacán, Puebla. It caused severe damage in the states of Puebla and Morelos (Singh *et* *al.,* 1999). In México City, about 200 km to the west of the epicenter, damage was only associated with non-structural elements (Singh *et al.,* 1999).

Figure 5 shows the north-south accelerograms from this earthquake recorded at hill zone sites and at station CUER in Cuernavaca, located o ver firm soil but outside the vailey (39 km south of station CU). The accelerograms show unusually large amplitude and high frequency contení. Singh *et al.* (1999) explained these observations due to source directivity toward the northwest and related these larger accelerations to greater damage observed toward the northwest of the epicenter. Figure 6 shows the Fourier amplitude spectra of both horizontal components of motion. The largest spectral amplitude is presented for the stations CUER, 34 and 74. Also shown in figure 6 are the average spectra of the southern and westera stations, ASW, and northern stations AN.

Spectral ratio between the southwestern and northern stations

Figure 7 shows the average north-south component Fourier spectra for both events for hill zone stations in the south, west and in the north. Notice the difference between the two types of earthquakes: for the 1999 intraplate earthquake the máximum spectral amplitudes occur at frequencies higher than 1 Hz, while in the 1995 subduction earthquake the máximum amplitude occurs at frequencies below 1 Hz. Similar to figures 4 and 6, figure 7 shows that for each earthquake spectral shapes are very similar, but strong variations exist in the amplitudes of both groups of stations. The difference is up to 3 times for both earthquakes. Although not shown previously, a similar response was observed for the east-west component for both earthquakes.

To compute ampliñcation factors for the southwestern stations with respect to the northern stations we eliminated the effects of azimuth, path, magnitude and depth by using the spectral ratio technique. Figure 9 shows these ratios with a practically constant ampliñcation of 3 for the 09/14/1995 earthquake and 3.5 for the 06/15/1999 earthquake. This difference between amplitudes of the two spectral ratios suggests that the incident wavefield character plays an important role in the excitation of the hill zone (Pacheco and Singh, 1995; Montalvo-Arrieta, *et al.* 2001).

A **DIPPING LAYER**

The amplification observed in figures 7 and 9 is almost constant for a wide range of frequencies. The puzzle is then to find a model that can reproduce and explain these observations. We assume that the configuration along the hill zone in the N-S direction (figure 1) can be a simple wedge as show in figure 2, where the Miocene lava is dipping south underneath the tuffs and lava flows. This may be a rough approximation but the response may give us some clues to interpret the observations.

The antiplane seismic response of a dipping layer when the moving basement is rigid has an analytical solution under certain geometry conditions (Sánchez-Sesma and Velázquez 1987). Indeed, for dip angles from the horizontal (free surface) of the form *is/2N,* where *N* ~ 1, 3, 5,..., the exact surface displacement, v, for a harmonio antiplane motion of the base (given by *exp(icot)*, where $i = \sqrt{-1}$, *co* is the circular frequency, and / is the time) can be written as

$$v=\sum_{J=0}^{M} \varepsilon_{M-J}(-1)^{j} \exp(-1 \text{ tacos } 0_j) \qquad (1)$$

where $M = (N-I)/2$, ε_n is the Neumann factor $\{e_n = 1$, if n = 0; $s_n = 2$ if n $\geq 1)$, $k = ay/fí$, is the shear wave number, *fi* is the shear wave velocity, *x* is the horizontal coordínate (figure 10), and *6j* is given by $0_j=(jV-2\_/-1)$^/ $_{Nr}$. The first ray, $j = 0$, represents the plañe wave emitted from the moving base. The image rays and the next reflections just come from boundary conditions. To widen the range of application of equation (1) we consider an elastic boundary, including reflection coefficients $\prod_{l=0}^{J}$ anc* anelastic attenuation (where *Q*, the quality factor, is included). The expression (1) becomes:

$$v=\sum_{7=0}^{M} \varepsilon_{M-j}\left(\prod_{l=0}^{j} R_l\right)\exp(-¿tacos 0_j) \qquad (2)$$

where *Ri* = 1 if / = 0, and

$$R_l = \frac{1-\dfrac{\rho_2\beta_2}{\rho_1\beta_1}\dfrac{\sqrt{1-\left(\beta_2/\beta_1\right)^2\sin^2\gamma_l}}{\cos\gamma_1}}{1+\dfrac{\rho_2\beta_2}{\rho_1\beta_1}\dfrac{\sqrt{1-\left(\beta_2/\beta_1\right)^2\sin^2\gamma_l}}{\cos\gamma_1}}, \gamma_l = \frac{\pi}{l}-\theta_j\frac{\pi}{2N}, \text{ if } l \geq 1$$

where /?, *p* are the shear wave velocity and density. Index 1 indicates the half-space, while 2 is for the material inside the wedge.

We calculated the transfer functions of stations located along the free surface from 3 to 16 km, They are equally spaced with a distance of 1 km with *pi* = 1500 m/s, *$2* = 700 m/s, *pi* = 2.0 g/cm^3, *p2* = 1.5 g/cm^3 and g = 250. These valúes were taken from Singh *et al* (1995) and correspond to the structure below western and northern stations. Figure 8 shows the transfer function forN = 51, 31 and 19 that correspond to dip angles of 2, 3 and 5 degrees, respectively. For JV = 19 the seismic response is on average 0.5 times smaller than the response for *N* = 31, and for A^ = 51 the seismic response at low frequencies is very different to the observed ratio since máximum thickness is approximately 500 m. We propose that *N* = 31 is adecúate because the máximum thickness for the dip layer with this angle correspond to a valué of approximately 900 m, which is similar to the one mentioned by Ordaz and Singh (1992). These authors found that station MD, located at the northern part of the hill zone, has no amplification compared to the motion at the southern and western stations. They suggested that the cause of the observed increase in the amplitudes of seismic waves at the other hill sites is related to a relatively shallow (<1 km) clay deposit.

Figure 9 shows the average transfer functions for all stations. This transfer function is almost constant with an amplification of 3. Figure 9 shows that the amplitude of surface displacements of the dipping layer (thick line) is very similar to the spectral ratio of 1995 subduction earthquake. These results suggest that the seismic response of the average of southwestern stations for the 1995 earthquake can be explained by a simple model of a dipping layer, where the total wavefield is represented by *SH* waves. Campillo *et al.* (1988) and Chávez-García *et al.* (1995) remarked that the incident wavefield in México City during the great 1985 Michoacán earthquake (Ms = 8.1) was dominated by guided propagation (higher modes), and the characteristics of *SH* waves (Love waves) and *SV* waves (Rayleigh waves) were very similar. Apparently, our *SH* model allows us to explain the seismic response for the September 1995 subduction earthquake. However, for the normal fault earthquake the agreement is not so good. This may be due to:

- The strong ground motion recorded for the June 1999 earthquake, in the hill zone sites, shows a high frequency content and large amplitude due to a probable enhancement of amplitudes of *P* and *SV* waves, which can not be reproduced using our simple *SH* model.
- The geological structure of the valley of México is very complex both in macro and local scales. These local heterogeneities may be yielding different amplification patterns when comparing different backazimuths of earthquakes, as has been shown in this work for waves arriving from the east (south-east, as the 1999 earthquake) comparing to those arriving from the south (the 1995 earthquake). For instance, volcanic structures like Cerro de la Estrella and Sierra de Santa Catarina (figure 1) may be scattering seismic waves coming from the south-east, an effect that only is visible at northern stations. On the other hand, in north-south direction the geologic structure of the valley is more homogeneous, so the differences between the north and south are not as large (figure 2).

CONCLUSIONS

Using data from two earthquakes, we have found that spectral ratios of the averages of the stations located in the southern and western part of the hill zone in México City show a relatively constant amplification larger man 3 in the 0.7 to 10.0 Hz frequency band, compared to stations in the northern part.

Considering that northern hill zone sites represent the basement, we assume that the configuraron along the hill zone in the N-S direction can be approximated by a simple dipping homogeneous layer. We computed the antiplane seismic response for this model, averaged and compared it with the spectral ratio obtained from strong ground motion data. The agreement is good and suggests that a smooth, large scale feature could amplify seismic ground motion in a broad frequency band for subduction zone earthquakes.

Finally, we have found new evidence that the northern sites in the hill zone appear to be free of site effects. We propose to use northern strong ground motion as a virtual reference site free of local an regional amplification effects, since a regional amplification factor of 3 times (in the frequency range of 0.7 to 10 Hz) for CU and other sites located at southem of the hill zone have been observed.

ACKNOWLEDGMENTS

We are grateful to Martin C. Chapman, Rob Graves and an anonymous reviewer for their critical comments on the manuscript This work was partially supported by DGAPA-UNAM, under grant IN104998 and by the National Council for Science and Technology (CONACyT).

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FIGURE LEGENDS

Figure 1. México City: accelerometric stations (dots), geotechnical zones (gray scale) and some streets and city sites. The accelerometric network consists of more than 90 free-fíeld digital accelerometers, 18 of them located at the hill zone (07, 13, 18, 21, 28, 34, 40, 50, 64, 74, 78, CH, CN, CU, ES, MD, TX and TY).

Figure 2. Geological cross section across the Valley from Sierra de Guadalupe (North) to Sierra de Chichinautzin (South). This cross section is based on field geology, geological infonnation and seismic reflection and refraction data (Mooser *et al.,* 1996).

Figure 3. Strong ground motion (north-south, component) at hill zone sites during the September 14, 1995 earthquake (Reinoso and Ordaz, 1999).

Figure 4. Fourier amplitude spectra of the September 14, 1995 earthquake: (a) north-south component and (b) east-west component (ASW: average of southwest stations; AN: average of north stations; CU, station at *Ciudad Universitaria* (Reinoso and Ordaz, 1999).

Figure 5. Strong ground motion (north-south, component) at hill zone sites during the June 15, 1999 earthquake.

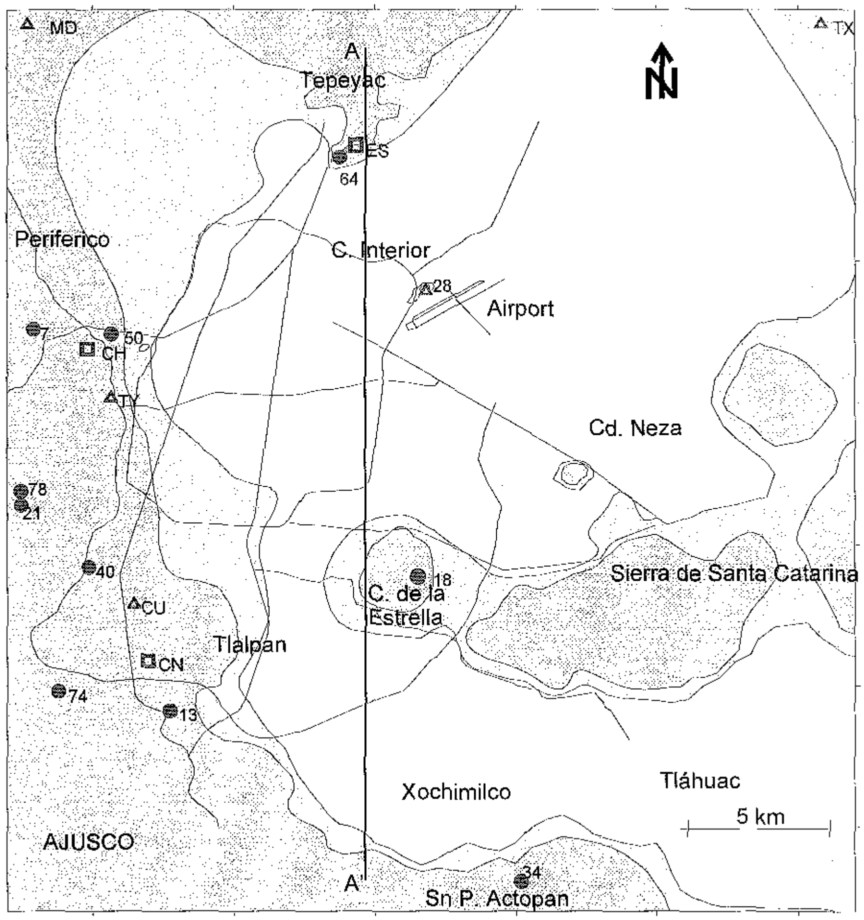
Figure 6. Fourier amplitude spectra of the June 15, 1999 earthquake: (a) north-south and (b) east-west components of motion (ASW: average of southwest stations; AN: average of north stations; CU, station at *Ciudad Universitaria;* CUER in Cuernavaca, located over firm soil but outside the valley).

Figure 7. Average Fourier spectra for the north-south component of motion for southwest and north stations: (a) June 15, 1999 earthquake and (b) September 14, 1995 earthquake.

Figure 8. Sensitivity of the seismic response of a dipping layer model for incidence of SH waves, for N = 31 and N = 19.

Figure 9. Spectral ratio for the September 14, 1995 and June 15, 1999 earthquakes, and seismic response of the dipping layer (thick line).

Figure 10. Dipping layer overlaying a moving rigid base. The complete family of rays is illustrated for the case of N = 5.



Hill zone Transition Zone Lake Zone

Figure 1

26

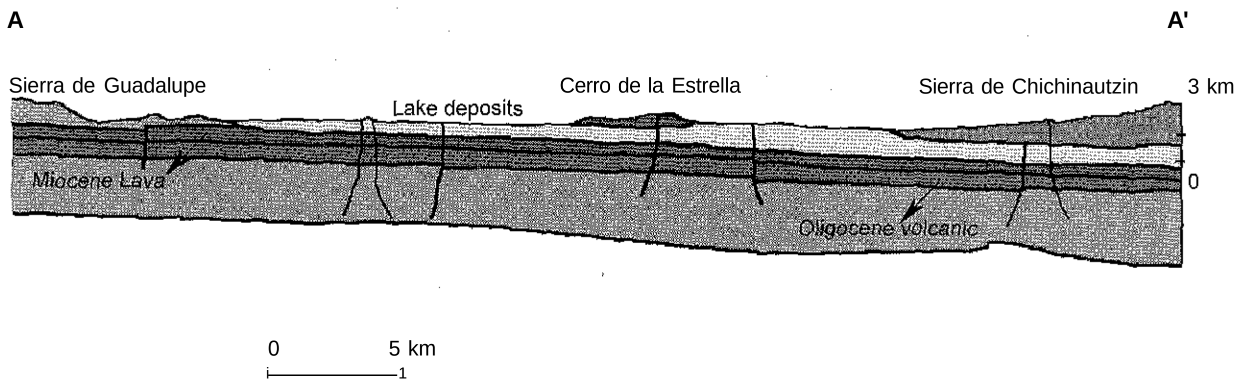

A
A'
Sierra de Guadalupe
Cerro de la Estrella
Sierra de Chichinautzin
3 km
Lake deposits
Miocene Lava
0
Oligocene volcanic
0
5 km

Figure 2

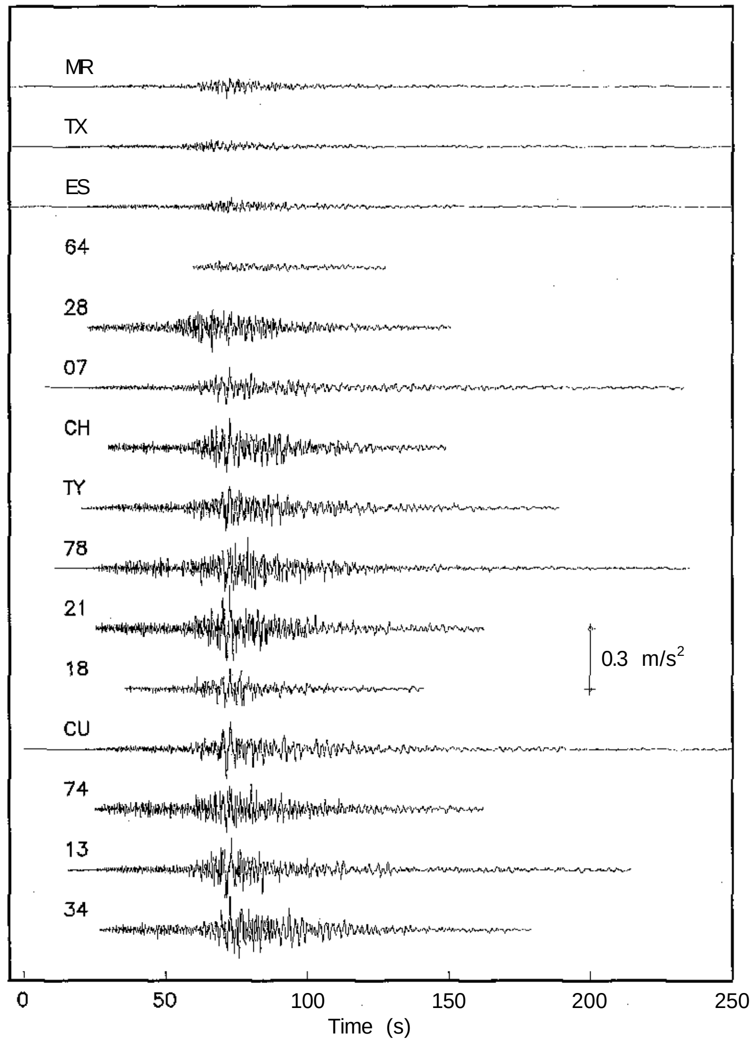

MR
TX
ES
64
28
07
CH
TY
78
21
18
CU
74
13
34
0.3 m/s^2
0
50
100
150
200
250
Time (s)

Figure 3

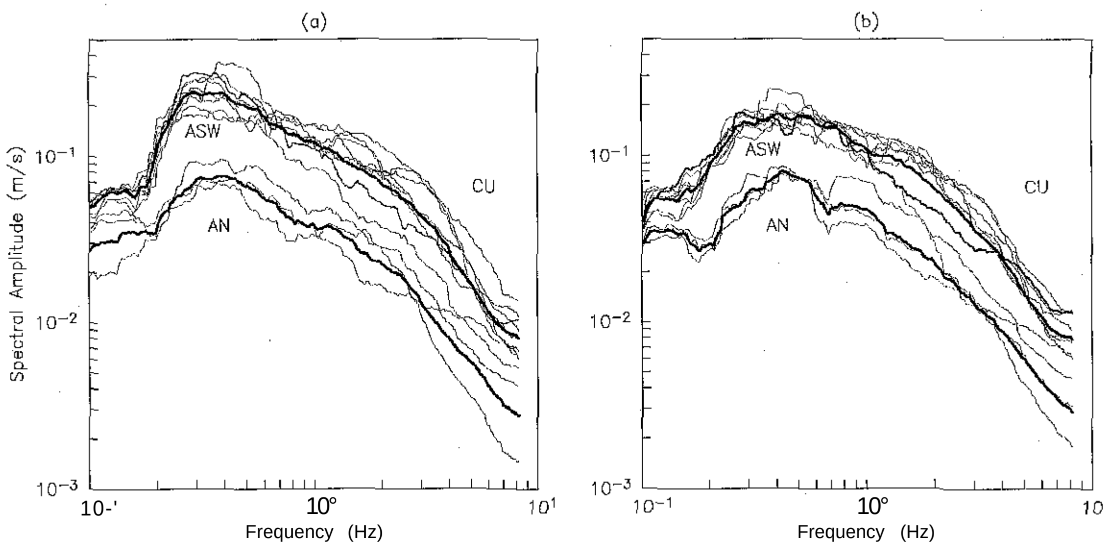

(a)
(b)
Spectral Amplitude (m/s)
10⁻¹
10⁻²
10⁻³
ASW
AN
CU
Frequency (Hz)
10⁻¹
10⁰
10¹

Figure 4

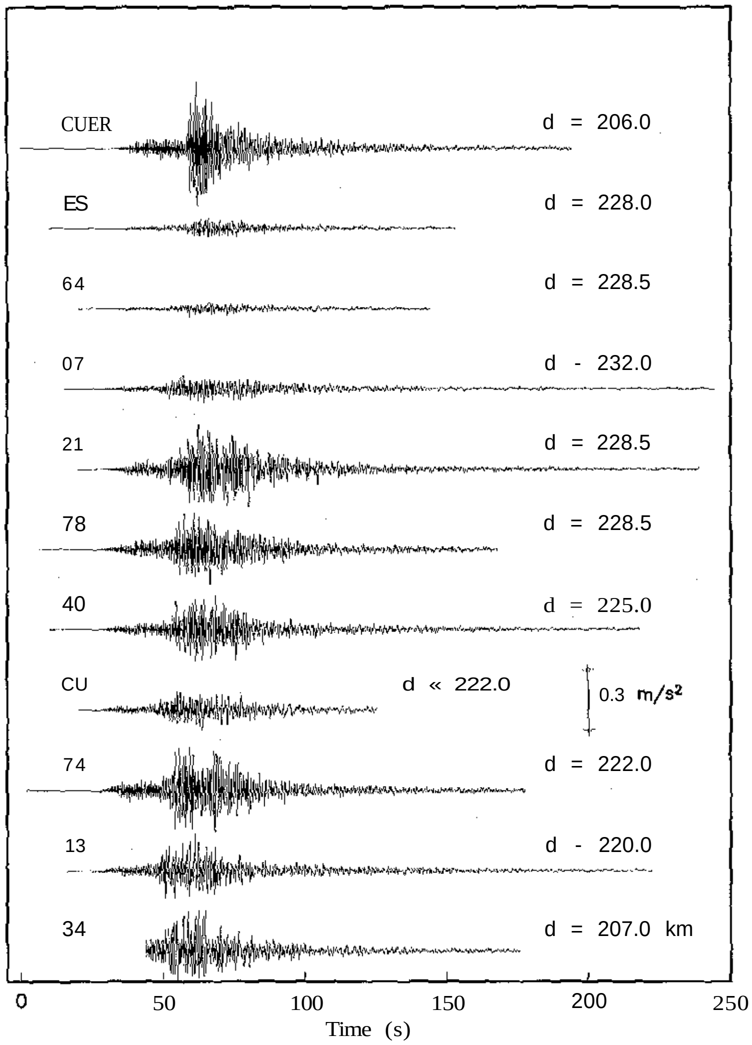

CUER
d = 206.0
ES
d = 228.0
64
d = 228.5
07
d - 232.0
21
d = 228.5
78
d = 228.5
40
d = 225.0
CU
d « 222.0
0.3 m/s²
74
d = 222.0
13
d - 220.0
34
d = 207.0 km
0
50
100
150
200
250
Time (s)

Figure 5

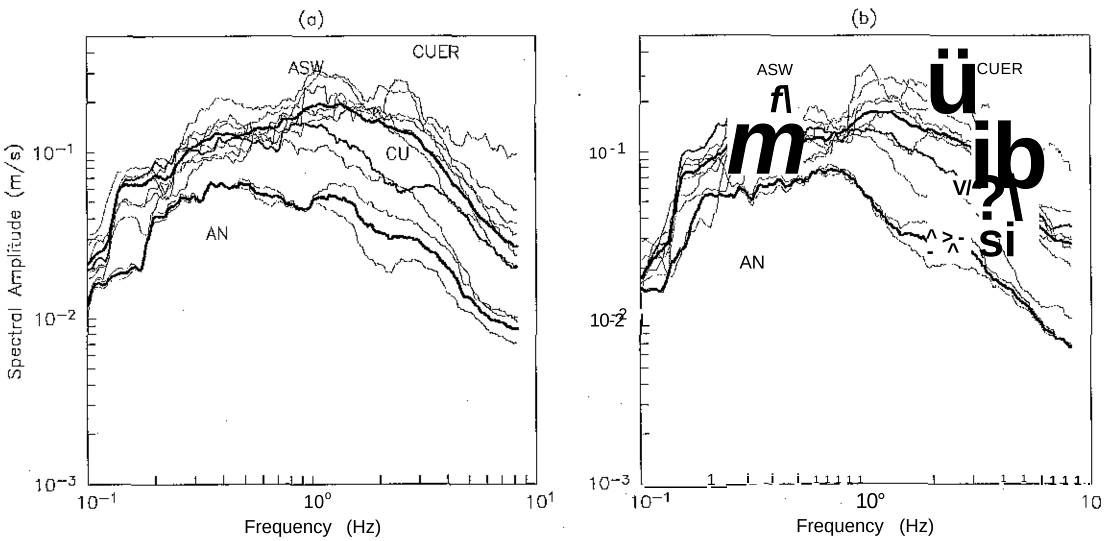

(a)
(b)
ASW
CUER
CU
AN
Spectral Amplitude (m/s)
Frequency (Hz)
10^{-1}
10^{-2}
10^{-3}
10^{0}
10^{1}

Figure 6

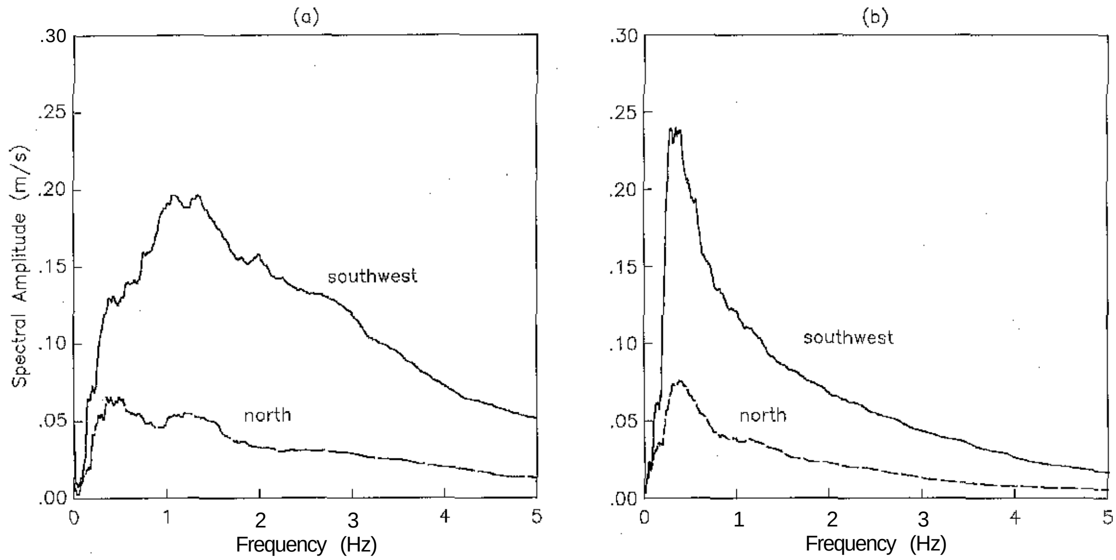

(a)
(b)
.30
.25
.20
.15
.10
.05
.00
Spectral Amplitude (m/s)
southwest
north
0
1
2
3
4
5
Frequency (Hz)

Figure 7

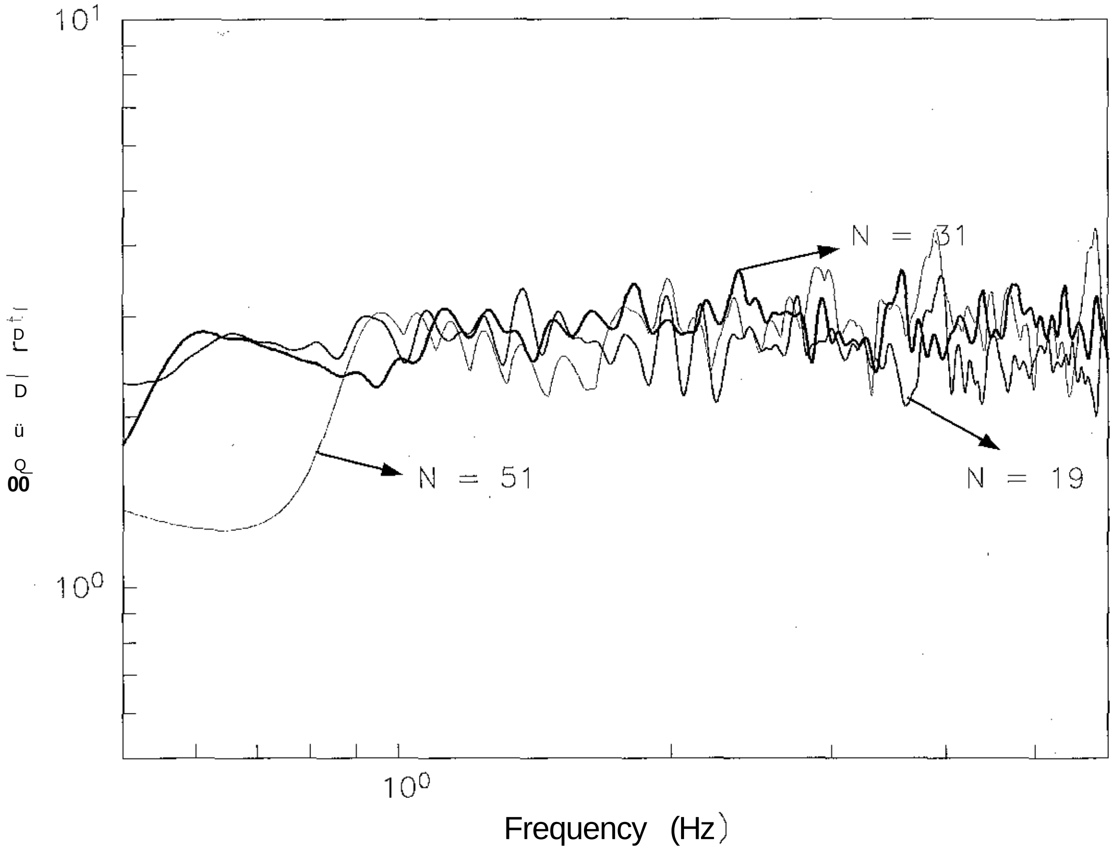
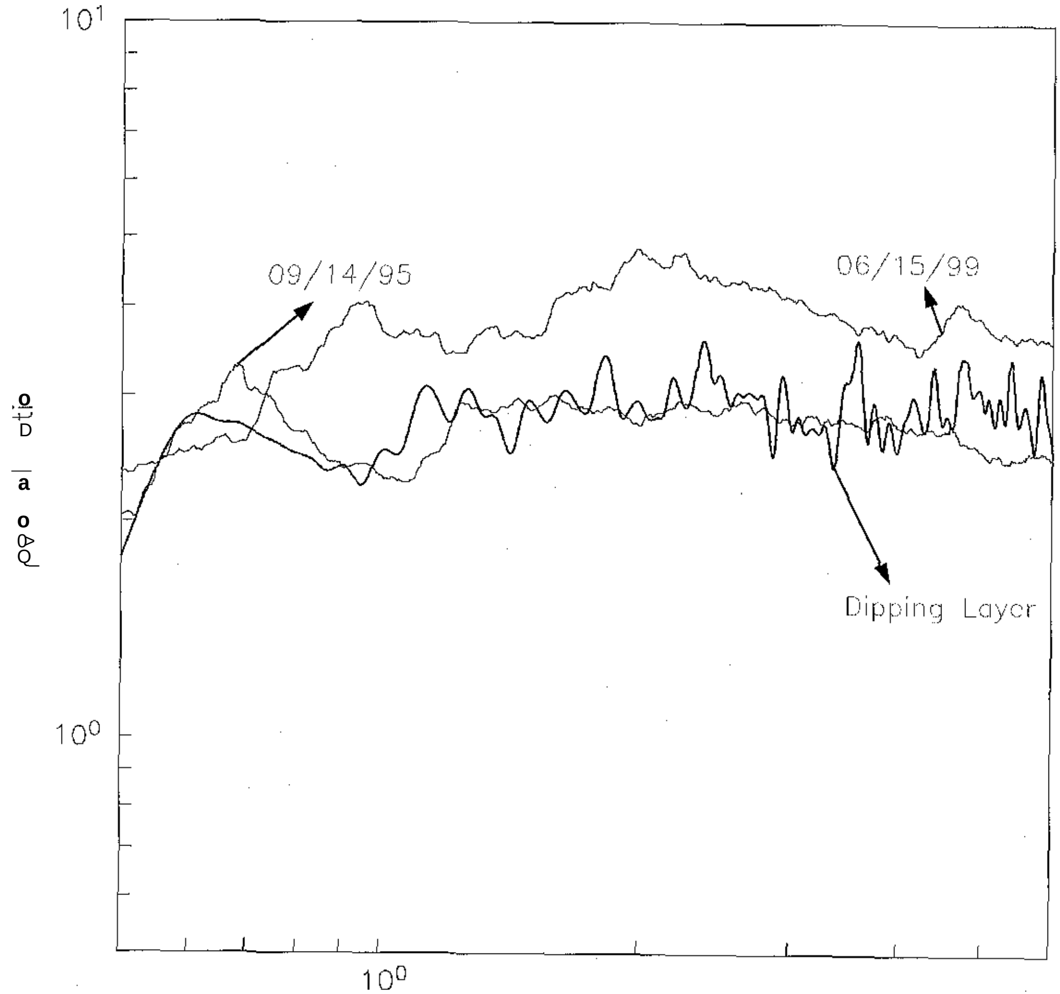


Figure 8



Frequency (Hz)

Figure 9

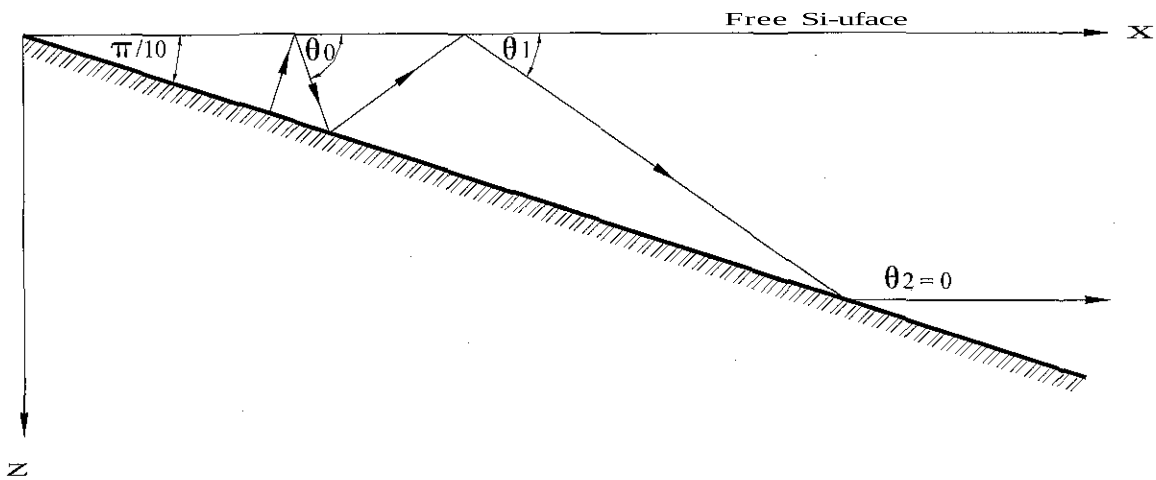


Figure 10

APÉNDICE III

THE SEISMIC RESPONSE OF DEEP SEDIMENTS IN MÉXICO CITY

(será sometido a la revista *Geophysical Journal Internationab* Í unio de 2002)

Research Note

THE SEISMIC RESPONSE OF DEEP SEDIMENTSIN MÉXICO CITY VALLEY

By

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Submitted to

Geophysical Journal International

June 2002

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THE SEISMIC RESPONSE OF DEEP SEDIMENTS IN MÉXICO CITY VALLEY

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SUMMARY

In recent studies it is mentioned the importance of the interaction between the propagation of waves guided by deep layers and the surñcial clay layer on the seismic response in the Valley of México. This relation may explain the long duration of strong motion and some ampliñcations of the seismic waves in the lake zone of México City. Chávez-García *et al.* (1995) mentioned that this deep layer lies between 1 to 4 km, and Shapiro *et al.* (2001) calculated the effect of this structure using ratios of the first two modes of Rayleigh and Love waves, they found that the dominant frequeney at 0.4 Hz represent this interaction. Shapiro *et al.* (2001), mentioned that if the higher-mode surface wave dominates the seismic response in the Valley of México, then it is necessary to know the deeper structure. Considering these observations, from the hypothesis of Montalvo-Arrieta *et al.* (2002b), we propose that seismic response of the north of the hill zone represents the wave field incident at the base of the deep deposits in the Valley of México.
Key words: Site effeets, transfer ñmetion, strong ground motion, Valley of México.

1INTRODUCTION

At present many studies of the seismic response in the Valley of México try to explain the origin of the long duration of recordings observed in México City *(e.g.* Singh & Ordaz, 1993; Chávez-García *et al.,* 1995; Shapiro *et a.,* 1997; Furumura & Kennett, 1998; Shapiro *et al.,* 2001) how a relation between the response of the basin and the regional structure (multipating, etc.). Chávez-García *et al.* (1995) & Shapiro *et al* (2001) mentioned that the long duration of strong motion is due to the interaction between the propagation of waves guided by deep layers and the surficial clay layer. This interaction is possible by the coincidence of the dominant frequeney of the uppermost layers and the frequeney of the deeply guided waves. In fact Shapiro *et al.* (2001) concluded that these guided waves corresponds to the higher mode of Rayleigh waves, in their work they calculated group and phase velocity dispersión curves for a crustal model that corresponds approximately to the Roma site, which is located at the lake-bed in México City, and compared with observations at this site. They mentioned moreover, that the evaluation of the site effeets in México City using ID models, only the very superficial clay layer has been taken into account. This approach successñilly reproduces the observed fundamental frequencies at many sites. However, a relatively low-velocity structure extends below the superficial clay layer to depths of a few kilometers (that is, a deep sedimentary basin and Quatemary volcanic rocks), further amplifying the seismic waves. Moreover, the resonance frequeney of tliis structure is relatively cióse to the fundamental frequeney of the clay layer. As a consequence of this coincidence, the dispersión curves and eigenfunctions of surface waves are expected to be strongly affected by the relative low-velocity structure below the clay layer. This observations may be important for understand the response of deep structures and their relation with surficial lake deposits.

In a recent work Montalvo-Arrieta *et al.* (2002b) proposed the use of the stations located in to the north of the hill zone in México City as the virtual reference site free of site effects. The geological conditions at the hill zone show that older Miocene age deposits are outcropping north of the City. They mentioned that northern sites represent the basement that underlay the lake deposits at approximately 1.0 - 2.0 Ion (Ordaz and Singh, 1992). In this paper we compared the spectral ratio between the station TXCR located in the north of the hill zone and the ROMA station located within the lake-bed, we used some of the events in Shapiro's work. The objective of this study is to show that stations located in the north of the hill zone represent the seismic response of the deep structure in the Valley of México.

2 DATA

Table 1 shows the events used in this study. Fig. 1 depicts the location of stations TXCR and ROMA, used for the calculation of the spectral ratio, these stations are located in the north of the hill zone and within the lake-bed zone, respectively. The Valley of México has been divided into three regions based on the geotechnical characteristics of its shallow layers: (1) the hill zone, formed by volcanic tuffs and lava flows, (2) the lake bed zone, formed by clays with thickness varying from 10 to 130 m; and (3) the transition zone, composed by alluvial sandy and silty layers, with scattered clay layers. Fig. 1 shows these geotechnical zones and some reference sites.

Fig. 2a and Fig. 2b shows the displacements of ground motion for the north-south component at TXCR and ROMA sites for event 2, whereas Fig. 2c depicts the displacement spectra of two stations respectively. In these figure we can see the enormous amplification between these two stations, and the monochromatic behaviour of the signal at ROMA site. Considering the hypothesis of Montalvo-Arrieta *et al.* (2002a,b) that the north of the hill zone represents a deep deposit of approximately 1000 - 2000 km below the lake-bed zone, we obtained the transfer ñnetion of ROMA/TXCR, this may represent the seismic response of the interaction of the deep layers and the surficial layer, like Shapiro's work. Chávez-García *et al.* (1995) mentioned that this deep layer is found between 1 to 4 km. For each event, we calculated average spectral amplitudes from two horizontal components at two stations. The average horizontal amplitudes obtained for ROMA site were divided by the average spectral amplitudes of TXCR site. Fig. 3 shows the average of the transfer ñmetion for all events, we fmd that the máximum amplitude in the írequeney domain is at 0.4 Hz, this peak corresponds to that found for Shapiro *et al.* (2001) using theoretical dispersión curves and eigenfunctions of the first two modes of Rayleigh and Loves waves. In fact, they mentioned that the higher-mode surface waves domínate in the response of the Valley of México, then a detailed knowledge of the deeper structure is requíred. Thus, we propose that seismic response of the north of the hill zone represent the wave field incident in the base of the deep deposits in the Valley of México.

3 CONCLUSIONS

In recent studies it is mentioned the importance of the interaction between the propagation of waves guided by deep layers and the surficial day layer in the seismic response in the Valley of México. This relation may explain the long duration of strong motion and some amplifications of the seismic waves in the lake zone of México City. Chávez-García *et al.* (1995) mentioned that this deep layer is found between 1 to 4 km, and Shapiro *et al.* (2001) calculated the effect of this structure using ratios of the fírst two modes of Rayleigh and Loves waves. Finally, from the hypothesis of Montalvo-Arrieta et al. (2002a), we propose that seismic response of the north of the hill zone represent the wave field incident in the base of the deep deposits in the Valley of México. We found that spectral ratio between TXCR and ROMA stations present a frequency dominant at 0.4 Hz like that Shapiro 's result.

ACKNOWLEDGMENTS

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FIGURE LEGENDS

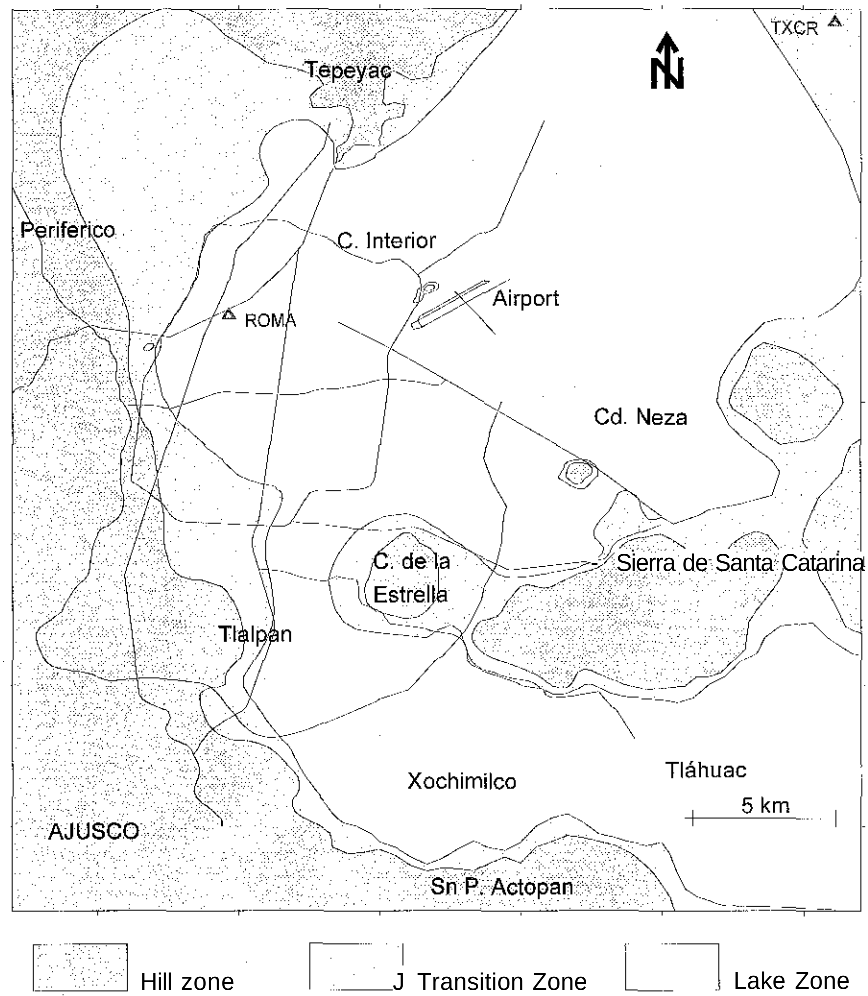
Figure 1. México city: Location of the TXCR and ROMA stations, geotechnical zones and some streets and city sites.

Figure 2. Ground motions recorded at (a) TXCR and (b) ROMA sites, and (c) displacement spectra.

Figure 3. Average of spectral ratio TXCR/ROMA, for all events used in this study.

Table 1. Parameters of earthquakes used in the study.

N	yy/mm/dd	Lat(°N)	Lon(°W)	Depth (Ion)	Mw
1	94/05/23	18.0	100.6	45	6.2
2	94/12/10	18.0	101.6	54	6.4
3	95/0914	17.0	99.0	22	7.3
4	95/10/09	18.8	104.5	10	8.0



TXCR
N
Tepeyac
Periferico
C. Interior
Airport
ROMA
Cd. Neza
C. de la
Estrella
Sierra de Santa Catarina
Tlalpan
Xochimilco
Tláhuac
5 km
AJUSCO
Sn P. Actopan
Hill zone
Transition Zone
Lake Zone

Figure 1

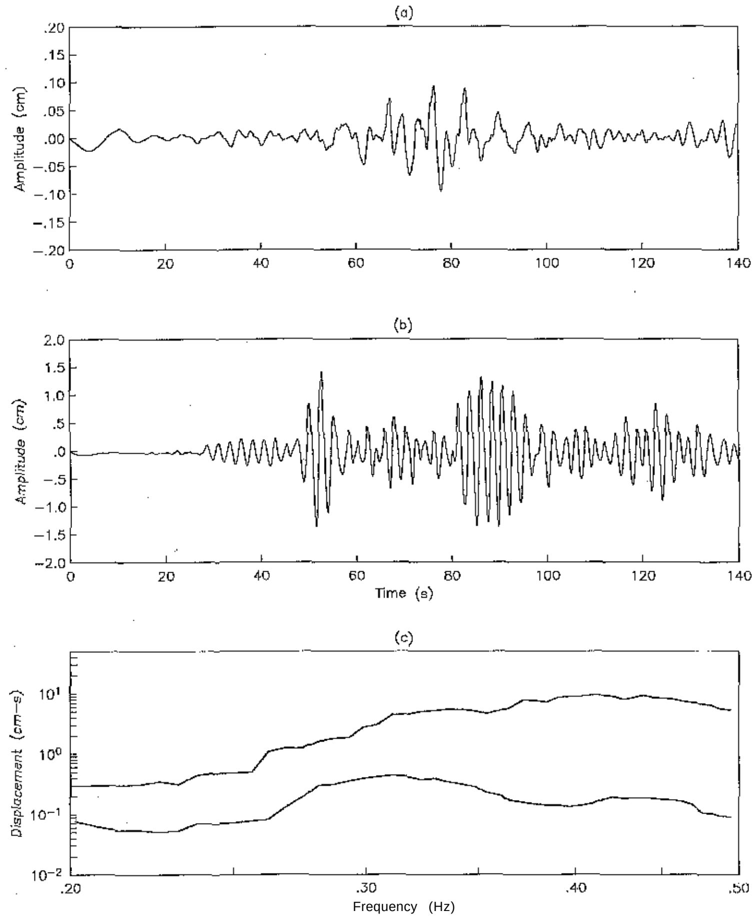

(a)
Amplitude (cm)
.20
.15
.10
.05
.00
-.05
-.10
-.15
-.20
0
20
40
60
80
100
120
140
(b)
Amplitude (cm)
2.0
1.5
1.0
.5
.0
-.5
-1.0
-1.5
-2.0
Time (s)
(c)
Displacement (cm-s)
10^1
10^0
10^{-1}
10^{-2}
.20
.30
.40
.50
Frequency (Hz)

Figure 2

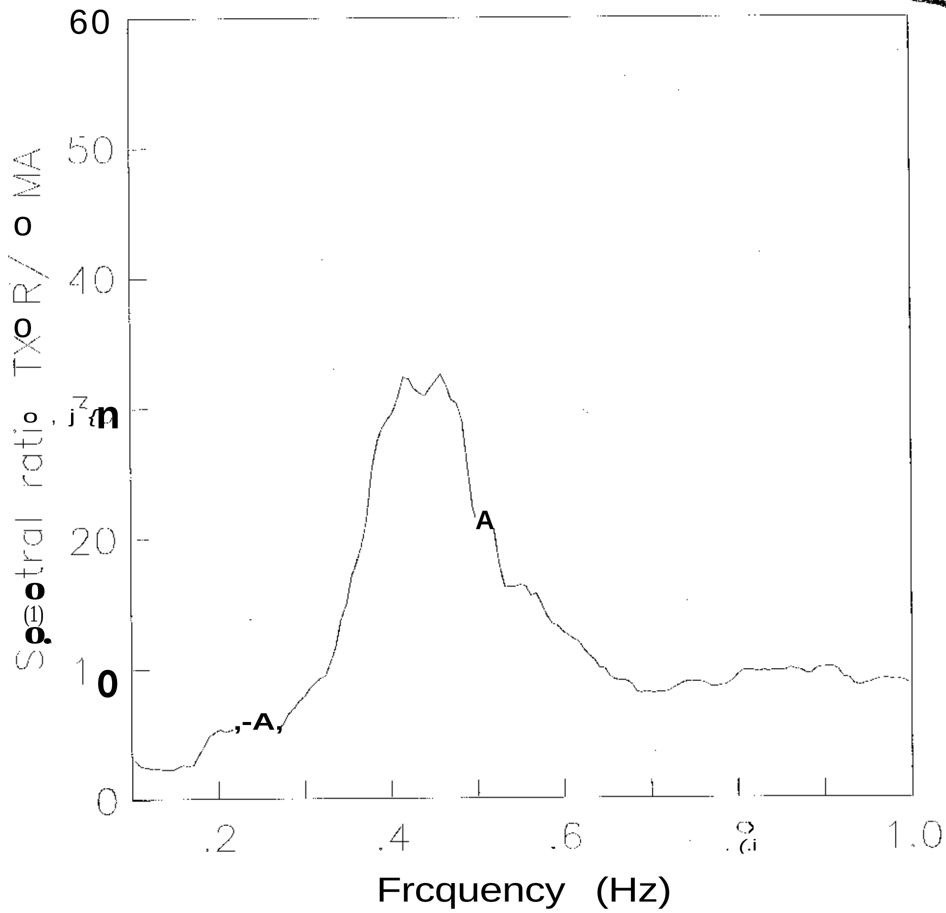


Figure 3

APÉNDICE IV

THE SEISMIC RESPONSE OF THE HILL ZONE IN MÉXICO CITY: A REVIEW AND NEW FINDINGS

(será sometido a la revista *Bulletin ofthe Seismological Society of America»* Íunio de 2002)

THE SEISMIC RESPONSE OF THE HILL ZONE IN MÉXICO CITY: A REVIEW AND NEW FINDINGS

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Submitted to

Bulletin of the Seismological Society of America

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THE SEISMIC RESPONSE OF THE HILL ZONE IN MÉXICO CITY: A REVIEW AND NEW FINDINGS

By

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ABSTRACT

The purpose of this paper is to make a review of the strong ground motion at *the hill zone* in México City. We examine all strong ground motion since 1964 data with Mw > 5.0 recorded at this zone considered as the seismic reference site in the city and evalúate the models existing, because it is important to try to explain the behavior of *the hill zone.* Additionally, we obtain a new attenuation laws to predict the seismic response for some dangerous earthquakes that affect the city.

INTRODUCTION

From a seismically view point, *the hill zone* in México City is considered the reference site. This región is characterized by lava flows (basaític composition) and volcanic tuffs (representing the hard soil). In the last 10 years, *the hill zone* has been growing at population and civil structures. A significant feature of this zone is the low risk to earthquakes that occur in the Mexican trench. In opposition, the lábelled lake zone has high vulnerability to distant, large, and frequent subduction earthquakes. However, the seismic risk in México City is not only related to coastal events. Exists another three types of sources that could cause damage in the city: inslab, continental and local earthquakes. These three last sources are the most dangerous for structures located at hill zone.

One characteristic of the coastal events is that the recurrence periods are low compared with the other types of sources. In general, the damage in the city has been associated to subduction earthquakes like the great Michoacan earthquake (September 1985; Mw = 8.0). Recently, in June 1999 and July 2001 occurred two inslab events: Tehuacan *(Mw* = 7.0) and Copalillo *(Mw* = 5.9). The former ones caused severe damage in the state of Puebla and Morelos, due to its proximity to epicenter (-100 km). In México City, at -200 km from the source, the damage was associated only to non-structural elements. For the Copalillo earthquake, no damage was reported in the México Valley. However, due tó the unusual nearness to the city (-137 km), acceleration records of very good quality were obtained. This situation has permitted to some authors scaled this event to Mw = 7.0 and estímate the seismic response in the valley. The results have shown that seismic risk is higher at hill zone in the frequency range of 1.0 to 3.0 Hz. For the others two types of sources (continental and local) the seismicity is low and scattered compared with coastal and inslab events. However, if earthquakes with M > 5.0 occurs in these regions, several damage could be produce in México City, mainly in *the hill zone.*

In México City the traditionally reference site is the CU station. The reason to use it is obvious, as since 1964 more than 60 modérate and strong earthquakes *(Mw* > 5.0) have been recorded. During the 1985 September 19 *(Mw* = 8.0) Michoacán earthquake three digital records were obtained at CU: MV, CU01 and CUIP located in an área of roughiy 2x2 km. At present, there are two three-component digital instrument operated by the Instituto de Ingeniería (Kinemetrics K2) and Instituto de Geofísica (broadband FBA23 + Quanterra of 24 bits), both at UNAM.

Añer the great 1985 Michoacán earthquake, the accelerometer array has grown considerably. ít now consists of more than 90 free-ñeld digital accelerometric stations, 18 of them in *the hill zone* (stations 07, 13, 18, 21, 28, 34, 40, 50, 64, 74, 78, ES, CH, CU, CN, TX, MD and TY). The stations 18 and 28 are at the Cerro de la Estrella and the Peñón outcrop. From all these stations is evident that the knowledge of the behavior of the seismic response of *the hill zone* will be better. In the early use of the reference site was assumed that the surface-rock-site record is equivalent to the input motion at the base of the soil layers. The ratios computed between lake and hill zone stations by Singh *et al.* (1988a,b) showed roughiy the same behavior during different earthquakes. These authors found that the spectral amplificaron at lake zone was higher that 50 times at 0.5 Hz, for the Michoacán earthquake. However, such ratios exhibited signiñcant variations if used to predict site amplification effects. Reinoso (1996) and Reinoso and Ordaz (1999) proved that these differences were smaller, and sometimes much smaller, when the average Fourier amplitude spectrum of south hill zone stations is used as the reference motion. The results by Reinoso and Ordaz (1999) showed that the average Fourier spectra are better and represent to general spectral shape of the incident wavefield at the base of the lake deposits at approximately 100 m. These authors compared spectral ratios with ID response of some sites at lake zone and found that are very similar. In the other hand Montalvo-Arrieta *el al.* (2002a,b,c) showed that northern stations located at hill zone represent the seismic response of the deep deposits (> 1000 m) in the Valley of México. These results show that the site response at lake zone is controlled by the effect of shallow and deep layers (Chavez-García, *et al.* 1995; Shapiro *et al.* 2001 and Montalvo-Arrieta *et al.* 2002c).

Other observations by Singh *et al.* (1988a, 1995); Ordaz and Singh (1992) and Sánchez-Sesma *et al.* (1993) pointed out that *the hill zone* in the Valley of México suffers important amplification with respect to that predicted by the attenuation laws for subduction earthquakes. This amplification could be as large as 10 for frequencies between 0.2 and 0.7 Hz. On the other hand, Pacheco and Singh (1995) studied several intermediate-dcpth normal faulting events recorded in México City. They found that *S* waves are amplified 2.5 times between 0.2 and 3.0 Hz. Montalvo-Arrieta *et al.* (2002a,b) mentioned that exist a clear dependence of the seismic response at hill zone sites on earthquake magnitude, azimuth, epicentral distance and depth. Subduction earthquakes concéntrate their energy at low frequencies (<1 Hz), while intraplate earthquakes have frequencies higher than 1 Hz. For subduction earthquakes, Pérez-Rocha (1998) found that exist a clear difference at the

Fourier spectra for each seismogenic zone at Mesoamerican trench, where Guerrero-San Marcos región represent to zone with greater risk to México City.

Severa! recent studies of México Valley reports significant differences between southwestern and northern hill zone stations (Reinoso and Ordaz, 1999; Montalvo-Arrieta *et al.,* 2002b). Ordaz and Singh (1992) found that strong ground motion at station MD, located in the northern part of *the hill zone,* was smaller when compared to the southwestern hill zone motion. Montalvo-Arríela *et al.* (2002b) try to explain this differences with 2D model and proposed that stations located in the north of this zone appear to be free of site effects and represent the geological basement of the Valley of México.

As mentioned before, the hill zone represent the reference site in México City. In this work we make a review and evalúate the models proposed in the literature to try to explain the behavior of this región for understand the seismic response of the México City Valley and we examine **ail** strong groun motion data recorded at CU site since 1964 with *Mw* > 5.0 to obtain new coefficients of attenuation laws to predict the seismic response for some dangerous earthquakes that affect the city.

PRINCIPAL SEISMIC SOURCES THAT AFFECT MÉXICO CITY

The seismic risk in México City is related to four types of sources: coastal, inslab, continental and local earthquakes (Figure 1); the three last seismogenic zones are the most dangerous for structures located at hill zone:

1) Subduction earthquakes have historically generated the most severe consequences in México City (located at -280 km from the near rupture área), mainly in *the lake zone* whereas *the hill zone* no damage is reported. The máximum magnitude recorded since 1800 is Ms = 8.4 for the 1932 Jalisco earthquake. However, the earthquakes originated near the coasts of Guerrero and Michoacán produce the most violent motions in the valley. These events have liad magnitudes smaller than Ms = 8.4. An analysis of the zones that can slip in a single great event shows that the most violent shock that one can reasonably expect in the Guerrero seismic gap, west of Acapulco, will be *Mw* = 8.2. The average computed time between great events is 59.3 year. The last such earthquake occurred in 1911 (Rosenblueth *et al.,* 1989). Although, the Continental and local earthquakes represents the most dangerous events to México City and mainly for *the hill zone;* the seismic risk is low due to the large recurrence period for this types of source (> 1000 year); in this case the inslab earthquakes represent the máximum perils to *the hill zone.*

2) intermedíate depth earthquakes. The inslab earthquakes in the subducted Cocos píate below south-central México cease to occur at depths of less about 80 km and well before reaching the Mexican Volcanic Belt (MVB) (Iglesias, *et al,* 2002). Rosenblueth *et al.* (1989) considered that the most dangerous earthquake of this group has *Mw* = 6.5 and a depth of 80 km. However, the magnitude of the Tehuacán earthquake of June 1999 *(Mw* = 7.0 and H = 60 km) (Singh *et al.,* 1999) is larger than the predicted by Rosenblueth's work. This earthquake caused severe damage in the state of Puebla and Morelos, due to its proximity to epicenter (-100 km). In México City, at -200 km from the source, the damage

was associated only to non-structural elements (Singh *et al,* 1999; Shapiro *et* a/., 2000). Iglesias, *et al.* (2002) and Montalvo-Arrieta, *et al.* (2002a) found that if occur an earthquake with *Mw* = 7.0 and epicentral distance ~137 km, several damage could be produce in México City, mainly in *the hill zone* in the frequency range of 1 to 3 Hz. García and Suárez (1996), compiled the effects of histórica! earthquakes felt in México, including México City and its hill zone. The earthquake of 1858 caused extensive damage in the state of Michoacán and in México City. Newspapers of that time describe that in Coyoacán "many houses suffer damages". This town is located in the border between the transition and hill zone. Singh *et al.* (1996) estimate a magnitude of 7.7 for this earthquake and suggest a normal fault origin.

3) Continental píate earthquakes. The activity in this región are characterized by shallow extensional seismicity. Large earthquakes in this área include the 1912 Ms = 6.7 Acambay and the 1920 Ms = 6.2 Jalapa earthquakes (Suter *et al.,* 1996). For the Acambay cvent these authors estimated a máximum intensity of VIII in México City. The seismicity in this región is related to numerous east-west striking normal faults that are characterized by pronounced scarps and displace Quaternary volcanic rocks (Suter *et al.,* 1996). The earthquakes that most seriously damage will produce to the valley is a Ms = 6.7 (Acambay type) that occurs to distance -100 Ion. However, the slip rate in Acambay región is nearly 0.4 mm/yr (Rosenblueth *et al,* 1989) and Suter *et al.* (2001) found that slip rates for Central México vary between 0.02 to 0.18 mm/yr. Rosenbíueth *et al.* (1989) suggests a return period of about 1500 years for magnitude 6.7.

4) Local earthquakes. These events are situated under the Valley of México. One characteristic of this zone is the shallow, extensional and low seismicity. The máximum magnitude registered instrumentally is Me = 3.9, in the Milpa Alta región (UNAM and CENAPRED Seismology Group, 1995). Mooser (1987) mentioned that máximum earthquake recorded in this century is M = 5.1. Rosenblueth *et al.* (1989) predict the seismic response of a event of M = 4.7 with focal distance of 11 km, this earthquake will produce the scenario must dangerous for the city.

Figure 2, show a map of southern México with epicenter locations (Tables 1 and 2) and rupture áreas of some of the earthquakes used in this study, that correspond to coastal and intraplate events with *Mw* > 5.0. Figure 3 depiets the average spectral amplitude of the southwest stations for a typical subduction and inslab events. As mentioned before, the coastal event concéntrate their energy at low frequencies (< 1 Hz), while intraplate earthquakes have frequencies higher that **1** Hz.

EFFECTS OF THE WAVE PATH IN MÉXICO CITY

The study of the effects of the incident wavefield in México City, is related to explain the origin of the long coda observed at strong ground motion recordings from subduction earthquakes (Campillo *et al.,* 1988; Sánchez-Sesma *et al,* 1988; Kawase and Aki, 1989; Sánchez-Sesma *et al,* 1993; Chávez-García, *et al,* 1995; Shapiro *et al.,* 1997; Furumura and Kennett, 1998 and Shapiro *et al,* 2001). The interest lies in to try to explain the relationship between local and regional structures (México basin, MVB and Mexican trench) in the seismic response. The presence of a low-velocity zone beneath the southern parí of the MVB produce that the seisrnic signal are amplified in the Central México (Shapiro *et al,* 1997). This structure was inferred from the observations of regional amplification by Ordaz and Singh (1992); Sánchez-Sesma *et al.* (1993); Singh *et al.* (1995). The results of Furumura and Kennett (1998), shown that long duration in México City could be reproduce by 2D regional model. In the other hand, recently, one important effect was observed, Shapiro *et al.,* (2000) found that the seismograms recorded in México City (CU station) reveal that amplitudes of seismic waves whose wavepaths pass below Popocatépetl volcano before reaching the city are diminished. Montalvo-Arrieta *et al.* (2002a); Iglesias *et al.* (2002) found same results. These authors mentioned that the high attenuation was attributed to the presence of magma and partial melting of rocks below the volcano. Shapiro *et al* (2000) mentioned that seismic waves are diminished by a factor of about one-third at f > 1 Hz, as compared to those that do not cross the volcano. This observations are for the June 15, 1999 *{Mw* = 7.0) earthquake, which epicenter was located southwest of Tehuacán, Puebla (figure 2). It cause severe damage to the states of Puebla and Morelos. In México City, about 200 km from the epicenter, damage was associated only to non-structural elements. The ground motion observed in México City during this earthquake was much smaller that the predicted one (Singh *et al.,* 1999; Shapiro *et al.,* 2000). The seismic risk in *the hill zone* from inslab earthquakes is higher. However, due to the high attenuation observed by the June 1999 event, Montalvo-Arrieta *et al.* (2002a) and Iglesias *et al.* (2002) from the Copalillo earthquake (July 21, 2000; *Mw* = 5.9) estimated the seismic response at CU for a postulated Mw = 7.0 earthquake, like the June 15, 1999 event, these authors found that the máximum acceleration and spectral amplitudes are larger than the June earthquake. This event was located to southeast of the Popocatépetl volcano while the Copalillo earthquake was located to southwest of this structure.

Recently, from May to August 2000 a small broad band network was installed around the Popocatépetl and Iztaccíhualt volcanoes. Preliminary analysis of two earthquakes are presented. Figure 4, shows the localization of the network around these volcanoes. The epicenter of the July 3, 2000 earthquake (M = 4.3), and the wavepath for all stations are presented. The trajectory to RFPP is the only one that crosses the volcanoes. The reference station is HUPP. Figures 5 and 6 shows the north-south and vertical components of the seismograms. At station RFPP there is a strong attenuation with respect to stations TLPP and PGPP, whose wavepaths do not cross the volcanoes. On the vertical component the effect is more enhanced. The *S* wave almost disappears in RFPP station.

To better understand the effect of Popocatépetl on wave propagation, we studied the Fourier spectra, SR(f), of ground motion of stations: HUPP, RFPP and PGPP. Since the hypocentral distances to RFPP and PGPP are lager than to HUPP, we reduced the observed spectra at RFPP and PGPP to the distance to HUPP. For this purpose, we used a regional spectral attenuation relation valid for south-central México, namely (Ordaz and Singh, 1992):

$$\frac{e^{-\pi f R/\beta Q_0(f)}}{R^{0,}} \qquad 0)$$

where f is the frequency, R is the hypocentral distance, *f*> is the S wave velocity, and $Q_0(f) = 273 f^{066}$ is the S wave quality factor. We take p = 3.5 km/s. Figure 7 shows the Fourier spectra corrected for the same distance for the stations HUPP, PGPP aiid RFPP. The Fourier spectra for the RFPP is strongly attenuated for frequencies larger than 1 Hz, while for the stations PGPP and HUPP the Fourier spectra are similar in the whole range of frequencies.

Figure 8 show the epicenter of the August 20, 2000 earthquake (M = 4.7) and the wavepath to stations. Figure 9 shows the seismograms (north-south component) for all stations including CUIG. It can be seen a strong attenuation at CUIG. This wave path crosses the Popocatépetl volcano, however at PGPP site there is an important attenuation. Figure 10 shows the Fourier spectra for RFPP, CUIG and HUPP sites, is clearly the strong attenuation in CUIG for frequencies larger than 1 Hz. This observations coincide with the result found by Shapiro *et al.* (2000) for the same range of frequencies.

Finally, one implication of the high attenuation observed is a decrease in the seismic hazard in the Valley of México from earthquakes which originated to the east of the Popocatépetl volcano.

GEOLOGICAL SETTING

The basin of México is located in the central part of the Mexican Volcanic Belt. Voícanic rocks form the geology of the basin. It is filled by lacustrine sedimenta, volcanic tuff, sands and gravéis. The Valley of México has been divided into three regions based on the geotechnical characteristics of its shallow layers: (1) *the hill zone,* formed by volcanic tuffs and lava flows, (2) the lake bed zone, formed by clays with thickness varying from 10 to 130 m; and (3) *the transition zone,* composed by alluvial sandy and silty layers, with scattered clay layers (Marsal and Mazari, 1959). Figure 11 shows these geotechnical zones together with the accelerometric stations and some reference sites.

The southern part of *the hill zone* is formed by recent Quaternary lava flows from the Sierra de Chichinautzin (sites CU, 13, 34, 40 and 74, figure 11) that overlie soft material (Delgado *et al,* 1999). The western portion is dominated by tuffs (sites CH, TY, 07 and 74) while the northern part is composed of Miocene lava (sites MD, TX, ES and 64) that overlies volcanic rocks from the Oligocene age (Mooser *et al.* 1996). This Miocene lava underlies the deposits from Sierra de Chichinautzin and the tuffs of the western poition. Figure 12 shows a geological cross section across the Valley in the north-south direction (Mooser *et al.* 1996), from Sierra de Guadalupe to Sierra de Chichinautzin (figure 11). In this cross section, it is clear that the horizon of Miocene lava is dipping to the south. Gutiérrez *et al.* (1994) and Flores-Estrella and Aguirre (2002) obtained geological sections for the south of *the hill zone* (site CU) using seismic refraction profiles and microtremors array measurements respectively, while in the north of *the MU zone,* Singh *et al.* (1995) show a geological model for MD and TX sites. Figure 13 depicts the 5-wave velocities of these profiles; we can see that 5-wave propagation is low compared with northern sites for the upper layers (< 100 m), while for thickness greater than 250 m, the *S* wave velocities are very similar.

DIFFERENCES AMONG HILL-ZONE SITES

As mentioned before *the hill zone* are characterized by variations in the soils deposits, the older and hard rocks outcrops are located in the north and the recently and soft materials are at south. Reinoso and Ordaz (1999) found a significant amplification (regional), for sites located to the southwest hill zone over sites located to the north of the city (figure 14). Stations located in the southwest área show higher amplification with respect to CU station (Montalvo-Arrieta *et al*, 2002a). The spectral ratio of the averages of the stations located in the south and west with respect to the northern stations show a relatively constant amplification of up to 3 times in the 0.7 to 10.0 Hz frequency band (Singh *et al,* 1995; Montalvo-Arrieta *et al,* 2002b).

From strong ground motion data, some authors had to try to evalúate the site effects in CU, using the north stations (TX, MD and ES) as reference site, the amplification factor is about 3 times (Singh *et al,* 1995; Montalvo-Arrieta *et al,* 2002a). Montalvo-Arrieta *et al.* (2002a) compared the spectral ratios of CU/ES and 74/ES, sites located around 3 km between them (over lava flows) with the response ID, these authors found that spectral ratio CU/ES is similar with the ID model, while the spectral ratio 74/ES present amplification that is not well reproduced by the ID model (figure 15). Delgado-Granad os and Mooser (2001, personal communication) mentioned that differences in the local amplification between CU and others sites of the south of *the hill zone* (74, 40 34), may be due to the presence of son material under the lava flows where the stations are located.

The geological conditions at *the hill zone* show that older, Miocene age deposits are located north of the city (figure 12). Montalvo-Arrieta *et al* (2002b) considered that northern sites represent the basement, they assume that the configuration along *the hill zone* in the N-S direction can be approximated by a simple dipping homogeneous layer. They computed the antiplane seismic response for this model, and averaged and compared it with the spectral ratio obtained from strong ground motion data from subduction earthquakes (figure 14). The agreement is good and suggests how a smooth, large scale feature could amplify seismic ground motion in a broad frequency band (figure 16).

THE INFLUENCE OF DEEP DEPOSITS IN THE SEISMIC RESPONSE

The origin of the long duration in the strong ground motion data in the Valley of México has been generated great controversy. Singh and Ordaz (1993) propose that this long coda is caused by multipathing between the source and the site (separated by more than 250 km for coastal earthquakes) and/or within the large basin on the Valley of México. Some authors try to explain this effect with 2D crustal models (Shapiro *et al,* 1997 and Furumura and FCennet, 1998). Recent studies mentioned the importance of the interaction between the propagation of waves guided by deep layers and the surficial clay layer in the seismic response in the Valley of México. This relation may explain the long duration of strong motion and some amplifications of the seismic waves in *the lake zone* of México City (Capillo *et al,* 1988; Sánchez-Sesma *et al,* 1993; Chávez-García *et al,* 1995 and Shapiro *et al,* 2001). Chávez-García *et al* (1995) mentioned that this deep layer is founded between 1 to 4 km. Shapiro *et al.* (2001) calculated the effect of this structure using ratios of the first two modes of Rayleigh and Loves waves; they found that the dominant frequency at 0.4 Hz represent this interaction. Shapiro *et al* (2001) mentioned that if the higher-mode surface wave domínate the seismic response in the Valley of México, then it is necessary to know the deeper structure. Considering these observations, Montalvo-Arrieta *et al.* (2002c) propose that the seismic response of *the hill zone* at the north may represent the wave ñeld incident in the base of the deep deposits in the Valley of México.

ATTENUATION LAWS FOR CU

One of the characteristics of earthquake ground motions of interest to engineers and seismologists alike are the prediction of the máximum ground motion. As mentioned before, the traditionally reference site in México City, is the CU station, as since 1964 more than 60 modérate and strong earthquakes (Mw > 5.0) have been record, and represent the site with more records in the city (including the great Michoacán earthquake of September 19, 1985; Mw = 8.0). The earliest attenuation laws obtained at CU for máximum acceleration are from Esteva and Villaverde (1973) and Bufaliza (1984) in Singh *et al* (1988). Singh *et al* (1988) obtained an attenuation relations for CUTP, from 16 coastal earthquakes with 5.6 < Ms < 8.1 and distances 282 < R < 466 *km.* Ordaz and Singh (1992) in order to quantify amplification of seismic waves in *the hill zone* of México City, obtained spectral attenuation curves from eight coastal earthquakes. Ordaz *et al.* (1994) applied a Bayesian regression technique to derive attenuation relations for Fourier spectra at station CU from 21 subdution earthquakes. Singh *et al* (1999) obtained an attenuation law for horizontal peak ground motion acceleration, a_{max}, as a function of focal distance for the June 15, 1999, Tehuacán earthquake (Mw = 7.0). They found that a_{max} valúes from normal fault event are higher than for shallow thrust events, the same observation is found for the Oaxaca normal fault event of September 30, 1999, Mw = 7.5 (Singh *et al,* 2000). In a recent study García *et al* (2001) obtained attenuation laws for a_{max} and spectral amplitude for intraplate earthquakes in Central México. Reyes *et al* (2001) obtained attenuation law for acceleration response spectra CU from 20 subduction earthquakes. Finally, Iglesias *et al.* (2002) obtained a_{max} valúes including CU from 15 inslab earthquakes.

One of the goals of this work is that we examine all strong ground motion data since 1964 with Mw > 5.0 to obtain new coefficients in the attenuation laws for station CU in México City. These expressions are for peak valúes (a_{maXi} v_{mflX}), spectral amplitude of the ground motion and response spectra of acceleration, velocity and displacement from 44 coastal and 14 intraplate earthquakes obtained at station CU from 1964 to 2000. Data cover magnitude 5.1 <Mw< 8.0 and distance 270 <R < 545 *km,* for subduction earthquakes and 5.2 <Mw< 7.4 and distance 136 < R < 435 *km,* for intraplate earthquakes respectively. We show the predicted seismic response for a hypothetical Mw = 8.2 earthquake that would take place in the Guerrero gap, with R = 280 *km* because it is considered as the región with more seismic risk to México City (mainly, *the late zone)* for the subduciton zone. We also show the predicted response for a Mw = 7.0 intraplate earthquake from the Copalillo región with R = 136 km, due to, represent the máximum perils to *the hill zone* and could cause heavy damage to small buildings in the city. This two kind of sources shows how the amplitude and spectral contení are very different for subduction and intrapíate earthquakes, so the damage associated to structures in México City is very different.

We ñt the strong-motion data by a linear regression (Joyner and Boore, 1981) using the equation:

$$\backslash ogy=a_0 + a_1 Mw + a_2 \backslash ogR + a_3 R \qquad (2)$$

where y is eifher peak horizontal acceleration, velocity, spectral amplitude of ground motion, response spectra of acceleration, velocity or displacement j=\)^ $\sqrt{(ns^2 + ew^2\backslash/}$ ¡A , $\bar{a}_i$, i = 1,...,4 are the regression parameters, Mw is the moment magnitude and R is the closest distance to rupture área in Ion. From equation (2) and data in Table 1 and 2, we obtained the coefñcients for each parameter.

a_{max}* and *v_{max}

Although, it is generally recognized that these characteristics of the ground motion are not necessarily the most important features of the damage potential of the motion, the máximum or peak acceleration and velocity provide a simple indicator of the intensity of shaking (Seed *et al.* 1976).

We used peak valúes of acceleration of ground motion of 44 coastal events. The accelerograms were integrated to obtain 40 peak valúes of velocity of ground motion, the remain 4 were not used because of their bad quality. Table 1 shows these results. Equations (3) and (4) are the regressions for peak valúes of a_{max} and Vm_{ax} :

$$\log a_{mK}=12.4+0.6Mw-6.7\log¿? + 0.0037.R \qquad (3)$$

$$\log v_{max}=3.3 + 0.7Aíw-3.11ogi?-0.0004i? \qquad (4)$$

Figure 17 show observed and predicted valúes of a_{mnx} for 44 events. Figures 18a and 18b show a_{mfiX} and v_{max} predictions 5.0 < Mw < 8.2 and 280 < R < 515 km, respectively. We obtained a máximum peale valúes of 70.11 cm/s^2~ and 9.54 cm/s that correspond to a predicted peale valué for a hypothetical Mw = 8.2 earthquake that would take place in the Guerrero gap, with R = 280 *km.* Figures 19 show the comparison of our results with those of Singh *et al.* (1988) for a_{max}. This figure shows that exist some variations at both predictions; however, at magnitude 7.0 the valúes of a_{max} are very similar, the differences observed in both predictions are probably due to the number of events used in the regression (16 for Singh *et al.* and 44 for our regression).

For intraplate earthquakes we used 14 inslab earthquakes (Table 2) for obtaining the predicted attenuation law for a_{max} and v_{max}:

$$\log a_{\max} = 3.2 + 0.9Mw - 3.6\log R + 0.0019R \qquad (5)$$

$$\log v_{imx} -0.3 + 1.lMw-3.41og£-0.0030i? \qquad (6)$$

Figure 20a show the observed and predicted valúes ofa_{max} for 14 events. Figure 20b depiets the comparison between the predicted a_{max} law for Iglesias *et al.* (2002) and our results It can see that there is a difference between the two curves; our fitting model is better because we only used data for CU site while Iglesias *et al.* used data of other stations located to the south of the México Valley, some of the stations used for these authors present minor site effect that CU station, this may produce lower amplitudes for $Mw > 6.0$ that our prediction. Figures 21a and 21b show a_{max} and $v_{mfíX}$ predictions for $5.9 < Mw < 7.0$ and $136 < R < 300$ km, respectively. We obtained máximum peak valúes of 67 cm/s^2 and 8 cm/s, that correspond to a predicted peak valué for a hypothetical $Mw = 7.0$ earthquake that would place in Copalillo región with $R = 136$ km, which represent to zone with great risk to liill zone (Iglesias *et al,* 2002).

Spectral amplitude

The seismic response of the soil in México City has successfully been characterized by empirical transfer functions (Singh *et al.* 1988; Reinoso and Ordaz, 1999) between soft sites and a reference hill-zone station (station CU, located in the University City). Since empirical transfer functions are reasonably well estimated for several sites within the Valley of México, Fourier acceleration spectra at these sites can be computed for a future earthquake if the Fourier spectrum at the reference station is known.

We obtain the coefficients for the predicted spectral amplitude of ground motion for frequencies from 0.1 to 10.0 Hz for coastal evetens. Figure 22 shows the predicted and observed spectral amplitude during 1985, $Mw = 8.0$, Michoacan earthquake; it can be seen that both spectra are very similar between 0.3 to 3 Hz. Figure 23a depicts the spectral amplitude for $5.0 < Mw < 8.0$ and $R = 280$ *km,* where there is a clear effect of the magnitude in the shape of the predicted spectra. Figure 23b shows the predicted spectrum of the hypothetical $Mw = 8.2$, $R = 280$ *km,* Guerrero earthquake compared with three recording at CU of the 1985, $Mw = 8.0$, Michoacan earthquake. The máximum amplitudes are for the predicted earthquake as expected, because it has a larger magnitude and shorter distance. These resulte imply that the ground motion experienced in México City during the 1985 event is not the maximun expected motion.

Figure 24a show the predicted and observed spectral amplitudes for the Copalillo earthquakes (2000/07/21, $Mw = 5.9$). Figure 24b show the predicted spectra amplitude of the Copalillo earthquake for $Mw = 5.9$, 6.5 and 7.0; it can be seen that there is a variation of the shape of the spectra due to the effect of the magnitude and that the frequency band are larger than the coastal earthquakes.

Response spectra

In earthquake engineering the response spectrum provides a convenient way of characterizing ground motion and their effects On structures. Response spectra is the base of the design spectrum, which provide the seismic coeñcients for the design of new structures, or the seismic safety evaluation of existing ones. Three types of response spectra are used in the evaluation of the seismic response of structures: the acceleration, velocity and displacement response spectra.

In this study we obtained attenuation laws for these three response spectra from 0.1 to 6.0 seconds, for a 5 percent damping ratio and for coastal and intraplate earthquakes.

Acceleration response spectra

Figure 25a shows the comparison of the observed and predicted response spectra for the 1985, Mw = 8.0, Michoacan earthquake. It can be seen that both curves are very similar. Figure 25b depicts the response spectra for 5.0 < Mw < 8.2 and R = 280 *km* for coast events. The predicted response amplitude for Mw 8.2 are the 140 cm/s between 1.0 to 3 second.

Figure 26a show the predicted and observed acceleration response spectra for the Copalillo earthquake, there is difference of two times in 0.3 s between the observed and the predicted. Figure 26b depicts the predicted response spectra for Mw = 5.9, 6.5 and 7.0.

Velocity response spectra

Figure 27a shows the comparison of the observed and predicted response spectra, for the 1985, Mw = 8.0, Michoacan earthquake. It can be seen that both curves are very similar. The maximun peales are between 1.5 and 3.5 s. Figure 27b depicts the response spectra for 5.0 < Mw < 8.2 and R = 280 *km*. ín this figure there is a clear effect of the magnitude in the shape of the predicted spectra.

Figure 28a shows the predicted and observed response spectra for the Copalillo earthquake. Figure 28b depicts the predicted response spectra for Mw = 5.9, 6.5 and 7.0. For magnitude of 7.0 there are a important band between 1 to 5 second that are not present in the coastal events, however the amplitudes of the response spectra for intraplate events are twice minor the coastal events.

Displacement response spectra

Figure 29a shows tbe comparison of the observed and predicted response spectra, for the 1985, Mw = 8.0, Michoacan earthquake, it can be seen that both curves are very similar. Figure 29b depicts the response spectra for 5.0 < Mw < 8.2 and R = 280 *km*.

Figure 30a show the predicted and observed response spectra for the Copalillo earthquake. Figure 30b depicts the predicted response spectra for Mw = 5.9, 6.5 and 7.0. For magnitude of 7.0 there are a important band between 1 to 6 second.

DISCUSSION AND CONCLUS1ONS

In this work has been studied the seismic response of *the hill zone* of México City. We make a review and examine all data since 1964 with Mw > 5.0 and model existing at this zone. The interest for this study is that, *the hill zone* represent the reference site in the Valley of México and because in the last 10 years this región has been in growing at population and civil structures.

One cause of the ausence of studies of the seismic risk in *the hill zone*, is that considered as low risk, due to no damage has been report from subduction earthquakes, this events, historically has cause great injures in the city, mainly in *the lake zone*. However, this kind of events is not the only sources that affect to México City. Exist another three types could cause damage: inslab, continental and local. These last sources are the most dangerous for structures located at hill zone; although, the two 1ast sources represent the most dangerous events to city, the seismic risk is low due to the large recurrence period (> 1000 year for continental), in this case the inslab earthquakes represent the máximum perils to *the hill zone.* In the last four years, two great intraplate events occur in Central-México. Tune 15, 1999 *(Mw* = 7.0) and September 30, 1999 (M=7.5) these earthquakes cause severe damages in some cities of the central part of México. Recently, in July 21, 2000, an intraplate event *(Mw* = 5.9) occur near of Valley of México (~ 130 km), in the city no injures was reported, however, some researches scaled this earthquake to *Mw* = 7.0, these authors mentioned that severe damage could be produce in México City, mainly in *the hill zone.*

In *the hill zone* exists significant differences in the seismic response between southwestern and northem stations. The spectral ratio of the averages of the stations of the south with respect to the north show a relatively constant amplifícation of up 3 times in the 0.7 to 10.0 Hz. Considering that northem sites represent the basement, some authors assume that the conñguration along *the hill zone* in the N-S direction can be approximated by a simple dipping homogeneous layer. The resuít between the comparison of the response of the 2D model and the spectral ratio, from strong ground motion data from subduction earthquakes is good and suggests how a smooth, large scale feature could amplify seismic ground motion in a broad frequency band. The site response in the *lake zone* is controlled by the effect of shallow and deep layers.
Another importan observation, is that the seismic risk in México City is low when the incident wave-fíeld cross the Popocatépetl volcano before to reach the city. The seismic waves suffer high attenuation by a factor of about one-third at f > 1 Hz.

Additionally, we obtain a new coeficientes for attenuation laws to predict the seismic response at CU, for peak horizontal acceleration, velocity, amplitude spectral of ground motion, response spectra of acceleration, velocity and displacement, using all recording at CU since 1964 with *Mw* > 5.0. We analize two dangerous scenarios (subduction and inslab earthquakes). We found that there exists in all prediction curves a clear effect of the magnitude in the obtained peak valúes, this produces a whole range of máximum valúes for greater magnitudes than for smaller magnitudes. This should be considered in the design spectrum. We show the predicted seismic response for a hypothetical *Mw* = 8.2 earthquake that would take place in the Guerrero gap, with *R* = 280 *km* and *Mw* = 7.0 intraplate event at ~130 km of distance. The results of the prediction for the subduction event imply that ground motion experíenced in México City during the great Michoacan earthquake, is not the máximum expected motion. The differences found between the subduction and inslab earthquakes, show that the damage in México City are variable with respect to kind of source.

ACKNOWLEDGMENTS

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Suter, M., Carrillo-Martínez, M., and Quintero-Legorreta O. (1996), Macroseismic study of shallow earthquakes in the central and eastern parts of the Trans-Mexican Volcanic Belt, México, *Bull. Seism. Soc. Am.* 86, 1952-1963.

Suter, M., López-Martínez, M., Quintero-Legorreta, O., and Carrillo-Martínez, M.. (2001), Quaternaty intra-arc extensión in the central Trans-Mexican volcanic belt, *Geol. Soc. Am. Bull.* 113,693-703.

IMAM and CENAPRED Seismology Group. (1995), The Milpa Alta earthquake of January 21, 1995, *Geofis. Int.* 34, 355-362.

FIGURE LEGENDS

Figure 1. Location of the mainly seismogenic regions that affect to México City (modified of Ronsenblueth *et al*, 1989).

Figure 2. Location of epicenters and rupture áreas (for $Mw > 5.0$) of subduction and normal-faulting earthquakes used in this study. The plot also show the data and magnitude of each event (modified of Reinoso and Ordaz, 2001).

Figure 3. Average of Fourier amplitude spectra for the north-south component of motion for southwest stations. (a) June 15, 1999 and (b) September 14, 1995 (modified of Montalvo-Arrieta *et al*, 2002a).

Figure 4. Location of the seismic broadband network (HUPP, PGPP, RFPP, TLPP and TEPP) around the Popocatépetl and Iztaccíhuatl volcanoes, the epicenter of the Juiy 03, 2000 ($M = 4.3$), and the wave-path for all stations.

Figure 5. Seismograms at HUPP, PGPP, RFPP and TLPP stations for the north-south component, for the July 03, 2000 ($M = 4.3$) earthquake.

Figure 6. Seismograms at HUPP, PGPP, RFPP and TLPP stations for the vertical component, for the July 03, 2000 ($M = 4.3$) earthquake.

Figure 7. Fourier spectral amplitude corrected for distance (Ordaz and Singh, 1992) for HUPP, PGPP, RFPP and TLPP stations for the north-south component, for the July 03, 2000 ($M = 4.3$) earthquake.

Figure 8. Location of the seismic broadband network (HUPP, PGPP, RFPP, TLPP and TEPP) around the Popocatépetl and Iztaccíhuatl volcanoes, and the CUIG station at UNAM, the epicenter of the August 08, 2000 (M = 4.7) earthquake and the wave-path for all stations.

Figure 9. Seismograms at HUPP, PGPP, RFPP, TLPP and CUIG stations for the north-south component, for the August 08, 2000 ($M = 4.7$) earthquake.

Figure 10. Fourier spectral amplitude corrected for distance (Ordaz and Singh, 1992) for HUPP, PGPP, RFPP, TLPP and CUÍG stations for the north-south component, for the August 08, 2000 ($M = 4.7$) earthquake.

Figure 11. México City: accelerometric stations, geotechnical zones and reference sites. The accelerometric array consists of more than 90 free-field digital accelerometric stations, 18 in the hill zone (07, 13, 18, 21, 28, 34, 40, 50, 64, 74, 78, CU, CH, CN_S ES, MD, TX and TY) (modified of Reinoso and Ordaz, 1999).

Figure 12. Geological cross section across the valley from Sierra de Guadalupe (north) to Sierra de Chichinautzin (south). This cross section is based on field geology, geological information and seismic reflection and refraction data (modified of Mooser *et al*, 1996).

Figure 13. Site pro file obtained for CU station using refraction proñles and microtremors array measurements (Gutiérrez *et al*, 1994 and Flores-Estrella and Aguirre, 2002) used to compute one-dimensional transfer function of site CU and site profile obtained for MD and TXbySinghe/a/., 1995.

Figure 14. Fourier amplitude spectra of the September 14, 1995 earthquake. (a) north-south, and (b) east-west (ASW, average of southwest stations, AN, average of north stations, and CU station).

Figure 15. Strong ground raotion at stations 74, CU and ES, of the June 15, 1999 earthquake east-west component. (a) Fourier amplitude spectra, (b) spectral ratios 74/Es, CU/ES, and ID response, (c) accelerometric data (Montalvo-Arrieta *et al,* 2002a).

Figure 16. Spectral ratio for the September 14, 1995 and the seismic response of a dipping layer (thick line, modified of Montalvo-Arrieta *et al*, 2002b).

Figure 17. Observed *versus* predicted valúes of a_{mnx} for 44 subductions earthquakes at CU station.

Figure 18. Predictions of máximum ground motion at CU station for subduction earthquakes, for distances of 280, 295, 300, 350, 400 and 515 km, up to down, respectively. (a) a_{max} and (b) v_{max}.

Figure 19. Comparison of our results and the predictions by Singh *et al,* 1988c, at CU station (a) a_{max} and (b) v_{mnx}.

Figure 20. (a) Observed *versus* predicted valúes of a_{max} for 14 inslab earthquakes at CU station with our attenuation law. (b) Comparison of our results and the predictions by Iglesias *et al*, 2002, at CU station.

Figure 21. Predictions of máximum ground motion at CU station for inslab earthquakes, for distances of 136, 200, 250, 300 and 350 km, up to down, respectively) (a) a_{max} and (b) v_{max}.

Figure 22. Comparison of the spectral amplitude of the great Michoacan earthquake of 1985 and the predicted with our attenuation law for a Mw = 8.0 and R = *295* km at CU station.

Figure 23. (a) Spectral amplitude predicted at CU station for 8 magnitudes (Mw = 5.0, 5.5, 6.0, 6.5, 7.0, 7.5, 8.0 and 8.2) and R = 280 km. (b) Comparison between the predicted spectral amplitude and the observed at CU for a Mw =8.2 and R ~ 280 km.

Figure 24. (a) Predicted and observed spectral amplitude for the July 21, 2000 Copalillo earthquake (Mw = 5.9 and R = 137 km), (b) Predicted spectral amplitude at CU station for Mw = 5.9, 6.5 and 7.0 and R = 137 km.

Figure 25. (a) Comparison between the predicted acceleration response spectra and the observed (September 19, 1985) at CU for *&Mw* = 8.2 and *R* - 280 km (0.1 to 6.0 s and 5 percent damping ratio). (b) Predicted response spectra for *Mw* = 5.0, 5.5, 6.0, 6.5, 7.0, 7.5, 8.0 and 8.2 and /? = 280 Ion.

Figure 26. (a) Comparison between the predicted acceleration response spectra and the observed (Copalillo earthquake) at CU for a *Mw* = 5.9 and *R* = 137 km (0.1 to 6.0 s and 5 percent damping ratio). (b) Predicted response spectra for *Mw* = 5.9, 6.5 and 7.0 and *R* = 280 km.

Figure 27. Same as figure 25 but for velocity response spectra.

Figure 28. Same as figure 26 but for velocity response spectra.

Figure 29. Same as figure 25 but for displacement response spectra.

Figure 30. Same as figure 26 but for displacement response spectra.

Table 1. Subduction earthquakes used in this work

Table 2. Inslab earthquakes used in this work.

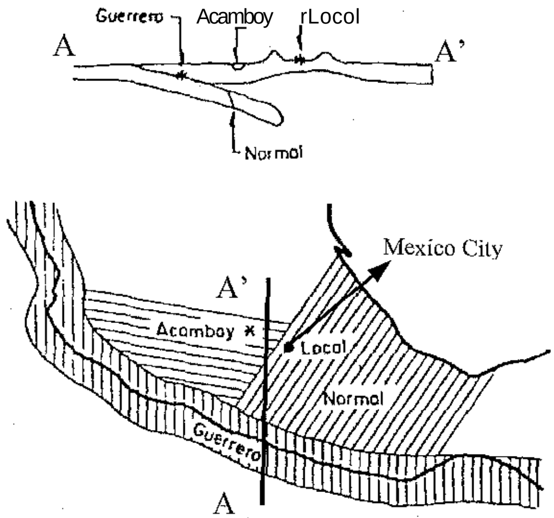

Guerrero
Acamboy
Local
A
A'
Normal
Mexico City
A'
Acamboy
Local
Normal
Guerrero
A

Figure 1

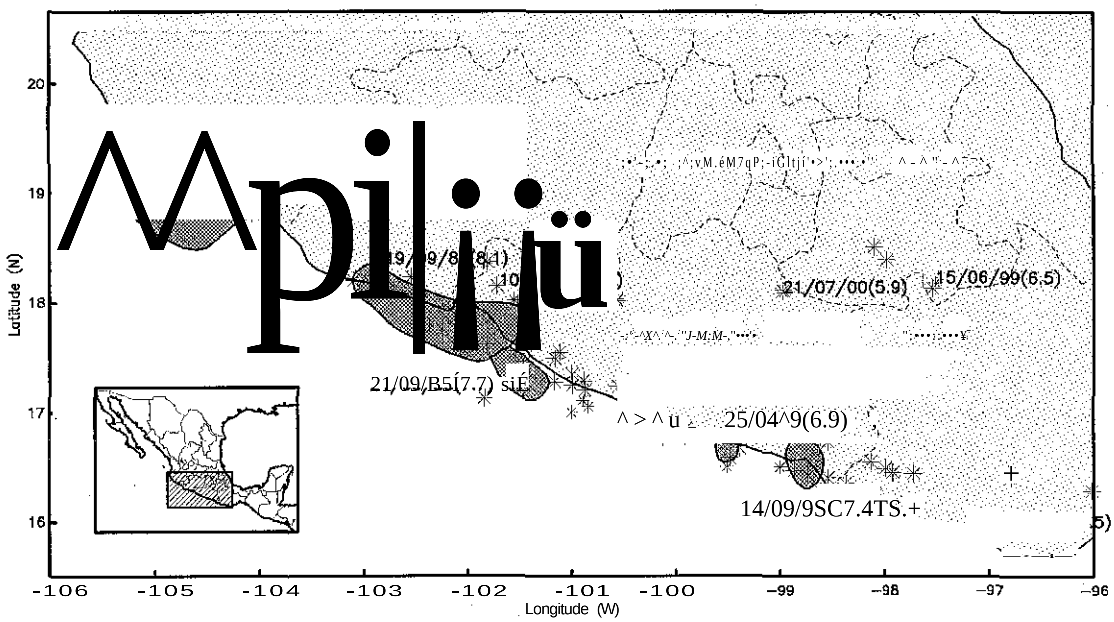

Latitude (N)
Longitude (W)
20
19
18
17
16
-106
-105
-104
-103
-102
-101
-100
-99
-98
-97
-96
21/07/00(5.9)
15/06/99(6.5)

Figure 2

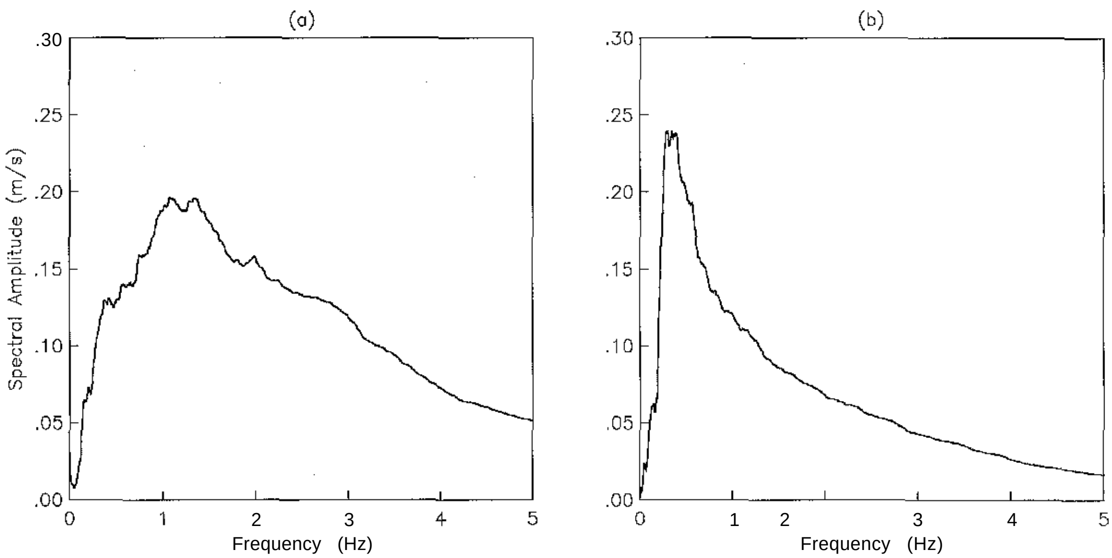

(a)
(b)
Spectral Amplitude (m/s)
.30
.25
.20
.15
.10
.05
.00
0
1
2
3
4
5
Frequency (Hz)

Figure 3

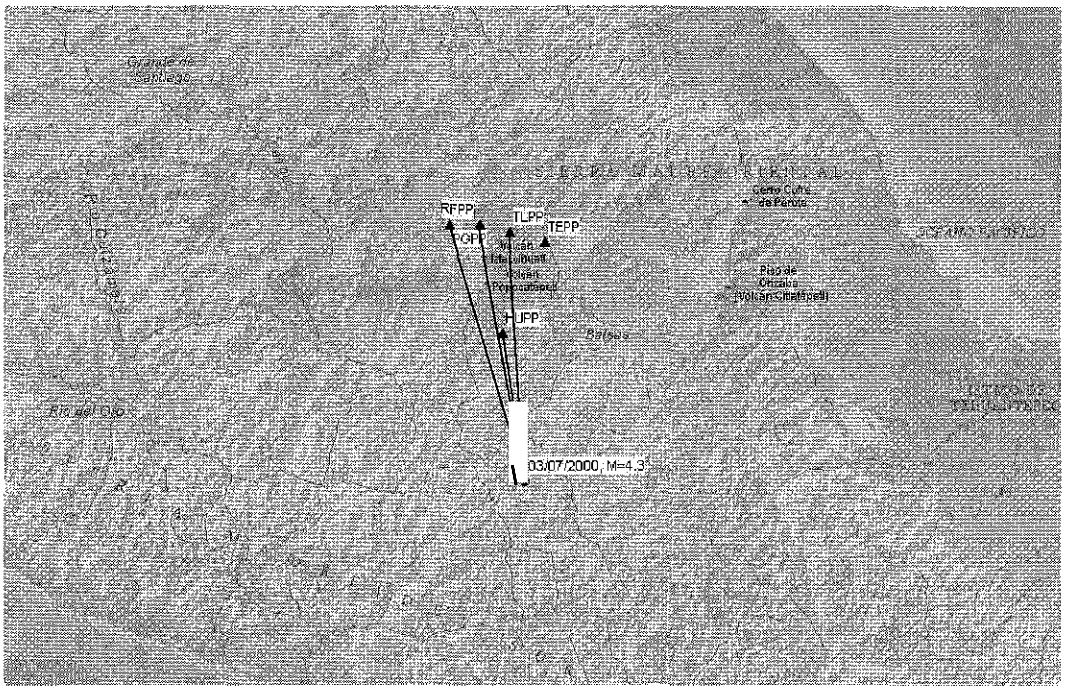


Figure 4

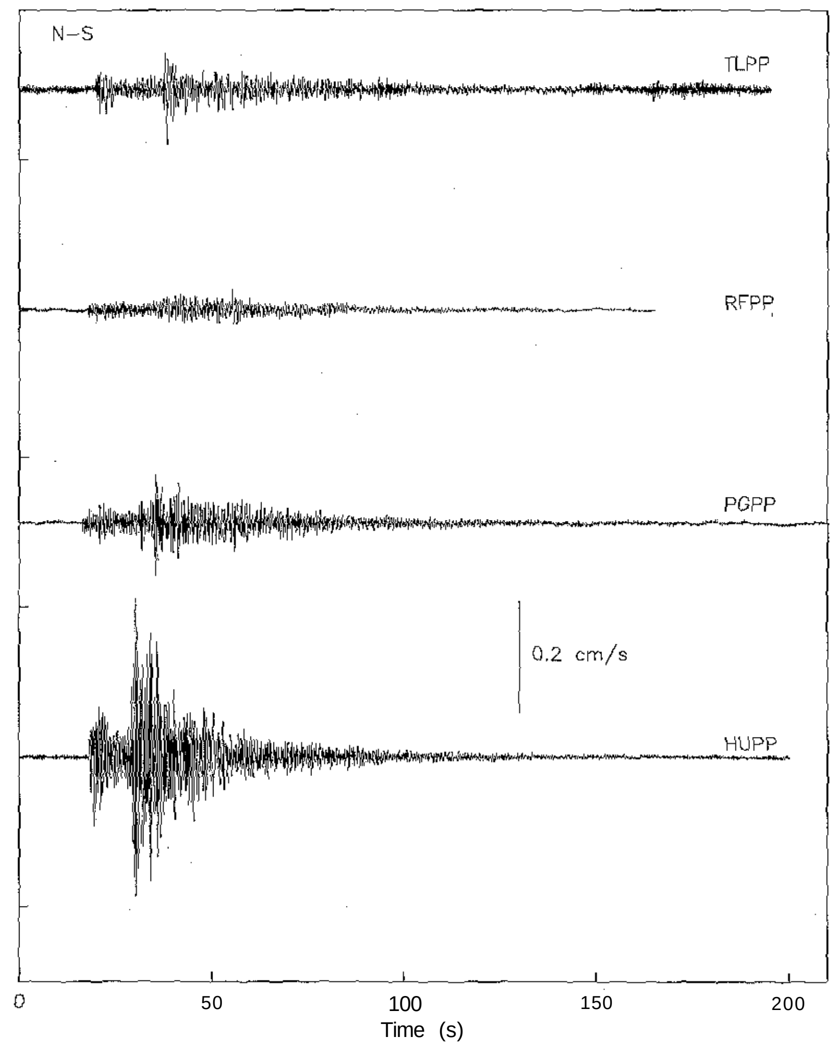


Figure 5

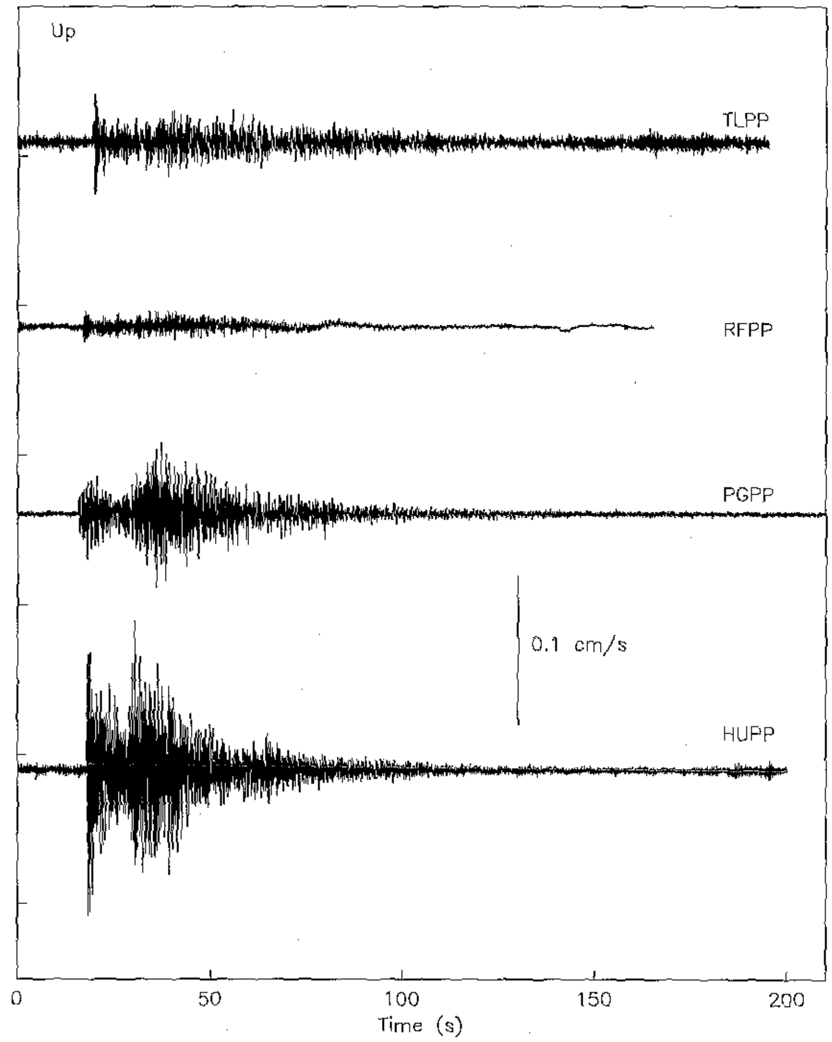


Figure 6

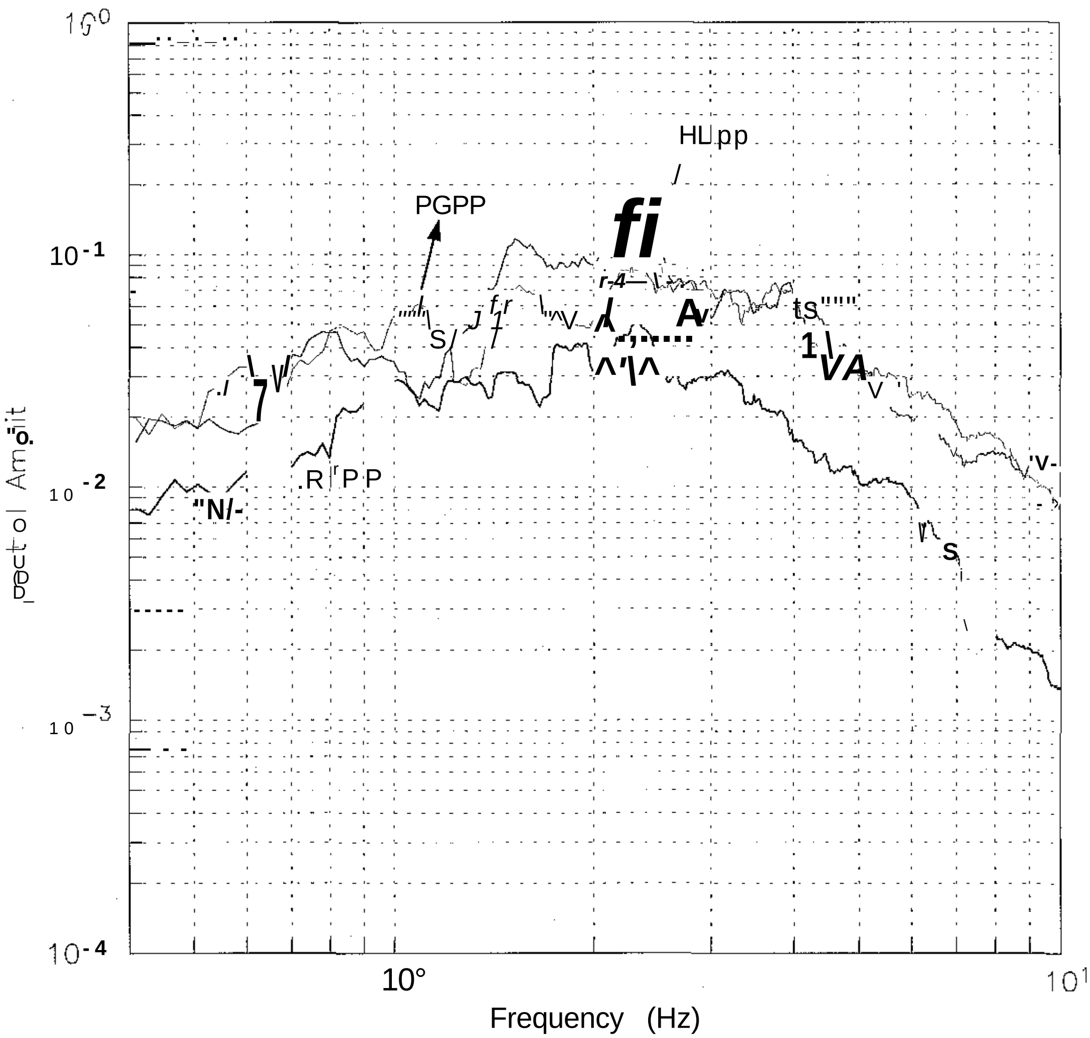


Figure 7

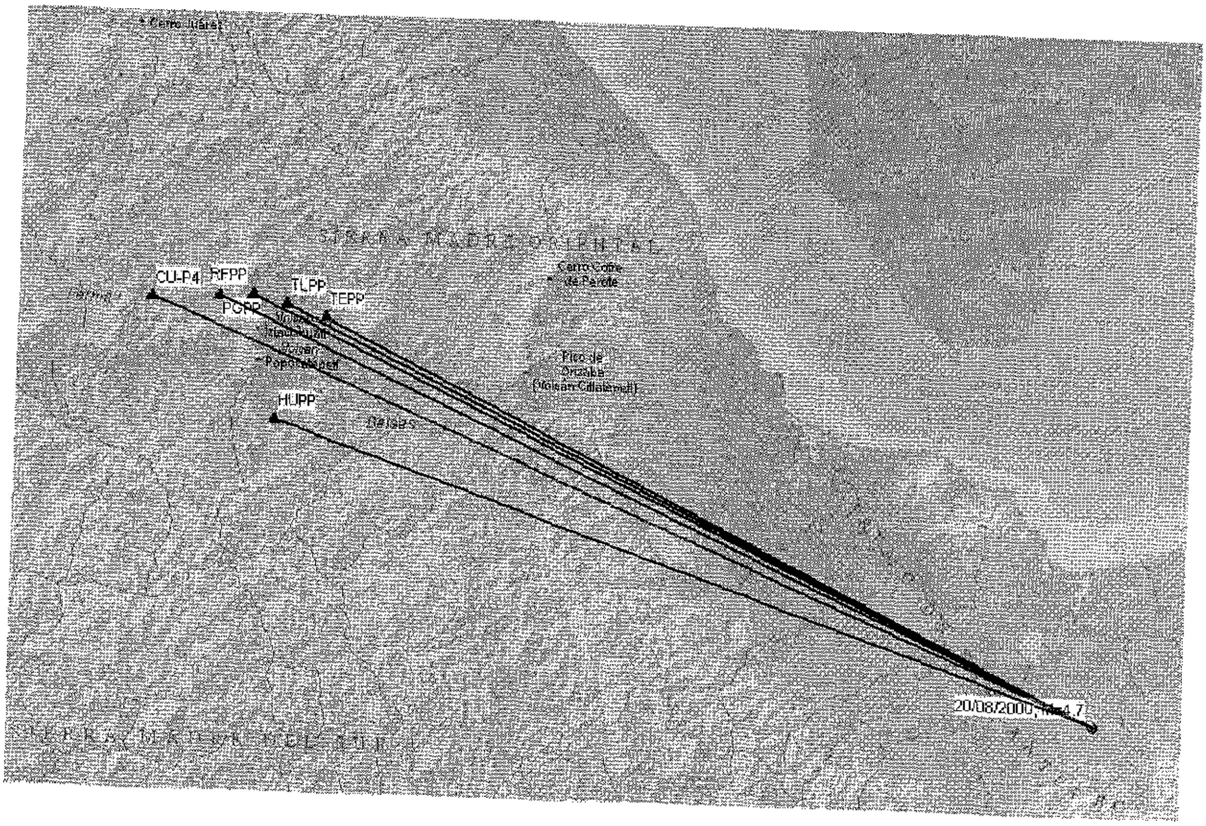

Cerro Juárez
SIERRA MADRE ORIENTAL
CU-P4
RFPP
TLPP
TEPP
PGPP
Cerro Cofre de Perote
Pico de Orizaba
HUPP
Balsas
20/08/2000

Figure 8

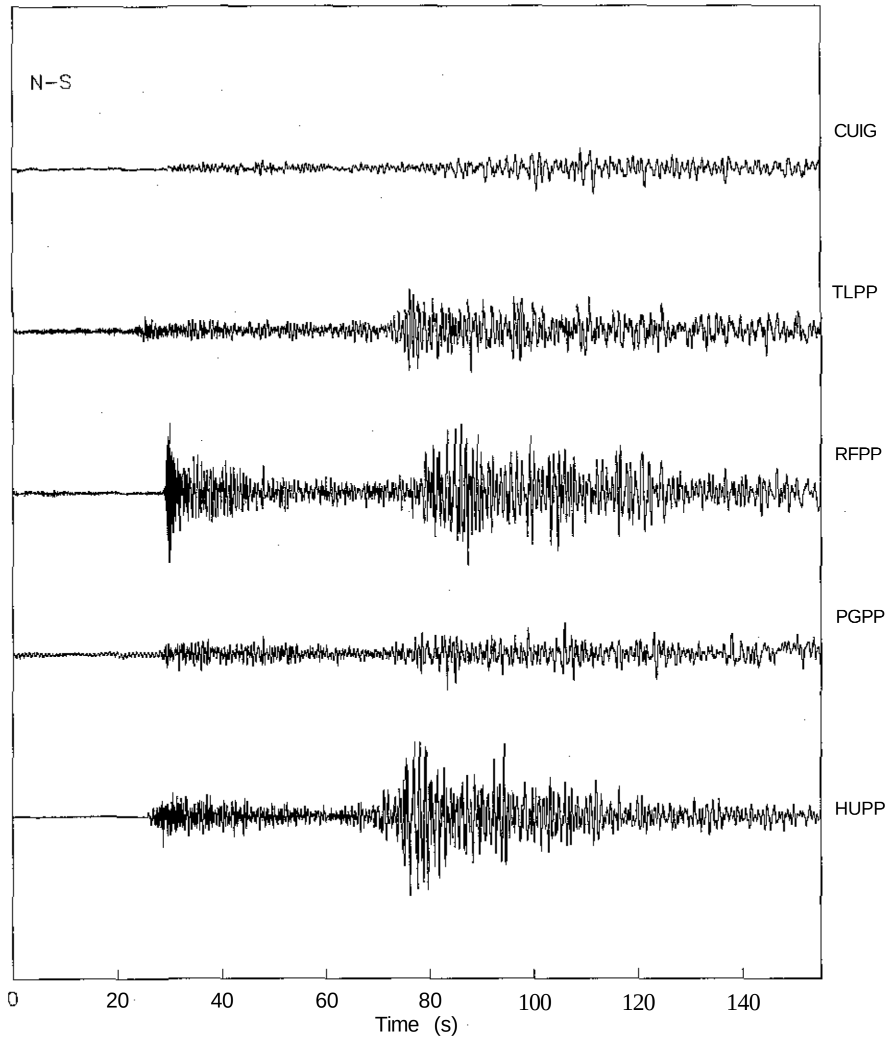
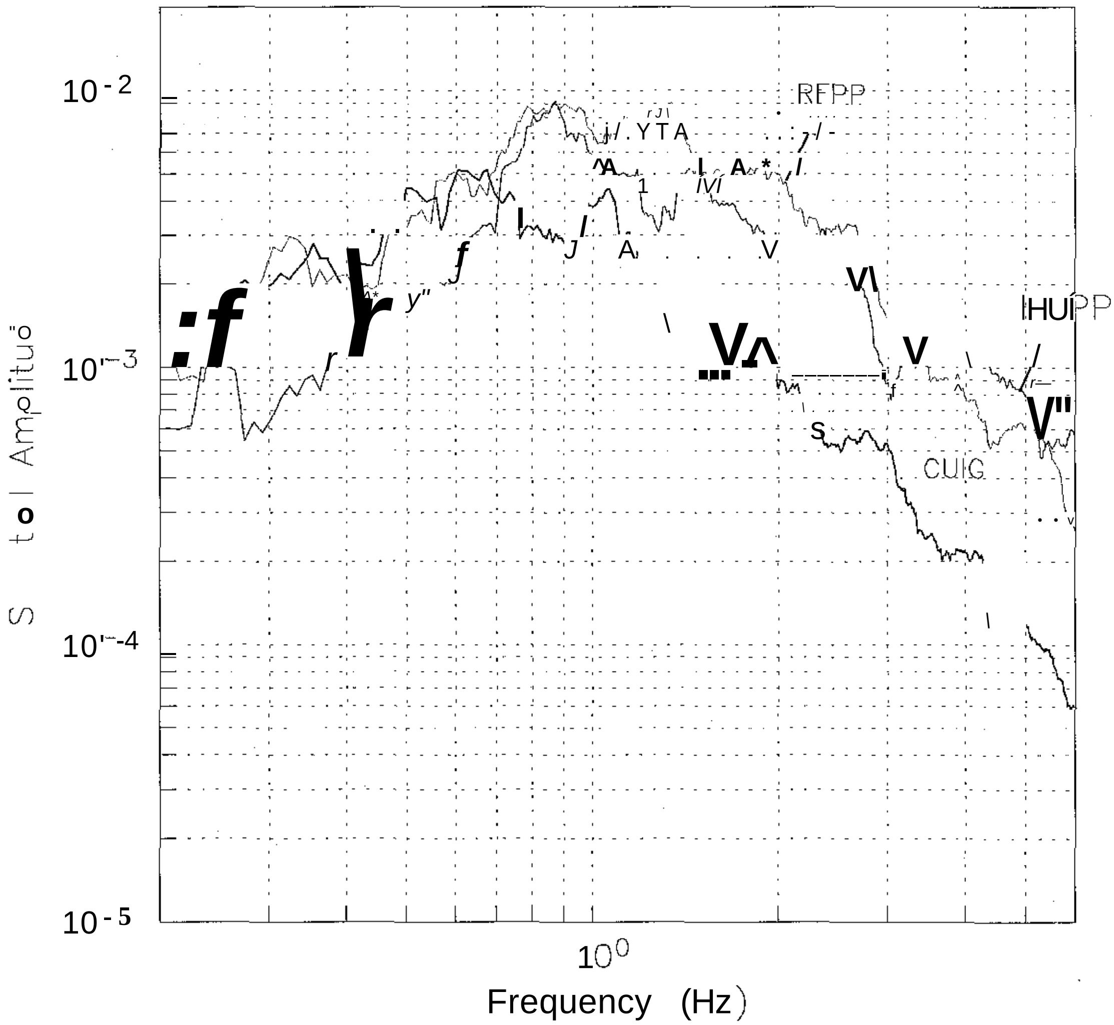


Figure 9



10-2
10-3
10-4
10-5
Spectral Amplitude
RFPP
YTA
HUPP
CUIG
10^0
Frequency (Hz)

Figure 10

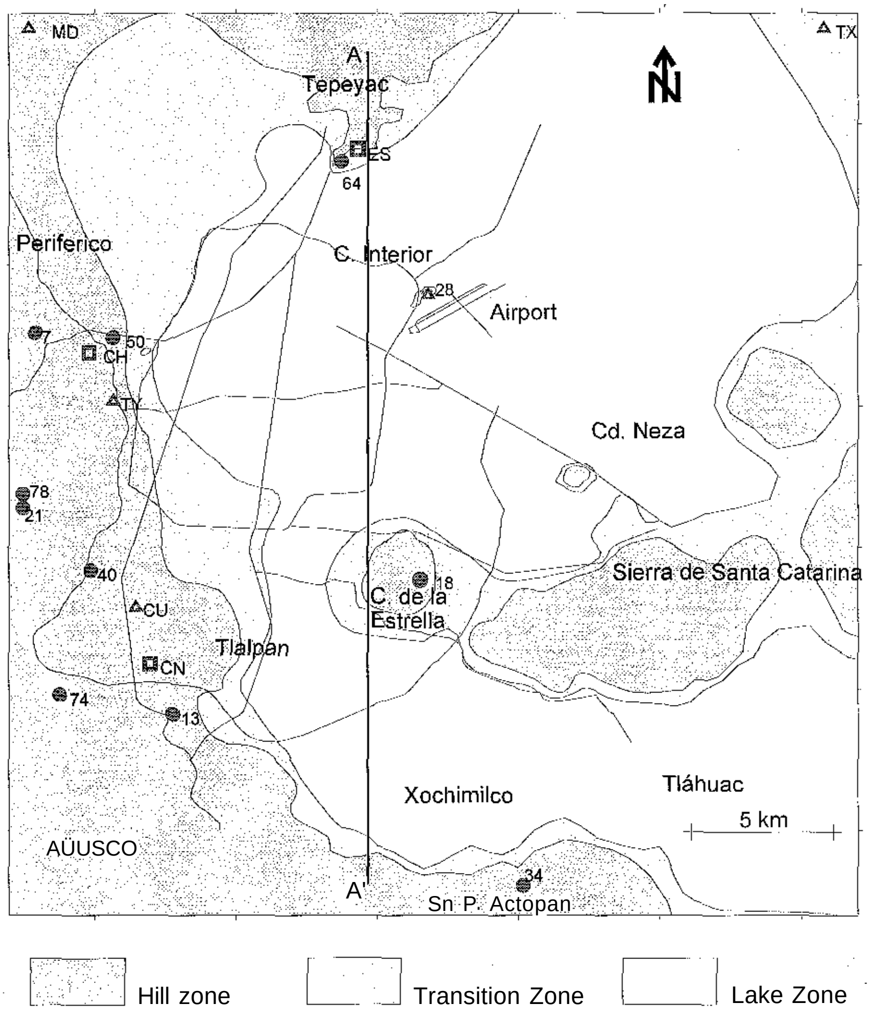

MD
TX
A
Tepeyac
N
ZS
64
Periferico
C. Interior
28
Airport
7
50
CH
TY
Cd. Neza
78
21
40
18
Sierra de Santa Catarina
C. de la
Estrella
CU
Tlalpan
CN
74
13
Xochimilco
Tláhuac
5 km
AÜUSCO
A'
34
Sn P. Actopan
Hill zone
Transition Zone
Lake Zone

Figure 11

31

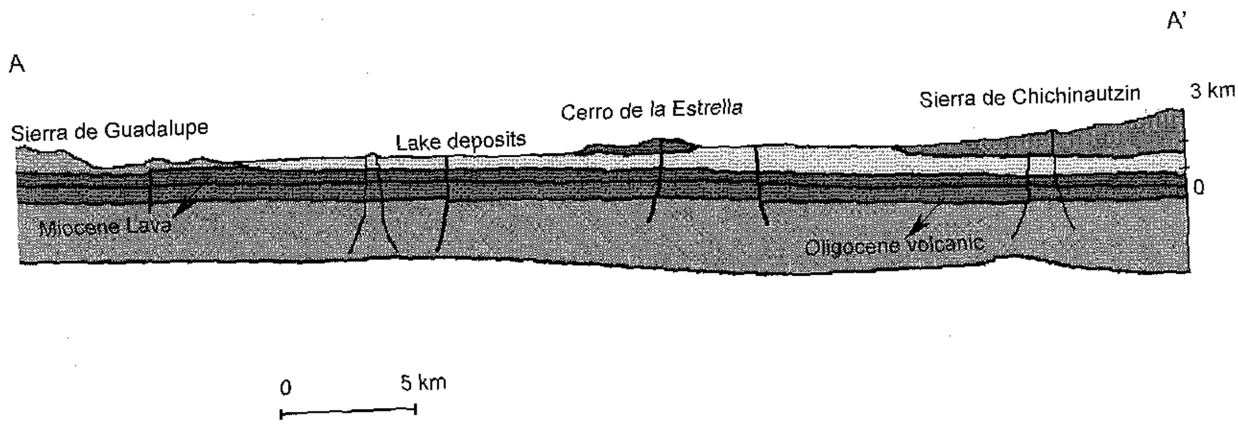

A
A'
Sierra de Guadalupe
Lake deposits
Cerro de la Estrella
Sierra de Chichinautzin
3 km
0
Miocene Lava
Oligocene volcanic
0
5 km

Figure 12

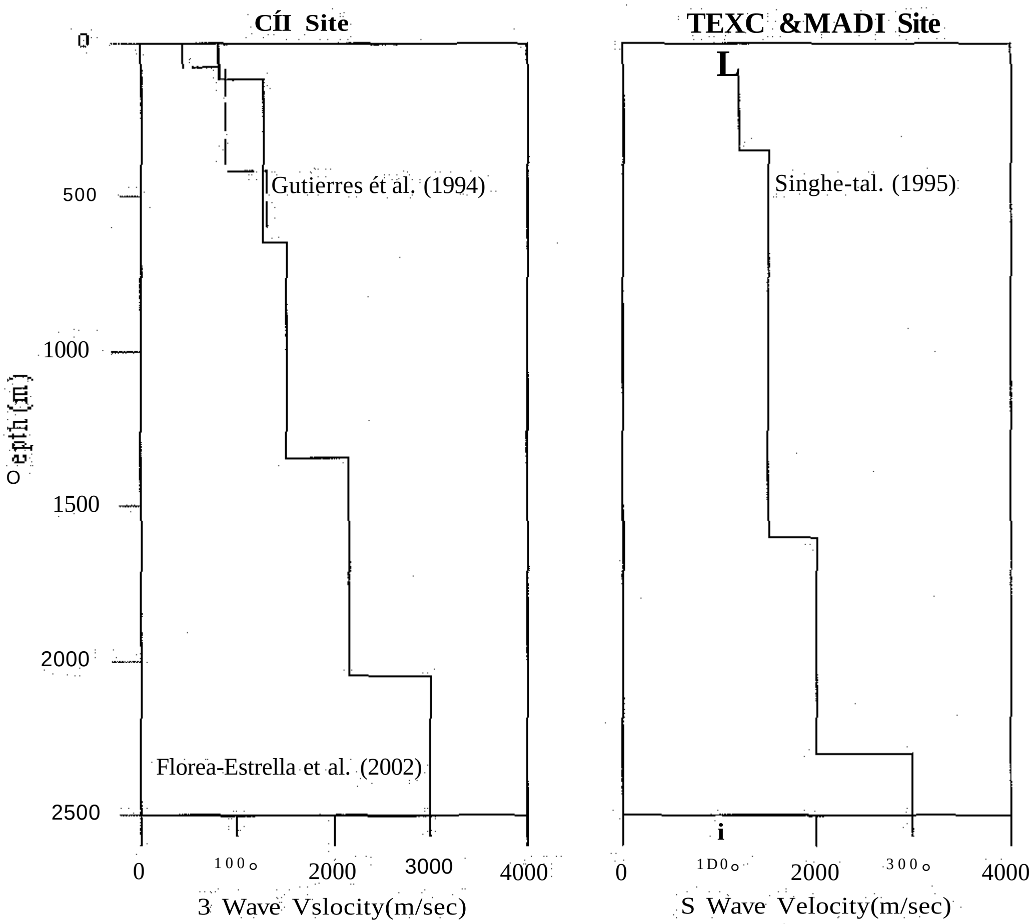

CÍI Site
TEXC &MADI Site
Gutierres ét al. (1994)
Singhe-tal. (1995)
Florea-Estrella et al. (2002)
epth (m)
0
500
1000
1500
2000
2500
0
2000
3000
4000
3 Wave Vslocity(m/sec)
S Wave Velocity(m/sec)

Figure 13

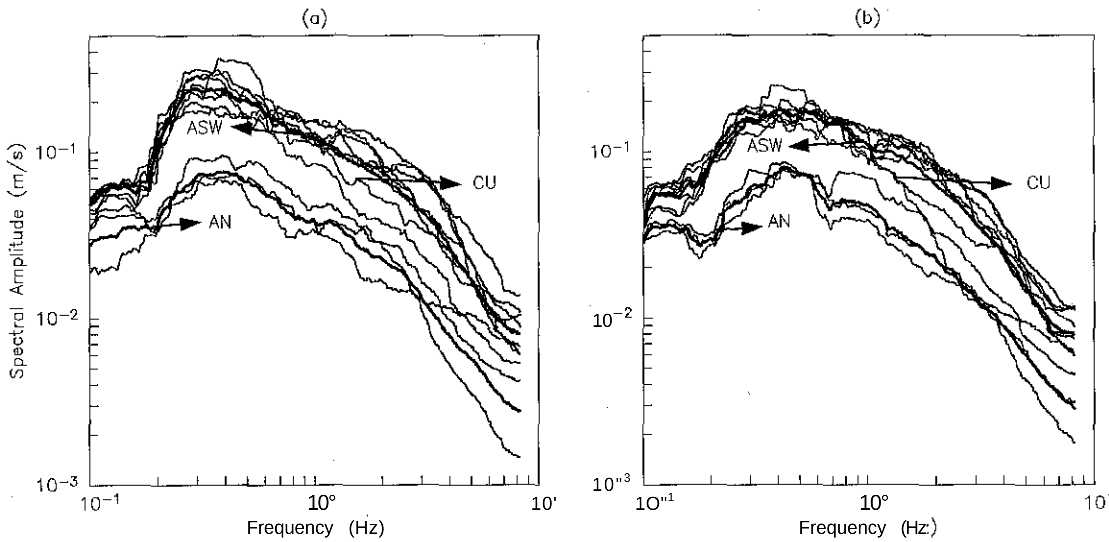


Figure 14

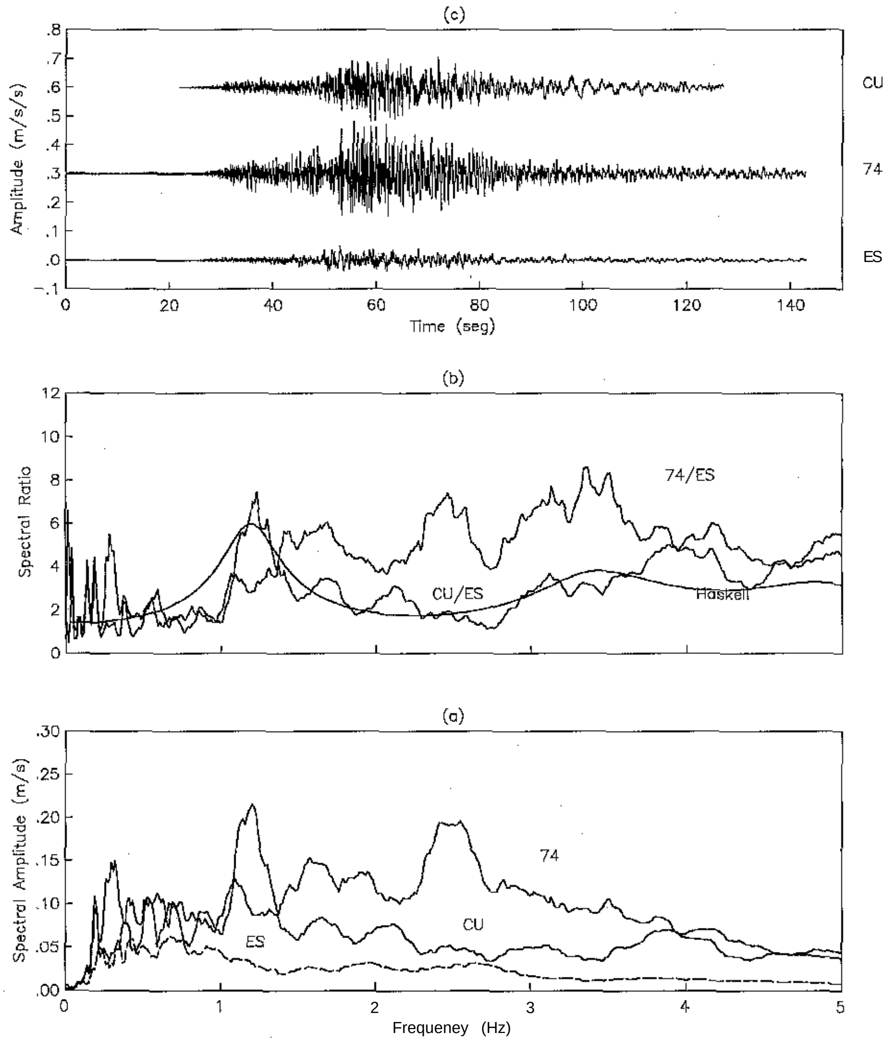

(c)
Amplitude (m/s/s)
.8
.7
.6
.5
.4
.3
.2
.1
.0
-.1
CU
74
ES
0
20
40
60
80
100
120
140
Time (seg)
(b)
Spectral Ratio
12
10
8
6
4
2
0
74/ES
CU/ES
Haskell
(a)
Spectral Amplitude (m/s)
.30
.25
.20
.15
.10
.05
.00
74
CU
ES
0
1
2
3
4
5
Frequeney (Hz)

Figure 15

35

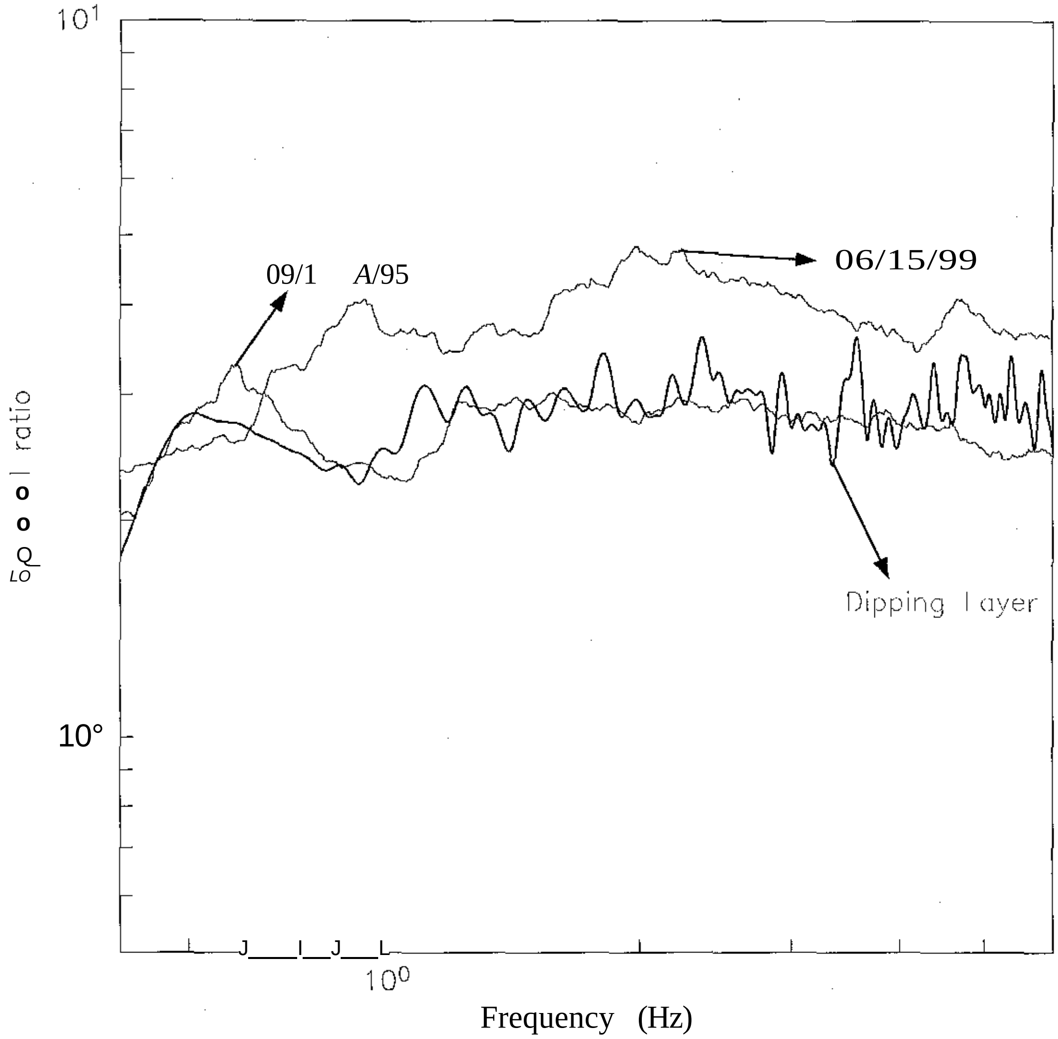


Figure 16

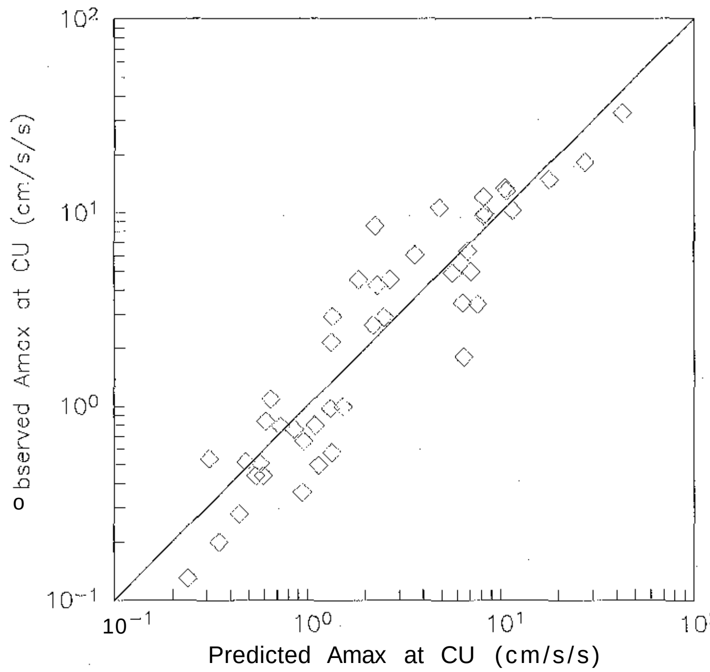


Figure 17

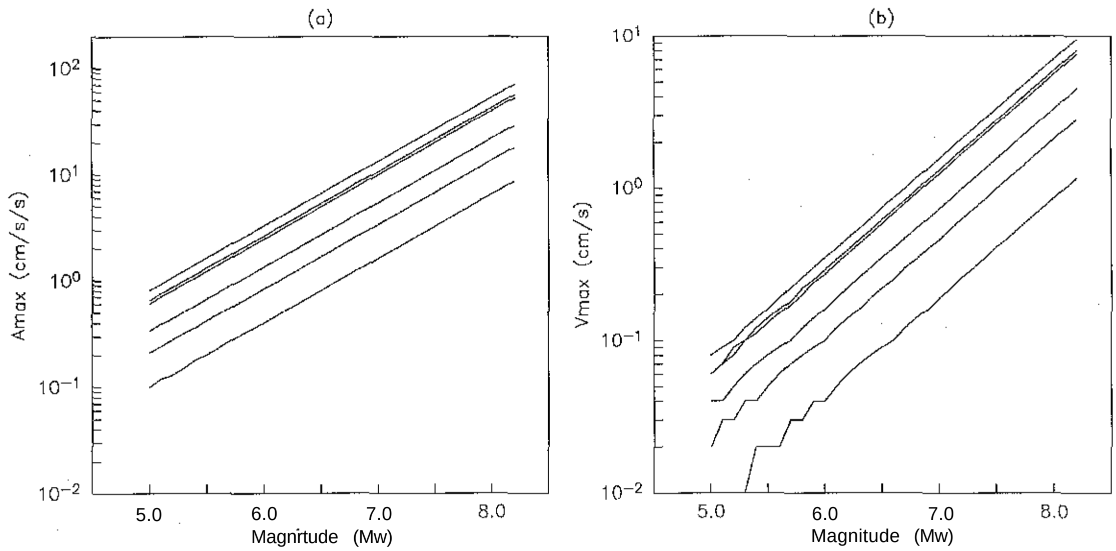
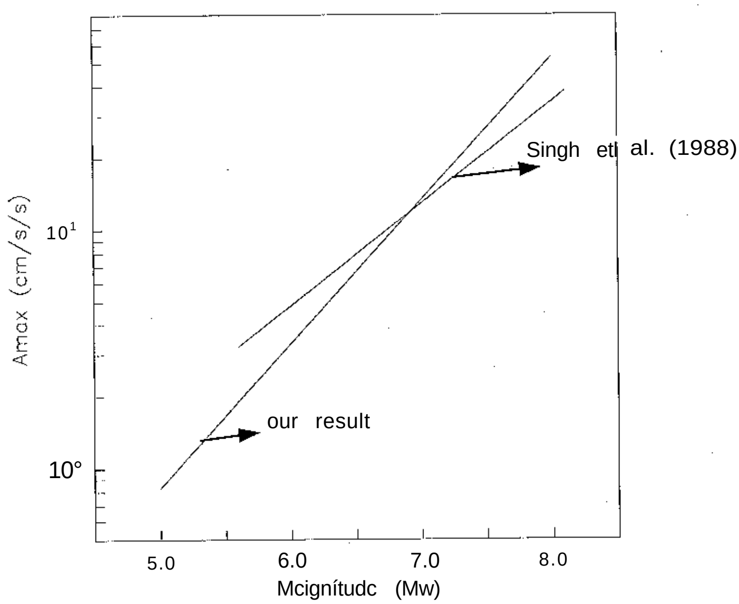

(a)
Amax (cm/s/s)
10^2
10^1
10^0
10^{-1}
10^{-2}
5.0
6.0
7.0
8.0
Magnitude (Mw)
(b)
Vmax (cm/s)
10^1
10^0
10^{-1}
10^{-2}
5.0
6.0
7.0
8.0
Magnitude (Mw)

Figure 18

38



Singh et al. (1988)
our result
Amax (cm/s/s)
10¹
10°
5.0
6.0
7.0
8.0
Mcignítudc (Mw)

Figure 19

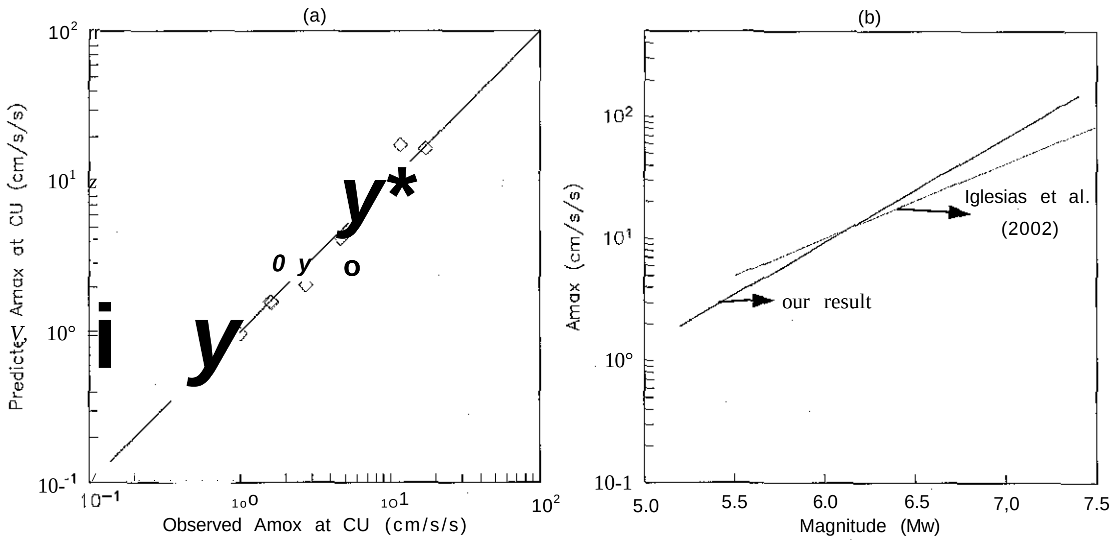
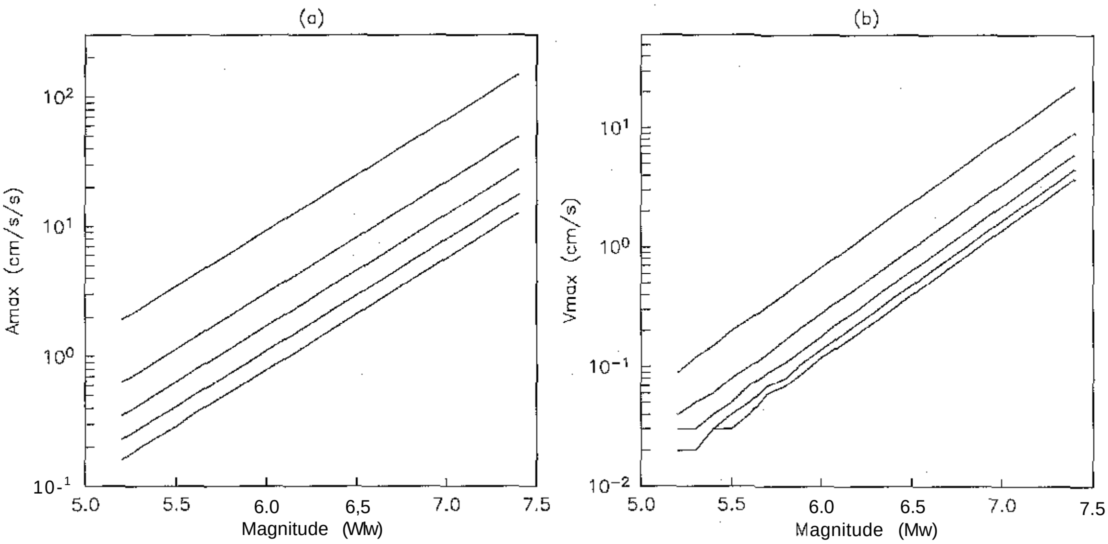

(a)
Predicted Amax at CU (cm/s/s)
Observed Amox at CU (cm/s/s)
(b)
Amax (cm/s/s)
Iglesias et al.
(2002)
our result
Magnitude (Mw)
5.0
5.5
6.0
6.5
7,0
7.5

Figure 20

40



(a)
Amax (cm/s/s)
10^2
10^1
10^0
10^{-1}
5.0 5.5 6.0 6.5 7.0 7.5
Magnitude (Mw)
(b)
Vmax (cm/s)
10^1
10^0
10^{-1}
10^{-2}
5.0 5.5 6.0 6.5 7.0 7.5
Magnitude (Mw)

Figure 21

41

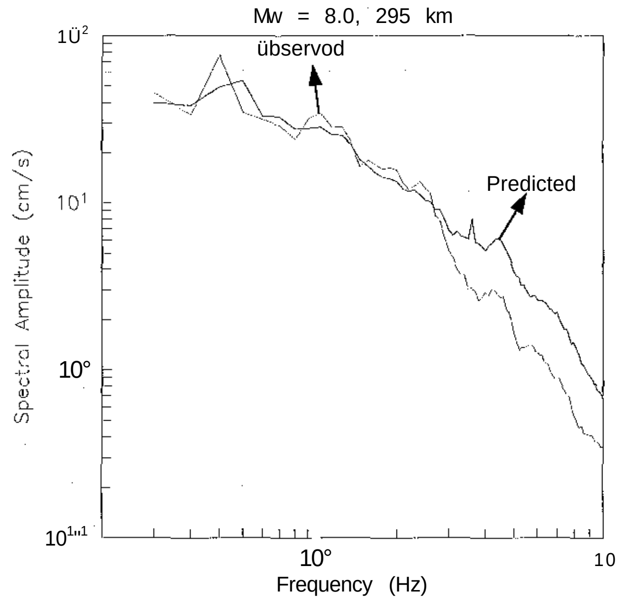
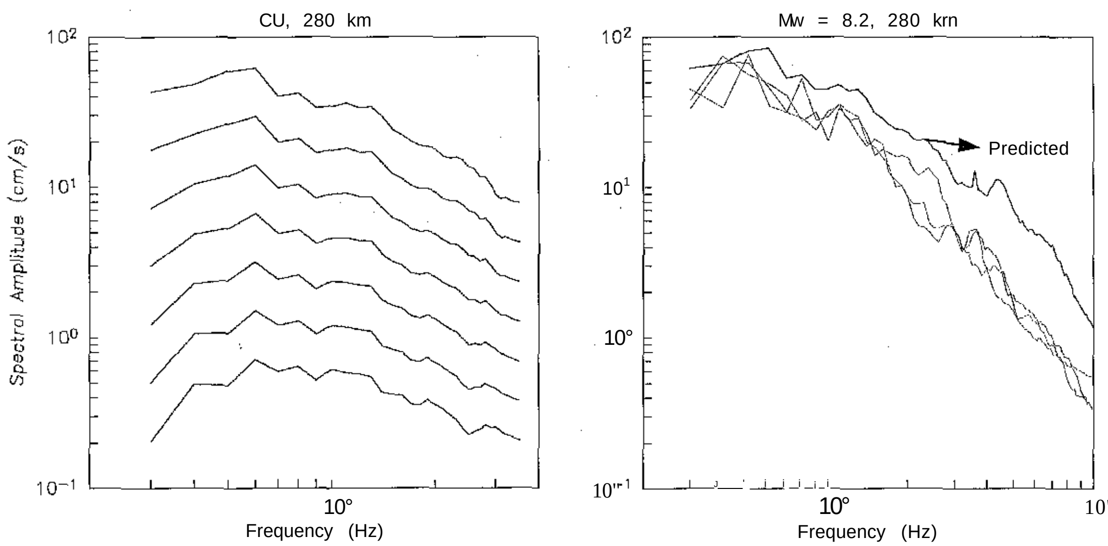

Mw = 8.0, 295 km
übservod
Predicted
Spectral Amplitude (cm/s)
Frequency (Hz)

Figure 22



CU, 280 km
Spectral Amplitude (cm/s)
Frequency (Hz)
Mw = 8.2, 280 krn
Predicted
Frequency (Hz)

Figure 23

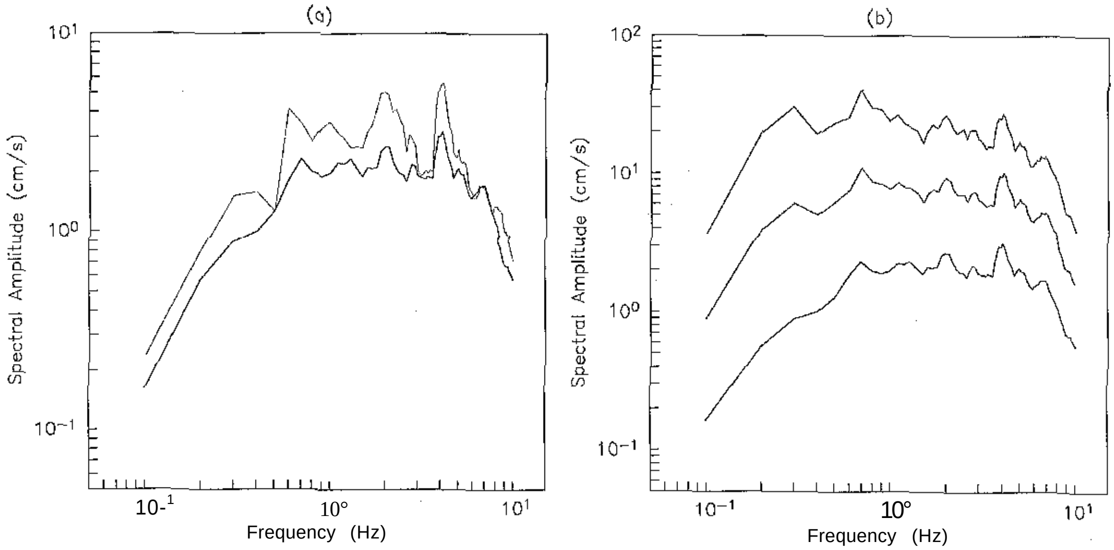

(a)
(b)
Spectral Amplitude (cm/s)
Frequency (Hz)
10-1
10°
10¹
10⁰
10²
10⁻¹

Figure 24

6/c

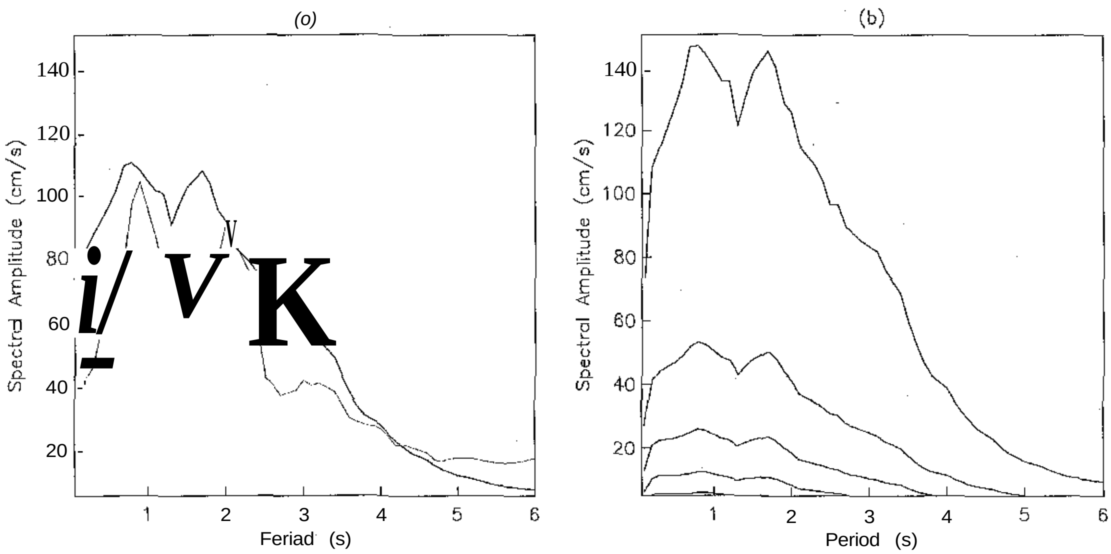

(o)
(b)
Spectral Amplitude (cm/s)
140
120
100
80
60
40
20
1
2
3
4
5
6
Feriad (s)
Period (s)

Figure 25

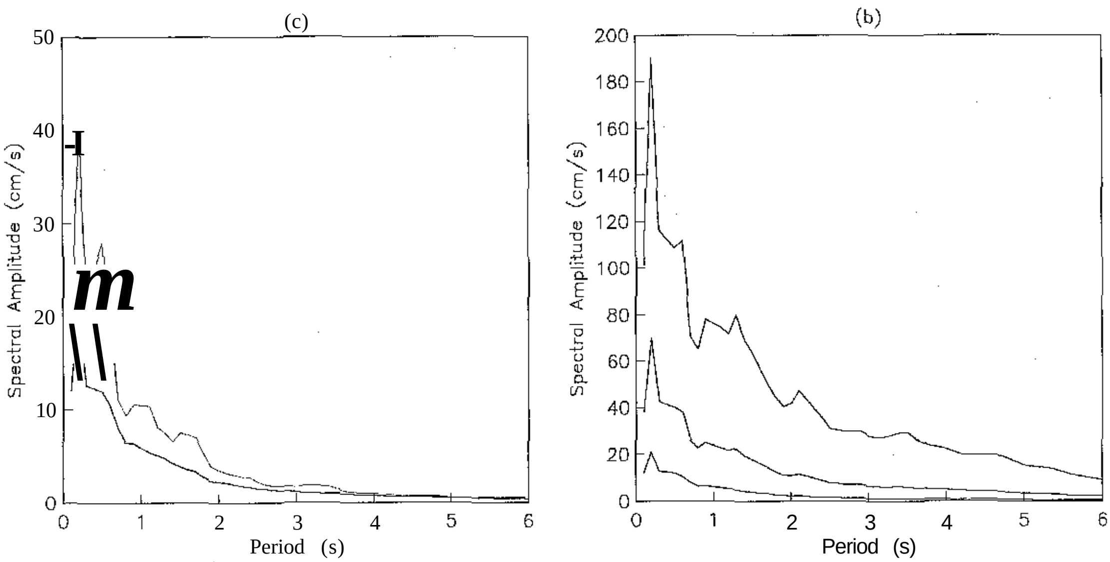

(c)
Spectral Amplitude (cm/s)
Period (s)
(b)
Spectral Amplitude (cm/s)
Period (s)

Figure 26

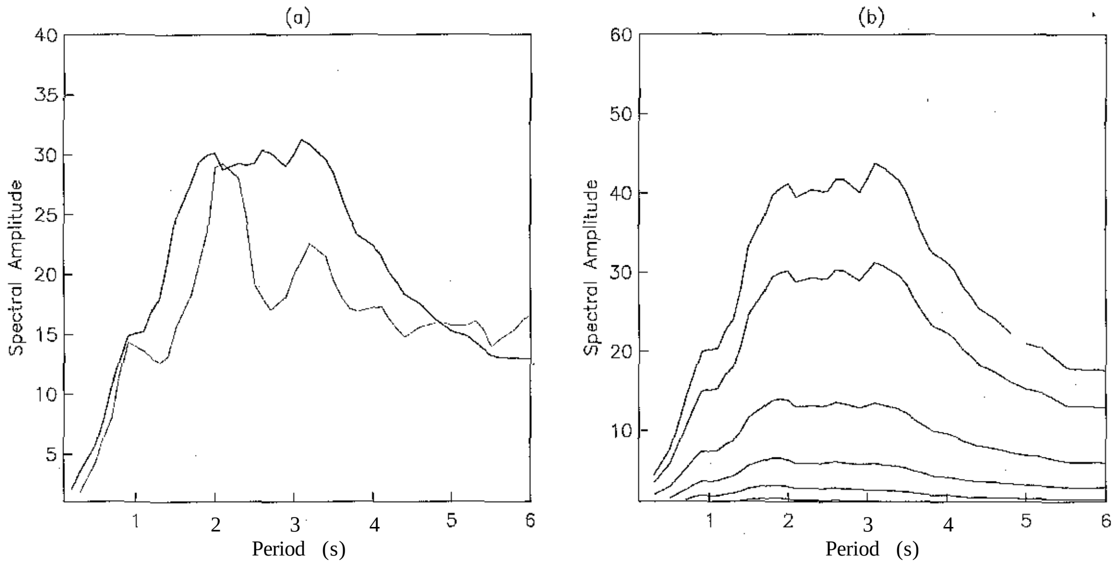
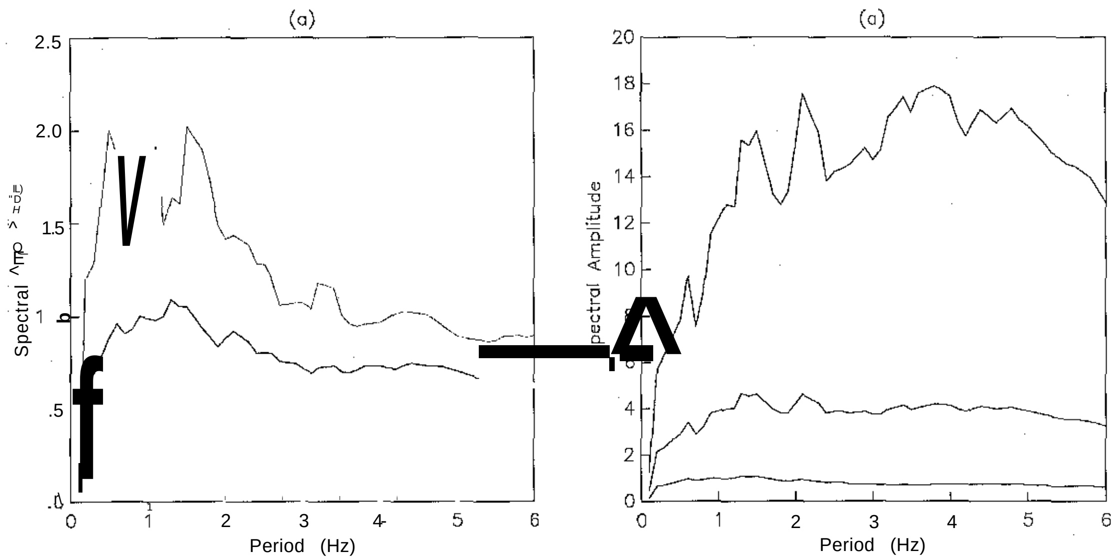
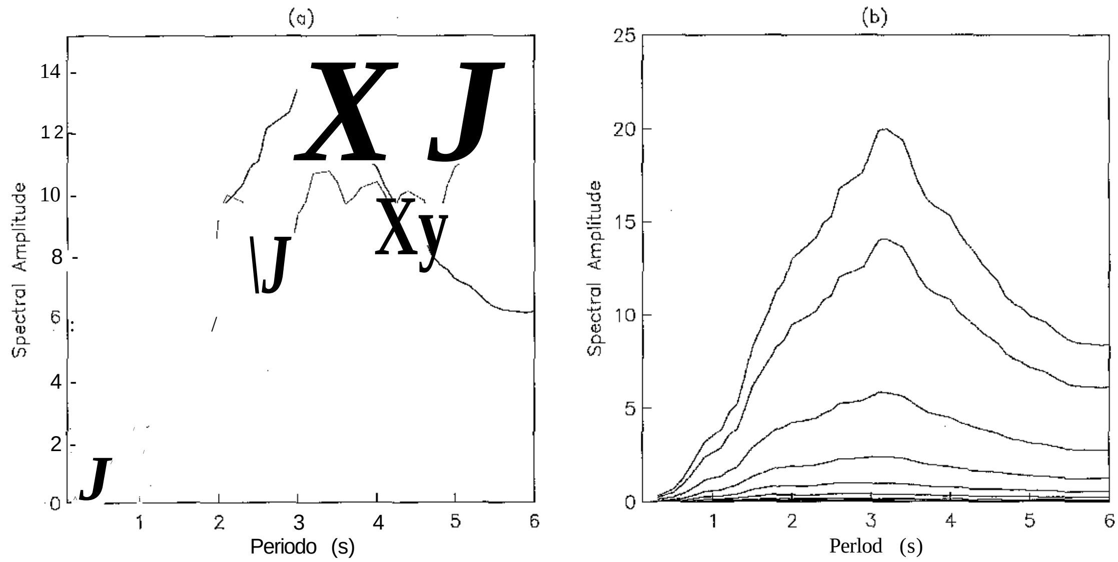

(a)
(b)
Spectral Amplitude
Spectral Amplitude
Period (s)
Period (s)
40
35
30
25
20
15
10
5
60
50
40
30
20
10
1
2
3
4
5
6

Figure 27



(a)
Spectral Amplitude
Period (Hz)
(a)
Spectral Amplitude
Period (Hz)

Figure 28



(a)
14
12
10
8
6
4
2
0
Spectral Amplitude
1
2
3
4
5
6
Periodo (s)
(b)
25
20
15
10
5
0
Spectral Amplitude
1
2
3
4
5
6
Perlod (s)

Figure 29

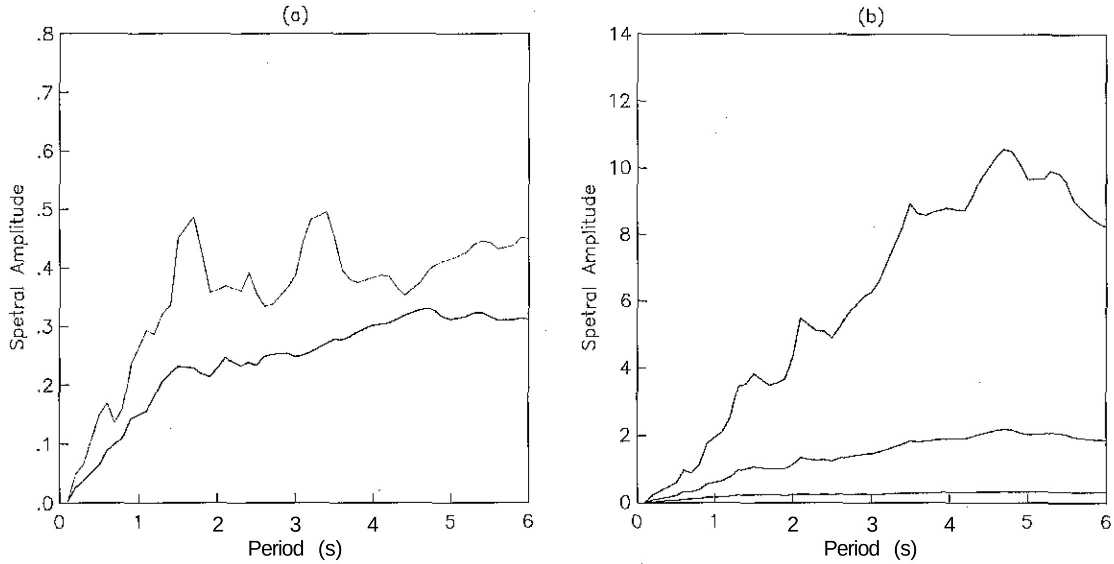

(a)
(b)
Spetral Amplitude
Period (s)
.8
.7
.6
.5
.4
.3
.2
.1
.0
0
1
2
3
4
5
6
14
12
10
8
6
4
2
0

Figure 30

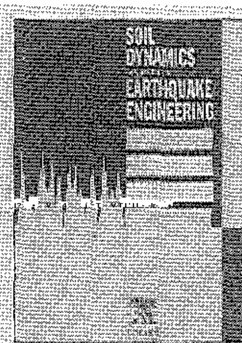
Event	Date	Mw		Amax		Vmax	Vmax	Lat. N.	Lon. W	Depth	RtoCU
	dd/mm/yy		n-s	V	e-w	n-s	e-w			(km)	(km)
1	230865	7.4	5.73	3.18	6.39	1.67	1.48	16.28	96.02	16	466
2	30268	5.9	7.73	0	9.36	1.79	1.51	16.67	99.39	16	297
3	20868	7.3	14.88	7.99	10.71	4.31	3.1	16.25	98.08	16	326
4	230475	6.2	2.9	1	2.9			16.47	98.86	16	320
5	10276	5.6	4.81	0	4.25	0.86	0.8	17.03	100.3	16	282
6	70676	6.4	13.45	3.74	6.66	2.76	1.08	17.2	100.88	16	292
7	190378	6.6	4.02	4.51	5.79	0.84	0.93	16.856	99.9	16	285
8	291178	7.6	5.06	0	7.43	1.65	2.12	16	96.69	18	414
9	260179	6.6	4.9	1	4.9			17.25	101	16	300
10	140379	7.6	19.45	8.84	17.02	3.89	3.25	17.46	101.46	20	287
11	251081	7.4	13.4	7	13.4	2.5	2.5	17.75	102.25	20	339
12	70682	6.9	11.86	4.35	6.79	3.41	1.33	16.35	98.37	20	304
13	70682	6.9	7.92	4.09	11.49	2.14	1.58	16.4	98.54	15	303
14	190985	8	31.4	21.06	34.07	9.31	9.45	18.14	102.71	16	295
15	210985	7.6	14.8	7	14.8	3.19	3.9	17.62	101.82	20	318
16	300486	6.9	4.54	2.65	4.54	1.43	1.15	18.2	103.1	16	409
17	80288	5.8	2.87	1.91	2.39	2.49	2.42	17.49	101.16	19	289
18	250489	6.9	13.4	5.74	10.53	4.63	2.95	16.58	99.48	17	304
19	110590	5.5	2.39	1.44	1.91	1.24	0.92	17.24	100.56	12	295
20	310590	5.9	5.07	1.34	3.21	0.84	0.45	16.77	100.12	16	295
21	310393	5.5	0.88	0.44	1.07	0.15	0.1	17.29	100.89	26	295
22	150593	6	2.78	1.38	3.08	0.34	0.21	16.45	97.92	20	349
23	241093	6.6	3.29	1.72	3.49	0.794	0.81	16.77	98.61	22	280
24	131193	5.7	0.42	0.2	0.29	0.134	0.142	16.28	98.61	17	344
25	131294	5.5	0.5	0.16	0.38	0.082	0.074	16.25	98.12	15	362
26	140995	7.3	12.42	3.8	7.65	3.32	1.89	16.73	98.54	22	320
27	301095	5.6	0.8	0.39	0.48	0.196	0.095	16.55	98.13	16	330
28	130396	5.1	0.56	0.29	0.43	0.077	0.057	16.93	98.86	29	269
29	200396	5.3	0.22	0.12	0.18	0.044	0.04	16.15	97.72	15	389
30	180796	5.4	0.65	0.53	0.92	0.07	0.115	17.35	101.002	26	298
31	161297	5.9	1.09	0.59	0.89	0.221	0.182	16.43	98.73	16	325
32	91095	8	2.22	1.92	1.28	1.03	0.75	19.34	104.8	15	515
33	121095	5.9	0.55	0.33	0.53	0.165	0.133	18.81	104.07	20	545
34	190396	5.8	0.55	0.19	0.3	0.165	0.142	16.4	96.79	15	419
35	270396	5.5	0.93	0.56	0.73	0.206	0.162	16.44	97.72	21	360
36	230496	5.5	0.54	0.24	0.5	0.096	0.135	17.13	101.84	20	383
37	150796	6.6	3.66	1.91	3.2	0.479	0.409	17.55	101.12	22	292
38	160197	5.5	0.29	0.17	0.27			18.23	102.55	25	393
39	210197	5.5	0.9	0.4	0.66			16.49	97.99	18	342
40	30298	6.3	1.05	0.49	1.13	0.255	0.263	15.92	96.22	24	501
41	160598	5.2	0.52	0.33	0.5	0.088	0.056	17.56	101.54	25	327
42	110798	5.4	0.81	0.35	0.71	0.119	0.107	17.28	101.17	24	317
43	120798	5.5	0.66	0.39	0.49	0.14	0.093	16.78	99.91	15	294
44	40499	5.4	0.12	0.05	0.14	0.049	0.076	15.89	96.94	15	456

ivent	Date	Mw		Amax		Vmax	Vmax	Lat. N.	Lon. W	Depth	R to CU
	dd/mm/yy		n-s	v	e-w	n-s	e-w			(km)	(km)
1	60764	6.7	18.3	11.99	15.74	2.12	2.4	18.03	100.77	55	179
2	50893	5.2	0.54	0.35	0.46	0.04	0.04	17.429	98.337	54	229
3	230294	5.8	1.04	0.62	0.95	0.07	0.07	17.75	97.27	75	267
4	60594	5.2	0.44	0.4	0.54	0.03	0.02	18.39	97.98	57	163
5	230594	6.2	4.97	2.88	4.28	0.41	0.4	18.02	100.57	50	206
6	101294	6.4	5.39	2.63	5.48	0.94	0.81	17.98	101.52	50	288
7	110197	7.1	4.15	3.08	5.86	1.99	2.11	18.37	101.82	35	374
8	30497	5.2	1.2	2.35	1.96	0.06	0.12	18.51	98.1	52	145
9	220597	6.5	1.48	1.65	2.08	0.29	0.32	18.37	101.82	54	298
10	200498	5.9	1.67	1.33	1.5	0.16	0.2	18.35	101.19	60	238
11	150699	7	11.4	7.51	11.85	2	1.72	18.13	97.54	61	218
12	210699	6.3	2.39	1.6	3.06	0.31	0.36	18.15	101.72	53	296
13	300999	7.4	7.66	5.12	7.8	2.21	2.72	16.03	96.96	47	435
14	210700	5.9	13.09	5.97	12.48	0.9	0.69	18.11	98.97	50	136

APÉNDICE V

HYBRID INDIRECT BOUNDARY ELEMENT - DISCRETE WAVE NUMBER METHOD APPLIED TO SIMÚLATE THE SEISMIC RESPONSE OF STRATIFIED ALLUVIAL VALLEYS

(Soil Dynamics and Earthquake Engineeringn aceptado)



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Dear Paco:

Enclosed please find the review comments concerning your paper "A hybrid indirect boundary element discrete wave number method….", submitted to SDEE for possible publication.

As you can see both reviewers are positive and recommend publication of your paper in [illegible]ion conforming to their comments. Thus I would appreciate receiving in triplicate a revised version of your paper in accordance with the comments of the reviewers together with your detailed response to those comments.

Thank you for your co-operation

Best regards

Dimitri

D. Beskos
Professor
Associate Editor : SDEE

A Hybrid Indirect Boundary Element - Discrete Wave Number Method Applied to Simúlate the Seismic Response of Stratified Alluvial Valleys

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Submitted to: Soil Dynamics and Earthquake Engineering

Revised versión

May, 2002

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ABSTRACT

A hybrid indirect boundary element-discrete wave number method is presented and applied to compute the ground motion on stratified alluvial valleys under incident plañe *SH* waves from an elastic half-space. The method is based on the single-layer integral representation for the waves diffracted into the half-space while refracted waves in the horizontally stratified región are expressed as a linear superposition of solutions for a specified set of discrete wavenumbers. These solutions are obtained in terms of the Thomson-Haskell propagators formalism. Boundary conditions of continuity of displacements and tractions along the common boundary between the half-space and the stratified región lead to a system of equations for the sources strengths and the coefficients of the plañe wave expansión. An example is given for a two-dimensional (2D) anti-plane problem of diftraction *ofSH* waves by a soft stratified inclusión in an elastic half-space. These results show the significant influence of locally generated surface waves in seismic response.

Key words: Indirect boundary element method; Discrete wave number method; 2D model; SH waves; Horizontal layers; Alluvial valleys; Seismic response.

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INTRODUCTION

Local geological conditions may genérate significant amplification of ground motion and concentrated damage during earthquakes. Tliere are important advances in the evaluation of such effects (e.g. Sánchez-Sesma [1]; Aki [2] and Sánchez-Sesma [3]). In most situations local amplification can be inferred reasonably using simple one-dimensional (ID) shear models. However, lateral heterogeneity may give rise to focusing and to locally generated surface waves; therefore, the estimates for local amplification using ID models may be wrong. Although the physics of this phenomenon is now reasonably well understood, most of the rigorous procedures require large amounts of computer resources. We believe there is a lack of practical methods for the assessment of local amplification.

Boundary Element Methods (BEM) have gained increasing popularity. Recognized advantages over domain approaches are the dimensionality reduction, the relatively easy fulfillment of radiation conditions at infinity and the high accuracy of results. Excellents surveys of the available literature on BEM in elastodynamics are by Manolis and Beskos [4]; Beskos [5,6]. One useful approach is the Indirect Element Methods IBEM, which formulates the problem in terms of forcé densities at the boundaries (e.g. Sánchez-Sesma and Campillo [7]; Sánchez-Sesma et al. [8]; Sánchez-Sesma and Luzón [9]).

On the other hand, in their pioneering work, Aki and Lamer [10] introduced a numerical method based on a discrete superposition of homogeneous and inhomogeneous plañe waves. Using this technique, Bard and Bouchon [11,12] studied alluvial valleys and pointed out the significant role of basin-induced surface waves in the valley's response and studied the resonant characteristics of these configurations as well. The Aki-Larner's technique is the departure of discrete wavenumber approximations. The combination of discrete wavenumber expansions for Green's functions with boundary integral representations has been successful in various studies of elastic wave propagation (e.g. Bouchon et al. [13]; Kawase and Aki [14]; Papageorgiou and Kim [15]).

The combination of a BEM-type approach with a discrete wavenumber formulation to compute the Green's function is particularly attractive: the singularities of Green's functions are not present in each one terms of the discrete wavenumber expansión (e.g. Kawase [16]; Kawase and Aki [14]). The integration along the boundary effectively malees the singularities to vanish and improves convergence as well. However, such procedures require considerable amount of computer resources. An alternative that may be welcomed for some applications could be the IBEM, with analytical Green's functions for the half-space, and a discrete wavenumber expansión for the layered part. A similar idea, that uses Une sources cióse to the boundary, propagators for the layered part and the least-squares method, was proposed by Bravo et al. [17].

In this work we extend the Bravo et al. [17] approach by placing the sources at the very boundary (e.g. Sánchez-Sesma and Campillo [7]) so that the singularities of the sources are dealt with analytically. Therefore, the IBEM is applied for the half-space. The sources strengths and the coefficients for the discrete wavenumber expansión equal in number the conditions (continuity of displacements and tractions) at the collocation points in the interface.

FORMULATION OF THE PROBLEM

Consider a horizontally stratiñed deposit in a homogeneous half-space. Elastic materials oceupy the entire domain. With reference to Figure 1, we desígnate by *S* the boundary between the región *R*, which is the stratified médium, and the exterior domain *E*. In Figure 1 we present also the cross-section of the stratified model and the hamionic incident Sffwave. In order to compute the displacement ííeld at the free surface, we treat the two regions separately. Continuity of displacement and traction at *S* will then be imposed.

Domain E

As far as the domain *E* is concerned, the ground motion comes frora the interferences of incoming waves with reflected and diffracted ones. The total motion in the half-space is the superposition of the free-field and the so called diffracted waves:

$$v^{E} = v^{(0)} + v^{(d)} \tag{1}$$

where u<> = free fíeld displacement, *i.e.* the solution in the absence of the irregularity. The free-field displacement at a point x = *(x,z)* is

$$v^{(0)}(\mathbf{x}) = e^{-i(kx-\eta z)} + e^{-i(!a+uz)} \tag{2}$$

where *k* = *(,co/¡3)* sin *y* and *r¡*— (OJ/JS) eos *y* are the horizontal and vertical wavenumbers, respectively, *a)* = angular frequeney, *p* = shear waves velocity in the domain *E*, *y* = incidence angle. We are dealing with antiplane *SH* waves, *i.e.* all of the displacement are perpendicular to the *x* - *z* plañe.

Let's discretize the surface *S* by JV straight segrnents of longitude A5' and cali ^ = (x/,z/) the midpoint of the segment /. Neglecting body forces, the diffracted field can be written as

$$v^{(d)}(\mathbf{x}) = \sum_{l=1}^{N} \phi(\xi_l) g(\mathbf{x},\xi_l), \quad (3)$$

where 0(4/) = unknown forcé density assumed to be constant (<\/>) along the corresponding *AS;*

$$g(\mathbf{x},\xi) = \int_{AS_l} G(\mathbf{x},\xi) dS_\xi \quad (4)$$

where *G(x,Q* = Green's function of a line source in a half-space, *i.e.* the displacement at point x due to the application of two unit forces at point £ = *(xt,z¡)* and at its irnage ï;' = *{xu~z¡)*. Therefore, one has

$$G(\mathbf{x},\xi) = \frac{1}{i4\mu}\left[H_0^{(2)}(r_l \tfrac{\omega}{\beta}) + H_0^{(2)}(r'_l \tfrac{\omega}{\beta}) \right], \quad (5)$$

where *Hfp(»)* = Hanlcel's function of the second kind and zero order, r_l and r'_l are the distances between the point x and the sources *%¡* and its image §J, respectively. The use of an image source guáranteos that the boundary conditions of nuil tractions at the half-space free surface are automatically satisfied in equation (3).

The integral in equation (4) is performed on the variable £; Gaussian integration is used. G(x,£) presents a logarithmic integrable singularity at x = §; where x is in the neighborhood of §, the second right term of equation (5) is treated numerically, while for the first one we obtained the analytical expressions from the ascending series for Hankel's functions (Abramowitz and Stegun [18]). Sánchez-Sesma and Campillo [7] presented an example for such expressions where only the leading terms of the series are retained. We consideredup to quadratic terms, which is enough if the number of segments per wavelength is larger than about ñve.

The quantities <fi(£,i)AS represent a forcé distribution at tlie boundary. Equation (3) is simply the discretized form of a single layer integral, that can be obtained from Somigliana identity (Sánchez-Sesma and Campillo [7]), and has been studied by Kupradze [19] from the point of view of potential theory. He showed that the displacement field is continuous across *S* if <Ӌ>(£) is continuous along *S*. The scalar 2D versión that we are using is also called the Kirchhoff-Helmhotz representation (Kouoh-Bille et al. [20]).

Likewise the displacement ñeld, we express the tractions in the half-space as the superposition of the contributions from the free-field and the diffracted one:

$$t^E = t^{(0)} + t^{(d)}. \tag{6}$$

Since we are dealing with an antiplane problem the traction vector can be represented by a scalar quantity, when a vector n lying in the plañe is specified. In fact, applying Hooke's law we have:

$$t^E = \mu \frac{\partial v}{do}, \tag{7}$$

where n is assumed to be specified. Developing equation (7) we fmd Cauchy's equation:

$t^E = \mu \frac{\partial v}{\partial n} = \mu($**f fr + f t) = Mf», + f».) = *w* + w-*** **(8)**

Therefore, the free-field traction is expressed by:

$$t^{(0)}(\mathbf{x}) = [\sim ikn_x + hin\wedge e\text{-}\wedge w) \_ [¿kn_x + \wedge n_z\sim]e\text{-}^{i\{kx+}\wedge\backslash \tag{9}$$

which clearly displays in the two groups of terms: the incident and reflected waves, respectively. In analogous terms the diffracted tractions can be computed by means of:

$$t^{(d)}(\mathbf{x}) = \sum_{l=1}^{N} \phi(\xi_l) t(\mathbf{x}, \xi_l), \tag{10}$$

where/^{w}(x) = traction at point x, *(j)(¿;i)* = *<Þ/,* /(x,£) is givenby:

$$t(\mathbf{x}, \xi) = \int_S T(\mathbf{x}, \xi) dS_\xi, \tag{11}$$

where *T(x,E,)* = traction Green's function, *Le.* the traction at point x on the boundary *S* with normal n(x) (assumed to be specified and pointing outside *E)* due to the application of two unit forces at point £ and at its image §'. Its formal expression is

$$T(\mathbf{x}, \xi_l) = \frac{i\omega}{4\beta}\left[\frac{\partial r_l}{\partial \mathbf{n}} H_1^{(2)}(r_l \tfrac{\omega}{\beta}) + \frac{\partial r_l'}{\partial \mathbf{n}} H_1^{(2)}(r_l' \tfrac{\omega}{\beta}) \right] \tag{12}$$

were *Hf\>)* = Hankel's function of the second kind and first order. The integral in equation (11) is also computed numerically using Gaussian integration. The traction Green's function present a singularity of the form 1/r when x_y = ^$_7$, in this case we obtain

$$t(\mathbf{x}_j, \xi_l) = 0.5\delta_{jl} + \int_{\Delta S_l} \frac{i\omega}{4\mu} \left[\frac{\partial r_l}{\partial n} H_1^{(2)}(r_l \frac{\omega}{\beta}) \right] dS_{\xi'} \qquad j = l. \tag{13}$$

δ_{jl} = Kronecker symbol (= 1 if j = l, = 0 if j * l). The formal contribution from the source t_l to the integral is nuil because the discretization segment is a straíght Une and ξ_l is the midpoint. In fact, it can be verified that, under this circumstance, such part of the integrand is a singular odd tunction on the segment. Therefore, its Cauchy's principal valué is zero. The contribution of singularity has been already taken into account in the fírst term of equations (11) and (13). This result has also been found by Kupradze [19]. In its scalar versión, the result appeared first in a paper by Fredholm in 1900 (Webster [21]).

Domain R

The ground motion in the regíón R is generated by waves refracted at S and at the interfaces of the layers. For the solution in the stratified médium, we combine Thomson-Haskell's propagator matrices (Thomson [22]; Haskell [23]) in the framework of a discrete wavenumber representation. In this way, the refracted field in the regíón R can be written as:

$$v^{(r)}(\mathbf{x}) = \sum_{m=-M}^{M} B_m l_1(k_m, z, \omega) e^{-ik_m x} \tag{14}$$

where B_m = unknown complex coefficients and $l_1(k_m,z,\omega)$ = first element of the motion-stress vector for Love waves (Aki and Richards [24]) for the horizontal discrete wavenumber k_m. The function $l_1(k_m,z,\omega)e^{-ik_m x}$ is solution of the 2D Helmholtz equation within the layered system. If, for example, we make

$$v(\mathbf{x}) = h(k,z,co)e^{-lkx}, \tag{15}$$

and define the traction at horizontal planes as:

$$\mu \frac{\partial v}{\partial z} = l_2(k,z,\omega)e^{-ikx}, \tag{16}$$

where *h* and *h* are continuous functions of z. Then it can be shown (Aki and Richards [24]) that, *in* a médium composed of horizontal homogeneous layers, the motion-stress vector at depth *z* can be written in terms of the propagator matrix and the motion-stress vector at depth zo as

$$\begin{bmatrix} U \\ h \end{bmatrix}_z = \mathbf{P}(z,z_0)\begin{bmatrix} \\ h \end{bmatrix}_{z_0}. \tag{17}$$

The propagator matrix for a homogeneous médium is given by

$$\mathbf{P}(z,z_0) = \begin{bmatrix} \text{cosí](z-z}_0\text{)} & \text{OííO}^{-}\text{ 'sin/j/z-zo)} \\ \text{--}r_j f.tsinr_j j(z\text{~}zo) & cosr](z - ZQ) \end{bmatrix}, \tag{18}$$

where $r_j = \overline{fer\text{-}^{2}/\beta^2 - k^2}$.

From the repeated application of equation (17), if we assume that zo = 0 corresponds to the free surface, we have

$$P(z,Z_0) = P(z,Z^\wedge t)P(zy_{-1},Z_{/\cdot 2})\ldots P(z,,í_0)_í \quad \textbf{(19)}$$

for $ZJ \geq z \geq z/$-ithat is, for z in *thejth* stratum. The form of equation (19) guarantees satisfaction of continuity of stress and displacement between adjacent layers. To satisfy the boundary condition at the free surface, we make $h = 0$ at $z = 0$. Without loss of generality, we can choose $U = 1$ at the free surface.

The displacement is directly related to the first component of the motion-stress vector (see equation (15)), while the traction associated to a vector n $= (n_x,n_z)^T$, considering equations (15) and (16), can be expressed as

$$t = fi\text{-}^\wedge = ft(jj_i;n_x + \text{f-Wz}) = -ipikhn_xe^{\sim ikx} + hn_ze^{\sim ik}\backslash \quad (20)$$

Coupling regions E and R

To guarantee perfect bonding between regions E and R we impose continuity of displacement and tractions, *Le.*

$$v^{(r)} = y(0) + v^{(d)} \qquad \text{at } S\text{, and} \quad (21)$$

$$t^{(r)} = t^{(0)} + t^{(d)} \qquad \text{at } S \quad (22)$$

These conditions can be expressed for a point x on (the common interface) S by means of

$$\sum_{l=1}^{N} Ti\wedge isfr,\xi_l) - \sum_{m=-M}^{M} B_m l_1(k_m,z,\omega)e^{-ik_m x} = -v^{(0)}(\mathbf{x}),\text{ and} \tag{23}$$

$$\sum_{l=1}^{N} \phi_l t(\mathbf{x},\xi_l) + \sum_{m=-M}^{M} B_m(-i\mu n_x l_1(k_m,z,\omega) + n_z l_2(k_m,z,\omega))e^{-ik_m x} = -t<\!\!\circledR(\mathbf{x}) \tag{24}$$

respectively. If we impose these conditions on a set of points along the boundary S we can speak of a collocation scheme. Assume we chose x = (XJ,ZJ), $j = l,...,N$, evenly distributed on S. Then in order to obtain a square system of equation we set $N=2M+1$. ít becomes clear that the total number of equations is 27V which is the same as the number of unknowns.

The choice of TVis guided by the following considerations. In the región E it is convenient to have at least fíve elements per wavelength. This criterion comes frotn previous experiences using IBEM (Sánchez-Sesma and Campillo [7]) and it is somewhat conservativa For shallow valleys one can approximately obtain $N > 5r_j$, where r_j = $((üá)l\{itjÍE)$ is the normalized frequency and a the half-width of the valley. If the sediments were to be treated with IBEM too, then an estímate of the lower bound for N in the layered región could be $N > 5rj(fÍE/pmm)_3$ which, for valleys with very soft layers may produce very large N and this gives unnecesarily dense sampling for both the half-space and the valley. This statement is clear of one considers the global nature of the discrete wavenumber expansión which is given in terms of $e^{\sim lkmX}$. Moreover, it is convenient to assume that $Ák$ is less than $nía.)$ to avoid spurious periodicities, and that only homogeneous waves are allowed in the softer layer (this means k_{miX} = co/pmm and implies that inliomogeneous waves can exist in the other layers). Then, we have $M > T_j(PE/Pm\backslash n)$, which gives A^> $2ri(f\$EfPmm)$ + 1- Obviously, this valué is less restrictive but still it can very easily dominate the choice of N. If the amount of allowed inhomogeneous waves in the softer layer is increased (this means $k_{ms\cdot x}$ > $co/pmm)$, N becomes larger.

For low frequencies, oversampling is not a serious problem. Actually, it is convenient and the solution is very stable. When frequency increases the different character of the involved representations requires care. A modérate amount of inhomogeneous waves may give good results for intermedíate frequencies. Sometimes imdersampling is a good solution. However, all this requires fiirther scrutiny. The formulation for the *P-SV* waves case, would be part of another research.

TESTING OF THE METHOD

The accuracy of our approach is tested here using a solution that has been verified extensively (Kawase and Aki [14]). These authors used direct boundary elements with the Green's function computed using the discrete wavenumber method. Their results have been tested by Ramos-Martínez [25] and Zahradnik [26], among others. Therefore, we regard these results as trustwortby.

Kawase and Aki [14] studied the trapezoidal valley depicted in Figure 2. Comparisons with their results are provided here for vertical incidence of plañe *SH* waves. The model response is computed in frequency domain and the synthetic seismograms are computed using the FFT algorithm. It was assumed an incident Ricker wavelet. The comparison in time domain is a good test as many frequencies have to be computed and synthesized to provide the results. The synthetic seismograms obtained from both methods for vertical incidence are displayed in Figures 3 and 4 for characteristic periods t_p = 4 and 2 *s*, respectively. The agreement is excellent. We computed also the synthetics for an incidence angle of 30 degrees with respect to the vertical and Figure 5 display the seismograms at the free surface for characteristic periods t_p = 4 and 2 *s*, respectively. These results clearly display the great significance of locally generated surface waves at the valley's edges.

We compared our approach also with the results by Bravo et al. [17]. These autbors studied the response of a stratified parabolic deposit depicted in Figure 6. Región *R* consists of two homogeneous strata, the top layer being 1. Máximum thickness are *H* = *a/6* and *Ui* = *a/3*, where *a* = half-width of the deposit. Shear-wave velocities and mass densities, referred to those of the half-space, are *p\/f$E* = 1/3, *f$i/pE* = 2/3 and *p\lpE* = 3/4, *pj/pE* = 0.85. Figures 7 and 8 shows the comparisons with their results. Three angles of incidence 7 = 0, 30 and 60 degrees were considered, and two normalized frequencies *r)* = 1.0 and 2.0, respectively. The agreement is excellent.

EXAMPLE

A Layered soft deposit

In order to illustrate a complet set of results we analyze the response of a stratifíed valley with irregular interface for three incident angles in both frequency and time domains. The model is depicted in Figure 9. The half-width is of 1 km and máximum depth is 0.35 km. The shear wave velocity assumed for the half-space is *fü* = *2* km/s, while the velocities of the four layers, from top to bottom, are 0.8, 0.6, 1.0 and 1.2 km/s, the corresponding thicknesses of the four layers are 50, 100, 100 and 100 m, respectively and the quality factor used are 50, 100, 150 and 200 respectively. Three angles of incidence *y* = 0, 30 and -30 degrees were considered. Figures 10, 11 and 12 display the surface displacement amplitudes for normalized frequencies *r*— 0.8, 1.0 and 1.5, respectively. Because of the properties of the model these frequencies have the same numerical valué in Hz. The various plots clearly illustrate the spatial variation of amplification and the effect of incidence angle.

In order to give a clearer idea of the response, Figures 13 to 15 show the contours of displacement amplitudes at surface receivers along the *x* axis, between *x* = -0.94 km and *x* = 0.94 km, against normalized frequency. These *f—x* diagrams display the transfer function (relative to the amplitude of incident waves) for all receivers and provide a good description of the frequency behavior of the valley, The plots also evince lateral resonances, which sometimes are clear when a series of peaks are distributed in space for a given frequency. This effect has beeri identified before (e.g. Sánchez--Sesma et al. [8]). The layering and the irregular profile tend to produce a complicated response. The resonant patterns vary with incidence angle. For instance, a resonant frequency of about 2.7 is clear for vertical incidence, but such a frequency is little excited by the oblique incidences.

The máximum valué of displacement amplitudes is of about 12 for a frequency of about 1 Hz. This amplification, relative to the horizontal free-fíeld surface displacement, is of about 6.5, nearly three times the average impedance ratio. On the other hand, the frequency is almost two times the ID shear resonant frequency for a fíat layer of the left side of the valley depth of 200 m). The same effects can be seen at the right side and can well be explained in terms increased stifíness due to latera] confmement. The ID model gives surface amplitudes of about 4. Thus, we are led to interpret that the large amplification is associated to the lateral effects. These in turn, are due to focusing and to locally generated surface waves.

From frequency domain results we computed synthetic seismograms using the FFT algorithm for a Ricker wavelet with characteristic period of 3.3 s and vertical and oblique incidences (0, 30 and -30 degrees). In Figures 16 to 18 such time series are plotted for 49 receivers equally spaced between x = -0.94 km and x = 0.94 km. The amplification effect seen in the frequency domain is also clear in the syntbetics. The traces show some "ringing" which is probably due to errors in the computation of transfer functions and/or to the fact that we interpolated frequency response to obtain the synthetics.

CONCLUSIONS

A hybrid indirect boundary element-discrete wavenumber method has been presented and applied to study the seismic response of liorizontally stratified alluvial valleys of arbitrary shape for incidence of plane S77 waves. The method is based upon the integral representation of scattered and diffracted elastic waves at the half-space in terms of single layer boundary sources. For this IBEM approach, the Green's function was selected to be the one for the half-space and the discretization is restricted to the contact between the regions. For the stratified región the field is constructed using a set of solutions for homogeneous and inhomogeneous plane waves that correspond to horizontal discrete wavenumbers (DWN). These solutions satisfy all the boundary conditions at the layers interfaces and are obtained in terms of Thomson-Haskell propagators formalism.

While simple and intuitively appealing, this hybrid approach may suffer of some drawbacks and particular care is required from the analyst. The two approximations are different in character. On one hand, the IBEM description is local (we seek for local physical quantities: the forcé densities), on the other, the DWN linear superposition is global (the needed coefficients represent harmonic patterns). This may give rise to large differences in the entries of the coefficient matrix that results from the collocation scheme for boundary conditions, thus producing large rounding errors. By limiting the amount of allowed inhomogeneous wave and sampling the horizontal wavenumber domain in such a way that spatial periodicities less than *2a* are avoided we may restrict the extent of numérica! errors. This may result for some problems a kind of undersampling. In order to fully automate this procedure some more scrutiny is required.

Some examples are given for the response of simple models of stratified alluvial valleys in an elastic half-space. We confirmed the appearance of complicated patterns for surface displacements even in simple cases. In the examples the generation of Love surface waves can easily be seen.

Synthetics for the irregular valley show a variety of complex patterns. Focussing of energy at low frequencies generally takes place at the deeper parts. Very large amplification was found in the sedimente; more than three times the TD prediction from the impedance contrast.

Our results suggest that variations in layer properties may produce important effects in the response. Strong impedance contrasts and the response at higher frequencies are among the issues that require attention.

The method is generally fast and accurate but it is not error-free and requires care from the analyst. Some more work is still needed for automatical choice of the calculation parameters.

ACKNOWLEDGEMENTS

We gratefully acknowledge the comments of two anonymous reviewers. Thanks are given to M. Ordaz and V. J. Palencia for his critical reviews and to M. Mucciarelli, F. Pacor and A. Méndez for their comments and suggestions. This work was partially supported by ISMES SpA of Bergamo, Italy and by DGAPA-UNAM, México, under project IN104998.

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FIGURE CAPTIONS

Figure 1. Half-space, E, with a horizontally stratifíed alluvial valley, R. The incidence of SH waves with angle y is depicted. The only discretization needed is along the interface S. In the free surface of the layered part and the fíat surface of the half-space the boundary conditions are satisfíed by construction.

Figure 2. Model studied by Kawase and Aki [14]. The regions E and R are shown and the coordinates are normalized in ternis of the half-width a = 5 km. Incidence of harmonic SH waves. Material properties are given in the text.

Figure 3. Synthetic seismograms at the surface of Kawase and Aki's model [14] for vertical incidence of plañe SH waves with time variation given by a Ricker pulse with characteristic period of 4 s. (a) Kawase and Aki [12] and (b) the presented approach.

Figure 4. Same as Figure 3 but with characteristic period of 2 s.

Figure 5. Synthetic seismograms at the surface of Kawase and Aki's model [14], computed with the presented approach, for oblique incidence (y = 30°) of plañe SH waves with time variation given by a Ricker pulse with characteristic period of (a) 4 and (b) 2 s.

Figure 6. Model studied by Bravo et al. [17]. Ratios of velocities of S waves and mass densities considered are: $P\backslash p_E$ = 1/3, *Pi/fe* = 2/3 *anápi/*p_E = 3/4, $p_2 f p_E$ = 0.85.

Figure 7. Comparison of displacement amplitudes on the surface of the parabolic stratified deposit of Figure 6. Normalized frequency is T_j = 1.0, and incident angles are y = 0°, 30° and 60°. Lines are for Bravo et al. and symbols are our approach.

Figure 8. Same as Figure 7 but for a normalized frequency of *t]* = 1.0.

Figure 9. Model of a layered valley. The coordinates are normalized in terms of the half-width a = 1 km. Incidence of harmonio SH waves. Material properties are given in the text.

Figure 10. Displacement amplitude at surface receivers at the layered valley along the x-axis for three incidences. Normalized frequency r_j = 0.8.

Figure 11. Displacement amplitude at surface receivers at the layered valley along the x-axis for three incidences. Normalized frequency *Í]* = 1.0.

Figure 12. Displacement amplitude at surface receivers at the layered valley along the x-axis for three incidences. Normalized frequency *i]* = 1.5.

Figure 13. Contours of displacement amplitude for surface receivers along the x-axis against frequency. Layered valley under vertical incidence of SH waves.

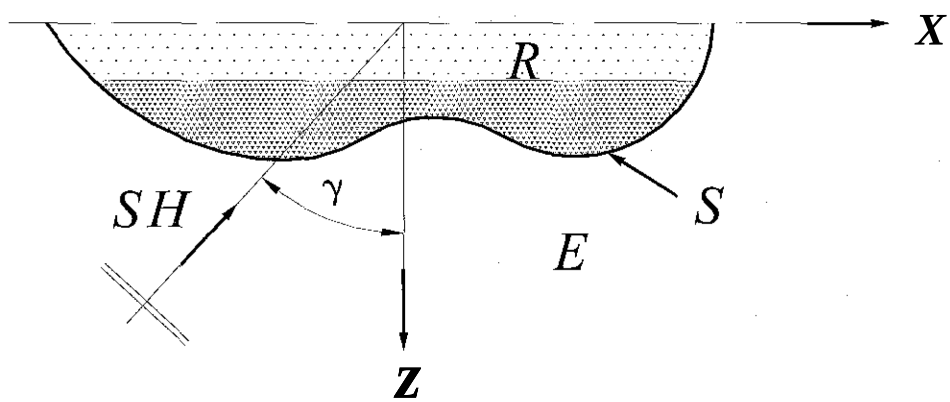
Figure 14. Contours of displacement amplitude for surface receivers along the x-axis against frequency. Layered valley under oblique incidence (7 = 30°) of SH waves.

Figure 15. Contours of displacement amplitude for surface receivers along the x-axis against frequency. Layered valley under oblique incidence *(y* = -30°) of SH waves.

Figure 16. Synthetic seismograms at the surface of the layered valley niodel, computed with the presented approach, for vertical incidence *(y* = 0°) of plañe SH waves with time variation given by a Ricker pulse with characteristic period of 3.3 s.

Figure 17. Same as Figure 13 but for oblique incidence *(y* = 30°).

Figure 18. Same as Figure 13 but for oblique incidence *(y* = -30°).



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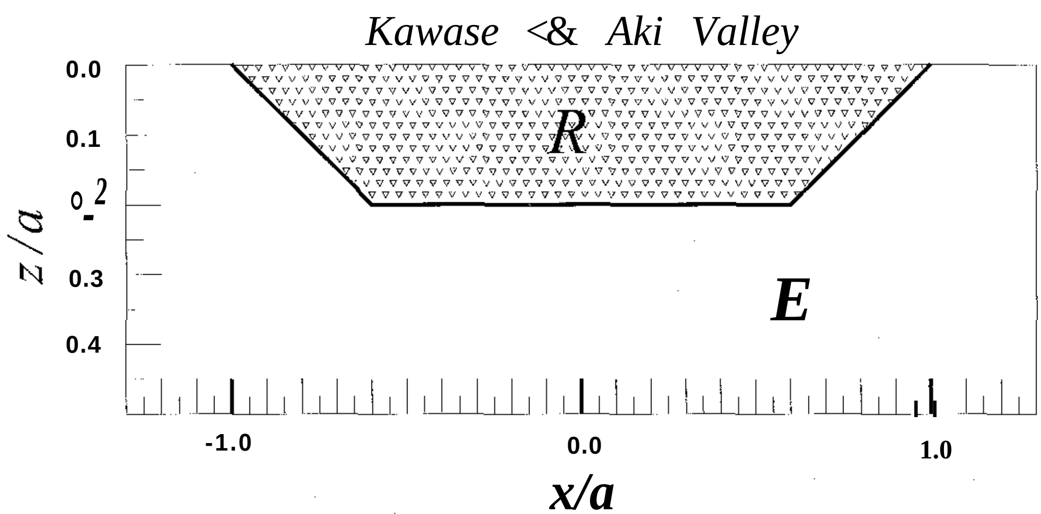

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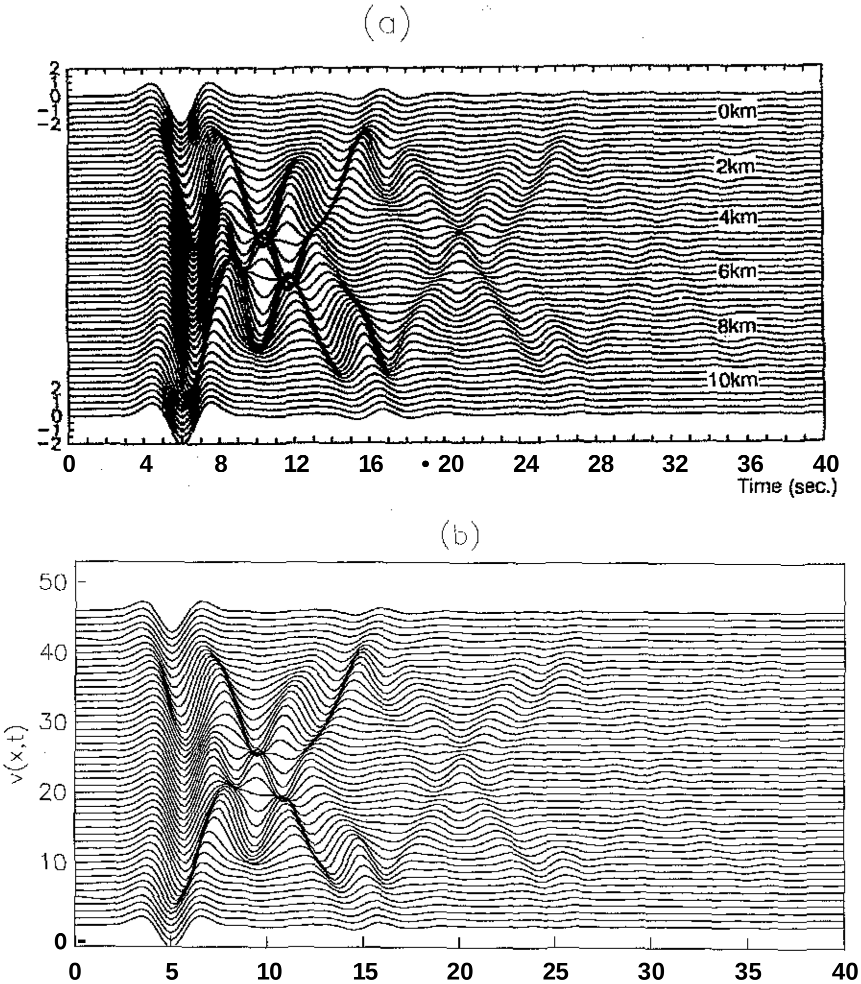
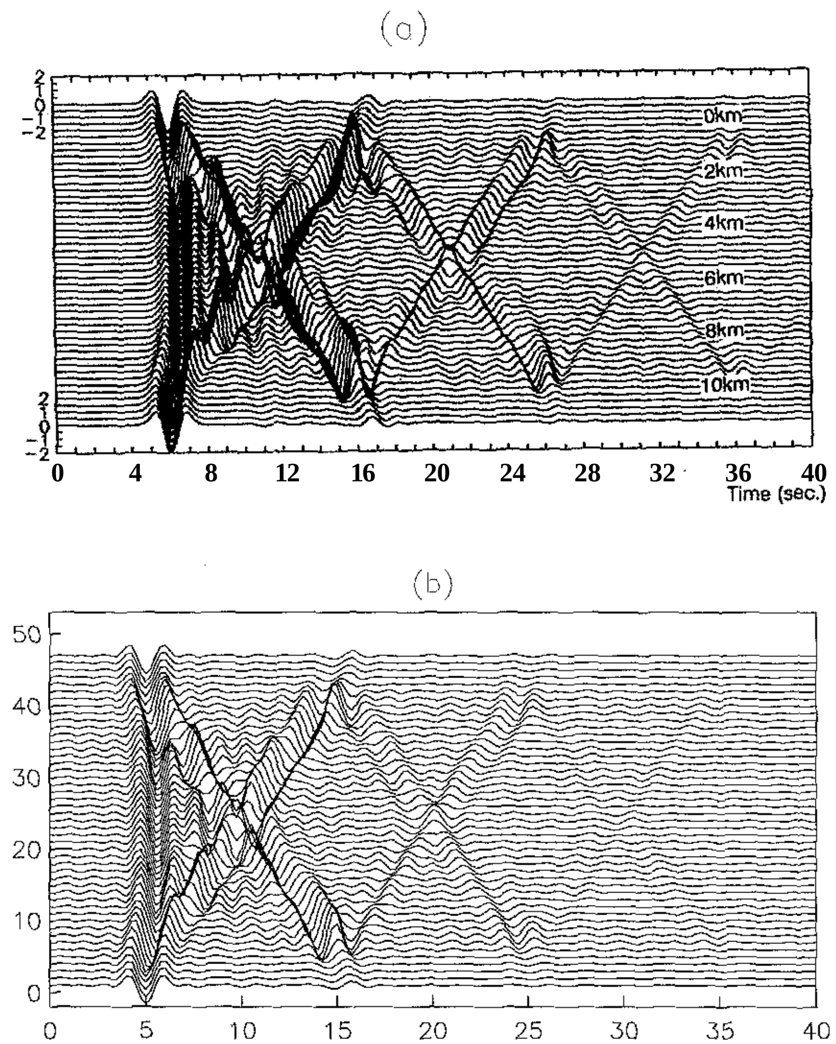

(a)
0km
2km
4km
6km
8km
10km
0 4 8 12 16 20 24 28 32 36 40
Time (sec.)
(b)
v(x,t)
0 10 20 30 40 50
0 5 10 15 20 25 30 35 40

Figure 3



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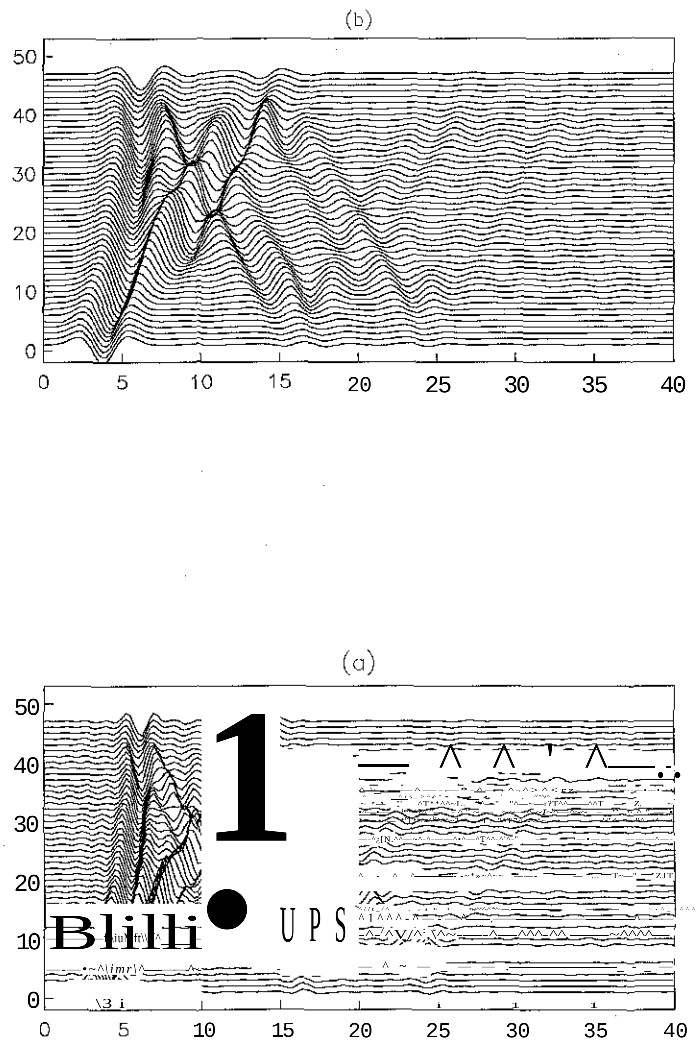

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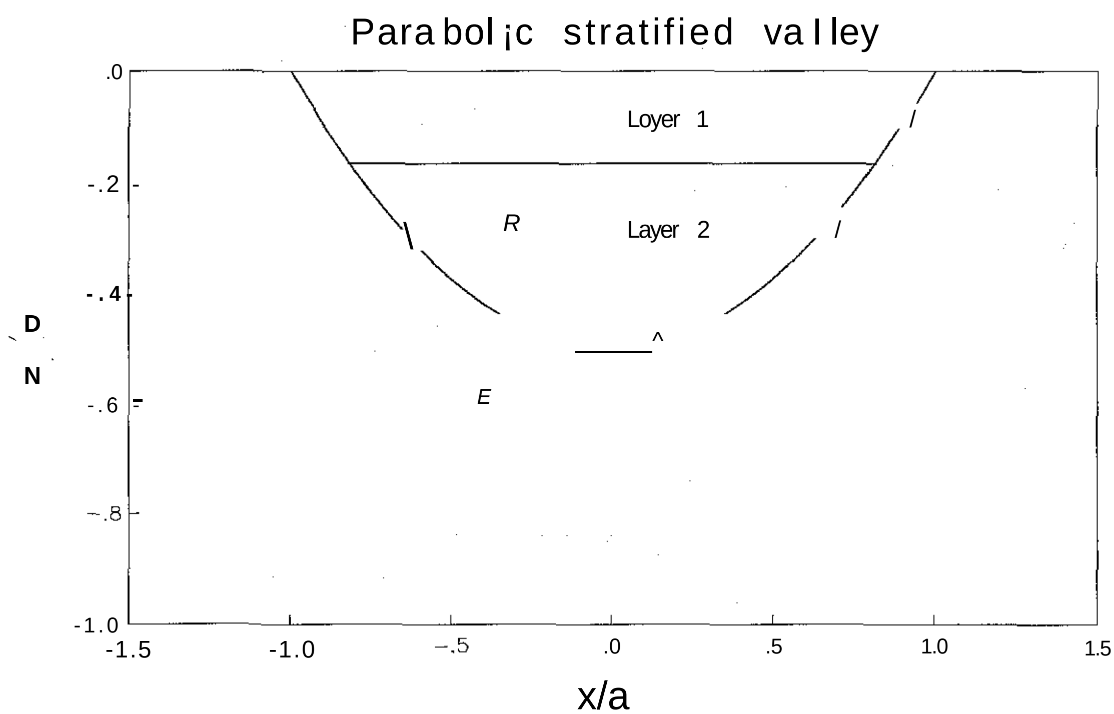
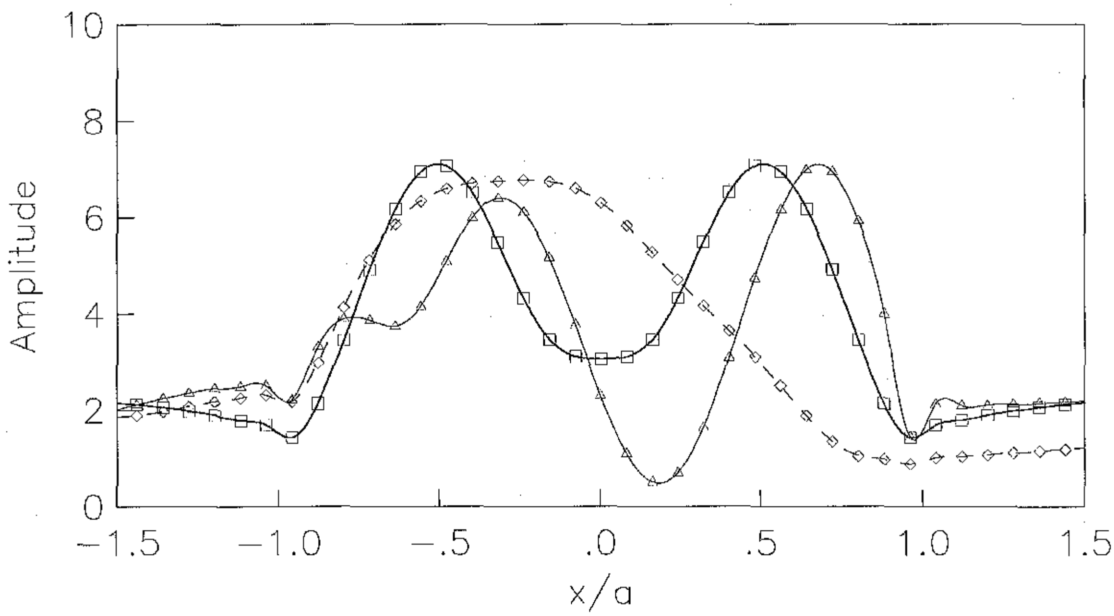

Parabolic stratified valley
Loyer 1
R
Layer 2
E
D
N
.0
-.2
-.4
-.6
-.8
-1.0
-1.5
-1.0
-.5
.0
.5
1.0
1.5
x/a

Figure 6



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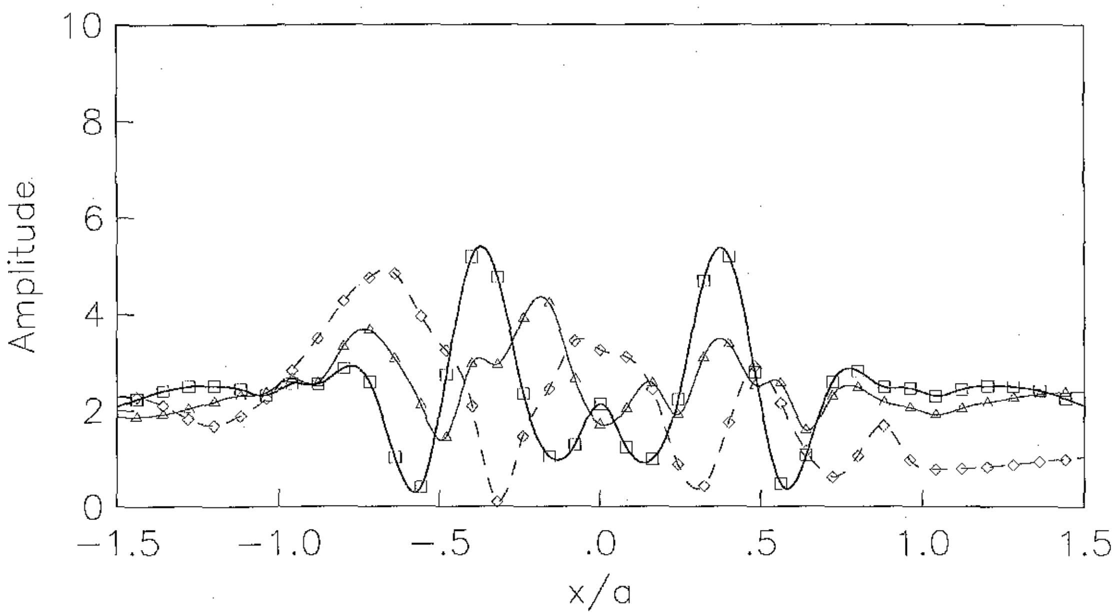

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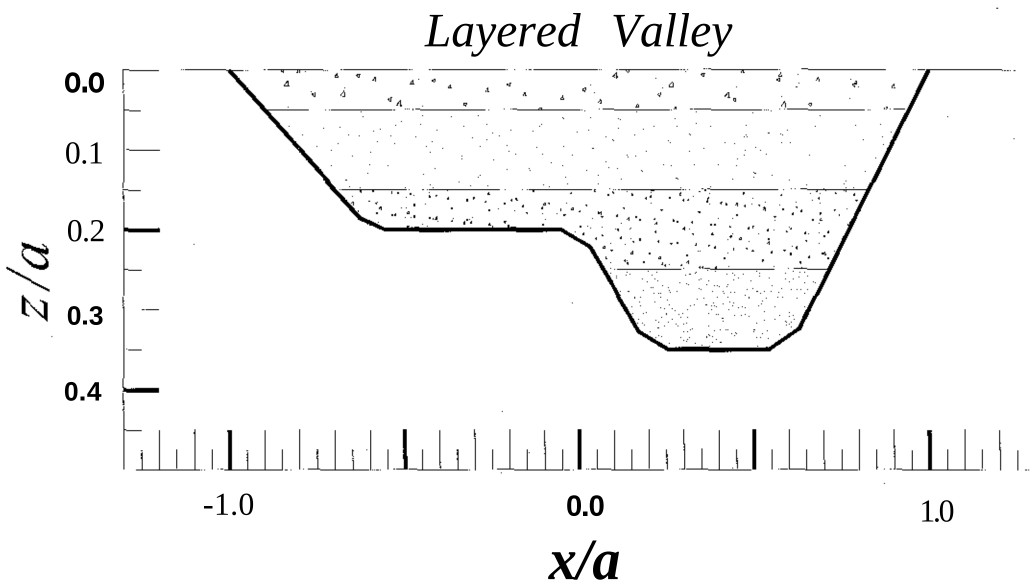

Figure 9

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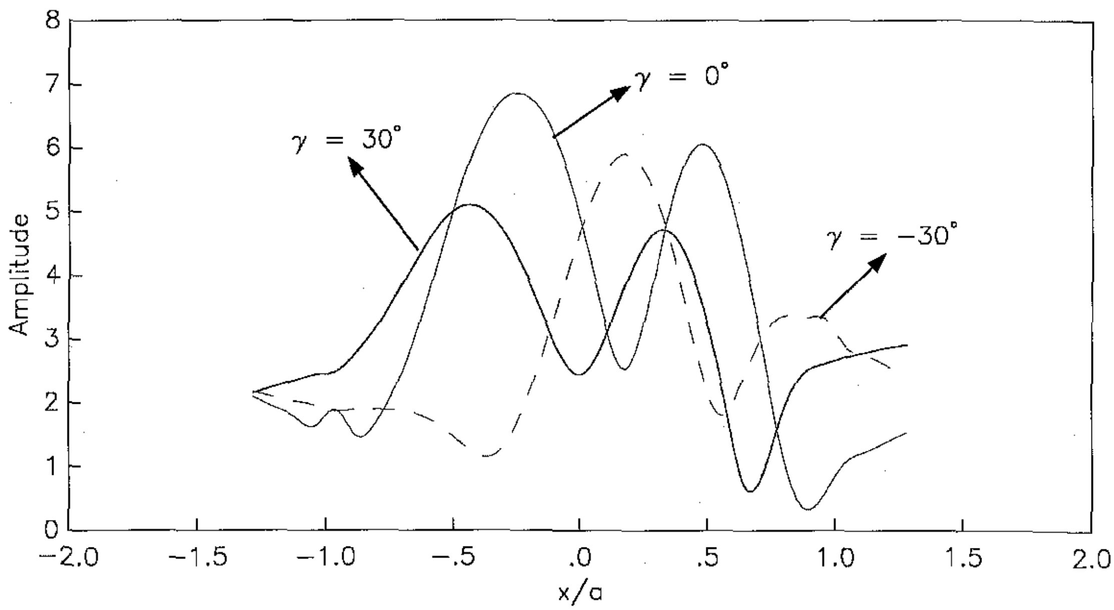
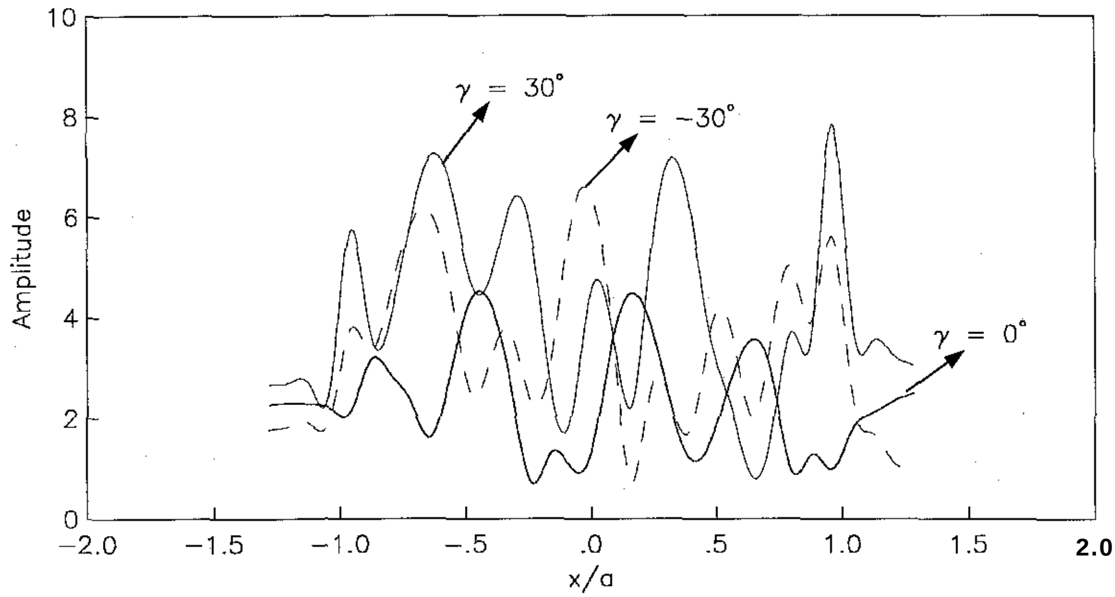


Figure 10



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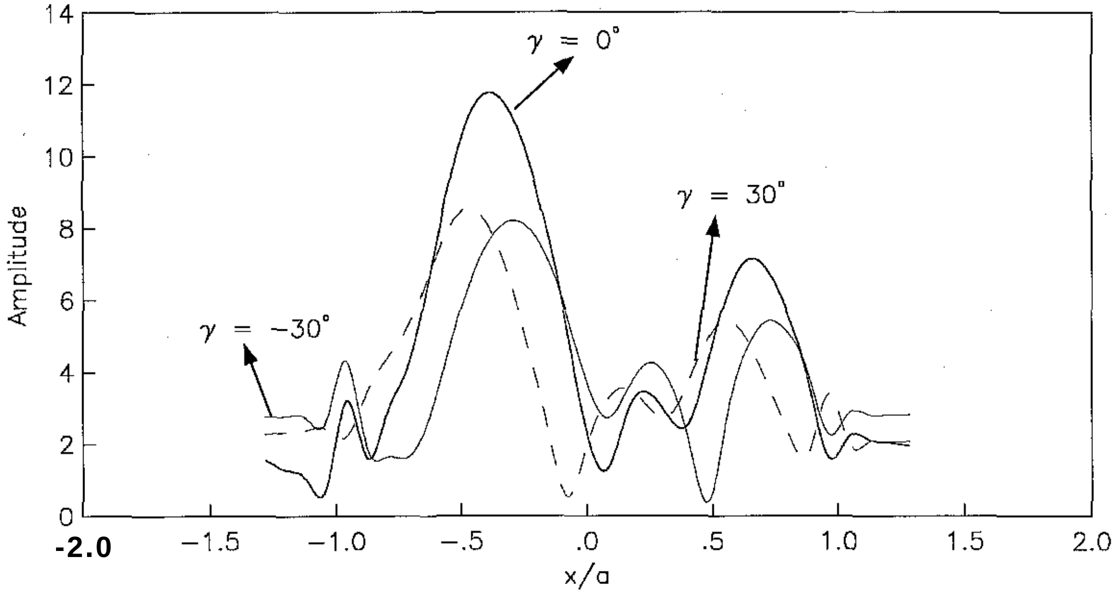


14
12
10
8
6
4
2
0
Amplitude
$\gamma = 0°$
$\gamma = 30°$
$\gamma = -30°$
-2.0
−1.5
−1.0
−.5
.0
.5
1.0
1.5
2.0
x/a

Figure 11

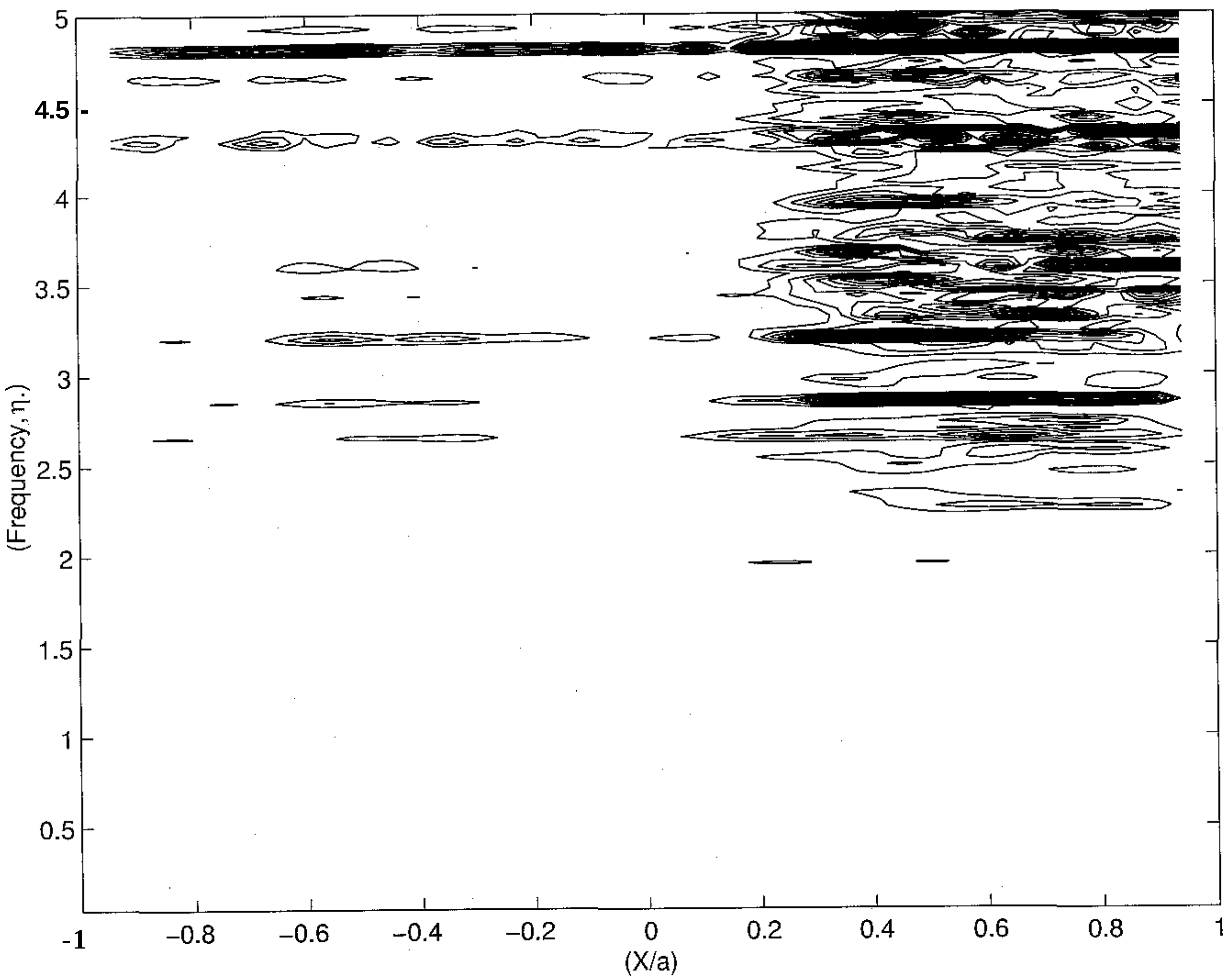
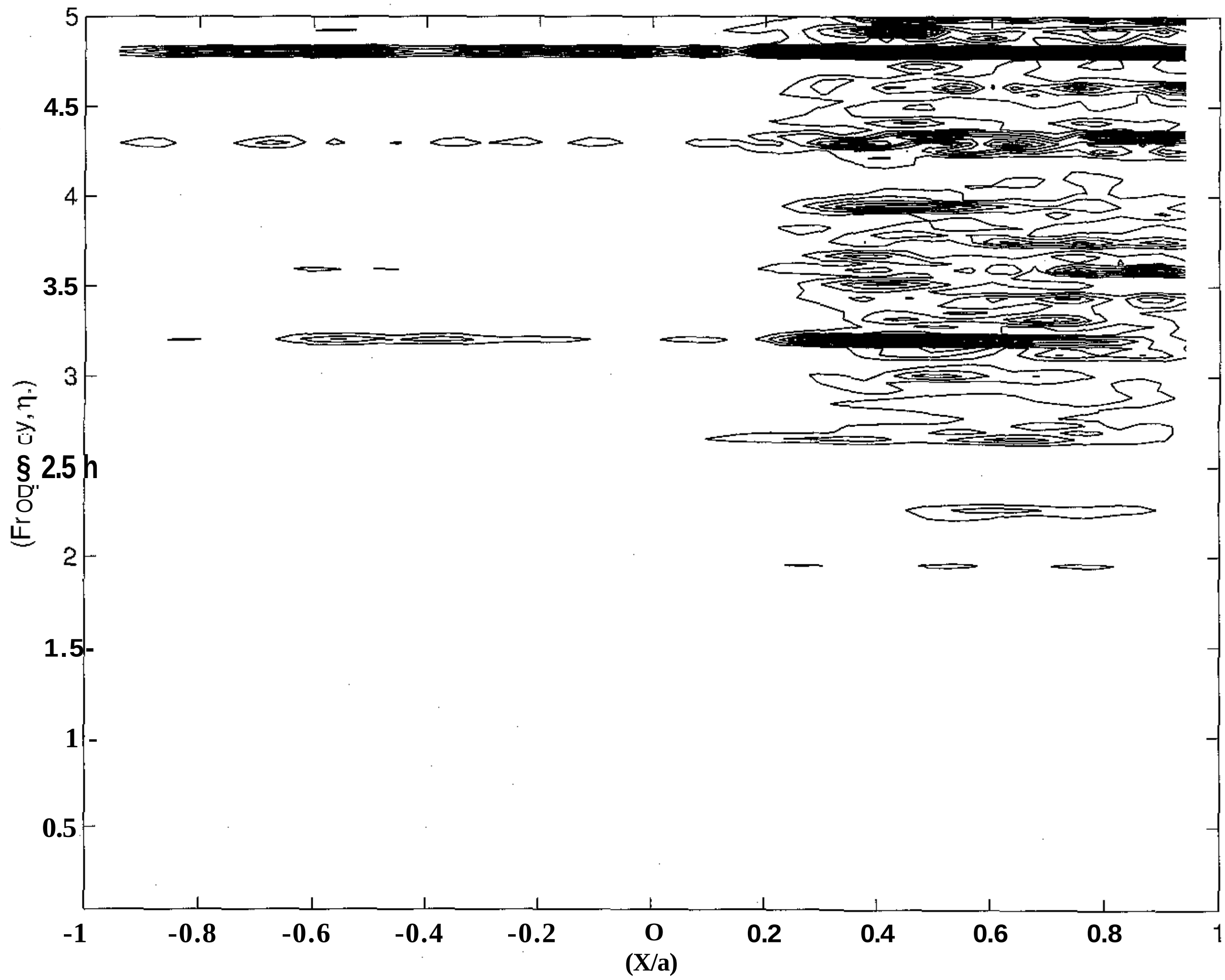


Figure 13



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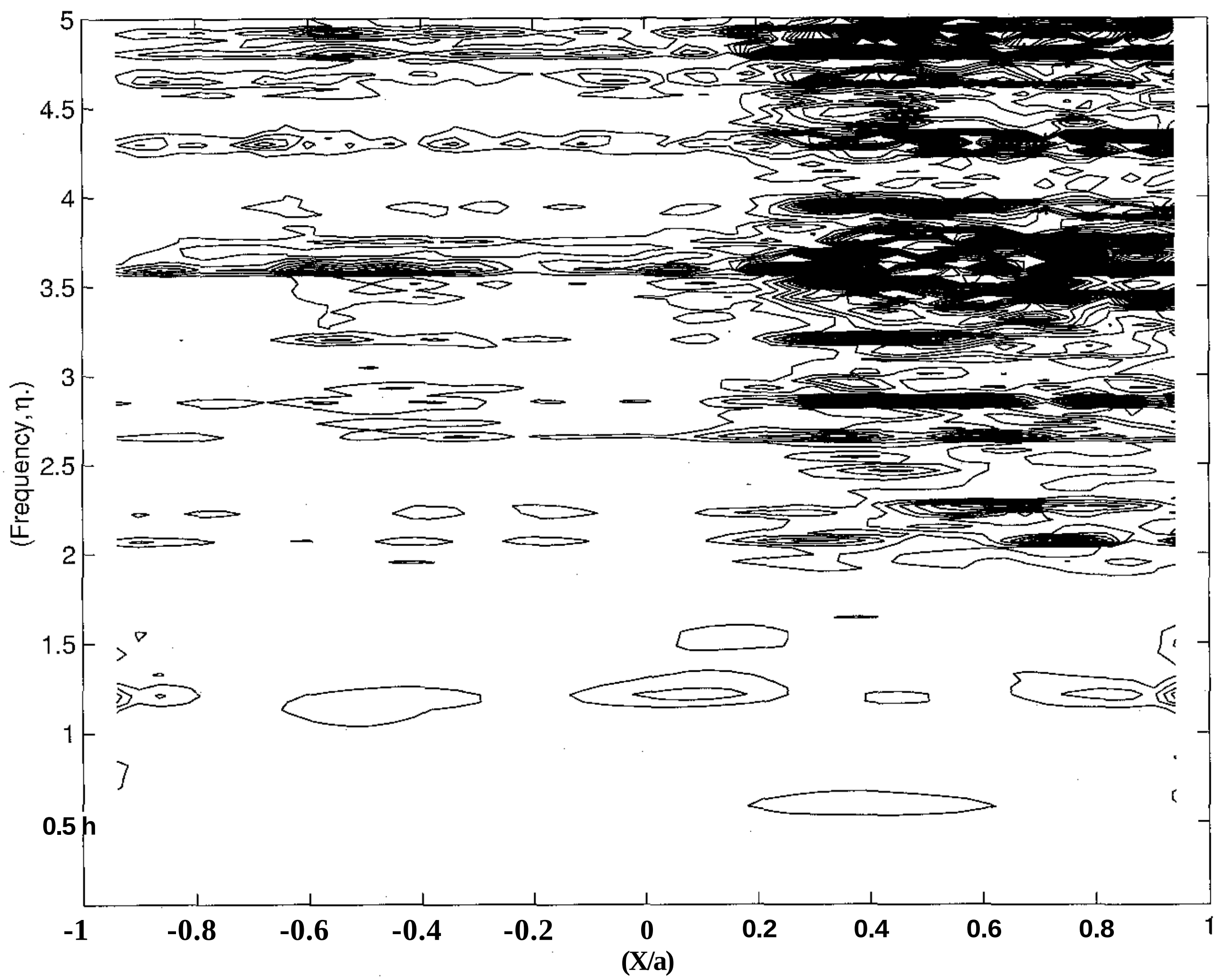

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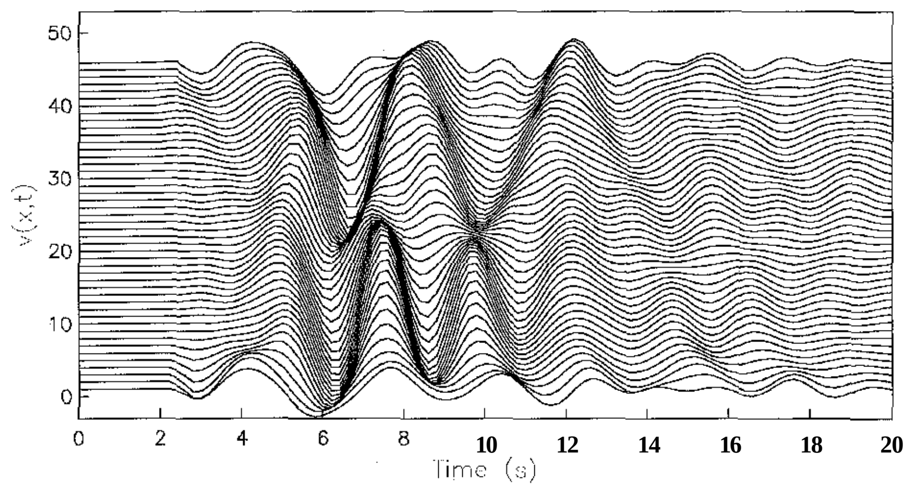

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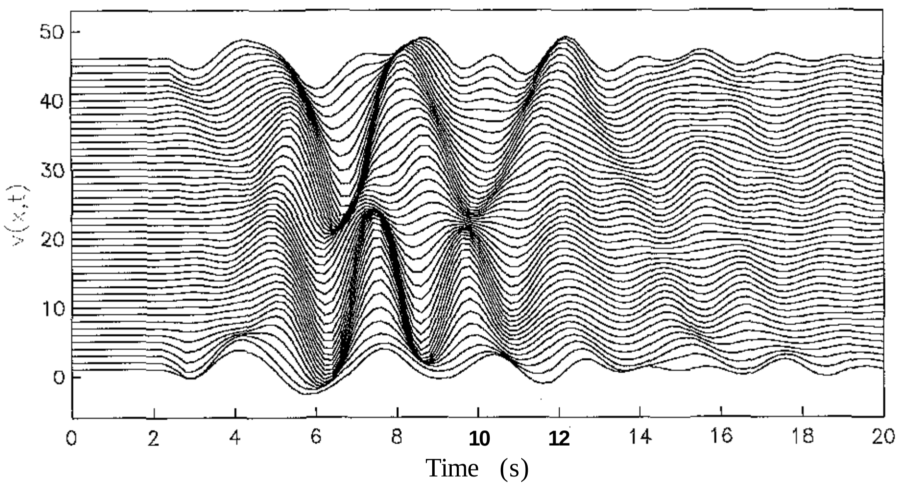

Figure 17

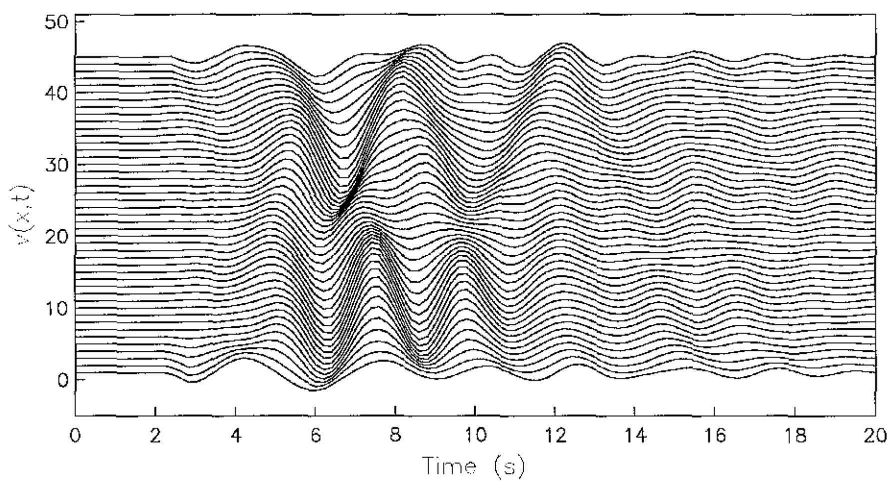

Figure 18