

Imagerie de la source sismique par inversion des mouvements forts et étude des variations de contraintes associées aux séismes: thèse soutenue sur un ensemble de travaux

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Guy Peneren

OBSERVATOIRE DE GRENOBLE et

LABORATOIRE DE GÉOPHYSIQUE ET TECTONOPHYSIQUE

THÈSE

présentée par

Fabrice COTTON

pour obtenir le titre de DOCTEUR de

L'UNIVERSITÉ JOSEPH FOURIER- GRENOBLE I

(Arrêtés ministériels du 5 juillet 1984 et du 30 mars 1992)

Spécialité : Géophysique-Géochimie-Géomécanique

Imagerie de la source sismique par inversion des mouvements forts et étude des variations de contraintes associées aux séismes.

Date de soutenance: le 20 décembre 1995

Composition du jury :

M. R. Madariaga	Président et rapporteur	Professeur de l'Université Paris VII
M. R. Archuleta	Rapporteur	Professeur de l'Université de Californie
M. M. Campillo	Directeur de thèse	Professeur de l'Université Joseph Fourier
M. J.P. Avouac	Examinateur	Ingénieur de recherche au C.E.A.
M. M. Cocco	Examinateur	Chercheur à l'I.N.G (Italie)
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Résumé

Cette thèse est organisée autour de deux thèmes complémentaires: l'analyse de la cinématique de la rupture sismique à l'aide de données de champ proche et l'étude des variations de contraintes associées à cette rupture dans le milieu entourant la faille.

Une méthode d'inversion des accélerogrammes de champs proche dans le domaine des fréquences est tout d'abord décrite. Cette méthode permet de trouver le glissement, la vitesse du front de rupture et la durée de la rupture en chaque point de la faille. La qualité de l'image de la rupture obtenue est ensuite analysée par une étude de résolution et la mesure quantitative de l'accord entre données et synthétiques. Une méthode d'inversion semi-globale (algorithme génétique) est enfin utilisée pour tester l'unicité ou la non-unicité de la solution obtenue. Cette méthode a été appliquée à trois séismes: le séisme californien de Landers (1992), le séisme d'Uttarkashi (1991) situé au nord de l'Inde et le séisme de Northridge en Californie (1994). Nos résultats montrent que les séismes étudiés sont complexes. La répartition spatiale du glissement est hétérogène et montre la présence de plusieurs sous-évènements. Dans le cas de Landers, une comparaison entre les propriétés de la rupture, la géométrie des failles et les résultats d'une tomographie locale est effectuée. La largeur de chaque sous évènement semble controlée par la segmentation de la faille. La durée de la rupture en chaque point est compatible avec un mécanisme de type fracture où la radiation d'un point de la faille se termine à l'arrivée d'une phase d'arrêt émise à partir des bords de chaque sousévènement.

Dans une deuxième partie, nous nous sommes interessés aux variations de contraintes associées aux séismes. Notre étude de la cinématique du séisme de Landers montre que le glissement a gardé une direction constante pendant la rupture alors que celle-ci est hétérogène. Nous interprétons ce résultat par une chute de contrainte partielle pendant le séisme. Une méthode de calcul du champ de contrainte dynamique radié par une source sismique dans un milieu à couches planes est ensuite décrite. Cette méthode est appliquée au séisme de Landers. La source est modélisée par le modèle trouvé par inversion dans la première partie. Les variations du critère de rupture de Coulomb sont calculées sur les plans de faille de deux séimes apparamment déclenchés par le séisme de Landers. Cette analyse montre que la résistance des faille est une fonction dépendante du temps. De faibles variations "statiques" des contraintes peuvent être corrélées avec le déclenchement des séismes alors que des variations "dynamiques" plus importantes mais de faibles durées n'ont eu aucun effet immédiat.

Introduction générale

L'application pratique de la sismologie est de prévenir ou de réduire les dégâts engendrés par les séismes. La première réponse possible à cette attente est phénoménologique. Elle consiste à utiliser les informations collectées lors des séismes passés pour prévoir les effets des événements futurs. Les spectres en accélération des forts séismes sont par exemple utilisés pour concevoir les structures des bâtiments. Les phénomènes inhabituels observés avant un séisme sont considérés comme des précurseurs possibles des événements futurs. Cette approche n'est pourtant pas satisfaisante car le nombre de données est encore insuffisant. En effet, les temps de récurrence des séismes dans une région donnée sont longs (plusieurs dizaines ou centaines d'années) alors que la majorité des observations sismologiques ont au plus une centaine d'années. Pour combler ce manque de données, une autre approche est possible. Il s'agit tout d'abord de comprendre plus en détail la physique de la rupture et notamment les différents phénomènes que sont l'initiation, la propagation et l'arrêt de la rupture pour ensuite en déduire des applications pratiques. Hélas, il n'est pas possible d'étudier directement les séismes dans la région où ils se développent. Depuis la surface du globe, le sismologue peut simplement observer les manifestations de ce phénomène. Il peut ensuite confronter les observations aux modèles simplifiés de la réalité développés de façon analogique ou théorique. La confrontation entre observations et modélisations permet alors d'éliminer les hypothèses non représentatives de la réalité. Elle permet aussi de tirer un meilleur parti des observations.

L'étude développée dans cette thèse participe à cet effort de compréhension de la source sismique. Parmi les observations collectées pendant un séisme, les enregistrements du mouvement du sol à proximité d'un séisme peuvent permettre de retrouver l'histoire du déplacement en tout point de la faille. Cette « imagerie » de la rupture sismique constitue un des thèmes principaux de ce travail. Après avoir rappelé les méthodes existantes et montré leurs limites, nous verrons quels sont les problèmes que se posent actuellement les sismologues et l'apport possible des résultats de l'imagerie. Dans une deuxième partie, nous décrirons une nouvelle méthode d'imagerie de la rupture à l'aide de données de champs proche. Cette méthode est développée et appliquée aux données exceptionnelles du séisme de Landers (1992). Elle est ensuite améliorée lors de l'étude des séismes d'Uttarkashi (1991) et de Northridge (1994). Cette description purement cinématique de la rupture est insuffisante et le sismologue aimerait aussi avoir une meilleure connaissance de l'histoire des contraintes associées au séisme. L'étude des variations de contraintes dans le milieu entourant la faille constitue ainsi le deuxième thème de ce travail. Nous verrons comment la cinématique de la rupture nous informe sur

l'état de contrainte absolu sur la faille. Nous verrons aussi comment le calcul des variations de contraintes associées à la rupture et l'étude de la sismicité déclenchée nous renseignent sur les mécanismes de l'instabilité sismique. Un dernier chapitre permettra enfin de faire une synthèse des résultats obtenus et de résultats publiés récemment.

Plusieurs parties de cette thèse (II-1, II-2, II-3, II-4, III-1, III-2) sont constituées d'articles publiés, sous-presse, soumis ou en préparation. Pour faciliter la lecture, nous avons essayé de garder une présentation cohérente dans tout le manuscrit. Les figures sont propres à chaque partie et chaque article publié a gardé sa bibliographie propre. Une bibliographie complète se trouve à la fin de la thèse.

Première partie Problématique

I. Problématique

1. Quelles sont les données disponibles lors de l'étude d'un grand séisme et quelles informations peut-on en tirer ?

Les données permettant l'étude de la rupture sont variées. Nous verrons dans ce travail comment l'analyse des sismogrammes enregistrés en champ proche est essentielle pour décrire les détails de la rupture. Nous verrons aussi que cette analyse suppose la connaissance de données complémentaires notamment pour estimer la géométrie de la faille ou le glissement moyen. Il est ainsi important de comprendre quel est l'apport de chaque ensemble de données sur l'étude des divers paramètres de la rupture.

A. Effets macroséismiques

Ce type d'observations comprend divers phénomènes difficilement quantifiables tels les effets ressentis par les témoins, les dommages aux bâtiments ou encore les glissements de terrains. Ces données sont souvent les seules dont le sismologue dispose pour étudier les séismes anciens. Elle permettent d'estimer approximativement la zone où a eu lieu le séisme et son importance.

B. Rupture de surface.

Pour certains séismes, la rupture atteint la surface du sol. On peut alors mesurer le décalage entre les deux lèvres de la faille et ainsi évaluer le glissement statique près de la surface. On peut aussi préciser de façon certaine la géométrie en surface du plan de faille. Par contre, les observations de surface ne peuvent pas être extrapolées en profondeur. De plus, bien souvent la rupture n'affecte pas une seule faille mais un système de faille voir une zone de cisaillement qui peut avoir plusieurs dizaines de mètres de large comme dans le cas de Landers (Johnson et al., 1993). Le glissement cosismique est alors partagé sur plusieurs failles ce qui rend sa mesure difficile.

C. Déformation finie du sol

La mesure de la déformation du sol permet quant à elle de préciser la géométrie de la faille, sa localisation et le glissement. Divers types de mesures sont utilisés (géodésie, profils de nivellement, triangulation, positionnement par satellite (GPS), interférométrie). Le principal problème provient de l'espacement en temps entre deux campagnes de mesure. Il est impossible de savoir si le glissement sur la faille déterminé par ce type d'observation est cosismique ou non.

D. Sismologie

Données télésismiques

L' analyse la plus simple d'un séisme par l'utilisation d'enregistrements sismologiques consiste à étudier les polarités et les temps d'arrivé du premier mouvement P et ainsi à déterminer le mécanisme au foyer du séisme. Le séisme est considéré comme étant un point source et les données proviennent de réseaux locaux ou mondiaux tels que les réseaux NARS ou Géoscope qui fournissent des signaux digitaux à très large spectre et grande dynamique. Connaissant le mécanisme au foyer, on peut calculer la fonction de Green théoriquement observée pour l'azimut, la profondeur et la distance correspondant au couple source-station. On corrige ensuite de l'atténuation et de l'instrument utilisé et on déconvolue pour obtenir une fonction temporelle du glissement qui permet d'avoir une image de la complexité de la source (Kikushi et Kanamori, 1982). Ce type d'inversion pose plusieurs problèmes. En premier lieu le mécanisme du séisme n'est pas toujours constant tout au long de la rupture surtout si celle-ci s'effectue sur plusieurs segments de faille ayant des directions différentes. De plus, l'extension de la source fait apparaître un facteur de directivité qui doit être pris en compte. Pour ces raisons, depuis ces premiers travaux l'utilisation des sismogrammes larges bandes à des distances télésismiques s'est perfectionnée (Ekström, 1989; Kikushi et Kanamori 1986, 1991; Nabèleck, 1984). Campos et al. (1994) utilisent par exemple les données enregistrées sur les réseaux IRIS et Géoscope pour préciser les variations spatio-temporelles de la rupture du séisme d'Iran de 1990. Cette étude montre bien toutes les possibilités de ce type d'inversion où par étapes successives le modèle se complique et se précise. Ainsi, une première inversion des ondes de volume dans un modèle de point source précise le mécanisme et la fonction temporelle du moment sismique. A l'étape suivante le mécanisme est fixé et on cherche la direction et la vitesse de la rupture. La source est ensuite modélisée par une ligne de sources ayant toutes le même mécanisme ce qui permet de montrer le caractère unilatéral de la rupture. Une dernière étape consiste à inverser le mécanisme de chacune des sources pour préciser les variations du mécanisme tout au long de la rupture. Pour donner une image encore plus fine de la rupture, les techniques d'imagerie telles que nous les décrirons plus loin peuvent elles aussi utiliser les données télésismiques (Das et Kostrov, 1990)

♦ Etude des répliques

Les répliques permettent de donner une image de la géométrie de la rupture et de localiser la zone du séisme. Elles permettent notamment de choisir l'un des deux plans nodaux déterminés par l'étude du mécanisme. L'origine en terme mécanique des répliques reste néanmoins mal comprise. (Scholz, 1990). De plus, la localisation de ces répliques peut aussi être imparfaite. Les erreurs de lecture de temps d'arrivée, les imperfections du modèle de vitesse utilisé pour la localisation ou une mauvaise géométrie du réseau peuvent engendrer des incertitudes qui rendent difficiles l'utilisation de ces répliques pour l'étude du séisme principal. En conclusion, les répliques

permettent d'avoir des idées sur les dimensions et le mécanisme du séisme principal. Par contre, elles ne permettent pas encore de contraindre fortement les paramètres de rupture.

Données de mouvement forts.

Les appareils dit de "mouvements forts" (strong motions) ne saturent pas lors de l'enregistrement de grandes magnitudes. Ils permettent donc d'étudier le mouvement du sol à des distances faibles des sources sismiques et parfois même pratiquement sur le plan de faille (station Lucerne lors du séisme de Landers, 1992). C'est le type de données que nous utiliserons dans la suite de ce travail et notamment dans le chapitre II pour évaluer la répartition spatiotemporelle du glissement sur la faille. Ces appareils sont la plupart du temps des accéléromètres et parfois des vélocimètres (Japon). On peut retenir que le développement de ce type de réseaux s'est accéléré ces dernières années. Ces appareils sont aujourd'hui présents près de nombreuses grandes failles (Californie, Mexique mais aussi Inde et Iran). On peut ainsi penser que les données permettant l'utilisation des techniques développées dans ce travail seront de plus en plus fréquentes les années à venir. Elles proviendront surtout de contextes tectoniques variés et pas seulement de Californie.

E. Conclusion

Seule la sismologie est capable de donner des informations sur l'aspect dynamique de la rupture telle que la vitesse du front de rupture sur le plan de faille ou la variation de l'énergie libérée au cours du temps. Les autres méthodes d'analyse (géodésie ou observations gélogiques) sont complémentaires et permettent notamment d'estimer la géométrie et les caractéristiques statiques de la rupture (déplacement final entre les deux lèvres de la faille par exemple)

2. Confrontation des observations et des modèles simples: la rupture est complexe et les modèles théoriques simples ne suffisent pas

Pour mieux comprendre la physique de la rupture de nombreux modèles théoriques ont été développés. Parallélement à ces développement théoriques les observations concernant la dynamique de la rupture se sont peu à peu accumulées.

A. Modèles simples de la rupture

Le type de modèle qui a été développé le plus tôt (Haskell, 1964) consiste à décrire avec un minimum de paramètres l'histoire du glissement sur le plan de faille. On définit ainsi les modèles cinématiques. Le modèle cinématique le plus simple est le modèle de dislocation qui définit l'histoire du glissement sur la faille avec 5 paramètres (longueur et largeur de la faille, temps de montée définissant la vitesse de glissement, déplacement final et vitesse de rupture). Ce type de modèle a été utilisé le premier pour modéliser numériquement le déplacement du sol au voisinage

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d'une faille (Aki, 1968). De nombreuses modélisations basse fréquence ont ensuite été effectuées essentiellement pour des séismes californiens (Bouchon, 1979; 1981). Pour permettre une simulation plus fine la faille peut être divisée en plusieurs dislocations caractérisées par des valeurs de déplacements, des vitesses de rupture ou encore de temps de montée différents. Historiquement, ce type de modélisation a ainsi été utilisé pour modéliser le séisme de San Fernando ou d'Imperial Valley (Bouchon, 1978; Hartzell et Helmberger, 1982, Archuleta, 1984). Nous verrons dans le chapitre II que ce principe (modélisation de la rupture par plusieurs dislocations discrétisant le plan de faille) est à la base de notre travail d'imagerie de la rupture.

Pour modéliser la rupture sans introduire d'hypothèses non physiques, les sismologues ont développé des modèles dits dynamiques où le glissement sur la faille et la cinématique du front de rupture sont déterminés par la contrainte tectonique et les caractéristiques des matériaux. Aux premiers développements analytiques de ce type de problèmes (Kostrov, 1964) ont succédés des approches numériques (équations intégrales et différences finies) dont Virieux (1986) fournit une liste exhaustive.

B. Les séismes semblent complexes quel que soit le contexte tectonique

Séismes de subduction

Le séisme d'Alaska (1964) fournit les premières observations montrant le caractère complexe de la rupture d'un séisme de subduction (Wyss et Brune, 1967). Depuis cette idée de complexité a été observée pour de nombreux séismes de subduction. En étudiant ce type de séisme, Kanamori et Stewart (1978) suggèrent l'idée d' "aspérité" pour expliquer les irrégularités du glissement sur le plan de faille. Papageorgiou et Aki (1983) préférèrent introduire le concept de "barrière". Les séismes de Michoacan et de Valparaiso en 1985 furent les premiers séismes de magnitude supérieure à 8 enregistrés en champ proche par des accélérographes. Nous verrons au chapitre II-2 comment les données concernant le séisme de Michoacan permettent de contraindre la physique de la rupture.

Décrochements

La littérature concernant la physique de la rupture de ce type de séisme est la plus complète. Le caractère complexe de la rupture a ainsi été tout d'abord observé pour quelques séismes en décrochements importants tel que le séisme d'Impérial Valley en 1940 (Trifunac et Brune, 1970) ou du Guatemala en 1976 (Kanamori et Stewart, 1978; Kikuchi et Kanamori 1982). Tout au long de cette étude nous verrons que les physiciens de la rupture lorsqu'ils confrontent données et modèles (par exemple Heaton, 1990) n'utilisent souvent que les observations obtenues lors de quelques séismes Californiens bien particuliers (Imperial Valley, 1979; Loma Prieta, 1989 ou encore plus récemment Landers 1992).

Chevauchements continentaux

Le nombre d'études de la rupture concernant des séismes de chevauchements continentaux est moindre. L'étude de ce type de séisme est néanmoins important car il concerne par exemple les grands séismes pouvant avoir lieu sur les grands chevauchements himalayens. L'étude du séisme d'Uttarkashi illustrera ce type de séisme.

Failles normales

Là encore le nombre d'études détaillées est faible. Les études les plus complètes concernent le séisme italien d'Irpinia en 1980. Malgré des données difficiles, toutes les études confirment la complexité de ce séisme en faille normale comprenant plusieurs sous-événements (Westaway et Jackson, 1987; Bernard et Zollo, 1989; Pantosti et Valensise, 1990).

Conclusions

Toutes ces études suggèrent la complexité de la rupture sismique. Mieux évaluer les caractéristiques de cette complexité est ainsi important pour valider les modèles théoriques de la rupture. On peut aussi souligner que même si les observations paraissent nombreuses, certains contextes tectoniques tels que les chevauchements himalayens ont été très peu étudiés.

3. Mieux évaluer la complexité: méthodes classiques d'imagerie de la source

A. Principes théoriques de l'imagerie: le théorème de représentation.

La base théorique de l'imagerie de la rupture est donnée par le théorème de représentation qui relie le glissement entre les deux lèvres d'une faille et la radiation élastique des ondes. D'après ce théorème (équation 3.2 de Aki et Richards, 1980; équation 3.2.18 de Kostrov et Das, 1988), le déplacement à une station située à une distance x peut s'exprimer en fonction de la distribution du glissement sur la faille par :

$$u_k(y,\tau) = \int_0^t dt \iint_{\Sigma} G_{ik}(y,x,\tau,t) \ a_i(x,t) \ dS$$

 $u_k(y,t)$, $a_i(x,t)$ et $G_{ik}(y,x,t,t)$ sont respectivement les composantes du déplacement du sol à la station, les composantes du glissement sur la faille et la réponse impulsionnelle du milieu dû à une dislocation au point source (x,t). Tout le problème de l'imagerie consiste à trouver le glissement $a_i(x,t)$ connaissant $G_{ik}(y,x,t,t)$ et $u_k(y,t)$. Pour résoudre ce problème, plusieurs méthodes ont été proposées. Certaines utilisent les enregistrements dans le domaine temporel (méthodes temporelles) et d'autres les spectres des enregistrements (méthodes spectrales).

B. Applications pratiques.

Méthodes temporelles

Multifenêtrage

La première méthode d'imagerie a été décrite par Olson and Apsel (1982). Ses principes sont à la base de nombreuse études et programmes encore utilisés aujourd'hui dans la plupart des méthodes d'imagerie de la source sismique. Pour cette raison et pour pouvoir mieux comparer avec la méthode développée dans cette thèse nous allons détailler le travail d'Olson et Apsel (1982). Cette méthode utilise les enregistrements sous leur forme temporelle. Dans cette paramétrisation, le plan de faille est discrétisé en J sous-failles. Chaque sous-faille peut glisser 2K+1 fois à des incréments de temps successifs dt. Le glissement s'effectue selon un vecteur glissement s_{jk} qui a deux composantes sur le plan de faille. Les sous-failles sont elles-mêmes représentées par un réseau de points sources localisés en x_i qui glissent tous selon le même vecteur s_{jk}. Avec cette paramétrisation, le glissement en chacun des points sources s'écrit :

$$a(x_{i},t) = \sum_{j=1}^{j} X_{j}(x) \sum_{k=-K}^{K} s_{jk} P_{k}(x_{i},t)$$
 (2)

où $X_j = 1$ si x_i est dans la sous faille j et $X_j = 0$ sinon.

Lors de chaque glissement k, la fonction temporelle du glissement à chaque point source est décrite par $P_k(x_i,t) = F(t-T(x_i)+k\delta t)$. La progression du front de rupture à l'intérieur de chaque sous-faille est prédite par T(x) qui est donné par la vitesse locale du front de rupture. Le glissement en chaque point a(x,t) est ainsi représenté par une suite de 2K+1 fenêtres de largeur dt. La forme et l'amplitude de la fonction temporelle du glissement dans chaque fenêtre sont respectivement données par la fonction F et par s_{jk} . Dans la suite de ce travail, cette paramétrisation sera appelée imagerie par multifenêtrage. La Figure 1 montre un exemple où la fonction F utilisée est une fonction dont la dérivée par rapport au temps est un triangle.

L'équation (1) devient

$$U^{i}(y,t) = \sum_{j=1}^{J} \sum_{k=-K}^{K} s_{jk}. g_{j}^{i}(y,t+k\delta t)$$
 (3)

 $g_j^i(y,t+k\delta t)$ est la fonction de Green pour la sous-faille j au point d'enregistrement y. Elle représente le déplacement du sol en y dû à un glissement s_{jk} unitaire. Si on connait suffisamment la géométrie et le mécanisme de la rupture pour déterminer s ou Σ , si de plus, on a une connaissance suffisante du milieu pour pouvoir calculer les fonctions de Green du milieu et ainsi $g_j^i(y,t+k\delta t)$ pour chaque sous-faille j, on peut résoudre le problème inverse qui consiste à trouver s_{jk} en connaissant U et $g_j^i(y,t+k\delta t)$. Pour trouver ces inconnues on résout de façon linéaire le système d'équation (3). Le nombre de paramètres à résoudre pour J sous-failles

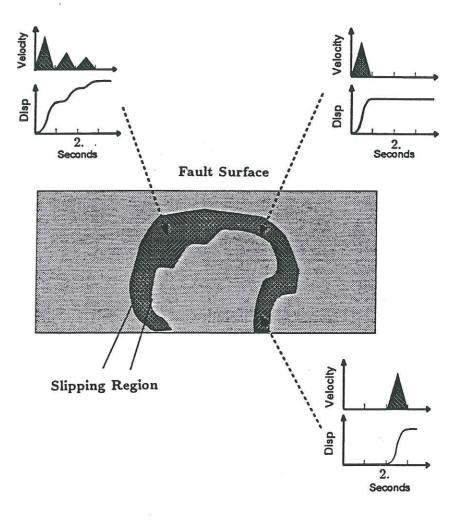


Figure 1. Schéma illustrant la méthode par multifenêtrage. La fonction vitesse de rupture en chaque point de la faille est représentée par une somme de triangles. Lorsque la durée de la rupture (temps de montée) en un point est importante, les 3 triangles ont une amplitude significative. Lorsque la durée de la rupture est faible, un seul triangle a une amplitude importante. La position du premier triangle ayant une amplitude significative donne le temps de rupture.

pouvant glisser 2K+1 fois est J(2K+1) si la direction du vecteur glissement est fixée. Ce nombre est égal à 2J(2K+1) si on inverse les deux composantes du vecteur glissement. Le nombre d'équations indépendantes est égal au nombre de stations, multiplié par le nombre de composantes et de points en temps utilisables pour chaque station. Le rapport entre le nombre d'équations et le nombre d'inconnues dépend des choix *a priori* de la paramétrisation.

Ainsi, selon les auteurs le nombre de fenêtres choisi et leur durée varient fortement. Ce choix est en effet difficile: si dt est petit et k grand cette technique a la flexibilité nécessaire pour pouvoir décrire précisément le temps du début de la rupture et la durée de la rupture en chaque point. Le début de la rupture est ainsi donné par la première fenêtre où s_{jk} est non nul. La durée de la rupture est égale au nombre de fenêtres où s_{jk} est non nul multiplié par la durée dt de chaque fenêtre. Ainsi, si on veut contraindre le moins possible les solutions possibles du problème inverse, il faut prendre un grand nombre de fenêtres et leur donner une largeur la plus courte possible. Ceci a alors l'inconvénient de faire exploser le nombre d'inconnues ce qui pose deux problèmes. D'une part, La taille des matrices décrivant ces systèmes d'équation est limitée par la puissance des machines permettant leur résolution. D'autre part, si le nombre de paramètres est trop important, l'inversion devient un problème sous-determiné dont la solution n'est plus unique. Il faut alors trouver des contraintes additionnelles qui permettent de choisir parmi toutes ces solutions.

En pratique, plusieurs astuces sont utilisées pour résoudre ces difficultés. La première consiste à réduire le nombre de paramètres en limitant le nombre de fenêtres utilisées. Ainsi, la plupart des inversions qui ont utilisé cette méthode ne considèrent souvent que 3 fenêtres (Hartzell and Heaton, 1983, 1986; Wald et al., 1990, 1991). Les solutions trouvées par ces inversions sont ainsi très contraintes a priori. Avec l'amélioration des moyens informatiques le nombre de fenêtres utilisées s'est accru (14 dans Mendoza et al., 1994). Pour choisir parmi toutes les solutions du système d'équation linéaire, les contraintes additionnelles sont variées. La plupart des inversions imposent une contrainte dite de positivité: la direction du glissement est fixe sur tout le plan de faille et le glissement ou la vitesse de glissement sont obligatoirement positifs (Hartzell et Heaton ,1983). Une autre méthode possible consiste à introduire des contraintes sur le moment sismique. Hartzell et Heaton (1983) retiennent ainsi la solution dont le moment sismique est minimum. Cette contrainte est due au fait que l'inversion par multi-fenêtrage tend à trouver des solutions dont le moment sismique est surestimé. Certains tests ont ainsi montré (Hartzell, 1989 ; Cohee et Beroza, 1994b).que lorsqu'une fenêtre est ajoutée pour donner une plus grande souplesse à l'inversion, le moment du modèle trouvé augmente. D'autres auteurs (Das et Kostrov, 1990) retiennent les solutions dont le moment est égal au moment obtenu par des estimations géodésiques ou des inversions ayant utilisé des signaux télésismiques tel que le CMT (Centroïd Moment Tensor). Une autre contrainte possible consiste à favoriser les solutions « lisses » pour obtenir des modèles le moins complexe possible. Lorsque la rupture atteint la surface comme dans le cas du séisme de Landers, on peut enfin contraindre le glissement des sous-failles situées au sommet de la faille à respecter les mesures du déplacement mesurées en surface.

Une des différences les plus importantes entre les diverses variantes de cette technique concernent aussi la forme de F(t). De nombreuses inversions ont été effectuées avec une fonction F(t) dont la

Fault cross-section:

$t_n : t_n$	1	1	!	1	1	1 /	1,j
4				-			2,j
			1				3,j
	1	1	;	1	1		4,j
```	1	1		1	1		5,j
1	1	1				N. N.	6,j

### Slip-velocity function:

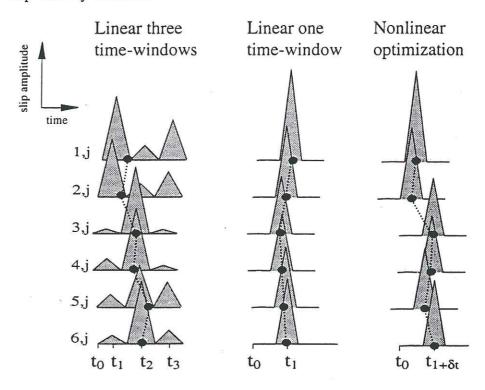


Figure 2. Schéma illustrant les méthodes d'inversion par simple et multifenêtrage. Le dessin du haut montre la faille divisée en sous-failles. La rupture se propage de droite à gauche. Le dessin du bas représente les fonctions vitesses de glissement sur les sous failles de la jème colonne. Les fonctions vitesses de glissement sont représentées par des triangles. L'inversion linéaire avec 3 fenêtres (« linear three time-window ») déjà illustrée figure 1 est comparée avec l'inversion simple fenêtrage. L'inversion simple fenêtrage comporte 2 étapes: la première est linéaire (« linear one-time window ») et consiste à trouver l'amplitude de la vitesse de glissement (amplitude du triangle). La seconde consiste à modifier (« nonlinear optimization ») les temps de rupture. D'après Cohee et Beroza, 1994b).

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dérivée temporelle est un triangle. Plus récemment, Das et Kostrov (1990) utilise une fonction F(t) linéaire avec le temps et dont la dérivée temporelle est une constante.

### Simple fenêtrage.

Cette méthode reprend la paramétrisation de la méthode par multifenêtrage mais chaque sous-faille ne rompt qu'une seule fois et la fonction source en chaque point est décrite par une seule fenêtre (Figure 2). Cette méthode est notamment utilisée par (Fukuyama and Irikura, 1986; Takeo, 1987; Beroza and Spudich, 1988; Hartzell and Iida, 1990). Elle comprend plusieurs variantes. Un premier type d'inversion (Beroza and Spudich, 1988) résout tout d'abord de façon linéaire le système d'équation (3). Le temps du début de chaque fenêtre est fixé en chaque point de la faille en utilisant une vitesse de rupture constante sur le plan de faille. Dans une deuxième étape, le temps du début de la rupture sur chaque sous faille (début de chaque fenêtre) est trouvé de façon plus précise en cherchant les perturbations à apporter à la vitesse de rupture constante prise dans la première étape. Un autre type d'inversion résout simultanément l'amplitude du glissement et le début de la rupture sur chaque sous faille (Hartzell, 1989). L'inconvénient majeur de cette technique réside dans le fait qu'elle ne peut permettre la résolution de la durée de la rupture en chaque point de la faille puisque la largeur de chaque fenêtre est fixée *a priori*.

### Méthodes dans le domaine des fréquences

Olson et Anderson (1988) proposent une approche originale de l'inversion. Leur idée consiste à utiliser non pas les traces en temps enregistrées à chaque station mais plutôt leur spectre. Dans le domaine spectral le théorème de représentation devient :

$$\hat{\mathbf{u}}_{k}(\mathbf{y},\mathbf{f}) = \iint_{\Sigma} \hat{\mathbf{G}}_{ik}(\mathbf{y},\mathbf{x},\mathbf{f}) \,\hat{\mathbf{a}}_{i}(\mathbf{x},\mathbf{f}) \, d\mathbf{S} \tag{4}$$

Le plan de faille est discrétisé par J points source à qui est attribuée une aire  $\Delta S$ 

$$\hat{\mathbf{u}}_{k}(\mathbf{y}, \mathbf{f}) = \sum_{j=1}^{J} \hat{\mathbf{G}}_{ik}(\mathbf{y}, \mathbf{x}_{i}, \mathbf{f}) \,\hat{\mathbf{a}}_{i}(\mathbf{x}_{j}, \mathbf{f}) \,\Delta \mathbf{S} \tag{5}$$

A chaque fréquence f le système des équations 5 est résolu. On détermine ainsi  $\hat{a}_i(x_j, f)$  sur tout le spectre. La transformée de Fourier de  $\hat{a}_i(x_j, f)$  permet finalement de trouver  $a_i(x_j, f)$ . Cette approche est très séduisante puisqu'elle permet de résoudre l'histoire temporelle de la vitesse de glissement (temps du début de la rupture, forme de la fonction source, vitesse de glissement maximale...) en chaque point de la faille et sans choix *a priori*. L'inconvénient majeur réside la aussi dans le nombre de paramètres à résoudre au cours du processus d'inversion. Dans un test Olson et Anderson (1988) estime ainsi que pour une faille de 10 km par 40 km et une inversion effectuée entre 0 et 2.5 Hz, le nombre de point source dicrétisant la faille est égal à 10000. Pour résoudre les deux composantes du glissement en chaque point, le nombre de paramètres est ainsi égal à 20000. Le nombre de stations 3 composantes utilisables dans ce type d'inversion est dans le meilleur des cas proche de 20. Le problème inverse résolu à chaque fréquence consiste donc à

trouver la valeur de 20000 paramètres avec 60 données. Le problème est alors clairement sous determiné et le choix de la solution nécessite là encore d'autres contraintes *a priori*. Olson et Anderson (1988) choisissent ainsi de retenir la solution la moins complexe, celle où la vitesse glissement varie le moins spatialement.

### C. Conclusions: limites des méthodes classiques.

Toutes ces méthodes quelles que soient leurs modalités pratiques souffrent des mêmes insuffisances. Avec ces méthodes le sismologue a le choix entre deux stratégies possibles. Le nombre de données étant limité, il peut d'une part choisir de chercher des modèles de source décrits par peu de paramètres et par exemple dans la méthode du multifenêtrage inverser avec peu de fenêtres. Le problème a alors une solution unique mais celle ci est très contrainte par la paramétrisation choisie (choix du début de chaque fenêtre par exemple). D'autre part, s'il veut donner une grande flexibilité à ses modèles, il doit les décrire avec de nombreux paramètres (fenêtres nombreuses et petites dans la méthode par multifenêtrage). Le problème inverse devient alors sous-déterminé, les solutions permettant d'expliquer les données sont nombreuses et la solution finalement obtenue dans l'inversion est fortement conditionnée par des contraintes additionnelles (positivité, solutions lisses, contrôle du moment sismique...) qui sont choisies a priori. Pour progresser et limiter l'influence des choix a priori dans le résultat final de l'inversion, il faut donc avoir une autre approche. Dans ce travail (chapitre II-1) nous avons ainsi essayé de trouver une paramétrisation qui permette de décrire des modèles de source dans un espace des paramètres très large (vitesses de rupture, amplitude de glissement et durée de la rupture non contraintes a priori) tout en gardant un nombre de paramètres limité pour ne pas avoir à résoudre des problèmes sous-déterminés.

Le second problème important lié à l'imagerie de la source réside dans le contrôle de l'image obtenue par l'inversion. En d'autres termes, quelle confiance pouvons-nous avoir dans les résultats de l'inversion. Toutes les inversions effectuées par des équipes différentes sur les mêmes données ont en effet montré des différences parfois significatives. C'est le cas par exemple pour le séisme d'Imperial Valley (Olson et Apsel, 1982; Hartzell et Heaton, 1983; Archuleta, 1984) ou le séisme de Loma Prieta (Beroza, 1991 ; Steidl et al., 1991 ; Wald et al., 1991). Pour juger la qualité de l'inversion, le lecteur n'a souvent à sa disposition que l'accord visuel entre les données et les synthétiques (d'ailleurs souvent représentés à des échelles différentes) et il est difficile de voir quel est le modèle qui donne le meilleur accord avec les données. De plus, comme Olson et Apsel l'ont déjà mentionné dans leurs travaux de 1982, les modèles obtenus par les méthodes d'imagerie comportent deux types de paramètres. Certains sont effectivement contraints par les données enregistrées. D'autres, au contraire, ne sont pas contraints par les données et leur valeur résulte simplement de choix a priori. Dans les images de la source obtenues dans les diverses études mentionnées ci-dessus les études de résolution sont rares (Steidl et al., 1991) et il n'est pas possible de déterminer quels sont les paramètres effectivement résolus par les données dans les diverses inversions. Nous verrons par la suite quelques propositions permettant d'évaluer les qualités de l'inversion et la validité de l'image.

### 4. Physique de la rupture: quelques problèmes intéressants

## A. Quelle est la cause de la complexité dynamique de la rupture?

Les premières approches suivies pour expliquer l'hétérogénéité de la rupture ont consisté à introduire des variations spatiales des propriétés de la faille avant le début de la rupture. Das et Aki (1977) suggèrent ainsi le concept de barrière: la faille contient des zones où la résistance à la rupture est forte et où le front de rupture ralentit ou même s'arrête complètement. Kanamori et Stewart (1978) proposent quasi simultanément le concept d'aspérité : la faille contient avant le séisme des zones qui n'ont pas cassé lors des séismes précédents et où la concentration de contrainte est importante. L'étude effectuée par Madariaga (1979, 1983) des propriétés de la radiation produite par des barrières ou des aspérités montre qu'il n'est pas possible de distinguer ces deux types d'hétérogénéités à partir de la seule étude des ondes sismiques. Les concepts de barrières et d'aspérités sont en fait des concepts complémentaires qui expriment le fait que les propriétés de la rupture sont conditionnées par le passé de la faille. Au cours de son histoire la faille a évolué de façon complexe créant des hétérogénéités (barrière ou aspérité) qui conditionneront les caractéristiques de la rupture. Pour comprendre la rupture présente du séisme nous devons donc d'une part comprendre comment évoluent les hétérogénéités sur le plan de faille (géométrie, contraintes, matériel) à diverses échelles de temps et de longueur. D'autre part, il faut essayer de modéliser la rupture sur des plans de failles hétérogènes (Day, 1982; Mikumo et Miyatake, 1983 ou Das et Kostrov, 1988).

Récemment, une nouvelle approche de l'origine de la complexité est apparue. Ces nouvelles hypothèses supposent que l'hétérogénéité de la rupture n'est pas due aux hétérogénéités des propriétés de la faille avant la rupture mais au processus dynamique lui-même. Ces nouvelles idées ont été suggérées par les travaux de Carlson et Langer (1989). En utilisant un modèle de blocs-ressorts imaginé par Burridge et Knopoff (1967) et une loi de friction dépendant de la vitesse de glissement, ils montrent qu'à partir d'un modèle de faille où les conditions initiales sont homogènes, une grande complexité dans la répartition finale des contraintes apparaît à cause de l'effet de la dynamique. Des modélisations numériques plus réalistes ont été poursuivies par quelques équipes (Okubo, 1989; Cochard et Madariaga (1994), Madariaga et Cochard (1994)) pour tester l'influence de loi de friction dépendant de la vitesse de glissement sur les propriétés de la rupture.

Déterminer si la complexité observée de la rupture (variation de la vitesse de glissement, présence de sous-événements...) est due à des variations spatiales des propriétés de la faille ou est engendrée par le processus dynamique lui-même n'est pas aisé. Il n'est en effet pas possible de mesurer les propriétés de la faille en profondeur et on ne possède pas encore de données de qualité de séismes successifs ayant eu lieu sur une même faille. Pour progresser, on peut néanmoins essayer de corréler les images cinématiques de la rupture obtenues avec les méthodes d'imagerie avec les variations de vitesse du milieu entourant la faille et obtenues par tomographie locale. Une revue

récente de ce type de travail (Chiarabba et Amato, 1994) met en évidence la corrélation entre les zones où la tomographie détecte des vitesses de propagation importante et les zones où le glissement sur la faille est le plus grand. Plus simplement, comme nous le ferons plus loin, on peut comparer l'image obtenue par notre inversion et la géométrie apparente du système de faille sur lequel s'est effectuée la rupture. Ceci permet de voir si les propriétés de la rupture sont corrélées avec les variations de la géométrie ou non.

L'imagerie de la rupture peut aussi permettre de comparer les propriétés de la rupture réelle et des modèles théoriques. Un point important concerne ainsi la durée de rupture (temps de montée) en chaque point de la faille. Les modèles numériques de la rupture utilisant des lois de friction où la friction diminue lorsque la vitesse de glissement augmente montrent que la complexité de la rupture peut être d'origine dynamique. Ils montrent aussi que le glissement en arrière du front de rupture s'arrête alors brusquement lorsque la vitesse de glissement diminue (Madariaga et Cochard, 1994; Campillo et al., 1995). Le temps de montée de ces modèles théoriques est ainsi très court ce qui correspond qualitativement à certaines observations dont Heaton (1990) fait la revue. L'évaluation du temps de montée à partir des données est ainsi importante pour comprendre si ces nouveaux modèles utilisant des lois de frictions dépendant de la vitesse correspondent à la réalité ou non. Nous aborderons plus précisément ce problème au chapitre II-2.

### B. Les failles sont elles «faibles»?

Fracture et glissement-arrêt (« stick-slip ») sous des tenseurs des contraintes fortement déviatoriques sont des phénomènes communément observés dans la vie de tous les jours et les expériences de laboratoire (crissement d'une craie sur le tableau, glissement de plaques de métal ou d'échantillons de roche avec une friction élevée). Peut on se satisfaire de ces analogies pour expliquer les glissements sismiques successifs sur les failles ?

La principale difficulté posée par ce type d'analogie réside dans le fait que les valeurs attendues du coefficient de friction (rapport entre la contrainte tangentielle et la contrainte normale au plan de faille) prédisent des valeurs de contrainte tangentielle d'un ordre de grandeur trop élevé pour être cohérente avec le flux de chaleur observé sur la faille de San Andreas (Lachenbruch et Sass, 1980, 1988). Le problème est devenu encore plus ardu avec les suggestions de Zoback et al; (1987) et Mount et Suppe (1987) proposant que la contrainte tectonique soit normale à la faille de San Andreas. Ce dernier argument confirme l'hypothèse d'une contrainte tangentielle faible sur le plan de rupture de San Andreas. La faiblesse de cette contrainte tangentielle implique la compréhension de la rupture sismique qui s'effectuerait avec des coefficients de friction bas (de l'ordre de 0.1). Une possibilité pour expliquer des coefficients de friction aussi bas et le fait que la faille soit ainsi "faible" (weak dans la littérature anglo-saxonne) consiste à introduire l'action de fluides maintenus à une pression proche de la pression lithostatique. Tout le problème consiste à comprendre comment une pression de pore peut rester élevée dans un milieu où la perméabilité de la zone faillée est supposée importante (Lachenbruch, 1980; 1991). Plusieurs mécanismes ont ainsi été proposés (Rice, 1992) et la discussion reste entière.

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L'analyse des mouvements forts peut-elle faire progresser le débat ? Classiquement la sismologie ne permet pas de trouver les valeurs des contraintes absolues sur le plan de faille, elle permet seulement une évaluation de la chute de contrainte moyenne pendant le séisme (Kanamori et Anderson, 1975) ou des variations de contraintes sur le plan de faille (Quin, 1990). Nous suggérons au chapitre III-1 que la variation spatiale la direction de glissement peut néanmoins permettre d'évaluer au moins qualitativement l'importance de la contrainte tectonique pendant le phénomène dynamique.

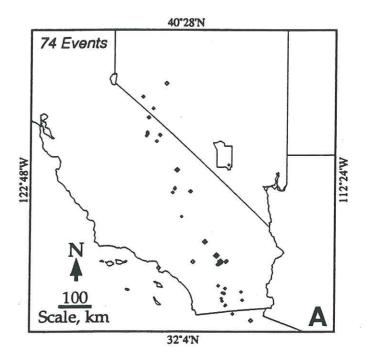
### C. Comment un séisme peut il en déclencher un autre?

Un phénomène important quant à la compréhension de la complexité de la rupture concerne les phénomènes de glissement « déclenché » (sur une faille ou une section de faille) par un glissement « déclencheur » sur une autre faille ou une autre partie de la faille.

Si le glissement est déclenché par des phénomènes dynamiques ayant lieu sur la même faille, on parle de séismes comprenant plusieurs sous-événements ou encore plusieurs barrières ou aspérités. L'étude de ce type de déclenchement se ramène à l'étude de la dynamique sur le plan de faille et par exemple aux travaux de Quin (1990).

Un deuxième type de sismicité déclenchée concerne les séismes déclenchés par les changements quasi-statiques des contraintes dans le milieu dues aux déformations engendrées par le glissement statique sur le plan de faille. Les exemples les plus connus concernent le séisme de Tennant Creek en Australie (Choy et Bowman, 1990) où plusieurs séismes importants eurent lieu sur une période de quelques jours. Un deuxième exemple concerne le déclenchement du séisme de Superstition Hill par le séisme de Elmore Ranch (Hudnut et al., 1989). Simpson (1988) décrit aussi des changements de vitesse de glissement asismique sur des failles proches d'un séisme. Ces observations confortent une idée déjà énoncée par Das and Sholz (1981): l'augmentation de contrainte statique produit par un séisme peut déclencher, après un certain délai, un séisme sur une faille proche. Dans le cadre du séisme de Landers, les changements de contraintes statiques dus au séisme ont été modélisés par Harris et Simpson (1992) et Stein et al. (1992). La corrélation entre la distribution spatiale des répliques situées en dehors de la zone de faille du séisme de Landers et les changements de contraintes statiques apparaissent clairement (Stein et al., 1992, King et al., 1994).

Un troisième type de glissement déclenché concerne les déclenchements à grande distance (plusieurs dizaines et même centaines de kilomètres) probablement dûs à des variations de contraintes engendrées par le passage des ondes. On parle alors de déclenchement dû à des changements de contraintes dynamiques. Ce type de déclenchement a été décrit plusieurs fois ces dernières années, notamment lors des séismes de North Palm Springs (William et al., 1988) ou de Loma Prieta (McClellan and Hay, 1990). Lors du séisme de Landers le déclenchement de la sismicité à des distances importantes du séisme principal fut particulièrement spectaculaire (Hill et al., 1993; Anderson et al., 1994). Un accroissement important du taux de sismicité a ainsi été observé dans la région située à l'est de la Sierra Nevada et à l'ouest du Basin and Range, sur les



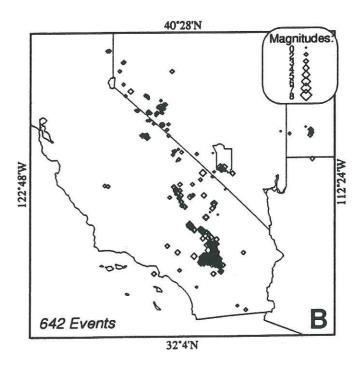


Figure 3. Séismes 24 heures avant le séisme de Landers (A) et séismes 24 heures après le séisme de Landers (B). D'après Anderson et al., 1994.

Etats du Nevada et de la Californie (Figure 3). Une étude statistique très complète a été effectuée par Anderson et al. (1994) pour démontrer la relation entre Landers et cette microsismicité. Cette microsismicité déclenchée se surimpose néanmoins à la sismicité normale de cette région et il est impossible de distinguer de façon absolue les séismes déclenchés des autres séismes. Essayons tout d'abord de regrouper les caractéristiques (Hill et al., 1993, Anderson et al. 1994, Gomberg et Bodin, 1994) de ces séismes déclenchés à grande distance.

Ils ont lieu à des distances telles de la faille que les variations de contraintes statiques peuvent y être tenues comme négligeables (Hill et al., 1993).

Leur distribution spatiale n'est pas corrélée avec la sismicité historique (Figure 3). Peu de séismes ont en effet été détectés dans la région de la faille de San Andreas où la sismicité historique est pourtant importante.

Les mécanismes sont surtout en faille normale ou en décrochement.

Leur répartition temporelle s'échelonne entre la fin de la coda du train d'onde du séisme de Landers et 80 jours après. Le nombre d'événements déclenchés diminue avec le temps selon la même loi que la diminution du nombre de répliques (Anderson et al., 1994).

Hill et al. (1993) mentionnent une corrélation spatiale entre ces séismes déclenchés et les zones où l'activité géothermale et volcanique est importante. Cette corrélation est un sujet controversé (Anderson et al., 1994). Il est vrai que dans la plupart des zones de volcanisme actif entourant Landers des séismes se sont produits dans les jours qui ont suivi Landers (Hill et al., 1993). Par contre, le taux de sismicité a lui aussi fortement augmenté dans des zones où le volcanisme et l'activité géothermale sont négligeables.

Plusieurs équipes ont essayé de mieux comprendre les causes de cette sismicité lointaine et les hypothèses sont nombreuses. Anderson et al. (1994) ont quantifié les déformations engendrées par le passage des ondes à plusieurs centaines de kilomètres du séisme. Leurs calculs montrent l'effet de la très forte directivité du séisme de Landers. La rupture unilatérale, en décrochement sur une grande distance (70 km), a focalisé les ondes vers le Nord. Les déformations à basse fréquence engendrées par le passage des ondes de surface ont ainsi été exceptionnelles si on les compare aux déformations habituellement créées par des séismes moins "directifs". Cette déformation très forte inhabituelle créée à basse fréquence par le passage des ondes de surface aurait déclenché des séismes sur des failles proche de leur seuil de rupture. Pour expliquer le délai entre le passage des ondes et le déclenchement du séisme, Anderson et al. (1994) pensent que la déformation dynamique aurait initié un glissement asismique dans la zone hypocentrale lui même responsable de l'instabilité quelques jours ou quelques heures plus tard. L'hypothèse du déclenchement par une variation des contraintes due au passage des ondes a aussi été envisagée par Gomberg et Bodin (1994). Ces derniers calculent les synthétiques en utilisant le fondamental et les premiers modes des ondes de surface. Par différentiation, ils obtiennent alors une estimation quantitative des déformations associées au passage des ondes. Ces calculs ont été effectuées pour estimer les variations de contraintes à Little Skull Mountain (LSK), où un séisme de magnitude Ms=5.4, a apparamment été déclenché (Anderson et al., 1994; Gomberg et Bodin, 1994). Ces calculs montrent que le mécanisme du séisme de LSK est favorablement orienté comparativement à la géométrie du tenseur des contraintes dynamiques dues au passage des ondes

de surface. L'amplitude de ces variations de contrainte est estimée à .2 Mpa Et les auteurs pensent que cette valeur est trop faible pour créer une fracture de type Mohr-Coulomb. Dans l'hypothèse d'un déclenchement dû aux variations dynamiques du champ de contrainte ces auteurs pensent donc que la faille était proche du seuil de rupture ou qu'un autre phénomène a été déclenché. Dans le même temps, Spudich et al. (1995) ont utilisé les données du réseau UPSAR (U.S. Geological Survey Parkfield Small Aperture Array) pour estimer les variations de déformations et de contraintes dynamiques à partir des déplacements enregistrés sur les capteurs de ce réseau. Ces calculs montrent que les variations de contraintes dynamiques à Parkfield et sans doute en de nombreux points de région de la faille de San Andreas ont été du même ordre de grandeur que ceux estimés à LSK. Dans l'hypothèse d'un déclenchement dynamique, il est alors difficile d'expliquer pourquoi il n'y a pas eu de déclenchement sur le système de faille de San Andreas où de nombreuses failles sont sans doute proche de la rupture.

Pour faire progresser le débat, nous proposons au chapitre II-2 une méthode de calcul des variations de contrainte à la fois statique et dynamique. En effet, nous avons vu que le calcul de contrainte dynamique a été effectué par Gomberg et Bodin (1994) loin de la faille de Landers (plus de 100km). Ces auteurs estiment les déplacements par sommation de modes d'onde de surface et obtiennent ensuite les variations de contrainte dynamique par différentiation. Plus près de la faille, cette méthode n'est pas valable et les évaluations de variations de contraintes demeurent statiques (Harris et al., 1992, King et al., 1994). Ces études statiques utilisent les développements analytiques de Okada (1992) et traitent le problème dans un demi espace homogène. La méthode de calcul proposée est plus complète. Ces formules analytiques permettent ainsi d'obtenir directement les variations de contraintes à la fois statiques et dynamiques dans un milieu réaliste (milieu stratifié plan) avec des sources sismiques réalistes (description cinématique d'une source étendue). Nous étudions ensuite quelques effets influant sur les variations de contraintes dynamiques (effet de la directivité et de la profondeur) et nous essayons enfin d'appliquer cette méthode au cas particulier de Landers en utilisant le modèle de source trouvé dans le chapitre II-1.

Deuxième partie Imagerie de la rupture

### Chapitre II-1

Inversion des mouvements forts dans le domaine des fréquences: application au séisme de Landers

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### II. Imagerie de la rupture

## 1. Inversion des mouvements forts dans le domaine des fréquences: application au séisme de Landers

### A. Résumé

Nous présentons une inversion dans le domaine des fréquences avec laquelle les mouvements forts sont utilisés pour contraindre les variations spatio-temporelles du glissement sur le plan de faille. Les fonctions de Green sont évaluées numériquement et les paramètres décrivant la rupture localement sont constitués par le temps de montée, le temps du début de la rupture et le glissement. Ces paramètres sont évalués sans contrainte de positivité ou lissage. Notre procédure permet de larges variations dans la vitesse de rupture. Le séisme de Landers (28 Juin 1992, Mw=7.3) a permis l'application de cette technique nouvelle à un séisme décrochant majeur. Nous modélisons l'évolution de la rupture en mettant en évidence les variations spatiales du glissement, de la durée et de la vitesse de rupture. Nos résultats sont en bon agrément avec les mesures du glissement en surface et d'autres inversions de données sismologiques ou géodésiques. Les différences majeures sont localisées en profondeur et à la fin de la faille de la Johnson vallée. Nous montrons que ces différences peuvent être dues à une résolution limitée dans ces régions. Notre modèle donne un moment égal à 0.90x10²⁰ N m. La distribution spatiale du glissement montre que cet événement est constitué par plusieurs régions où le glissement est significatif (sousévénements) séparées par des régions où le glissement est relativement moins important. Environ 50% du moment est relâché sur la faille de la Homestead vallée. Dans cette région où le glissement est important, la vitesse de rupture est bien contrainte et est égale à 3.0 km/s en profondeur et 2.5 km/s en surface. Notre inversion favorise l'hypothèse que la durée du glissement en chaque point est de l'ordre de la durée de la rupture de chaque sous-événement.

### Frequency domain inversion of strong motions: Application to the 1992 Landers earthquake

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Abstract. We present a frequency domain inversion in which the observed earthquake strong ground motions are used to constrain the space-time dependence of slip on a fault. Green's functions are numerically evaluated and the parameters describing the rupture are the local slip, rupture time and rise time. These parameters are simultaneously evaluated without additional constraints. This procedure allows for large variations in the local rupture velocity. The June 28, 1992 Landers earthquake ( $M_W = 7.3$ ) offers an exceptional opportunity to apply this technique to a major strike-slip event. We model the rupture evolution, including local differences in slip durations and variations in rupture velocity. Our final results are in good agreement with other inversion studies, geodetic and surface observations. The main discrepancies occurred at depth and at the end of the Johnson Valley fault. We show that a relatively low resolution could be an explanation for these differences. Rupture velocity and slip are extremely heterogeneous, both along strike and with depth. A moment of 0.90x10²⁰ N m was found. The slip distribution shows that this event consists of a series of regions of high slip (subevents) separated by regions of relative low slip. Approximately 50% of the moment was released on the Homestead Valley fault; in this region of large slip, the rupture velocity inferred by our inversion is well constrained and is equal to 3.0 km/s at depth and 2.5 km/s near the surface. Our inversion favors the hypothesis that the duration of the slip at each point of the fault is of the order of the duration of rupture of each subevent.

#### Introduction

Seismic records in the epicentral region are the most reliable source of information on the history of the development of a rupture and on the time function that describes locally the slip on a fault. The installation of strong motion broad band accelerometers near major faults in several places around the world now makes possible the collection of high-quality data at distances near to the source. Slip distributions on faults have been estimated for several earthquakes using strong ground motion records. Studding the 1979 Imperial Valley earthquake, Olson and Apsel [1982] and Hartzell and Heaton [1983] used a linear least squares inversion of the local strong motion to obtain the slip that occurred within each of many fault segments during several prescribed time intervals. In those linear inversions, the rupture velocity was allowed to vary only slightly, and the models considered involved a great deal of subjective decision making. To avoid this difficulty, other nonlinear approaches have been proposed which invert for both slip amplitude and rupture time [Beroza and Spudich, 1988, Beroza, 1991; Hartzell, 1989; Fukuyama and Irikura, 1986; Takeo, 1987; Steidl et al., 1991; Wald and Heaton, 1991]. These studies are time domain inversions,

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Paper number 94JB02121. 00148-0227/95/94JB-02121\$05.00 most of which are constrained by requiring that the slip is everywhere positive and by minimizing the differences between adjacent subfaults. Olson and Anderson [1988] have investigated the use of a linear frequency domain inversion in which the spatial dependence of the slip function at each frequency is related to the spectral amplitudes of ground motion at that frequency. Each frequency is inverted independently, and the total fault motion is finally obtained by a Fourier transform leading in theory to a complete description of the slip.

This paper proposes a new linearized frequency domain inversion technique. The problem is parametrized using a model in which the fault is represented by subfaults of equal area. We introduce the local slip, rupture time, and rise time as parameters to be directly and simultaneously evaluated without constraints through the inversion of the records. This intentional absence of constraints offers the possibility to study the quality of solution in terms of resolution. We apply this method to the magnitude 7.5 Landers earthquake of June 1992, 28, which was the largest earthquake to strike California in 40 years. Figure 1 shows a map of the rupture breaks together with the location of the epicenter of the Landers earthquake. Discontinuities of the fault trace clearly indicate the complexity of the faulting process. One goal of this paper is to check if the rise time and the rupture velocity can be as heterogeneous as the slip so that strong motion inversions have to take into account the possibility of spatial variation of those two parameters in addition to the slip amplitude.

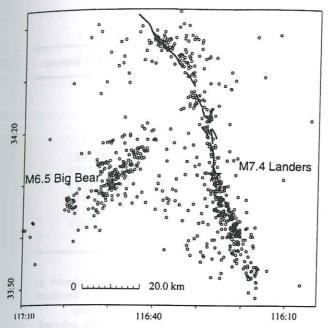


Figure 1. Epicenters of the 1992 M 7.4 Lander and the M 6.5 Big Bear (stars) earthquakes with their aftershocks (circles). Heavy solid lines represent the mapped surface rupture of the Landers earthquake (K. Sieh, California Institute of Technology, written communication, 1992).

#### **Inversion Procedure**

#### Parameterization of Strong Motion Modeling

In this procedure we divide the fault plane into small rectangular regions of equal area called subfaults in the following, and each subfault is allowed to slip once. With this parameterization, the ground motion V at a given station i and a given frequency  $\omega$  can be represented as a linear sum of n subfault contributions, each appropriately delayed in time to account for front propagation:

$$V_i(\omega) = \sum_{k=1}^{n} \operatorname{slip}_k \exp[-i\omega t_k] u_{ki}(\omega) S_k[R_k, \omega], \qquad (1)$$

where  $u_{ki}$  represents the ground motion for a unit constant slip on the subfault k with a given source mechanism,  $\omega$  is the angular frequency. Slip_k,  $t_{k}$ , and  $S_k$  are the slip, the rupture time, and the source function of the k th subfault depending on a single variable: the rise time  $R_k$  This simple parametrisation allows to limit the number of parameters of the model with respect to the technique proposed by Olson and Anderson [1988].

From a practical point of view, each subfault is represented by an array of point sources separated by a distance of less than one sixth of the shortest wavelength. The subfault contributions  $u_{ki}$  are obtained by summing the response of these point sources appropriately delayed in time to include the travel time difference due to the propagation of the rupture front across each subfault. A local rupture velocity is therefore assumed for the integration on a subfault and will be discussed later. This local rupture velocity will not be changed in the inversion process. We therefore assumed also that the rupture was unilateral during the earthquake. It is not formally excluded that a series of subfaults break sequentially in a direction opposite to the local rupture propagation on each

subfault. The technique can be refined to avoid this problem in a general case by recomputing the subfault contribution at each step of the inversion. The point source Green functions are calculated for a layered velocity model using the discrete wavenumber integration method [Bouchon, 1981] associated with the reflection transmission matrix method [Kennett, 1983].

Therefore given a layered crustal model and assuming on each subfault a constant focal mechanism and rupture velocity,  $V_i(\omega)$  can be considered as a function of  $\mathrm{slip}_k$ ,  $t_k$  and  $R_k$  if a functional form of the slip function is prescribed (in the following section the influence of this function will be discussed). In this study, the rise time, the slip, and the rupture time on each subfault are inverted simultaneously using the spectral components of the records from several stations in the vicinity of the fault according to the inversion procedure presented in the next section.

#### Inversion Procedure

The parameter vector p and data vector d are related by the function model vector f as d=f(p). The vector p consists of  $slip_k$ ,  $t_k$  and  $R_k$  which appear in equation (1). The elements of d consists of complex three-component spectra from all stations. The f function is a nonlinear function of  $t_k$  and  $R_k$ . If we assume an initial parameter vector  $p_0$ , we can get the iterative solution  $p_i$  by linearization of f around  $p_0$  at the first iteration and around  $p_{i-1}$  at each subsequent iteration i. Using the observed data vector  $d_0$  and an inversion algorithm based on the work by T arantola and V alette [1982],  $p_{i+1}$  is given by

$$p_{i+1} = p_i + b(A_i^l C_d^{-1} A_i + C_p^{-1})^{-1}$$

$$.[A_i^l C_d^{-1} (f(p_i) - d_0) + C_p^{-1} (p_i - p_0)].$$
(2)

Here  $A_i$  is the Jacobian matrix of  $f(p_i)$  (the lmth element of  $A_i$  is  $\delta f_i(p_i)/\delta(p_i)_m$ , b is a damping constant between 0 and 1 used to prevent divergence, and  $C_p$  and  $C_d$  are the covariance matrices for p and d. Because a functional form of the slip is assumed, all derivatives are evaluated analytically. Since this problem is intrinsically nonlinear the final results depend on the starting model chosen a priori.

### Quality of the Solution: Analysis of Error and Resolution

The fit to the data at each iteration i is evaluated using the misfit function S [Tarantola and Valette, 1982]:

$$S(p_i) = \frac{1}{2} \left[ (f(p_i - d_0)^t C_d^{-1} (f(p_i) - d_0) + (p_i - p_0)^t C_p^{-1} (p_i - p_0) \right].$$
 (3)

Following Cohee and Beroza [1994], we also evaluate the variance reduction between theoretical and observed seismograms defined as

$$\Delta\sigma^{2} = 1 - \frac{[d_{0} - f(p_{\infty})]^{i} C_{d}^{-1} [d_{0} - f(p_{\infty})]}{d_{0}^{i} C_{d}^{-1} d_{0}} . \tag{4}$$

This variance reduction can be computed for all the data but also for each frequency and station component.

The resolution matrix Res illustrates how well the inverse problem can be solved with our data.

$$Res = (A_{\infty}^{i} C_{d}^{-1} A_{\infty} + C_{p}^{-1})^{-1} A_{\infty}^{i} C_{d}^{-1} A_{\infty} .$$
 (5)

Station	N. t.				
	Network	Abbreviation	Latitude, deg	Longitude, deg	CSMIP Station
Goldstone Pasadena Seven Oak Dam Pinyon Flat Observatory	Тетга Тетга Тетга Тетга	GSC PAS SVD PFO	35.300 34.148 34.104 36.169	-116.804 -118.172 -117.10 -121.378	
Amboy Barstow Baker Desert Hot Spring Fort Irwing Joshua Tree Lucerne	CSMIP CSMIP CSMIP CSMIP CSMIP CSMIP Edison	AMB BAR BAK HOT FOR JOS LUC	34.560 34.887 35.272 33.962 35.268 34.131 34.558	-115.74 -117.047 -116.07 -116.509 -116.68 -116.314 116.612	21081 23559 32075 12149 24577 22170

The farther the resolution operator is from the identity, the worse the resolution is. If the diagonal element of the resolution matrix corresponding to a parameter is equal to 1, the parameter is perfectly resolved by the data set. The trace (sum of the diagonal elements) of Res indicates the number of parameters effectively resolved by the data. The values of the resolution are dependent on the a priori choice of  $C_p$  and  $C_d$ . For this reason, the resolution matrix in this study is used in a relative way to compare the relative resolution of different parts of the fault or to compare results obtained with different starting models but the same a priori variances.

### An Application of This Inversion Procedure to the Landers Earthquake

#### Data

The strong motion data used in this study come from three sources:1) accelerograms from the California Division of Mines and Geology network (California Strong Motion Instrumentation Program) which are standard SMA-1 analog recordings [CSMIP, 1992]; 2) force balance accelerometer

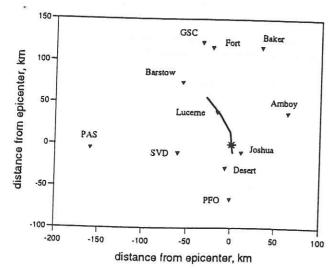


Figure 2. Locations of the stations used in this study. The stations are listed in Table 1. The solid line represents the surface projection of the fault model.

digital records from the TERRAscope network [Kanamori et al., 1991]; and 3) an SMA-2 type record from the Southern California Edison Company network in Lucerne Valley.

For each azimuth range, the closest stations were selected. Several stations located in the Los Angeles urban area to the southwest of the epicenter have been removed from the data set. The records given without absolute time were also ignored in the inversion. Table 1 gives the 11 stations used and their locations. The station distribution is shown in Figure 2, which also shows the map view or surface projection of the model fault plane. Each of the 11 stations recorded three components of ground acceleration, and all the components are used in the final inversion.

The accelerograms are band-pass filtered between 20 s and 2.0 s then are doubly integrated in the fourier domain to obtain particle displacements. The synthetics shown later were filtered in the same way. When the time is available, synthetic and observed waveforms are shown in absolute time. Since there is no absolute time available for Lucerne station, the records are not used in this inversion. Nevertheless, the fit between the synthetics and observed waveshapes at this station is shown.

In the following, 60 frequencies are used. The inversion converges in about 100 iterations (with a damping factor b equal to 0.1). The process is stopped when the decrease of the misfit function is less than 0.2 % from one iteration to the next.

#### Crustal Model

The model used in this study (Table 2) is the model described in a direct approach of the Landers earthquake rupture done by

Table 2. Crustal Structure Used in Calculation of Green's Function

Depth , km	Vp , km/s	Vs , km/s	Density, Mg/m ³	Qp	Qs
0.0	4.1	2.3	2.5	300	300
2.0	5.5	3.2	2.8	500	500
4.0	6.3	3.65	2.9	500	500
26.0	6.8	3.9	3.1	500	500
32.0	8.2	4.7	3.2	500	500

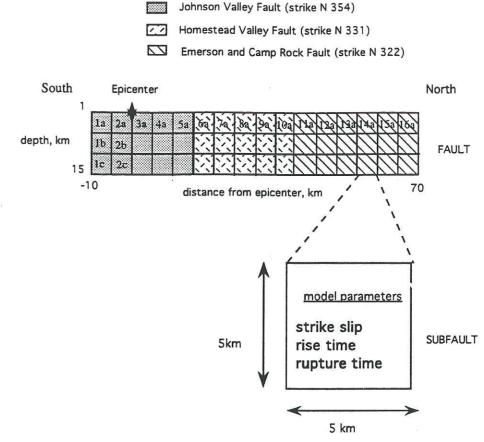


Figure 3. Parameterization of the fault model used in the inversion of strong motion records. The fault is divided in 96 subfaults for which the strike-slip amplitude, rise time, and rupture time are inverted. The epicenter defines the axis origin and is denoted by a star.

Campillo and Archuleta [1993]. This crustal model is adapted from the crustal model for southern California by Kanamori and Hadley [1975] by introducing a surficial low-velocity layer. It consists of four layers over a half-space. Campillo and Archuleta [1993] showed that this lower surficial material is necessary to model the Love waves especially at GSC.

#### Fault Model

We assume that the surface offset gives an image of the actual fault at depth. The surface breaks show that the Landers rupture involved three different fault segments with different strikes (Figure 1). This observation leads us to consider a model which consists of three distinct segments (30 km, 20 km, and 30 km long, respectivly) oriented with different strike azimuths. The three segments represent, from south to northwest, the Johnson Valley, the Homestead Valley, and the Emerson and Camp Rock faults. The fault model extends from a depth of 1 km to 16 km. The fault model is represented in Figure 3. The Landers earthquakes occurred on a series of stepping fault segments, but because of the width of the shear zones (50-200 m according to Johnson et al. [1993]) as well as the proximity and the similar strikes of the overlapping segments, it is difficult to resolve the slip on each of the overlapping segments at the periods used here (2 s and greater); we therefore idealize the faults as non overlapping planes. The onset of moment release is delayed by 3 s with respect to the origin time given by the high-frequency first

arrivals to account for a foreshock in the hypocentral region [Campillo and Archuleta, 1993]

#### Choices

In the starting model, a slip value of 4 m is prescribed on each subfault located between 0 and 60 km north of the epicenter. The others subfaults have no slip. The initial rise time given for each subfault is the same on all the fault plane.

We assume that off-diagonal element of C, and C are equal to 0. The diagonal elements of C, and C are given by the data and parameter variances. As mentioned by Hartzell [1989], the estimation of a priori variances in strong motion inversion is not easy. In our inversion we give a greater weight to the TERRAscope stations which have acquisition system of higher dynamics by giving smaller variances (C_d=0.25) to those stations compared to the others  $(C_1=1.0)$ . The final results are comparable to those presented in this paper if the same value of C₄ is assumed for all the stations. According to Spudich and Frazer [1984] the slip distribution is strongly dependent on the rupture-front velocity. This effect is clearly shown by Cohee and Beroza [1994] in their inversion of Landers strong motions: the slip distributions found with different rupture front velocities between 2.2 and 3.2 km/s are very different. Since the most important parameter to reproduce is the time at which each point ruptures, we choose to give the rupture time a greater weight in the inversion than either the slip or the rise time of each subfault. We then

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Table 3. Moment Estimates and Fit to Data Using Different Parameterizations

Number of Subfaults	Starting Velocity of the Rupture Front, km/s	Initial Rise Time on Each Subfault, s	Slip Function	Mean Resolution, %	Moment, x10 ²⁰ N m	Variance Reduction, %
32 125 48 48 48 48	3.0 3.0 2.5 3.0 3.0 3.0 3.0	4.0 4.0 4.0 4.0 4.0 3.0 2.0	smooth smooth smooth smooth rough smooth	68 36 64 65 68 67 69	0.87 0.88 0.89 0.88 0.78 0.79	65 72 71 70 63 68

assume values of smaller a priori variances for the rise time and the slip  $(C_p=81.0)$  than for the time of rupture of each subfault.  $(C_p=169.0)$ . Like Fukuyama and Irikura [1986] and Fukuyama and Mikumo [1993] these diagonal values of  $C_p$  and  $C_d$  were found after several inversion tests to find the best convergence (with a damping constant b=0.1).

The observed and synthetic amplitude spectra at each station are equally normalized by the maximum observed amplitude spectra of the three components. This means that for example, the weight given to each station in our inversion is independent of the distance of the station. Indeed, the directivity effect and decay with distance remain present in our proposed physical model.

#### Tests

Since the inversion is not purely linear, the final results are dependent on the initial parameterization of our model. This dependence leads us to test the influence on our results of some important parameters like the number of subfaults, the shape of the slip function, and the rupture front velocity of our starting model. All those tests have been summarized in Table 3. To compare the quality of the final results obtained with different starting models, the following criteria are used: 1) value of the final misfit, 2) value of the final resolution, and 3)

Table 4. Moment Estimates

Source	Estimates, x10 ²⁰ N m	Method
Sieh et al. [1993]	0.9	geological estimation
Kanamori et al. [1992]	1.1	inversion of teleseismic surface waves
Dziewonski et al. [1993]	1.1	CMT
Murray et al. [1993]	0.8	geodesy (EDM data)
Freymueller et al. [1993]	0.9	geodesy (GPS data)

CMT. Centroid Moment Tensor; EDM. Electronic Distance Measurement; GPS. Global Positioning System.

comparison of the seismic moment obtained from the inversion results with other reliable estimates. The estimates of the moment using long-period data or geological estimates yields values between 0.8 and 1.1 10 ²⁰ N m (Table 4).

To evaluate the influence of the number of subfaults, three different inversions were done with 32, 48, and 125 subfaults, respectively. As the number of subfaults increases, the resolution greatly decreases while the variance reduction remain quite the same (Table 5). Therefore the use of a finer mesh may not resolve the rupture process more accurately. In the following discussion, the model consists on 48 subfaults, each of which has a dimension of 5 km by 5 km. To compute the radiation of each of this subfaults, we used 121 point sources that corresponds to 9 points per shortest wavelength.

The influence of the value of the rise time chosen for all the subfaults in our starting model was evaluated by performing several inversions with a priori values of 2.0, 3.0, 4.0, and 5.0 s (Table 3). The best fits to the data are obtained for rise time of 3.0 and 4.0 s. Since the inversion with an a priori value of 4.0 s leads also to a moment closer to the other independant estimates (Table 4), we choose to consider a value of 4 s in the starting models of the two inversions discussed in the following.

All considerations about the value of the rise time must take into account the specific function used to describe the slip. To test the importance of this factor, we considered two different functions to represent the slip history. They are shown in Figure 4 for a rise time of 4 s. One of the functions is a simple smooth ramp while the other shows a rapid evolution of the beginning of the rupture and a progressive decay of the slope Although these functions may be regarded as roughly similar, they correspond to quite different slip rates. Therefore they lead to different inversion results, particularly in considering the seismic moment. In the following, we choose to use the function corresponding to a smooth ramp which leads to results that give a better variance reduction and a moment closer to that estimated with long-period data.

#### Results

We present two models which give similar results in terms of misfit, resolution and moment. The first one was obtained with an initial rupture velocity of 2.5 km/s (model A) and the second one was obtained with an initial rupture velocity of 3.0

Table 5. Values of the Parameters Obtained for Each Subfault Using an Initial Rupture Velocity of 2.5 km/s (Model A)

							Subfa	ult								
	1a	2a	3a	4a	5a	6a	7a	8a	9a	10a	11a	12a	13a	14a	15a	16a
Rise time, s Slip, m Time rupture, s	1.29	1.77 0.77 -1.16	0.69		3.55	1.78	2.44	3.34 3.65 10.65	3.95			4.66	4.48	0.07	1.34	0.49
							Subfa	ult								
	1 b	2b	3 b	4b	5ъ	6b	7b	8 b	9Ъ	10b	11b	12b	13b	14b	15b	16b
Rise time, s Slip, m Time rupture, s		3.77 0.26 -3.70	1.40		2.06	2.32	4.15		2.85	3.03	2.04	-0.70	4.13 4.64 22.67	-3.92	-0.60	0.6
							Subfa	ult								
B	1c	2c	3с	4c	5c	6с	7c	8c	9c	10c	11c	12c	13c	14c	15c	16c
Rise time, s Slip, m Time rupture, s		2.16 3.01 1.74		3.03	4.46 3.11 1.97	2.00	4.13 2.07 7.99	1.44 3.31 9.66	2.33	4.65	0.95	4.04 5.29 22.82		1.48 -3.10 22.47	1.11	-1.1

km/s (model B). The mean variance reduction at each frequency has been calculated with the contribution of all the stations (Figure 5), and this shows that the frequency range over which we fit the data is 0.1-0.3 Hz. Therefore our inversion has a limited spectral resolution, and only the coherent part of the process is taken into account.

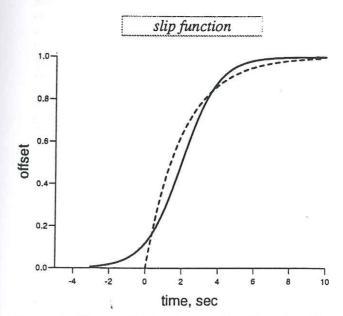


Figure 4. The two slip functions used in this study with a rise time equal to 4 s. For a given rise time R the analytical forms of those functions represented with a solid and a dashed line are respectively  $S(t) = 0.5[1 + \tanh((t+R/2.0)(R/2.0))]$  and  $S(t) = H(t)(1 - \exp(2t/R))$ , where H(t) is the Heaviside step function.

Table 5 and 6 presents the values of the parameters found for each subfault, while Figure 6a and 7a are smoothed images of the distributions of those parameters. Since our inversion is performed in the frequency domain, the modulus of the synthetic and data spectra of all the station and are shown in Figures 6b and 7b. To present a usual view of the actual fit, the corresponding observed and synthetic time histories are shown in Figure 6c and 7c. The data are the solid lines and the synthetics are plotted as dotted lines at the same scale. For both models, the horizontal components of the TERRAscope stations (PFO, PAS, SVD, and GSC) are very well matched

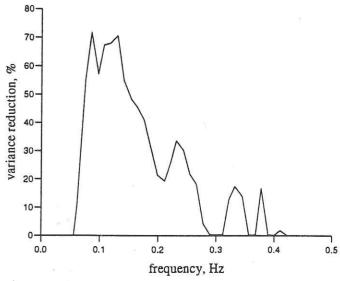


Figure 5. Mean variance reduction computed for each frequency between data of all the stations and synthetics computed with the model described in Table 5.

Table 6. Values of the Parameters Obtained for Each Subfault Using an Initial Rupture Velocity of 3.0 km/s (Model B)

							Subf	ault								
	1a	2a	3a	4a	5a	6a	7a	8a	9a	10a	11a	12a	13a	14a	. 15a	16a
Rise time, s Slip, m Time rupture, s	0.82	2.34 0.98 -1.58	0.53	1.51 1.35 1.62	3.91 -0.04 3.63	1.89 1.52 3.70	5.62 2.63 5.24	3.36 4.27 10.37	2.76 3.30 711.92	3.93 1.66 210.76	3.04 6.07 5.13.66	3.38 3.28 015.2	3 .31 3 -2.89 5 17.94	4.55 2.40 420.3	3 .73 3 -1.22 3 20.32	3.87
							Subfa									
	1 b	2ъ	3ъ	4 b	5ъ	6Ъ	7b	8 b	9b	10ь	116	12b	13b	14b	15b	16b
Rise time, s Slip, m Time rupture, s	, 0	U L	0.02	4.77	1 (1)	2.81 2.53 7.24	1 70	4 4 4				3.73 -0.30 14.90	2.26 1.17 16.84	1.53 2.44 17.50	1.98 1.89 16.91	4.54 1.40 24.65
		,					Subfau	ılt								
	1c	2c	3с	4c	5c	6с	7c	8c	9c	10c	11c	12c	13c	14c	15c	16c
Rise time, s Slip, m Fime rupture, s	3.06 1.76 -0.34	0.21	4.39 0.82 0.12	3.31 2.33 0.37	4.86 3.05 1.97	4.88 2.53 4.04	5.15 4.38 3.31	1.95 4.44 9.86	3.72 2.63 9.51	1.26 3.80 12.83	1.54 2.07 12.37	3.90 -0.30 16.86	1.56 1.17 15.39	3.83 2.44 18.27	0.54	2.66 1.40

both in amplitude and shape. The other horizontal components and even the Lucerne station, which is not considered in our inversion, fit pretty well. Each component is weighted proportionally to the maximum of the spectra of the three components, so the small -amplitude vertical components show the smallest variance reduction.

large, specially at depth (more than 3km/s). The rupture front velocity decreases in the second part of the Johnson Valley fault. This relatively fast rupture near the hypocenter is confirmed by Cohee and Beroza. [1994]. Their interpretation is that fast rupture could be facilitated by the foreshock

### Rupture Evolution

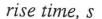
As mentioned above, a strong trade-off is expected between the rupture front velocity and the slip distribution, and our model was allowed to have strong variations of rupture front velocities. In both examples presented, one of the most striking features is that rupture velocity varies strongly with position on the fault. To have a more precise idea of this variation, the times of initial rupture of the subfaults situated at the top, middle, and bottom of the fault are compared for the two final models (Figures 8a, 8b, and 8c). The straight dashed and solid lines represent the time of rupture for a constant rupture front velocity of 3.0 and 2.5 km/s, respectively. For both models, the velocity of the final result significantly differs from the starting value. We find for all the models that a mean value of the rupture front velocity is not representative for the entire length of the fault. The most important characteristic of the two solutions is a common rupture time and rupture velocity where the slip is large (over the first 10 km from the epicenter of the Johnson Valley fault at depth, over the Homestead Valley fault between 25 and 40 km and over the top of the Camp Rock fault between 45 and 50 km near the surface). Along the other regions of the fault, the final value of the rupture velocity and thus the associated slip distribution depend on the a priori rupture velocity. Let us consider only the parts of the fault where the two inversions give the same results. Near the hypocenter the velocity is very

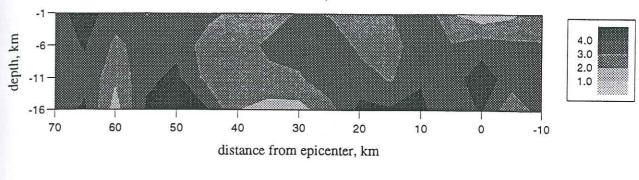
large, specially at depth (more than 3km/s). The rupture front velocity decreases in the second part of the Johnson Valley fault. This relatively fast rupture near the hypocenter is confirmed by Cohee and Beroza. [1994]. Their interpretation generating dynamic stresses that would bring the fault closer to failure just as the mainshock rupture was beginning to propagate northward. According to our inversions, it is difficult to say something about the rupture front velocity between 10 and 25 km from the epicenter. The transition of rupture between the Johnson Valley and the Homestead Valley faults is then poorly constrained. The analysis of rupture evolution along the Homestead Valley fault (between 25 and 45 km) is interesting. The results obtained with our two inversions are convergent, and it is the only part of the fault where we can compare the rupture velocities at depth and near the surface since large slip occurred over the entire depth of the fault. At depth (Figure 8b and 8c), the rupture is faster (3.0 km/s) than near the surface (Figure 8a) where the velocity is only of the order of 2.5 km/s. Such a decrease may be related to the shallow slow velocity layer and seems to confirm the idea that the rupture velocity could be proportional to the shear wave velocity. At the end of the Homestead Valley fault and on the transition with the Emerson/Camprock fault (between 40 and 45 km), we observe a deceleration of the rupture velocity at depth where the rupture ends and an acceleration of up to 3.0 km/s near the surface where the slip remains large (between 45 and 50 km).

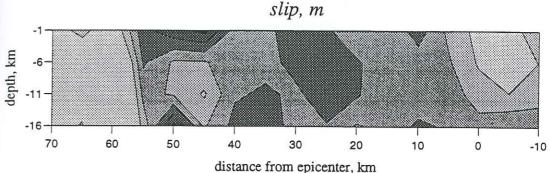
#### Slip Distribution

The total moments of both models (Table 4) fall between 0.8 and  $0.9 \times 10^{20}$  N m which is in good agreement with other

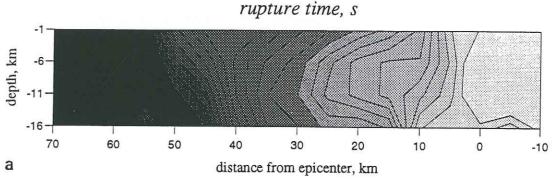
### COTTON AND CAMPILLO: FREQUENCY DOMAIN INVERSION











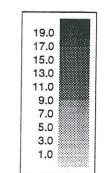


Figure 6. (a) Maps of rupture time, slip, and rise time obtained by interpolation of the results of Table 5 (model A). The 2-s contours are plotted on the map of rupture time to show the evolution of the rupture front. (b) Modulus of the observed spectra (solid line) and synthetic spectra (dotted line) obtained with the fault rupture model described in Table 5 (model A). The maximum in the quantity of mxs of each data amplitude spectrum is shown to the right of each pair. In each case, synthetics and observations are plotted at the same scale. (c) Strong motion seismograms of the Landers earthquake (solid line) compared with synthetic seismograms (dotted line) calculated using the model described in Table 5 (model A). Each pair of data and theoretical seismograms is plotted at the same amplitude scale with the variance reduction shown to the right of each pair.

estimates (Table 3). In both models, significant slip occurred on all three segments. The Johnson Valley fault has about 20% of the total moment, the Homestead Valley fault has 50%, and the Emerson-Camp Rock fault has 30%. Near the hypocenter, in the first part of the Johnson Valley fault, the slip occurred at depth. At the end of the Johnson Valley fault, according to the previous velocity discussion, we believe that the kinematic results are poorly constrained by the strong motion data. Most of the energy radiated by the earthquake comes from the Homestead Valley fault which is situated between 20 and 40 km from the epicenter. The Homestead Valley fault is the only part of the fault where the slip occurred on the entire width of the fault. Slip is important on the Camp

Rock fault, but in this part of the fault, the rupture is shallower. In both map views of the two models presented here, we can see a double peak in the predicted displacement (at about 25 and 45 km) with values of more than 5 m. If we focus our attention on the slip of the subfaults which are at depths of 1-6 km, the maximum slip occurs at a distance between 40 and 45 km from the epicenter. One can notice the precence of negative slip on the edge of the actual rupture zone. In spite of the absence of positivity constraint, the nonphysical negative slip remains small. The absence of positivity constraints allows to perform a resolution analysis that we consider to be important to achieve in strong motion inversions.

N-S

C

E-W

U-D

0.21

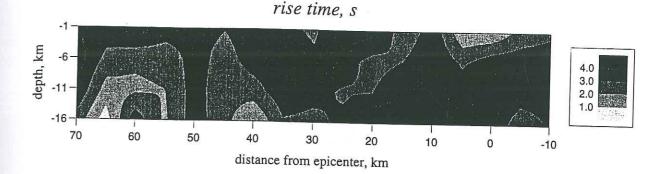
0.12

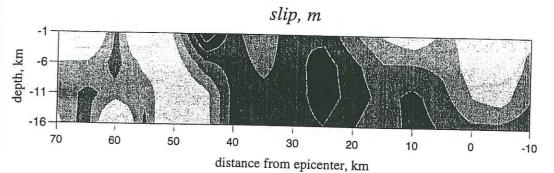
U-D

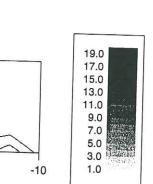
0.25 frequency, Hz

90 seconds

Figure 6. (continued)







5.0 4.0 3.0 2.0 1.0

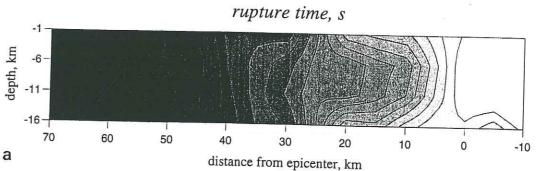


Figure 7. (a) Maps of rupture time, slip, and rise time obtained by interpolation of the results of Table 6 (model B). The 2-s contours are plotted on the map of rupture time to show the evolution of the rupture front. (b) Modulus of the observed spectra (solid line) and synthetic spectra (dotted line) obtained with the fault rupture model described in Table 6 (model B). The maximum in the quantity of mxs of each data amplitude spectrum is shown to the right of each pair. In each case, synthetics and observations are plotted at the same scale. (c) Strong motion seismograms of the Landers earthquake (solid line) compared with synthetic seismograms (dotted line) calculated using the model described in Table 6 (model B). Each pair of data and theoretical seismograms is plotted at the same amplitude scale with the variance reduction shown to the right of each pair.

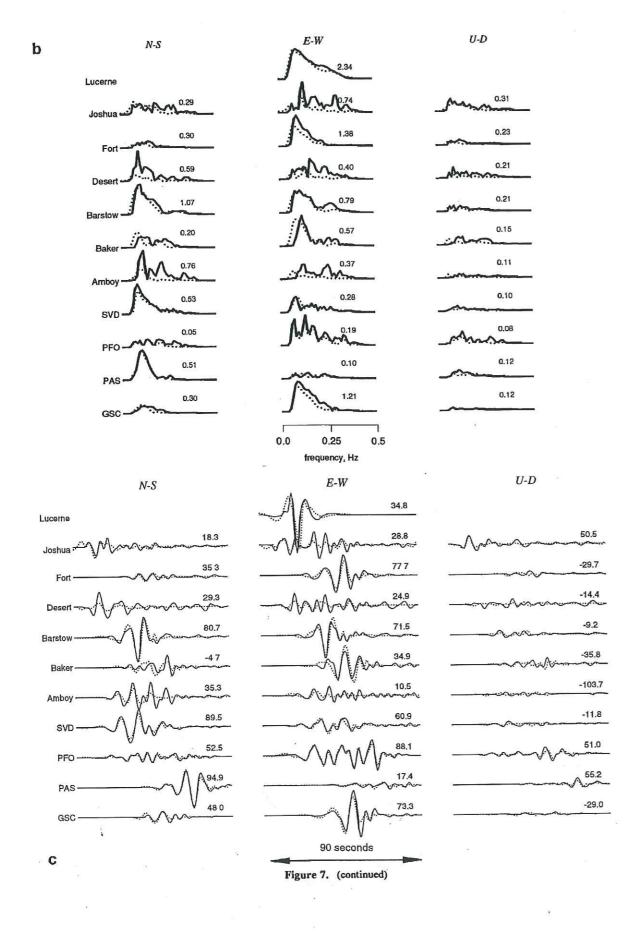
### Rise Time Distribution

An important feature of our inversion method is that the rise time on each subfault is directly inverted. The rise time distribution is heterogeneous, and if we examine the values of the rise time obtained where the slip is more than 3 m, the final values of the rise time vary from 1.1 s to 5.2 s with most of the values distributed between 3.0 s and 4.0 s The rise time is larger near the hypocenter and along the Johnson Valley fault than on the Homestead Valley fault and Camp Rock fault. In these regions of high moment release, the mean values of the rise time is between 3.0 and 3.5 s.

We already noticed that values of rise time of about 3-4 s lead to realistic values of the seismic moment. Nevertheless, it is important to check if the inversion of our data alone gives a

Table 7. Final Mean Values of the Rise Time Obtained with Different a Priori Values

Initial Rise Time on Each Subfault, s	Mean Rise Time at the End of the Inversion (Subfaults With Slip>3.0 m)
1.0	2.74
2.0 3.0	2.62
4.0	2.83
5.0	4.01



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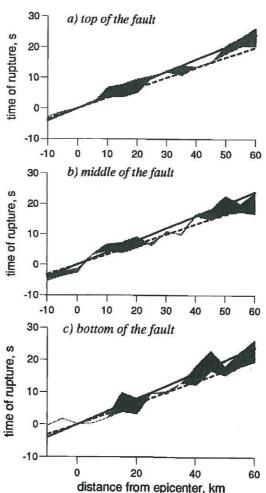


Figure 8. Comparison between the starting values of the time of rupture (straight lines) and the final one obtained by our inversion for both inversions described in the text: the dashed line represents a initial rupture velocity of 3.0 km/s and the solid line represents a starting rupture velocity of 2.5 km/s. The final results of both inversions limit a gray area. Our results are well constrained where this area is thin. (a) Subfaults situated at the top of the fault (depths from 1 to 6 km). (b) Subfaults situated at the middle of the fault (depths from 6 to 11 km). (c) Subfaults situated at the bottom of the fault (depths from 11 to 16 km).

strong constraint on the value of the rise time. We performed a series of inversion with initial values of the rise time between 1 s and 5 s. The results obtained are summarized in Table 7 where the final mean values of the rise time on the zones of high moment release are given. When considering a priori low values as 1 s or 2 s, the inversion converges to larger final values close to 3 s. On the contrary, when considering a priori values of 4 s or 5 s, the inversion tends to decrease the rise time. Indeed, the a priori value of the rise time gives the final value in the regions of low slip where the resolution is weak. This test shows that the value of the rise time is clearly constrained by the data used in this study.

The distance over which rupture is occurring simultaneously is given by the product of rupture velocity by rise time.

According to our results, this distance is at least 9 km (if we consider a mean velocity of at least 3 km/s on the region of high slip release as seen above). The regions of high slip (subevents) have characteristic dimensions between 10 and 20 km which is 50-100% of the typical distance on which the rupture is developing. One can notice that the rise time is larger in the hypocentral area.

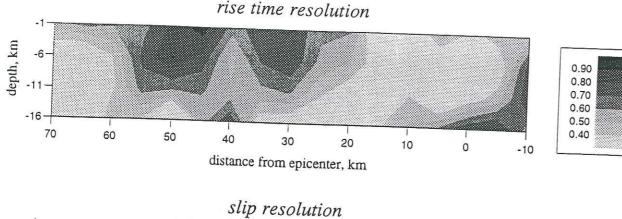
#### **Discussions and Conclusions**

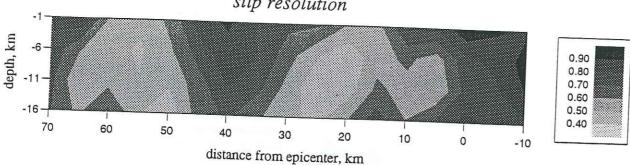
Resolution of the inversion for the case with an initial velocity of 2.5 km/s is shown in Figure 9. This resolution is computed at the last iteration. The resolution has only small variations from one iteration to another, indicating that the problem is not strongly nonlinear and therefore the iterative linearization is reasonable. This resolution map gives us an idea of the regions of the faults where our inversion is well constrained by the data. The most striking characteristic is that the resolution at depth is less than the resolution at the top of the fault which indicates that the waveforms of strong motions are more dependent on the slip of the subfault situated near the surface than at depth. This result corroborates the study of Mendoza and Hartzell [1989], who found that the strong motion records of the September, 19, 1985, Michoacan earthquake recorded at the station Caleta de Campos were relatively insensitive to slip motion across subfaults located at depth. From our imaging, we see that the resolution of the rupture time is relatively poor on the Johnson fault. In this part of the fault, our inversions show a significant difference from the inversions of local records, teleseismic data, or even geodetic data which have been performed using different approaches by Cohee and Beroza [1994] or Wald and Heaton [1994], and most of the differences between all the proposed model concern the Johnson Valley fault and the adjacent part of the Homestead Valley fault (0 to 25 km from the epicenter). The lack of resolution could be an explanation for those discrepancies. On the other hand, a common characteristic of all the models is a high moment release on the Homestead Valley fault and a shallow rupture on the Camp Rock fault, these are the faultt areas where the resolution is high.

Our models show a good agreement with others studies. For example, we found that the fault extends to a greater depth south of the epicenter. The study of guided waves trapped in the fault zone of *Li et al.* [1994] leads to a similar conclusion. According to geological studies [Sieh et al., 1993], the maximum surface slip occurred about 40 km north of the epicenter; our inversion shows similar results near the surface where the maximum slip occurs between 40 and 45 km from the epicenter. Figure 10 shows a complete comparison between the slip of the top subfaults (models A and B) and the mapped surface slip. The agreement can be regarded as a demonstration of the accuracy of our inversion.

The existence of a discontinuity at the fault bend situated at the end of the Johnson fault (20 km from the epicenter) is supported by several seismological results [Kanamori et al., 1992; Campillo and Archuleta, 1993; Li et al., 1994] and the surface offsets mapped in the field [Sieh et al., 1993]. The rupture models inferred from strong motions are more complicated than a simple two-sources model but they all show a region of relatively small slip at the end of the Johnson fault [Cohee and Beroza, 1994] or at the beginning of the Homestead Valley fault [Wald and Heaton, 1994]. Our inversion shows a similar feature with a relatively lower slip

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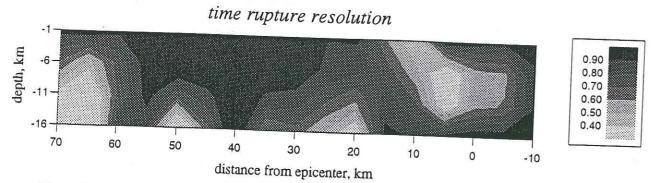


Figure 9. Maps of rise time, slip, and rupture time resolution. These maps are obtained by interpolation of the resolution matrix diagonal value calculated for each inverted parameter. Since those values depends on the choice of the a priori variance of each parameter, these maps are used for relative comparisons of the resolution

at the end of the Johnson Valley fault. Nevertheless the localization of this barrier is not perfectly constrained.

An important goal of strong motion inversion studies is to try to find a causal relation between rupture velocity variations and the slip amplitude distributions. Several studies of previous earthquakes show that the rupture velocity is not constant [Hartzell and Iida, 1990; Zeng et al., 1993] but no simple relation with the slip distribution was found. Cohee and Beroza [1994] suggest that the rupture front slows down as it encounters high-slip regions. According to Wald and Heaton [1994], there is a decrease in the rupture velocity as the rupture propagates trough the shallow regions and slows at it nears a fault step over. In this study, we show that the rupture velocity is well constrained only in regions of high moment release (bottom of the beginning of the Johnson Valley fault, Homestead Valley fault, and top of the Camp

slower at the top of the fault (2.5 km/s) than at depth (3.0 km/s), and the rupture slows down on the northern Homestead Valley fault before jumping to the top of the Emerson-Camp Rock fault. A similar feature was proposed by Campillo and Archuleta [1993]. This type of behavior for dynamic rupture from one segment to another is predicted by numerical simulation [Harris et al., 1991].

An important result of our inversion concerns the rise time distribution. The time duration of slip at a given point is a key piece of information because it depends on the type of rupture process. Two types of scenarios for the rupture process are proposed. The dislocation-type process was originally proposed by Aki [1968] for kinematic model. With this type of process, recently invoked to describe actual earthquakes by Heaton [1990] and Brune et al. [1990], the duration of the slip is small with respect to the total duration of the rupture Rock fault). On the Homestead Valley fault, the velocity is process, and there is no large interaction of the different parts

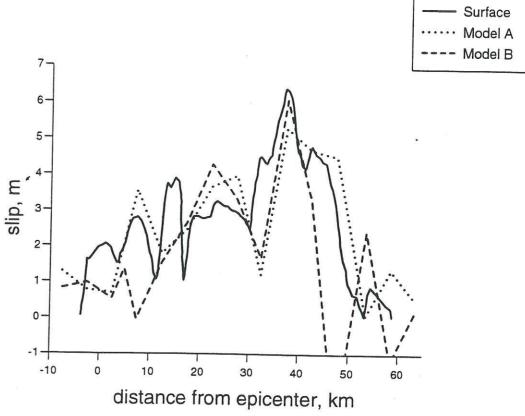


Figure 10. Comparison of fault slip solutions for the subfaults situated at the top of the fault (models A and B) and the mapped surface slip.

of the fault during the dynamic process. On the other hand, in a crack model the slip at each point depends on the slip on the entire fault and consequently the duration of the slip is on the order of the duration of the rupture process [Kostrov, 1964; Madariaga, 1976; Das and Aki, 1977]. According to our results, most of the values of the rise time are between 3.0 and 4.0 s in region where significant slip occurred. The same feature was found by Wald and Heaton [1994], who mention a duration of the order of 4 s in the regions of high slip (near 30-35 km at middepth and for the shallow slip peak at about 40-50 km along strike). The duration of slip is then of the order of the rupture duration of a subevent which have width of about 10 km and is coherent with a crack-type process occurring during each subevent. For such a large earthquake, the slip duration is however smaller than the total rupture duration .

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### Chapitre II-2

Fracture ou dislocation ? Etude de la source par analyse des mouvements forts

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## 2. Fracture ou dislocation? Etude de la source par analyse des mouvements forts

### A. Résumé

Nous présentons les résultats de deux analyses du séisme de Michoacan et de Landers. Ces deux études ont permis de contraindre la cinématique du processus de rupture. L'exemple du séisme de Michoacan montre qu'un modèle de fissure décrit de façon plus appropriée le déplacement au voisinage de la faille qu'un modèle de dislocation. Nous montrons aussi que ce point n'est pas contradictoire avec le fait que la rupture apparaisse complexe. Nous attribuons cette complexité aux instabilités dans la croissance de la fissure. Le séisme de Landers (28 juin 1992) offre l'opportunité exceptionnelle d'appliquer la méthode d'inversion décrite au chapitre II-1 à un séisme décrochant majeur. Les résultats du Chapitre II-1 sont complétés par divers tests sur la forme de la fonction source et la durée du temps de montée. La discussion sur la durée du temps de monté est approfondie et nous faisons une synthèse des résultats récents concernant les valeurs obtenues par ce paramètre afin de compléter la revue d'Heaton (1990). Nos résultats montrent que le séisme de Landers est composé de plusieurs sous-événements et favorisent l'hypothèse que la durée de la rupture est de l'ordre de la durée nécessaire à un processus de type fissure sur chaque sous-événement. Pour un séisme de cette importance, la durée de la rupture est néanmoins beaucoup plus petite que la durée totale du séisme.

### Application of seismogram synthesis to the study of earthquake source from strong motion records

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#### Abstract

We present the results of an analysis of the Michoacan and Landers earthquakes to constrain the kinematic description of the rupture process. The example of the Michoacan earthquake shows that a crack type model is better than a dislocation model to describe the displacement in the vicinity of the fault. We also show that this point is not in contradiction with the fact that the faulting appears to be a complex process. We attribute this complexity to instabilities in crack growth. The June 28, Landers earthquake offers an exceptional opportunity to apply a new inversion technique to a major strike slip event. We model the rupture evolution including local differences in slip durations and variations in rupture velocity. The slip distribution shows that this event consists of a series of regions of high slip (sub-events) separated by regions of relative low slip. Our inversion favors the hypothesis that the duration of the slip at each point is of the order of the duration of the rupture of each sub-event and is consistent with a crack type process occurring during each sub-event. For such a large earthquake, the slip duration is however smaller than the total rupture duration.

Key words strong motion – earthquake source – rise-time – inversion – Landers – Michoacan

### 1. Introduction

We compute synthetic seismograms to study the rupture process during large earthquakes. The uncertainties concerning the state of stress in the Earth and the friction on the faults are so important that we cannot use a universal source model that could be applied to particular cases. So, our purpose is modest and consists in testing the fundamental hypothesis on the rupture process. A key observation should be the time history of slip at a given point on the fault. In this respect, two scenarios are proposed:

 the dislocation type process in which the displacement on the fault is basically independent of the position and for which the duration of the slip at a point is small with respect to the total duration of the rupture process. Originally proposed for kinematic models (e.g. Aki, 1968), this type of process has recently been invoked to describe actual earthquakes. Brune (1970), Heaton (1990) and Brune et al. (1990) suggested physical explanations for such a behavior. For example, Brune's «normal motion hypothesis» should explain why there is no large scale interaction of the different parts of the fault during the dynamic process;

- the crack type process that corresponds basically to the development of an idealized crack in an elastic medium assuming Coulomb's friction law. In this case, the slip at each point depends on the slip on the entire fault and consequently the duration of the slip is of the order of the duration of the rupture process on the fault (Kostrov, 1964; Madariaga, 1976; Das and Aki, 1977). The end of the slip occurs at arrival at the point of the healing front that propagates from the periphery at the shear wave velocity. A refinement of

this model consists in taking into account the heterogeneity of the properties of the fault (Das and Kostrov, 1988) and a friction law deduced from laboratory experiments in place of Coulomb's friction law (e.g. recently Madariaga and Cochard, 1992). We do not separate these two types of scenarios into kinematic or dynamic models since short rise time should be the consequence of a slip-rate dependent friction as suggested by Heaton (1990).

To find to which category the actual earthquake processes belong is important because it has implications on the determination of friction laws and on the origin of the complexity of the seismic signals.

We present two approaches to this problem using data from two large earthquakes: the 1985 Michoacan, Mexico earthquake (Mw = 8.1) and the 1992 Landers, California earthquake (Mw = 7.4). In both cases, we compare synthetic seismograms, computed in flat layered media for different rupture models, with observed ground motions. In a layered medium, Green's function may be computed very accurately. In the following, we use a discrete wave number representation (Bouchon, 1981) combined with the reflection-transmission matrix method (Kennett, 1983) to compute theoretical Green's functions. The extended sources are represented by arrays of point sources whose distance is less than one sixth of the smallest wavelength.

### 2. The 1985 Michoacan earthquake

On September 19, 1985, a large subduction earthquake occurred beneath the Mexican states of Michoacan and Guerrero. This earthquake produced huge damage in Mexico City, about 350 km away from the epicenter. This disaster was due to the large amplification of 2 to 4 s period waves by the sediments and the clay of the Mexico City Valley (e.g. Sanchez-Sesma et al., 1988). Ripples with a period around 3 s superimposed on the long period source signals can be seen on local, regional and teleseismic records (Singh et al., 1990; Campillo et al., 1989). These ripples were

emitted a few seconds after the start of the rupture. To identify the origin of these waves is a key problem for the assessment of seismic risk in Mexico City. At the same time we have here a clear example of rupture complexity during a large earthquake with a similar signature on different types of data.

We have an exceptional opportunity to study the slip history with the displacement records at the accelerometric station Caleta de Campos (Anderson et al., 1986). This station is close to the epicenter and shows the effect of the propagation of the rupture front beneath the station. As the subduction is gently dipping at an angle of about 15 degrees, the vertical component of displacement is close to corresponding directly to the transverse component with respect to the fault plane. As seen in fig. 1, the observed vertical displacement wave form is characterized by a smooth ramp reaching the static value. The wave form expected from a crack model and a dislocation model are different since the singularities of slip velocity at the rupture front are different for the two models (Aki and Richards, 1980). We used numerical modeling to test if a difference can be seen in the configuration corresponding to the record at Caleta de Campos. We built up a homogeneous dislocation model with a ramp function.

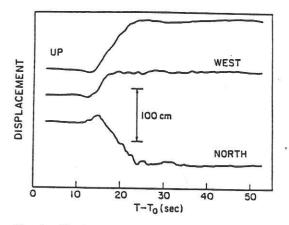


Fig. 1. Displacement records observed during the great Michoacan earthquake at the station Caleta de Campos located close to the epicenter and just above the fault zone (after Anderson *et al.*, 1986).

The rise time was chosen to be 4 s. The self similar crack is a good approximation of a finite crack until the rupture stops, as shown by Madariaga (1976). To model the passing of the edge of the crack below the station, we set up a model depicted in fig. 2a, in which the slip distribution at each time is the one of a self similar circular crack. The center of the self similar crack is moving from the hypocenter to its final position when the crack stops. This model is useful only because the discontinuity of slip velocity at the tip is the one of a crack. Both models have a seismic moment equal to the one of the first subevent obtained from teleseismic records by Houston and Kanamori (1986). The source-receiver configurations are shown as surface view in fig. 2a together with the slip along the fault for the crack model. In the two models the rupture front propagates at a constant velocity equal to 75% of the shear wave velocity. Figure 2b shows the synthetic displacements obtained for the two models at the locations indicated in fig. 2a. As suspected, the displacement produced by the dislocation model is very different from that of the crack model. The dislocation model implies an overshoot that is not present in the data. On the contrary, a model of crack growth shows the characteristics observed at Caleta de Campos. This means that at least the first part of the rupture process behaves as the growth of a large smooth crack. A similar conclusion was proposed by Yomogida (1988). This approach to analyze the slip function at the crack tip is limited to the cases where the integration of the accelerograms is sufficiently accurate to retrieve the static displacements and when the station is well located to record the passing of the rupture front.

Our conclusion that the first part of the rupture developed as a large crack may seem to contradict our previous statement that rupture complexity might have occurred to explain the existence of the 3 second period ripples. If the crack-like rupture required by the displacement records ruled out the hypothesis of a large heterogeneity of the slip on this part of the fault, it does not preclude a complexity of the history of the development of the crack. In order to check this hypothesis, we considered a self

similar crack model whose rupture velocity changes. In this case the complexity of the ground motion is entirely due to the kinematics of the rupture front while the slip distribution is the same as for a constant velocity crack. In fig. 3, we present the synthetic velocities obtained for three different kinematics of the rupture front together with the observed vertical displacement at Caleta de Campos. The traces exhibit a complexity that is the image of the kinematics of the rupture front. In particular, the synthetics show the presence of ripples similar to the ones on the observed record. Therefore, changes in the rupture velocity of a large crack is a possible explanation for the complexity of the wave forms.

In conclusion, the displacement recorded at Caleta de Campos, i.e. just above the rupture, requires that the rupture developed as a large crack. This means, in terms of rupture physics, that the process involves interactions over a large area. The complexity of the observed seismograms at a shorter period is not in contradiction with this first conclusion since instabilities in the growth of a large crack, such as rupture velocity jumps, produce wave forms similar to the observed ones. The use of the shape of transverse displacement is restricted by the fact that the double integration of the accelerograms may not be as robust as the standard processing leading to band-passed velocities. Nevertheless, it provides a «direct» view of the phenomenon that cannot be achieved with inversion techniques that rely on remote records that are strongly affected by propagation. In the following, we shall present the use of an inversion technique to investigate the source process of a large earthquake.

### 3. The 1992 Landers, California earthquake: a direct approach

The Landers earthquake of June 28, 1992 produced large surface offsets in the Mojave Desert. These surface traces indicate at least two major directions of faulting (fig. 4). According to these observations we built up a preliminary rupture model (Campillo and Archuleta, 1993) that we constrained with the

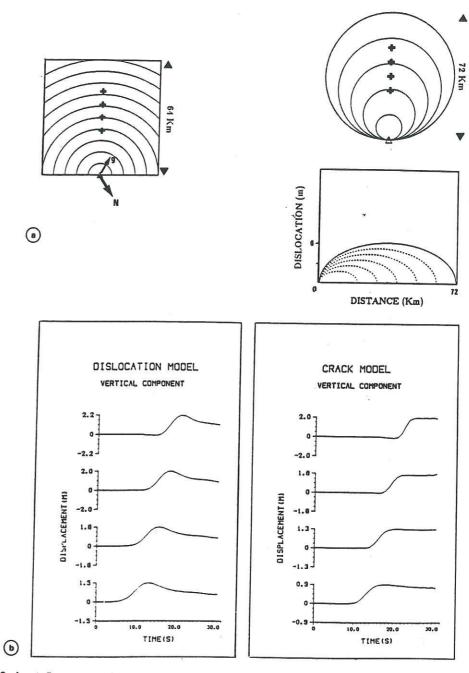


Fig. 2a,b. a) Source models used for computing synthetic seismograms: dislocation model (left) and crack model (right). The lines indicate the position of the rupture front at equal time intervals. The arrows N and g denote the direction of the north and of the slip. The crosses show the locations of the receivers at the earth surface. This figure shows the slip on the fault at equal time intervals for the crack model. b) Synthetic vertical displacements computed from these two models (adapted from Campillo et al., 1989).

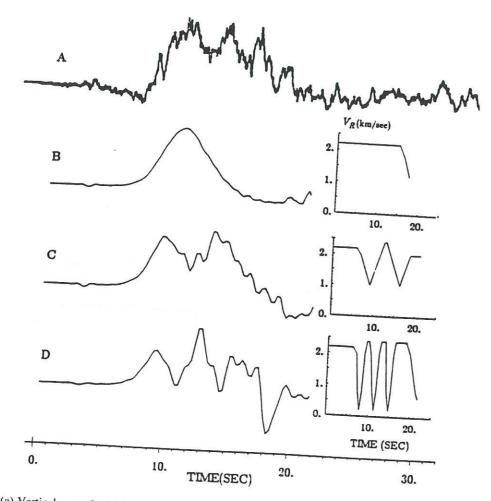


Fig. 3. (a) Vertical ground velocity recorded at Caleta de Campos, (b), (c) and (d) synthetic ground velocities computed for a crack model with variable rupture velocity. The time dependance is shown on the right side of each synthetic (after Campillo *et al.*, 1989).

displacement records obtained at the TERRAscope stations PAS, SVD, PFO and GSC. We consider a fault model that consists of two segments. We assumed a uniform slip and a uniform rupture velocity (fig. 5) on each segment. The synthetics are computed in a crustal structure (table I) derived from the model proposed for Southern California by Kanamori and Hadley (1975). The presence of dispersed Love waves on certain records requires taking into

account the superficial low velocity layer. Using a simple model characterized by a very small number of parameters we obtain a satisfactory agreement between the synthetics and the observations (fig. 6). In this case we assume a dislocation model, that is on each point of the fault segment the slip function is a ramp function with a constant final slip. The rise time and final slip are 4 s and 3.5 m for the first segment and 4 s and 5.2 m for the second segment. In this

### 1992 Southern California Earthquake Sequences

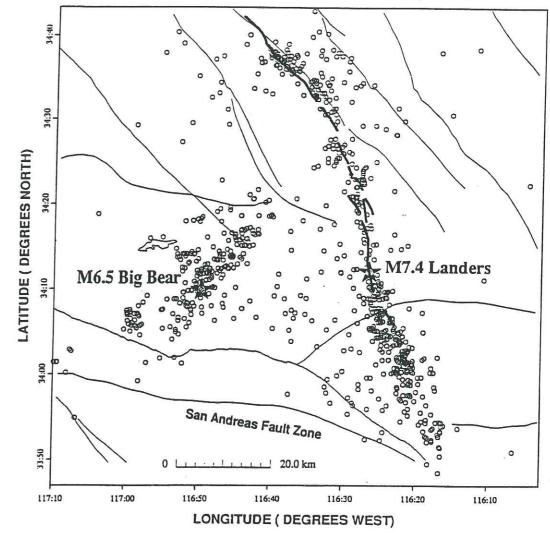
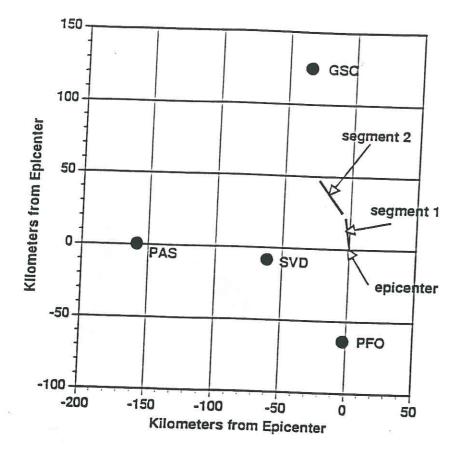


Fig. 4. Mapped surface ground rupture of the Landers earthquake (heavy lines). The stars indicate the epicenters of the Landers and Big Bear earthquakes while the circles represent the locations of the aftershocks.

model, there is a delay of 1 s between the end of the rupture on the first segment and the initiation of the rupture on the second one. This test indicates the accuracy of the numerical modeling. The misfit between simulation and observation obtained here with a small number of parameters

is the upper bound of what can be expected from further more detailed studies. Large number of records were obtained in the epicentral region during the Landers earthquake which is probably the best case to achieve an inversion of the rupture process.



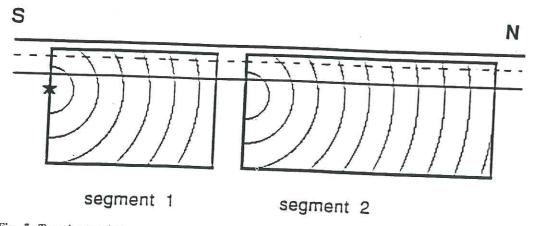
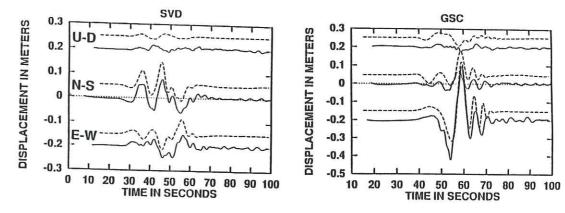


Fig. 5. Top: the two fault segments used to build a preliminary model and the location of the TERRAscope stations. Bottom: view of the fault segments indicating the propagation of the rupture front. The horizontal lines represent the interface in the flat layered crustal model used for the simulation (after Campillo and Archuleta, 1993).

Table I. Crustal structure used in calculation of Green's function.

Depth (km)	Vp (km/s)	Vs (km/s)	Density (g/cm ³ )	Qp	Qs
0.0	4.1	2.3	2.5	300	300
2.0	5.5	3.2	2.8	500	500
4.0	6.3	3.65	2.9	500	500
26.0	6.8	3.9	3.1	500	500
32.0	8.2	4.7	3.2	500	500



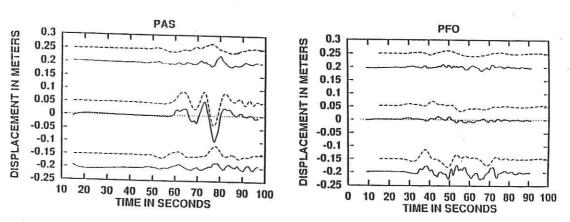


Fig. 6. Comparison between observed displacements (solid lines) and synthetics (dashed lines) (after Campillo and Archuleta, 1993).

### 4. The 1992 Landers, California earthquake: an inverse approach

### 4.1. Parameterization of a complex faulting model

In order to obtain a more detailed model of the rupture process, we used a linearized inversion of the seismograms in the frequency domain (Cotton and Campillo, 1995). In this procedure we divide the fault plane into small rectangular regions of equal area called subfaults in the following. Each subfault is allowed to slip once. With this parameterization, the ground motion (V) at a given station i and a given frequency  $\omega$  can be represented as a linear sum of n subfault contributions, each appropriately delayed in time to account for front propagation.

$$V_i(\omega) = \sum_{k=1}^n \operatorname{slip}(k) \exp^{-i\omega t(k)} U_{ki}(\omega) S_k(R_{(k)}, \omega)$$
(4.1)

where  $U_{ki}$  represents the ground motion for a unit constant slip on the subfault k with a given source mechanism,  $\omega$  is the angular frequency. Slip (k), t(k) and  $S_k$  are the slip, the rupture time and the source function of the k-th subfault depending on a single variable: the rise time R(k). This simple parameterization limits the number of parameters of the model with respect to the frequency domain technique proposed by Olson and Anderson (1988).

From a practical point of view, each subfault is represented by an array of point sources separated by a distance of less than one sixth of the shortest wavelength. The subfault contributions  $U_{ki}$  are obtained by summing the response of these point sources appropriately delayed in time to include the travel-time difference due to the propagation of the rupture front across each subfault. A local rupture velocity is therefore assumed for the integration on a subfault. This local rupture velocity will not be changed in the inversion process. We therefore assumed also that the rupture was unilateral during the earthquake. The point-source Green functions are calculated for

the layered velocity model described in table I, using the discrete wave number integration method (Bouchon, 1981) associated with the reflection transmission matrix method (Kennett, 1983). In this study the distance between 2 point sources is 500 m.

Therefore, assuming on each subfault a constant focal mechanism and rupture-velocity,  $V_i(\omega)$  can be considered a function of slip(k), t(k) and R(k) if a functional form of the slip function is prescribed (in the following section the influence of this function will be discussed).

### 4.2. Inversion procedure

The parameter vector p and data vector dare related by the function model vector f as d = f(p). In the sections 4.3. and 4.4., the value of the rise time is fixed and the vector p consists of slip (k) and t(k) which appear in eq. (4.1). In section 4.5. the value of the rise time is determined through the inversion procedure and p consists of R(k), slip (k) and t(k). The elements of d consist of complex 3-component displacement spectra from all stations of table II. We used the data from 12 stations: 4 TER-RAscope stations (Kanamori et al., 1992), 7 CSMIP accelerometers (CSMIP, 1992) and Lucerne Edison Electric Company accelerometer. The accelerograms are bandpass filtered between 20 s and 2.0 s then doubly integrated in the Fourier domain to obtain particle displacements. The synthetics shown later were filtered in the same way. Synthetic and observed wave forms are shown in absolute time. Since there is no absolute time available for Lucerne station, the records are not used in this inversion. Nevertheless, the fit between the synthetics and observed waveshapes at this station is shown.

The f function is a non-linear function of t(k) and R(k). If we assume an initial parameter vector  $p_0$ , we can obtain the iterative solution  $p_k$  by linearization of f around  $p_0$  at the first iteration and around  $p_{k-1}$  at each subsequent iteration k. Using the observed data vector  $d_0$  and an inversion algorithm based on the

Table II. Stations.

Station name	Network	Abreviation	Latitude	Longitude	CSMIP station number
Goldstone	TERRAscope	GSC	35.30	-116.804	
Pasadena	TERRAscope	PAS	34.148	-118.172	
Seven Oak Dam	TERRAscope	SVD	34.104	-117.10	
Pinyon Flat Obs.	TERRAscope	PFO	36.169	-121.378	
Amboy	CSMIP	AMB	34.560	-115.74	21081
Barstow	CSMIP	BAR	34.887	-117.047	23559
Baker	CSMIP	BAK	35.272	-116.07	32075
Desert Hot Spring	CSMIP	HOT	33.962	-116.509	12149
Fort Irwing	CSMIP	FOR	35.268	-116.68	24577
Joshua Tree	CSMIP	JOS	34.131	-116.314	22170
Lucerne	Edison	LUC	34.558	116.612	22237

work by Tarantola and Valette (1982),  $p_{k+1}$  is given by:

$$p_{k+1} =$$

$$= p_k + b \left( A_k^t C_d^{-1} A_k + C_p^{-1} \right)^{-1} \left( A_k^t C_d^{-1} (d_0 - f(p_k)) + C_p^{-1} (p_0 - p_k) \right)$$
(4.2)

Here  $A_k$  is the Jacobian matrix of  $f(p_k)$  (the ij th element of  $A_k$  is

$$\frac{\partial f_i\left(p_k\right)}{\partial\left(p_k\right)_i}\,,$$

b is a damping constant between 0 and 1 used to prevent divergence, and  $C_n$  and  $C_d$  are the covariance matrices for p and d. Because a functional form of the slip is assumed, all derivatives are evaluated analytically. Since this problem is intrinsically non-linear the final results depend on the starting model chosen a priori. In the following, 60 frequencies are used. The inversion converges in about 100 iterations (with a damping factor b equal to 0.1). The process is stopped when the decrease of the misfit function is less than 0.2% from one iteration to the next. We assume that off-diagonal element of  $C_d$  and  $C_n$  are equal to 0. The diagonal elements of  $C_n$  and  $C_n$  are given by the data and parameter variances. As men-

tioned by Hartzell (1989) the estimation of a priori variances in strong-motion inversion is not easy. In the inversion described in sections 4.4 and 4.5 we give a greater weight to the TERRAscope stations which have acquisition systems of higher dynamics by giving smaller variances ( $C_d = 0.25$ ) to those stations compared to the others ( $C_d = 1.0$ ). The final results are comparable to those presented in this paper if the same value of  $C_d$  is assumed for all the stations. According to Spudich and Frazer (1984) the slip distribution is strongly dependent on the rupture-front velocity. This effect is clearly shown by Cohee and Beroza (1994) in their inversion of Landers strong motions: the slip distributions found with different rupture front velocities between 2.2 and 3.2 km/s are very different. Since the most important parameter to reproduce is the time at which each point ruptures, we chose to give the rupture time a greater weight in the inversion than either the slip or the rise time of each subfault. We then assume values of smaller a priori variances for the rise time and the slip  $(C_p = 81.0)$ than for the time of rupture of each subfault  $(C_n = 169.0)$ . Like Fukuyama and Irikura (1986), Fukuyama and Mikumo (1993) these diagonal values of  $C_n$  and  $C_d$  were found after several inversion tests to find the best convergence.

The observed and synthetic amplitude spectra at each station are equally normalized by the maximum observed amplitude spectra of the three components. This means that, for example, the weight given to each station in our inversion is independent of the distance of the station. Indeed, the directivity effect and decay with distance remain present in our proposed physical model.

### 4.3. Quality of the solution: analysis of error and resolution

The fit to the data at each iteration k is evaluated using the misfit function S (Tarantola and Valette, 1982):

$$S(p_k) = \frac{1}{2} \left[ (f(p_k) - d_0)^t C_d^{-1} (f(p_k) - d_0) + (p_k - p_0)^t C_p^{-1} (p_k - p_0) \right]$$
(4.3)

Following Cohee and Beroza (1994), we also evaluate the variance reduction between theoretical and observed seismograms defined as

$$\Delta \sigma^2 = 1 - \frac{(d_0 - f(p_\infty))^t C_d^{-1}(d_0 - f(p_\infty))}{d_0^t C_d^{-1} d_0}$$
 (4.4)

The resolution matrix Res illustrates how well the inverse problem can be solved with our data.

Res = 
$$(A_{\infty}^{t} C_{d}^{-1} A_{\infty} + C_{p}^{-1})^{-1} A_{\infty}^{t} C_{d}^{-1} A_{\infty}$$
 (4.5)

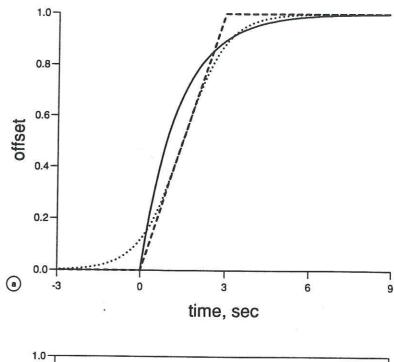
The farther the resolution operator is from identity, the worse the resolution is. If the diagonal element of the resolution matrix corresponding to a parameter is equal to 1, the parameter is perfectly resolved by the data set. The trace (sum of the diagonal elements) of Res indicates the number of parameters effectively resolved by the data. The values of the resolution are dependent on the a priori choice of  $C_p$  and  $C_d$ . For this reason, the resolution matrix in this study is used in a relative way to

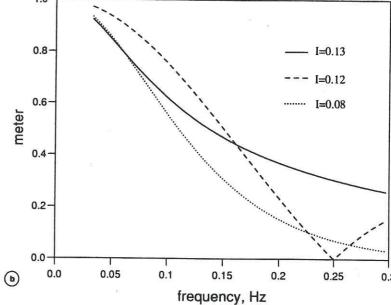
compare the relative resolution of different parts of the fault or to compare results obtained with different starting models but the same a priori variances. One can note the absence of positivity constraint in our inversion. This absence of positivity constraints allows a resolution analysis to be performed which is important in strong motion inversions.

### 4.4. Importance of the a priori slip function

All considerations about the value of the rise time must take into account the specific function used to describe the slip. To test the importance of this factor, we considered three different functions acceptable to represent the slip history. As a first stage, we tried to invert the slip distribution and the rupture time of each subfault using the TERRAscope records only and a simple model of the fault. We assumed a given value of the rise time on each subfault and we inverted for slip and rupture time of 16 subfaults 5 km long and extending in depth from 1 km to 16 km. In the following inversions, the misfit in the frequency domain is reduced by a factor of 3 with respect to the simple model proposed by Campillo and Archuleta (1993).

We present these three functions in fig. 7a,b together with the Fourier spectra of their derivatives in the frequency band selected for filtering our displacement data. Although the function used in those inversions are roughly similar, their spectra show significant differences in the frequency range relevant to the ground displacement modeling. As a test, we performed the inversion of the slip and rupture time distributions assuming a rise time of 3 s with the three different slip functions. The results are shown in fig. 8. The different slip functions lead to different absolute values of the slip while the relative distributions are quite similar. A simple explanation of this dependence is the fact that the far-field seismic radiation is proportional to the derivative of the slip function. In spite of the complex configuration of this problem (the vicinity of an extended source), the dependence of the displacement on the slip velocity remains visible. To





**Fig. 7a,b.** a) Slip functions used in this study with a rise time equal to 3 s. For a given rise time R the analytical forms of those functions represented with a solid, dotted and dashed line are respectively  $S(t) = 1/2 [1 + \tanh(t + R/2.0) * R/2.0)]; S(t) = H(t) (1 - \exp^{(-2*t)/R})$  (where H(t) is the Heaviside step function) and a ramp function. b) Spectra of the derivatives of the 3 slip functions depicted in (a).

Johnson Valley Fault (strike N 354)

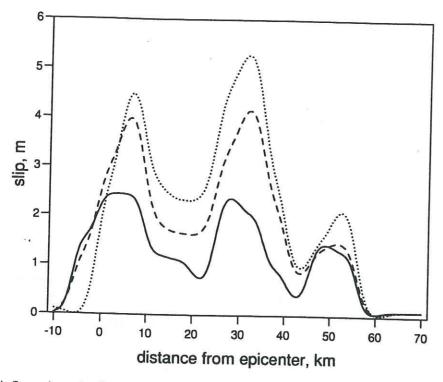


Fig. 8. This figure shows the slip distribution obtained by inversion of the TERRAscope data with the a priori slip functions shown in fig. 7a,b. In this inversion procedure, the fault has been divided into 16 subfaults. Each subfault is 5 km long and 15 km large. The curves we show are given by the interpolations of the 16 values obtained through the inversion.

give a simple quantitative consideration, we computed the integral of the slip velocity function spectra in the frequency range relevant to our inversion. The values for the three functions considered are indicated in fig. 7b. The largest slip distribution is obtained when using the slip function associated with the smallest velocity spectrum integral. In the following paragraph, we chose to use the function corresponding to the smooth ramp (denoted by a dotted line on the figures). This function leads to inversion results that are closer to the field observations than the other two tested functions.

### 4.5. Test on the rise time

In order to specify the possible values of the rise time, we performed a series of inversions

of the slip and rupture time with different rise times between 1 and 5 s. In those inversions the rise time is constant during the inversion procedure. We inverted for a more complex model with 48 subfaults as described in fig. 9. The location of the stations used in this study is shown in fig. 10. A rise time of 3.0 s leads to the smallest misfit (fig. 11). Values of 3.5 or 4.0 s give similar misfit. On the opposite values of 1.0, 2.0 or 5.0 s give significantly greater misfit. Since different a priori values of the rise time give different slip distribution, it is important to check the value of the corresponding seismic moment. The seismic moment is known from other studies (fig. 11) to be between 0.8 * 10²⁰ N-m and 1.1 * 10²⁰ N-m (Kanamori et al., 1992; Sieh et al., 1993; Murray et al., 1993; Freymueller et al., 1994). Figure 11 also shows the moment obtained from

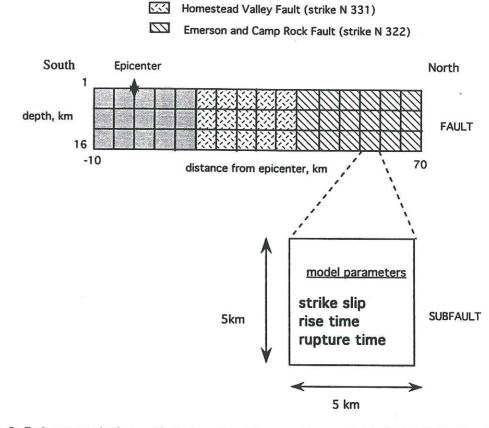


Fig. 9. Fault parameterization used in the inversion of strong motion records. The fault is divided into 48 subfaults for which the model parameters (strike-slip, rupture time and rise time) are inverted. The epicenter defines the coordinate origin and is marked by a star.

our inversion with different rise times. The moment corresponding to the smallest misfit is smaller than the values deduced from long period studies. To make a better agreement between the two evaluations of the moment, we ought to use an even larger rise time in this model. On the contrary, a small rise time value would lead to a larger gap between the two evaluations of the moment. It is important to notice that the choice of one of the two other slip functions described in fig. 8 (dashed line and solid line) would have required an even larger rise time to agree with the known seismic moment.

### 4.6. Simultaneous inversion of the rise time

We introduce now the local rise time as a parameter to be evaluated directly through the inversion of the records. The value of the rise time chosen in our starting model is 4 s. The influence of this a priori value on the final result of our inversion will be discussed in the following.

Figure 12a gives the maps of rupture time, slip and rise time obtained by interpolation of the results of the inversion. Since our inversion is performed in the frequency domain, the modulus of the synthetic and data spectra of all

the stations are shown in fig. 12b. To present a usual view of the actual fit, the corresponding observed and synthetic time-histories are shown in fig. 12c. The data are the solid lines and the synthetics are plotted as dashed lines at the same scale. For both models, the horizontal components of the TERRAscope stations (PFO, PAS, SVD and GSC) are very well matched both in amplitude and shape. The other horizontal components and even the Lucerne station which is not considered in our inversion, fit pretty well. Each component is weighted proportionally to the maximum of the spectra of the three components, so the small amplitude vertical components show the smallest variance reduction.

As mentioned above, a strong trade-off is expected between the rupture front velocity and the slip distribution and our model was allowed to have strong variations of rupture front

velocities. The most striking feature is that rupture velocity varies strongly with position on the fault. We showed in Cotton and Campillo (1995) that where the slip is large (over the first 10 km from the epicenter of the Johnson Valley fault at depth, over the Homestead Valley fault between 25 and 40 km and over the top of the Camp Rock fault between 45 and 50 km near the surface) the rupture front velocity is well constrained and near 3.0 km/s.

The total moment of our model is  $0.9*10^{20}$  N-m which is in good agreement with other estimates. Significant slip occurred on all three segments. The Johnson Valley fault has about 20% of the total moment, the Homestead Valley fault 50% and the Emerson-Camp Rock fault 30%. Near the hypocenter, in the first part of the Johnson Valley fault, the slip occurred at depth. Most of the energy radiated by the earthquake comes from the Homestead Valley

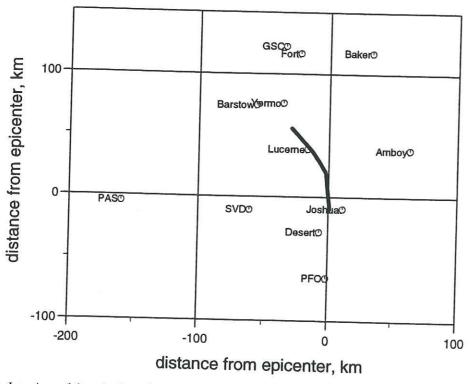


Fig. 10. Locations of the seismic stations used in this study. The stations are listed in table II. The solid line represents the surface projection of the fault model.

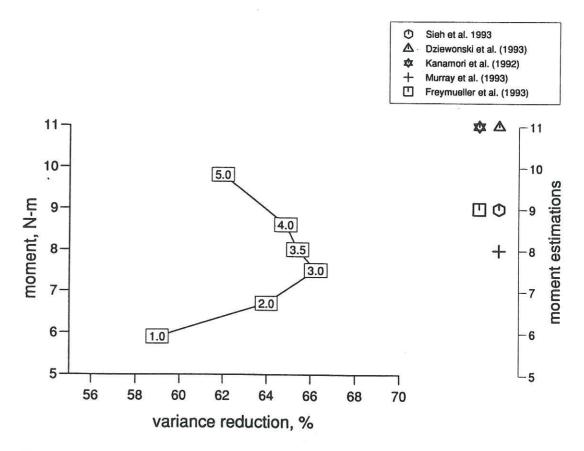


Fig. 11. Misfit and moment deduced from the rupture models computed with different rise times (in boxes) between 2 s and 6 s (left). Moment estimations (right) given by geological studies (Sieh *et al.*, 1993), geodesical studies (Freymueller *et al.*, 1994; Murray *et al.*, 1993) and long period wave studies (Kanamori, 1992; Dziewonski, 1993).

fault which is situated between 20 and 40 km from the epicenter. The Homestead Valley fault is the only part of the fault where the slip occurred on the entire width of the fault. Slip is important on the Camp Rock fault, but in this part of the fault, the rupture is shallower.

An important feature of our inversion method is that the rise time on each subfault is directly inverted. The rise time distribution is heterogeneous. The rise time is larger (more than 4 s) near the hypocenter and along the Johnson Valley fault than on the Homestead

Valley fault and Camp Rock fault. In these regions of high moment release, most of the values of the rise time are between 3.0 and 4.0 s

Since our inversion is non linear, we have to test the influence of the choice of the rise time in the starting model on our final results. We already noticed that values of rise time of about 3-4 s lead to realistic values of the seismic moment. Nevertheless, it is important to check if the inversion of our data alone gives a strong constraint on the value of the rise time.

We performed a series of inversions with initial values of the rise time between 1 s and 5 s. The results obtained are summarized in table III where the final mean value of the rise time on the zones of high moment release are given. When considering a priori low values as 1 s or 2 s, the inversion converges to larger final values close to 3 s. On the contrary, when considering a priori values of 4 s or 5 s, the inversion tends to decrease the rise time. Indeed, the a

priori value of the rise time gives the final value in the regions of low slip where the resolution is weak. This test shows that the value of the rise time is clearly constrained in a region of high slip by the data used in this study.

The distance over which rupture is occurring simultaneously is given by the product of rupture velocity by rise time. According to our results, this distance is at least 9 kilometers (if

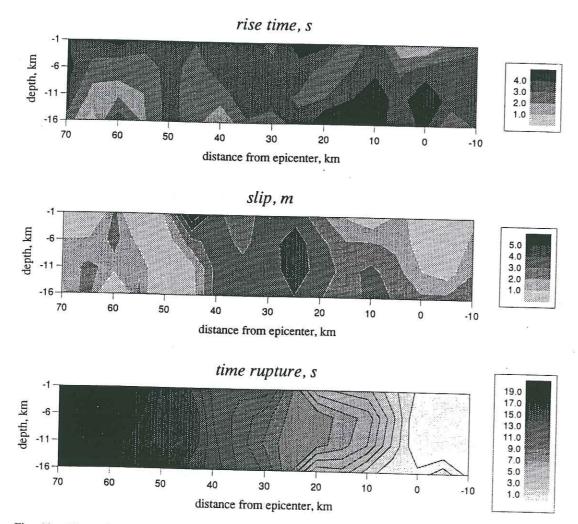


Fig. 12a. Maps of rupture time, slip and rise time obtained by interpolation of the inversion results. The 2 s contours are plotted on the map of rupture time to show the evolution of the rupture front.

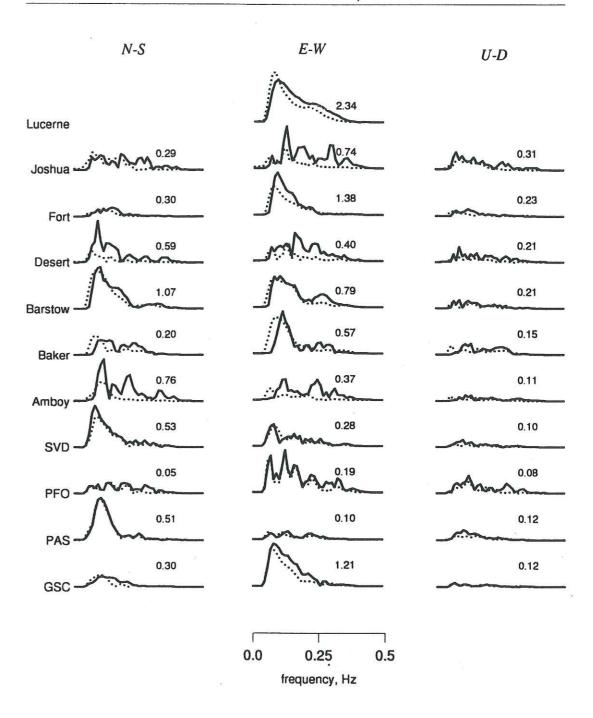
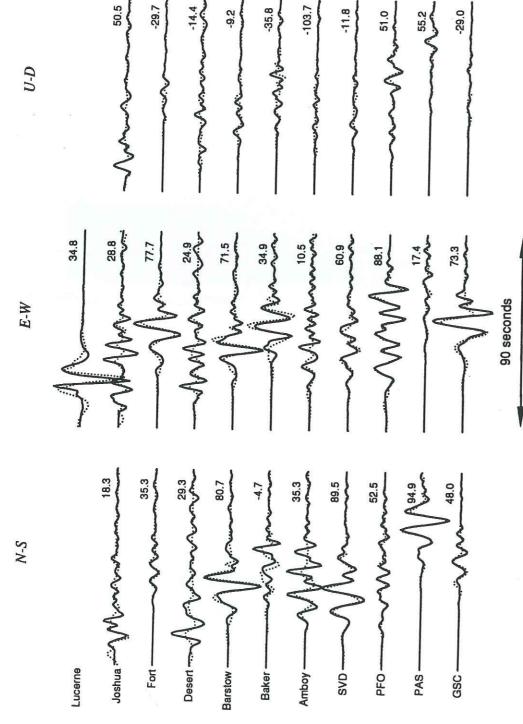


Fig. 12b. Modulus of the observed spectra (solid line) and synthetic spectra (dashed). The maximum in ms of each data amplitude spectrum is shown to the right of each pair. In each case synthetics and observations are plotted at the same scale.



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Fig. 12c. Strong motion seismograms of the Landers earthquake (solid line) compared with synthetic seismograms (dashed). Each pair of data and theoretical seismograms are plotted at the same amplitude scale with the variance reduction shown to the right of each pair.

Table III. Final mean values of the rise time obtained with different a priori values.

Initial rise time on each subfault (s)	Mean rise-time at the end of the inversion (subfaults with slip > 3.0 m)
1.0	2.74
2.0	2.62
3.0	2.83
4.0	3.37
5.0	4.01

we consider a mean velocity of 3.0 km/s on the region of high slip release as seen above). The regions of high slip (sub-events) have characteristic dimensions between 10 and 20 kilometers which is 50% to 100% of the typical distance on which the rupture is developing. One can note that the rise time is larger in the hypocentral area.

# 4.7. Control of our results: resolution analysis and comparison with other observations

Resolution of the inversion is shown in fig. 13. This resolution map gives us an idea of the regions of the fault where our inversion is well constrained by the data. The rise time resolution is poor on the Johnson Valley fault. On the other hand, the resolution is good in the region of high slip. In these regions of high moment release the mean value of the rise time ranges between 3.0 and 4.0 s. The same feature was found by Wald and Heaton (1994) who mention a duration in the order of 4 s in the regions of high slip (near 30-35 km at mid-depth and for the shallow slip peak at about 40-50 km along the strike). Nevertheless they found a smaller value of the rise time (2 s) in other parts of the fault where our inversion shows a poor resolution.

Our source model shows a good agreement with those resulting from other studies. We present in fig. 14 the distribution of slip at the surface measured directly by the geologists in

the field (Sieh et al., 1993) and the surface slip inferred from our inversion. The agreement is surprisingly good owing to the fact that our values are integrated over 5 km. Geodetic analysis indicates that in the northern end of the fault, the slip occurred essentially near the surface (Bock et al., 1992; Blewitt et al., 1992). A similar pattern can be seen in our results. We found that the fault extends in depth south of the epicenter. The study of guided waves trapped in the fault zone of Li et al. (1994) leads to a similar conclusion. Inversions of local, teleseismic data or even geodetic data have been performed by Cohee and Beroza (1994) or Wald and Heaton (1994). Most of the differences between all the proposed models occur on the Johnson Valley fault and the adjacent section of the Homestead Valley fault. On the other hand, a common characteristic of all the models is a high moment release on the Homestead Valley fault and a shallow rupture on the Camp Rock fault. A possible explanation for those discrepancies is the spatial variation of the resolution of those three different inversions. For example, our inversion shows a low resolution at depth for the Johnson Valley

# 4.8. Comparison with other inversions of the rise time

Dislocation rise time is one of the most difficult parameters to extract from the data. As demonstrated above, the consideration of this parameter has to take into account the fit to the data but also the respect for the total moment. This point was already noted by Hartzell and Mendoza in their study of the 1978 Tabas, Iran, earthquake (1991). He compared a short (0.7 s) simple triangular source-time function with a modified Kostrov function with a duration of 2.9 s. Although the wave forms used in this inversion cannot resolve between these two time functions, the function with longer rise time gives a moment estimate and maximum fault displacement more consistent with long period surface waves and field observations (table IV). Another important point is the spatial variation of the rise time. Hartzell and Heaton

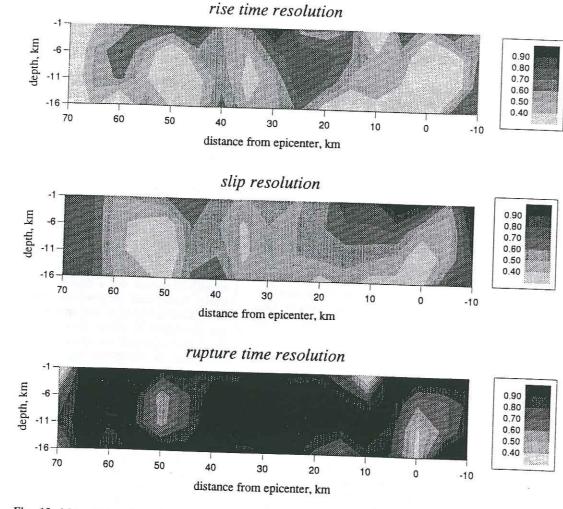


Fig. 13. Maps of rise time, slip and rupture time resolution. These maps are obtained by interpolation of the resolution matrix diagonal value calculated for each inverted parameter. Since those values depend on the choice of the a priori variance of each parameter, these maps are used for relative comparisons of the resolution in different parts of the fault.

(1983) showed that inversions that assume constant rupture velocity and rise time can lead to false results with slip patterns around the hypocenter and low total moment. In recent years, several studies have tried to specify the values of this parameter through inversion procedures which allow significant spatial varia-

tions of the source time function along the fault plane. Different procedures have been used to include the rise time as a free parameter in the inversion:

- The first approach is a discretization of the total rupture duration of each subfault

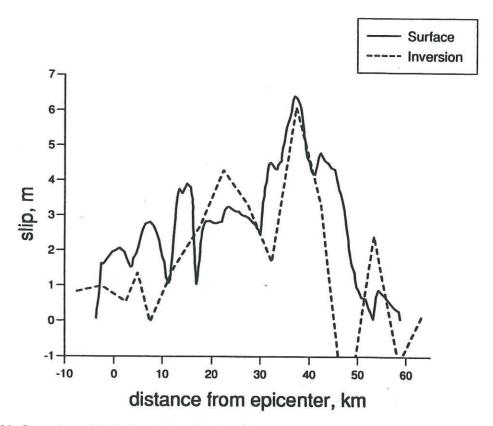


Fig. 14. Comparison of fault slip solutions for the subfaults situated at the top of the fault and the mapped surface slip.

(Hartzell and Heaton, 1983, 1986; Wald et al., 1990, 1991; Hartzell and Mendoza, 1991; Hartzell and Langer, 1993; Mendoza et al., 1994; Wald and Heaton, 1994). An elementary source-time function is used for each discrete rupture interval of each subfault. Each subfault is allowed to slip several times, allowing the possibility of a longer slip duration. In this way, general rise time functions can be constructed that vary as a function of the position on the fault. In this procedure, a maximum rupture velocity needs to be chosen that gives the earliest possible time at which any point on the fault may fail, the subfault is required to fail within the allowed time window (which also gives the maximum rise time) and the number of unknowns increases with each additional time interval added to the discretized rise time. In earlier studies using this time window parameterization (Hartzell and Heaton, 1983, 1986; Wald et al., 1990, 1991) three time windows were employed. With the improvement of computational capabilities the number of time windows has increased (14 in Mendoza et al., 1994) and more freedom has been given to the rise time and the rupture velocity.

- The frequency domain inversion method of Olson and Anderson (1988) and Mendez et al. (1990) was used to study the Michoacan earthquake (Mendez and Anderson, 1991). This approach places no a priori restrictions on the source rise time or rupture velocity. However, good data coverage is required to obtain a stable solution (Olson and Anderson, 1988).

Table IV. Rupture parameters.

Earthquake	Moment (10 ²⁵ dyne-cm) LP ^a INV ^b	l km	w km	l _a km	V _r km/s	$\frac{2 l_a}{3 V_r}$	Type of the dat used		Rise time (s) mean high slip region	Reference
3/3/85 ( $M_w = 8.0$ ) Central Chile	<b>1200</b> <i>1500</i>	220	90	50	3.0	11.1	NF° TEL ^d	multiple windowing	14	Mendoza et al. (1994)
15/9/78 ( $M_s = 7.4$ ) Tabas, Iran	<b>150</b> 52	90	40	12	2.5	3.2	NF TEL	trial error (no spatial variation allowed)	0.7	Hartzell and Mendoza (1991)
15/9/78 ( $M_s = 7.4$ ) Tabas, Iran	150 130	90	24	12	2.5	3.2	NF TEL	trial error (no spatial variation allowed)	2.9	Hartzell and Mendoza (1991)
3/10/74 ( $M_w = 8.0$ ) Peru	<b>1500</b> <i>1500</i>	250	100	50	3.0	11.1	TEL	multiple windowing	18.0	Hartzell and Langer (1993)
9/11/74 ( $M_s = 7.1$ ) Peru	63	40	20	10	3.0	2.22	TEL	multiple windowing	2.0	Hartzell and Langer (1993)
$24/11/87$ $(M_s = 6.6)$ Superstition Hill	<b>7.2</b> 4.8	20	5	7	2.4	1.9	NF	multiple windowing	1.0	Wald et al. (1990)
19/9/85 ( $M_w = 8.0$ ) Michoacan	1100 1100	170	80	50	2.8	11.9	NF	frequency domain inversion (Mendez et al., 1990)	8.0	Mendez and Anderson (1991)
18/10/89 ( $M_s = 7.1$ ) Loma Prieta	<b>20-34</b> ^e 35	47	18	10	3.0	2.22	NF	trial-error (no spatial variation allowed)	1.0	Steidl <i>et al.</i> (1991)
$18/10/89$ ( $M_s = 7.1$ ) Loma Prieta	<b>20-34°</b> 30	40	20	10	2.7	2.46	NF TEL	multiple windowing	1.0 <	Wald et al. (1991)
20/02/90 (Mjma = 6.5) Izu-Oshima, Japan	4.0	19	14	7	2.5 ?	1.86	NF	calculated from the slip functions found by the dynamic inversion	7.0	Fukuyama and Mikumo (1993)
$28/06/92$ ( $M_s = 7.3$ ) Landers	110 70	70	15	15	2.7	3.7	NF TEL	multiple windowing	2.0 <b>4.0</b>	Wald and Heaton (1994)
$28/06/92$ ( $M_s = 7.3$ ) Landers	110 90	70	15	15	3.0	3.33	NF	directly found through the inversion	3.5	Cotton and Campillo (1994)

^a Moment estimated with long period study; ^b moment calculated from the slip distribution deduced from the inversion; ^c near-field data; ^d teleseismic data; ^e for a review see Wald *et al.*, 1994.

– Another approach was proposed by Fukuyama and Mikumo (1993): a waveform inversion has been applied to strong motion data using a dynamic shear crack model. The slip time function at each subfault has been calculated from the final dynamic rupture model. From these slip functions, they calculate the rise times which are defined as the time until the slip reaches 95% of the final slip.

- In this paper, the value of the rise time is directly evaluated through an inversion procedure performed in the frequency domain. Strong spatial variations are allowed and the number of unknowns is limited by specifying the a priori form of the slip function.

In table IV we present the results of all those inversions. Following Heaton (1990), we give for each inversion a rise time derived from this inversion (mean rise time or rise time in region of high moment release) and rise times that would be expected if the duration of the slip is comparable to the time required to rupture the dominant asperity, in which case we would expect the dislocation rise time to be approximated by (Day, 1982; Heaton, 1990):

$$T_S \approx \frac{2 \, l_a}{3 \, V_r}$$

 $l_a$  and  $V_r$  are the characteristic dimensions of the dominant asperity and the mean rupture velocity. The approximate rupture length l and the approximate rupture width w are also given and for each earthquake we try to give a comparison between the moment calculated from the slip model deduced from the inversion and the moment evaluated with long period studies. We can note the very small duration (less than 1 s) of the rise time found for the Loma Prieta earthquake (Steidl et al., 1991; Wald et al., 1991; Beroza, 1991). In this case the rise time is much lower than the time required to break the dominant asperity (2.5 s). On the contrary, other studies with the same inversion procedure as the studies of the 1985, Central Chili, earthquake (Mendoza et al., 1994) or the 1974. Peru, earthquake (Hartzell and Langer, 1993) show long rise time (respectively 14 s and 18 s) consistent with the time to rupture the local asperity. Those recent results, together with our study of the Landers earthquake indicate that the rise time could be of the same order as what we expected from the breakage of an asperity.

#### 5. Conclusions

The example of the Michoacan earthquake shows that a crack type model is in this case more appropriate than a dislocation model to describe the displacement record just above the fault. We also showed that this point is not in contradiction with the fact that the faulting appears to be a complex process. We attributed this complexity to instabilities in crack growth.

The Landers earthquake was recorded by numerous stations with an acceptable azimuthal coverage. Therefore we used an inversion process to describe the slip history. Our final results are in a good agreement with other reliable observations. According to our resolution analysis the rise time values seem to be constrained only in a region of high moment release. In those regions of high slip the mean value of the rise time is between 3.0 and 4.0 s.

The typical distance on which rupture is developing simultaneously is given by the product of rupture velocity by rise time. According to our results, this distance is around 10.0 km in the regions of high moment release, that is roughly the dimension of these regions. The large earthquake studied here consists in three subevents. The analysis of the displacements favors the hypothesis that the duration of slip at a point of the fault is of the order of the rupture time of a subevent. This behavior corresponds to a crack type process, as we defined it in the introduction of this paper, for each subevent. However, the slip duration is much smaller that the total duration of the earthquake as previously argued by Heaton (1990).

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# Chapitre II-3

Histoire de la rupture et sismotectonique du séisme himalayen d'Uttarkashi

Article sous presse dans Tectonophysics

# 3. Histoire de la rupture et sismotectonique du séisme himalayen d'Uttarkashi.

## A. Résumé

Le séisme d'Uttarkashi (19 Octobre 1991) a eu lieu dans la zone où la plaque indienne est chevauchée par la plaque eurasienne. Avec une magnitude de moment egale à 6.8, ce séisme est représentatif de la déformation actuelle sur ce système en chevauchement. Nous examinons ce séisme en utilisant divers types de données. Dans un premier temps, la modélisation des enregistrements télésismiques montre que le séisme est un chevauchement peu penté et superficiel (profondeur entre 10 et 15 km). A proximité de la source, le séisme a été enregistré par un réseau d'accélérographes. Six stations trois composantes ont déclenché dans un rayon de moins de 60 km. Deux d'entre elles sont situées à courte distance du plan de faille. Une modélisation directe de la rupture montre que le séisme s'est propagé vers l'ouest. Cette modélisation nous donne la possibilité de confirmer la localisation hypocentrale et d'évaluer les temps absolus des données. Les signaux sont ensuite inversés pour retrouver la distribution du glissement sur le plan de faille. Le glissement maximum s'est produit 10 km à l'Ouest et 15 km au Sud Ouest de l'hypocentre. La variation du taux de moment sismique relaché dans le modèle obtenu avec les enregistrements de champ proche est similaire à la fonction source obtenue avec les enregistrements télésismiques. Le maximum du relachement du moment sismique a lieu 4 secondes après le début du séisme. Les relations entre la distribution spaciale du glissement, la localisation des failles et des répliques sont ensuite discutées et nous proposons finalement une interprétation tectonique du séisme d'Uttarkashi. Le séisme d'Uttarkasi a probablement eu lieu le long du détachement situé au sommet de la plaque indienne subductée. Ce détachement plonge avec une pente faible sous le Bas-Himalaya au sud du chevauchement de Vaikrita. Ce chevauchement marque la limite séparant la partie sismique du détachement (sur laquelle des séismes comme le séisme d'Uttarkashi peuvent avoir lieu) de la partie plus pentue et asismique. Cette conclusion est importante pour estimer correctement le risque sismique dans la région d'Uttarkashi.

# Rupture history and seismotectonics of the 1991 Uttarkashi, Himalaya earthquake.

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#### Abstract

The 19 October 1991 Uttarkashi, India, earthquake occurred in the Himalaya main thrust zone. With a moment magnitude of 6.8, this event is characteristic of the present-day motion on the thrust fault system. We examine this earthquake using different sets of data in order to understand better the faulting process of a major earthquake in the Himalayan region. Firstly, the modeling of the teleseismic records indicates that the mechanism is similar to the published CMT and indicates a shallow (between 10 and 15 km depth) low angle thrust event. In the vicinity of the source, the earthquake was recorded by a network of accelerometers run by the University of Roorkee. Six three components accelerometers were triggered within a radius of 60 km. Two of them were very close to the surface projection of the fault. Forward modeling of those records shows that the rupture propagated toward the west. This forward modeling gives us the possibility to confirm the epicenter location and to evaluate the timing of the accelerograms. The accelerogram records are inverted to obtain the distribution of slip on the fault plane. The results show a complex rupture process. The slip maxima (1.5 m) occurred 10 km West and 15 km South-West of the hypocenter. The slip source function obtained with near-field data is similar to the function obtained from teleseismic records and shows a low moment release at the beginning of the rupture and a maximum rate of moment release 4 seconds after. The relation between the slip distribution obtained by inversion, isoseimals, mapped faults and the aftershocks location is then discussed and we finally propose a seismotectonic interpretation of this earthquake. The Uttarkashi earthquake probably occurred along the detachment surface which coincides with the upper surface of the subducting Indian lithosphere. This detachment surface is gently dipping under the lesser Himalaya and south of the Vaikrita thrust. The Vaikrita thrust marks the line separating the very shallow-dipping detachment (along which big earthquakes like the Uttarkashi earthquake could occur) from the steeper-dipping, aseismic, basement thrust. This observation is

#### Uttarkashi, Himalaya, India

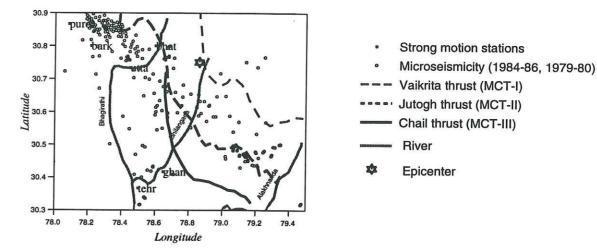


Figure 1 . Map of the Uttarkashi region showing the strong motion stations used in this study, the main thrusts of this region and the microseismicity recorded by the University of Roorkee in 84-86 and 79-80.

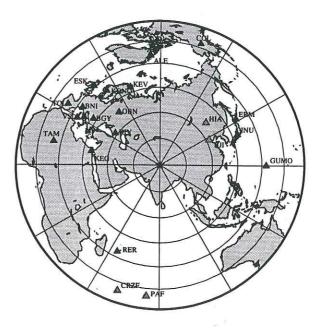


Figure 2. Azimuthal coverage of the source around the epicenter (full star in the center) by the broad-band digital stations used in this study. Full triangles are for VBB records, open triangles for BRB records.

important for correctly estimating the seismic hazard in the Uttarkashi region.

#### B. Introduction

The 20 October 1991 Uttarkashi earthquake (Ms=7.0) killed over a thousand people and caused important economic loss in the Garhwal, Himalayan region. The earthquake occured in the Himalaya mountains, north India at the plate boundary between the colliding Indian and Asian plates. The earthquake also generated a great deal of interest because it is located within a distance of 50 km from the site of the Tehri dam under construction to the south. In Garhwal region, several earthquakes are known to have occurred in the historical times (Rastogi, 1995). The most important one, the event of 1803 with intensity IX, was highly destructive (Smith, 1843). The study of this 1991 earthquake is then important for the assessment of hazards in this region and more generally in seismically active collision environments.

This earthquake is the largest well-recorded earthquake in the Himalaya to date. We examine this earthquake using all available types of seimological data (aftershocks, teleseismic, strong-motions records) in order to understand better the faulting process in the Himalayan collision environment. In a first step, we have used a waveform inversion of teleseismic data to obtain the focal mechanism and the time history of moment release of the Uttarkashi earthquake. This earthquake had also the unique chance to be recorded by a network of accelerometers in the near-field. These strong-motion records represent the most complete set of accelerograms for a thrust earthquake in the Himalayan region (Chandrasekaran and Das, 1992). Methods for estimating slip distributions and rupture characteristics using waveform inversions of near-source seismograms have been also developed and applied to many earthquakes in California (Olson and Apsel, 1982; Hartzell and Heaton, 1983, Beroza and Spudich, 1988; Cohee and Beroza, 1994; Wald and Heaton, 1994; Cotton and Campillo, 1995), in subduction (Fukuyama and Irikura, 1986; Takeo, 1987; Mendoza et al., 1994) and intraplate zones (Hartzell and Mendoza, 1991) but never in a collisional tectonic context like the Himalayan orogen. These strong motion records give us the opportunity to apply a newly developed, deterministic, inversion technique (Cotton and Campillo, 1994, 1995) to a major thrust himalayan earthquake.

This event is characteristic of the tectonic processes currently affecting the Garhwal province. It provides information on the present-day motion on the thrust fault system and the relative roles of the different faults in accomodating the plate tectonic motion. In this paper, we examine the relationship between the different faults, the main shock and the aftershocks of the Uttarkashi earthquake.

## Seismotectonic setting

Uttarkashi lies near the plate boundary between the Indian and the Asian plates. From the pioneering work of Argand (1924), numerous authors discussed the collision between these two plates and the formation of the Himalaya (for example: Gansser, 1964; Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1977; Le Fort, 1975; Seeber et al., 1981; Ni and Barazangi, 1984; Lyon-Caen and Molnar, 1983). The total convergence between India and Asia has been estimated

Table 1. Stations used in the body-waves teleseismic inversion. Type of record concern the frequency band: VBB .005-10Hz, BRB .06-2Hz

Code	^a Institution	Type	Azimut	Distance	Used wave
ALE	IRIS	VBB	354.8	65.4	P,S
AQU	MEDNET	<b>VBB</b>	302.0	52.6	P
BGY	<b>MEDNET</b>	<b>VBB</b>	304.4	47.2	P
ВЛ	CDSN	BRB	62.5	31.6	P,S
BNI	<b>MEDNET</b>	VBB	306.4	56.9	P,S
COL	GSN/USGS	<b>VBB</b>	18.5	78.0	P,S
CRZF	GEOSCOPE	BRB	198.5	80.6	P
ERM	IRIS	<b>VBB</b>	18.5	78.0	P,S
ESK	IRIS	<b>VBB</b>	319.6	60.9	P,S
GUMO	GSN	<b>VBB</b>	90.0	62.7	P
HIA	CDSN	BRB	47.1	35.9	P
INU	<b>GEOSCOPE</b>	<b>VBB</b>	68.3	48.4	P,S
KEG	<b>MEDNET</b>	<b>VBB</b>	281.2	40.3	P
<b>KEV</b>	<b>DWWSSN</b>	<b>VBB</b>	338.6	48.5	P,S
KIV	IRIS	<b>VBB</b>	305.1	31.3	P P
KONO	GSN	VBB	323.9	53.6	P
OBN	IRIS	VBB	321.7	38.5	P,S
PAF	<b>GEOSCOPE</b>	BRB	185.7	80.1	P P
RER	GEOSCOPE	VBB	206.1	56.1	P,S
<b>TAM</b>	GEOSCOPE	VBB	164.9	66.0	P,S
TOL	GSN	VBB	303.3	66.0	P,S
VSL	MEDNET	VBB	299.3	56.2	P P

^aCDSN: Chine Digital Seismograph Network; GDSN: Global Digital Seismograph Network; GSN: Global Seismographic Network; GEOSCOPE: Project Geoscope; IRIS: Inorporated Research Institutions for Seismology; MEDNET: Mediterranean Network

Table 2. Focal parameters of the Uttarkashi earthquake

Source	localisation	Moment (*10 ¹⁹ N-m)		Strike	Dip	Slip	Depth (km)
^a IMD	30.75 N		***************************************	~~~~			12.0
	78.86 E						12.0
CMT							
(Harvard)	30.22 N	1.8	NP1	317°	14°	115°	15.0
	78.24 E			112°	78°	84°	13.0
CT TO GO					, 0	01	
°USGS		8					
	30.78 N	1.8	NP1	296°	5°	90°	10.0
	78.77 E		NP2	116°	85°	90°	10.0
This study		1.2		318°	11.0°	114°	14.0

to be about 2000-3000 km. About 300-500 km could have occurred along the Himalaya by repeated overthrusting of Indian Crust on itself. These thrusts separate geological units with characteristic geology (Gansser, 1964). The southernmost unit, named the outer Himalaya, consists of the folded and faulted Siwalik molasse sediments of Miocene age. These molasse sediments form the low hills that rise in front of the plains of the Sindhu-Ganga basin. The Outer Himalaya is separated from the Lesser Himalaya in the North by the Main Boundary Thrust (MBT). The most northerly subprovince is the Great Himalaya which is made up of crystalline metamorphic and igneous rocks. The slab of the great Himalaya is thrusting southward along the Main Central Thrust (MCT) over the Lesser Himalaya. In the Uttarkashi region, the problem is more complex because it is difficult to locate precisely the MCT. There are three major sets of northward dipping thrusts which are called, from north to south the Vaikrita Thrust (MCT-I), the Jutogh Thrust (MCT-II) and the Chail Thrust (MCT-III) (Figure 1). The inclination of these thrust planes varies between 30 and 40 degrees (Valdiya, 1980). Seeber et al. (1981) postulated a gently dipping thrust under the lesser Himalaya. They called this plane the detachment surface which coincides with the upper surface of the subducting Indian lithosphere. The MBT and MCT which dip steeply (30-40 degrees) near the surface flatten out at depth and merge with this surface. Downdip from the interplate thrust zone the detachment presumably becomes steeper and eventually bends back to merge with a subhorizontal aseismic interplate zone beneath the northern Tethyan Himalaya and Tibet.

The distribution of seismicity throughout the Himalaya appears to form a belt of moderate-sized events (mb<6) beneath the Lesser Himalaya between the MCT and the MBT (Molnar et al., 1973, 1977; Chandra, 1978; Seeber et al, 1981, Ni and Barazangi, 1984). Most of those events are located just south of the MCT. Medium-sized earthquakes with fault plane solutions and well determined focal depths determined by Ni and Barazangi (1984) define a simple planar zone from about 10-km to 20-km depth, with an apparent dip of 15 degrees. This planar zone apparently defines the detachment that separates the underthrusting Indian plate from the Lesser Himalaya crustal block and along which, according to Seeber et al. (1981) and Ni and Barazangi (1984) occurred the great Himalayan earthquakes which have struck the Himalayan arc during the past 90 years. Four magnitude 8 earthquake have occurred since 1897 on the Himalaya plate boundary. The Uttarkashi earthquake occured near the northwestern part of a seismic gap called the Central gap by Khattri (1987).

# D. Study of the rupture using teleseismic data

The use of full waveform modeling of long period teleseismic data has improved our knowledge of the rupture process of large earthquakes with teleseismic data (Nábèlek, 1985, Mendoza and Hartzell, 1989; Campos et al., 1994). In a first step, we have used a full waveform inversion of teleseismic data to determine the focal mechanism and the time history of moment release of the Uttarkashi earthquake. We have selected 34 broad-band body wave (22 P and 12 S) records from IRIS/GSN, MEDNET and GEOSCOPE networks to obtain the seismic moment tensor, the mean focal depth and the general features of the source in time as seen from distant

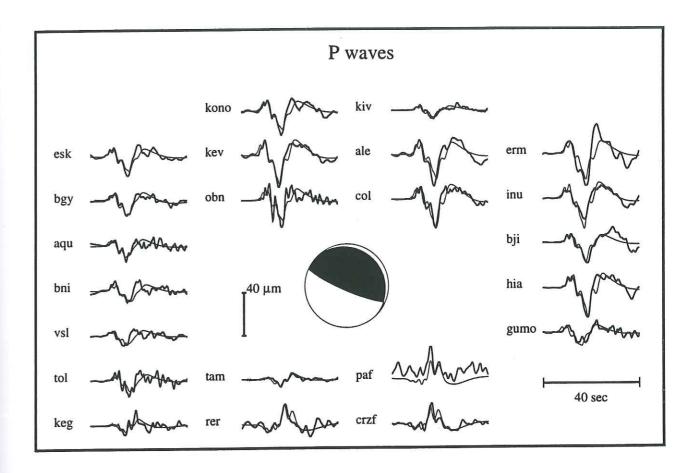


Figure 3. Comparison of observed P wave records (thick line) in displacement and synthetic seismograms (fine line) for the best point source model of the 1991 Uttarkashi earthquake found by our teleseismic body-wave inversion. All of them are filtered between 0.02 and 2 Hz, normalized to a recording distance of 40 degrees and plotted at the same amplitude scale. The focal mechanism (strike=318°, dip=11°, rake=114°, depth=14 km, seismic moment=1.2 10¹⁹ Nm) and the source-time function of this model are shown in the middle and the right top of the figure.

stations. The station distribution at distance range from 30 to 80 degrees is good (Table 1 and Figure 2). Since the VBB coverage is poor in the lower hemisphere, we are required to consider together BRB and VBB stations. All the records are numerically converted to the same response curve and displacements are inverted in the frequency range of 0.02 to 2 Hz. The procedure proposed by Nábèlek (1984) is used in the inversion. This procedure allows a multiple source inversion and solves simultaneously for focal mechanism, centroïd depth and source time function. The first inversion was performed assuming a single point source associated with a single mechanism and a source duration of 14 seconds composed of two seconds elementary triangle sources. The results give a good fit to the data with a centroid depth of 14 km, a global seismic moment of 1.2 1019 N-m, a strike of 318°, a dip of 11° and a rake of 114°. The source time function (Figure 3) shows 3 sharp sub-events with a total duration of 8 seconds. A small precursor could be observed on seismograms at stations where the signal to noise ratio is low like stations HIA and BJI, but is not determined by the inversion procedure. The results clearly show a complex source time. The main seismic moment release occurred after an initial pulse of 2 seconds duration. The source function of this main event shows three sharp sub-events of a total duration of 8 seconds. The comparison between observed and calculated seismograms in displacement is quite good in the initial part of the signal and deteriorates later when crustal reverberations appear that are not taken into account in our modelling.

We tried in a second step to determine a source directivity, assuming that the 3 different parts of the source are associated with different point sources, with slightly different mechanisms along the N318 direction. The resolution of the teleseismic data for such details is poor, and none of our results was convincing enough to imply more on the source mechanism process with this data set.

The results obtained for the focal mecanism are quite comparable to those obtained by other teleseismic analyses (Table 2). These teleseismic inversions give values of the moment between 1.2 10¹⁹ N-m (our study) and 1.8 10¹⁹ N-m (CMT determination)

# E. Study of the rupture using strong motion data

#### ♦ Data

13 SMA-1 strong motion instruments were triggered by the Uttarkashi earthquake in the distance range of 25 to 150 km. Among these 13 stations, 6 are close enough to have sufficient amplitude to be usable in the waveform inversion (Table 3). Each of those stations recorded three components of ground acceleration and the three components were used in this study. Because of the difficulty in modeling high frequencies the acceleration records have been band-passed filtered from 0.17 to 0.6 Hz. The filtering is done to remove frequencies outside the bandwidth for which Green's functions could be accurately evaluated. In this frequency band, the latest part of the signal is made up of low velocity surface waves that propagate in the shallow layers. These late arrivals do not give information on the rupture process and can not be modeled in a simple plane layered crustal model. The information on the rupture process is given by the first P and S

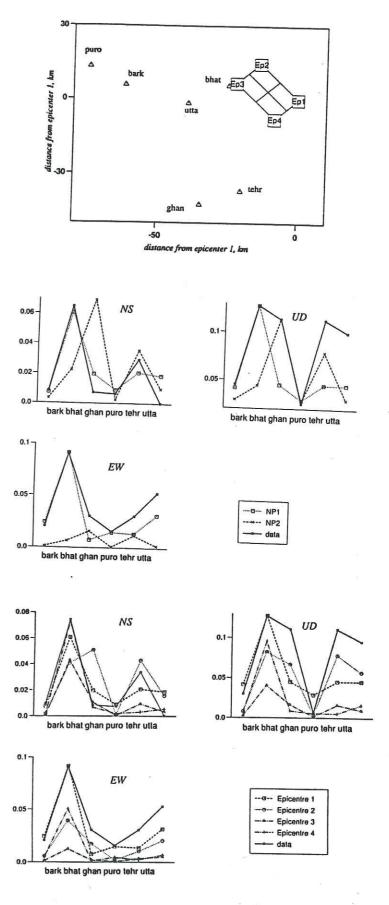


Figure 4. a) Location of the fault plane (NP1) and the hypocenters used in the forward modeling of the Uttarkashi earthquake.

- b) Comparison of the data maxima of the bandpassed accelerogramms (0.2-0.6 Hz) and the maxima of synthetics computed for the fault plane NP1 of the Harward CMT (Dziewonski, 1992) and different hypocenters.
- c) Comparison of the data maxima of the bandpassed accelerogramms (0.2-0.6 Hz) and the maxima of synthetics computed for the two fault planes NP1 and NP2 of the Harward CMT

waves that have energy in a large frequency band. Therefore, to eliminate the low frequency late arrivals, the traces filtered between 0.1 and 0.6 Hz have been multiplied by the normalized envelopes of the signal filtered between 0.1 and 5.0 Hz in order to select the body wave part of the signals.

#### Forward modeling of strong ground motion

The details of the method used have been presented by Cotton and Campillo (1994, 1995). Here we introduce this method briefly. In this procedure, we divide the fault plane into small rectangular regions of equal area called subfaults. Each subfault is allowed to slip once. With this parameterization, the ground motion (V) at a given station i and a given frequency  $\omega$  can be represented as a linear sum of n subfault contributions, each appropriately delayed in time to account for front propagation.

$$V_{i}(\omega) = \sum_{k=1}^{n} \operatorname{slip}(k) \exp^{-i\omega t(k)} U_{ki}(\omega) S_{k}(R_{(k)}, \omega)$$
 (1)

where  $U_{ki}$  represents the ground motion for a unit constant slip on the subfault k with a given source mechanism,  $\omega$  is the angular frequency. Slip(k), t(k) and S_k are the final slip, the rupture time and the source function of the k-th subfault depending on a single variable: the rise time R(k). This simple parameterization allows to limit the number of parameters of the model with respect to the technique proposed by Olson and Anderson (1988).

From a practical point of view, each subfault is represented by an array of point sources separated by a distance of less than one sixth of the shortest wavelength. The subfault contributions  $U_{ki}$  are obtained by summing the response of these point sources appropriately delayed in time to include the travel-time difference due to the propagation of the rupture front across each subfault. A local rupture velocity is therefore assumed for the integration on a subfault. The point-source Green functions are calculated for a layered velocity model using the discrete wavenumber integration method (Bouchon, 1981) associated with the reflection transmission matrix method (Kennett, 1983). In this study we used the regional seismic model (Table 4) given by Khattri (1992). We expect that site effects are of limited importance in the frequency band of our study. Therefore given a layered crustal model and assuming on each subfault a constant focal mechanism and a local rupture-velocity,  $V_i(\omega)$  can be considered as a function of slip(k), t(k) and R(k).

# Inverse modeling of strong ground motion

The parameter vector p and data vector d are related by the function model vector f as d=f(p). In the case of the Uttarkashi earthquake, the number of data (6 stations) is much lower than the number of stations used in previous inversions with the same technique, such as the inversion of the 1992, Landers earthquake (11 stations, Cotton and Campillo, 1995). For this reason, we have chosen to restrain the number of parameters describing the Uttarkashi earthquake.

Table 3. Stations used in the strong motion inversion

Stations	code	latitude	longitude
Barkot	bark	30°48' N	78°13' E
Bhatwari	bhat	30°48' N	78°35' E
Ghansiali	ghan	30°25' N	78°39' E
Purola	puro	30°52' N	78°05' E
Tehri	tehr	30°22' N	78°30'E
Uttarkashi	utta	30°44'N	78°27'E

Table 4. Reference regional velocity model

Thickness (km)	Vp (km/s)	Vs (km/s)	Density (g/cm ³ )	Qp	Os
0.4	3.5	2.00	1.80	50.0	25.0
1.0	5.0	2.86	2.40	80.0	50.0
15.0	5.2	2.97	2.60	4000.0	2000.0
30.0	6.0	3.43	2.90	4000.0	2000.0
∞	8.33	4.83	3.30	100.0	500.0

Table 5. Misfit and moment deduced from the rupture models computed with different rise times.

Rise time (sec)	Variance reduction	Moment (*10 ¹⁹ N-m)
0.5	0.43	1.02
1.0	0.42	1.55
1.5	0.35	2.0

Table 6. Variance reductions (a value of 1.0 means that the synthetic match perfectly the data)

	bark	bhat	ghan	puro	tehr	utta
N-S	0.04	0.90	-0.8	0.09	0.30	-0.30
E-W	0.31	0.78	0.21	0.20	0.12	0.49
U-D	0.43	0.67	070	-0.12	0.50	0.81

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In the following inversions, slip is constrained to be positive, and the rise-time is constant on all the subfaults. The parameters of the inversion are slip(k) and t(k) which appear in equation (1). Slip(k) is by definition positive, so taking a gaussian function to model the a priori information is not appropriate because a gaussian function gives a non-null probability to negative values. To avoid this problem, we choose to inverse the logarithm of this parameter, which defines a new (unbounded) parameter whose a priori density is gaussian and for which our standard inversion technique applies (Tarantola, 1987). The elements of d consists of complex 3-component spectra from all stations. The f function is a non-linear function of t(k) and t(k). If we assume an initial parameter vector t(k)0, we can get the iterative solution t(k)1 we have a priori density iteration and around t(k)2 are a get the iterative solution t(k)3. Using the observed data vector t(k)4 and an inversion algorithm based on the work by Tarantola and Valette (1982), t(k)5 given by:

$$p_{k+1} = p_k + b(A_k^t C_d^{-1} A_k + C_p^{-1})^{-1} \left( A_k^t C_d^{-1} (d_0 - f(p_k)) + C_p^{-1} (p_0 - p_k) \right)$$
 (2)

Here  $A_k$  is the Jacobian matrix of  $f(p_k)$  (the ij th element of  $A_k$  is  $\frac{\partial f_i(p_k)}{\partial (p_k)_j}$ ), b is a damping

constant between 0 and 1 used to prevent divergence, and  $C_p$  and  $C_d$  are the covariance matrices for p and d. Because a functional form of the slip is assumed, all derivatives are evaluated analytically.

The fit to the data at each iteration k is evaluated using the misfit function S (Tarantola and Valette, 1982):

$$S(p_k) = \frac{1}{2} \left[ (f(p_k) - d_0)^t C_d^{-1} (f(p_k) - d_0) + (p_k - p_0)^t C_p^{-1} (p_k - p_0) \right]$$
(3)

Following Cohee and Beroza (1994), we also evaluate the variance reduction between theoretical and observed seismograms defined as

$$\Delta \sigma^2 = 1 - \frac{(d_0 - f(p_{\infty}))^t C_d^{-1}(d_0 - f(p_{\infty}))}{d_0^t C_d^{-1} d_0}$$
(4)

### ♦ Application of forward modeling.

The modeling of ground acceleration in the 6 available stations was used to test the accuracy of two important pieces of information found with teleseismic data.

The teleseismic study of the Uttarkashi earthquake provides us two focal planes (NP1 and NP2). One of them is the plane containing the fault, so the first problem is to eliminate this ambiguity between those 2 nodal planes. Because of the effect of fault finiteness and directivity,

# Uttarkashi, Himalaya, India

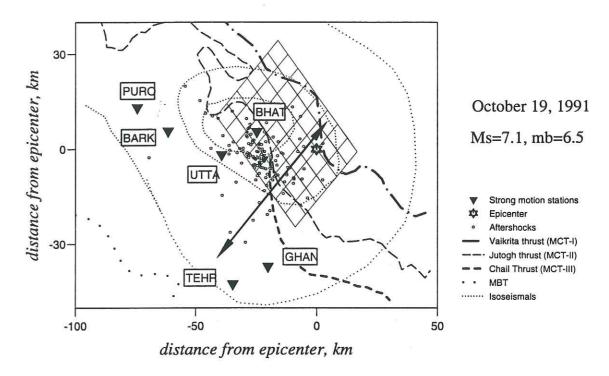


Figure 5. Location map showing the fault plane (divided in 48 subfaults) used in the strong motion inversion and the aftershocks of the Uttarkashi earthquake. The arrow shows the location of the cross section of Figure 10.

the peak amplitude of strong motions contains information that allows us to discriminate between those two planes. We have computed forward models with the two focal mechanisms NP1 (strike 317°, a dip 14°, slip 115°) and NP2 (strike 112°, dip 78°, slip 84°) found by the Centroid Moment Determination (CMT) of Harvard (Dziewonski et al., 1992). As already shown, this determination is more or less consistent with our long period study (Table 2). The long period study gives a rupture duration of 8 seconds. Since rupture velocities generally remain in the range of 0.8 to 0.95 of the shear wave speed in the source region, this source duration gives information about the source dimension, which is modeled as a simple planar fault 20 km long and 12 km wide (Figure 4a). The slip chosen is uniformly distributed on the fault with a constant value of 3.0 m (this value of slip gives a moment of 1.6*10¹⁹ N-m compatible with the moments determined from long period studies) and the hypocenter has the location given by the India Meteorological Department (IMD). In Figure 4b we compare the observed maxima with the results of the simulation for the two focal planes. The amplitude maxima of the synthetics obtained in this parameterization with NP2 show a clear disagreement with the data while the maxima found with NP1 are relatively close to the observations. In the following, NP1 is chosen to be the fault plane.

The epicentral determinations of the Uttarkashi earthquake, as reported by three different agencies are listed in Table 2. It is likely that the IMD solution which additionally used local short period and regional data is the closest of the actual hypocenter. In the near field, the effect of fault finiteness and directivity can be quite important and the study of relative peak amplitude of ground motion at different strong motion stations provides information about the direction of the rupture. Because of directivity, varying the hypocenter causes significant effects on ground acceleration in the near field domain so that strong motion data gives an indication of the direction of the rupture propagation. Figure 4a illustrates four different extreme models of rupture initiation at the far end of the fault. In these forward models, the source was modeled as a simple planar fault (20 km long and 12 km wide) consistent with (NP1). We chose a uniformly distributed slip on the fault with a constant value of 3.0 m. Figure 4c shows the maxima of the bandpassed (0.2-0.6 Hz) three components accelerograms for the different models compared to the maxima of the bandpassed data. Hypocenter 1 and a rupture propagation toward the west give the best agreement with the data. This hypocenter corresponds to the location of IMD and our modeling of the directivity effect confirms the accuracy of this determination.

In order to study the rupture process we need to know the absolute triggering time of each accelerogram. To solve this problem, the triggering time was computed by first matching by intercorrelation the first seconds of the observed waveforms with the first seconds of synthetic seismograms computed in a forward waveform model with a simple model (fault plane NP1, 20 km long and 12 km large, hypocenter IMD, uniform slip: 3m, uniform rise-time: 2s). This procedure gave us a first estimate of the absolute times.

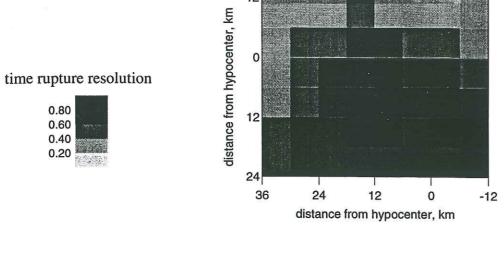
# Application of our inversion method to find the slip distribution.

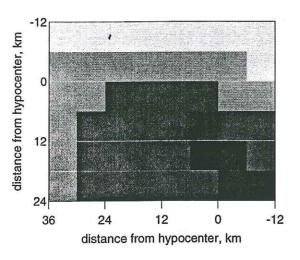
Our method operates through an iterative perturbation of the solution using locally determined gradients of the misfit surface. This method requires knowledge of an adequate starting solution. To chose this starting model we use some a priori information taken from seismotectonic studies and from our teleseismic inversion. The extreme limits of the rupture surface are estimated on the basis of the aftershocks and epicenter locations. According to this information we have chosen a fault plane of 48 km by 36 km divided into 48 subfaults (6 km long by 6 km wide). Rupture velocities values generally remain in the range of 0.8 to 0.95 of the shear wave speed in the source region and we take the initial rupture velocity (2.7 km/sec) to be 90% of the local shear wave speed in the source region. As already mentioned above, the source duration found by the teleseismic study is 8 seconds and it means that the fault surface defined by the 48 subfaults probably did not rupture entirely. In the starting model, a slip value of 0.2 m is prescribed on each subfault in the border of the fault. On the other subfaults, a value of 1.0 m is chosen. The seismic moment of this starting model is 1.2 1019 N-m. The orientation of the fault plane and the focal mechanism on each subfault is given by NP1 of the Harvard CMT. The localization of the hypocenter is the localization given by IMD. Figure 5 summarizes this configuration.

The observed and synthetic amplitude spectra at each station are equally normalized by the maximum observed amplitude spectra of the three components. This means that, for example, the weight given to each station in our inversion is independent of the distance to the station. Indeed, the directivity effect and decay with distance remain present in our proposed physical model.

We assume that off-diagonal elements of  $C_d$  and  $C_p$  are equal to 0. The diagonal elements of  $C_d$  and  $C_p$  are given by the data and parameter variances. As mentioned by Hartzell (1989) the estimation of a priori variances in strong-motion inversion is not easy. We assume a value of 25.0 for the a priori variances of the rise time, the slip and the time of rupture of each subfault and a value of 2.0 for the data variance. Like Fukuyama and Irikura (1986), Fukuyama and Mikumo (1993), and Cotton and Campillo (1995) the diagonal values of  $C_p$  and  $C_d$  were found after several inversion tests to find the best convergence (with a constant of damping b=0.1).

For each station, an inversion was first performed using only the data of this station. In this inversion the preliminary absolute times found by crosscorrelation with a forward modeling were used. This inversion procedure with only one station shows the time at which the rupture had to begin on the subfault containing the hypocenter in order to match the first arrivals at that station. This time of the beginning of the rupture gave us the correction to apply to the preliminary absolute time. The other information given by this type of inversion (values of slip and rupture time on the other subfaults) are not useful. Such timing of accelerograms gives the possibility to account for hypocenter errors and to accommodate slight departures of the chosen velocity model from the actual crustal structure.





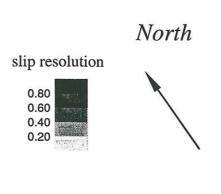


Figure 6. Resolution matrix diagonal value calculated for each inverted parameter. Since these values depend on the choice of the a priori variance of each parameter, these maps are used for relative comparisons of the resolution in different parts of the fault.

# Influence of the choice of the starting model on the final solution: resolution study.

The information taken from seismotectonic and teleseismic studies do not give any information on the possible values of the rise time. As noted by Hartzell and Mendoza (1991) or Cotton and Campillo (1994, 1995), dislocation rise time is one of the most difficult parameters to extract from the data. The consideration of this parameter has to take into account not only the fit to the data but also the value of the total moment. In order to specify the possible values of the rise time, we have performed a series of inversions with different values between 0.5 and 1.5 seconds. The different values lead to similar misfit but to significantly different seismic moment (Table 5). A simple explanation is the fact that the far-field seismic radiation is proportional to the derivative of the slip function. In spite of the complex configuration of the vicinity of a extended source, the dependence of the displacement on the slip velocity remains visible. Table 5 shows that a value of 1.0 second gives a good fit and an acceptable seismic moment. This value will be chosen in the inversion study described below.

Solutions of strong motions inversions can be divided into parts: a stable and an unstable part (Olson and Apsel, 1982). The unstable part of the solution consists of distributions of slip which have little or no effect on the data. In other words, they are not resolved by these data. Since the inversion is not purely linear, the final values of the parameters which are not resolved by the data are mostly dependent on the a priori values chosen in the starting models. The trace (sum of the 96 diagonal elements) of the resolution matrix (Tarantola and Valette, 1982) indicates the number of parameters (49.3) that were effectively resolved by the data. This resolution information is lower for the slip (35%) than the rupture time (53%). The resolution maps of Figure 6 give us an idea of the regions of the fault where our inversion is well constrained. These maps show us that the resolution is particularly bad in the northeast portion of the fault. This bad resolution in the North-east of the fault is explained by the absence of stations in this direction. Moreover, the stations are located south-west of the epicenter. Rupture on subfaults located northeast of the epicenter is then antidirective for the stations. Because of this directivity effect, even a large moment release in that region of the fault will not have a strong influence on the synthetic produced in the strong motions used in the inversion: the final slip distribution is then strongly dependent on the choice of the starting model in this region.

#### Results of the inversion

The synthetic waveform traces predicted by the slip distribution found by our inversion are compared with the observed seismic data in Figure 7. Quantitative fit to the data using a measure of the variance reduction between the theoretical and recorded accelerograms for each station is given in Table 7. The mean variance reduction is 0.40. The two closer stations

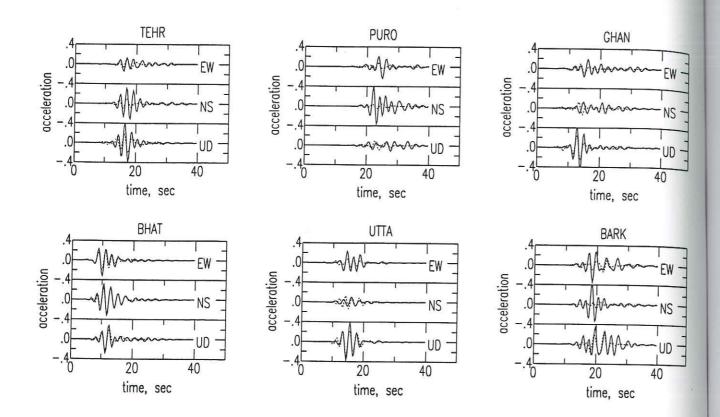


Figure 7. Strong-motion accelerogram observations (solid lines) and synthetics (dashed lines) for the rupture model of Figure 8. All have a common scale and have been band-passed in the same way (0.2-0.6 Hz).

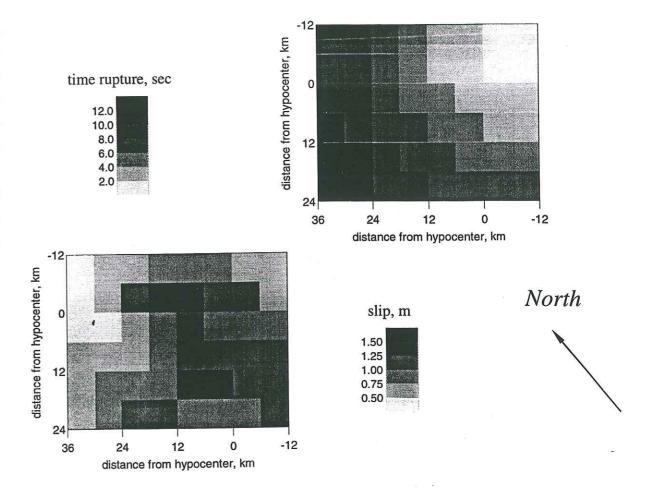


Figure 8. Slip and rupture time distributions on the fault found by our strong motion inversion. The seismic moment of this model is  $1.5 * 10^{19}$  N-m.

(Uttarkashi and Bhatwari) and the vertical components of all the stations show a good fit. On the other hand, horizontal components at the other stations are significantly lower than predicted. Yu et al. (1995) noticed that the velocity structure that worked for the two nearest stations and that we used in this inversion was not the best to predict the strong ground motion at more distant stations. Their trial-and-error modifications of the velocity model emphasize the potential role of unknown site and path effects, which could explain our poor fit of the horizontal components of distant stations.

The slip and rupture time distributions found by our inversion are shown in Figure 8. The earthquake ruptured mainly to the south-west of the hypocenter for a distance of about 25 km; its total rupture duration is about 10 s. The slip distribution presents two maxima where slip is near 1.5 m: one is located 10 km west of the epicenter and occured 4 seconds after the beginning of the rupture, the other one is located 15 km southwest of the epicenter and occured about 7 seconds after the beginning of the rupture. The total seismic moment of this rupture model is 1.5*10¹⁹N-m. This value is consistent with the values found from long period teleseismic studies (Table 2). The final low slip distribution found through the inversion in the North-east of the fault is strongly dependent on the choice of the slip value in the starting model in that region. In other words, whatever the value of the slip in this region is, the fit to the strong motion data will be the same. However, our model with no slip in that region gives a seismic moment which is in the range given by long period studies. If more slip is added in this region, the moment infererred from our inversion will no more be in agreement with the one obtained from teleseismic records and CMT.

### F. Discussion and conclusions

#### ♦ Source-time function

Because of the lack of absolute time, the imperfect azimuthal coverage and lack of knowledge of the structure, the set of strong motion data used in this study is not ideal, and the resolution of our inversion is not as good as that of inversions recently performed with Californian earthquakes like the Landers earthquake of 1992 (Cohee and Beroza, 1994; Wald and Heaton, 1994; Cotton and Campillo; 1995). For this reason, we have tested the coherency of our strong motion inversion with the work done with teleseismic data. Figure 9 shows the far-field source-time function obtained by integrating our inferred slip distribution across the fault and the source time function found with our teleseismic study. Both inversions show that the Uttarkashi earthquake began with a small subevent. Most of the moment was released between 2 and 8 s of the rupture with a maximum rate of moment release about 4 seconds after the beginning of the rupture

# ♦ Spatial correlation of the rupture zone with aftershocks and isoseismals

The isoseismal map and the damage distribution observed in the field after the earthquake

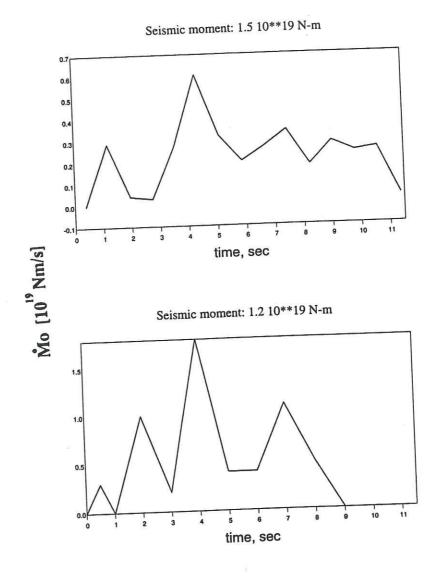


Figure 9. Comparison of source time functions obtained from our point source analysis of teleseismic data and our strong motion inversion

are described by Rastogi and Chadha (1995) and Kumar and Mahajan (1994). The region of most numerous aftershocks and the center of the damaged zone are located about 20 km west of the IMD hypocenter. This discrepancy between the localization of the hypocenter and the area of maximum damage was already noticed by Kayal et al. (1992) and Rastogi (1995). Our inversion study shows that the hypocenter area is not the zone of maximum slip. A zone of high moment release lies 15 km south-west of the hypocenter, which is closer to the center of the aftershocks and damage area. The observation of the aftershocks distribution (Figure 5) shows that several aftershocks are not located on the fault zone but a little more southward. It is possible that aftershock distribution does not define the area where slip occurred during the earthquake, but rather defines the zones where rupture stopped. This conclusion was already reached for Californian earthquakes (Mendoza and Hartzell, 1988; Hartzell and Iida, 1990). This effect is expected since it was shown that, in the case of a crack, stresses increase at the edges of the rupture area (Das and Scholz (1982)).

#### ♦ Tectonic interpretation

The Uttarkashi earthquake provides information on the tectonic processes currently affecting the Garhwal province. In this region, various sedimentary and metamorphic units are repeated several times as a result of imbricated thrusting along faults like the Jutogh and the Chail thrusts (Valdiya, 1980). This earthquake is not the result of slip along on one of these imbricate thrusts, but rather it ruptured a low angle thrust fault that extends south of the Vaikrita thrust. Microearthquake monitoring during 1979-80 and 1984-86 indicated a concentration of the seismicity north of Uttarkashi (Khattri et al., 1989). The majority of the 252 earthquake epicenters determined with this array lie close to or south of the surface projection of the MCT-II (Figure 1) with depth less than 10 km. Figure 10 presents a cross section of the Uttarkashi region showing the fault plane of the earthquake, the location of the aftershocks and the fault trace of the main thrusts near the surface.

We propose in Figure 11 an interpretative cross section of the Uttarkashi region: the region south of the Vaikrita thrust forms an accretion prism between the block of the High Himalaya and the Indian crust. At the base of this wedge, a detachment releases strain energy in moderate to large earthquakes such as the Uttarkashi earthquake. Between two major events on the detachment, the seismic activity is weak: aftershocks and microseismicity occur on the surounding faults like the Chail (MCTIII) of the Jutogh (MCTII) thrusts. Farther north the detachment fault probably steepens beneath the Higher Himalaya and become aseismic. In this region, the greatest seismic hazard is then from these large detachment earthquakes and one important point question is the location of the northern limits of the seismic part of the detachment. Our study of the Uttarkashi earthquake shows that this earthquake occurred north of the Jutogh (MCTII) and south of Vaikrita thrust (MCTI). The region on the detachment where big earthquakes can develop extends in the north at least to the Vaikrita thrust. This conclusion is important for correctly estimating the seismic hazard in the Uttarkashi region.

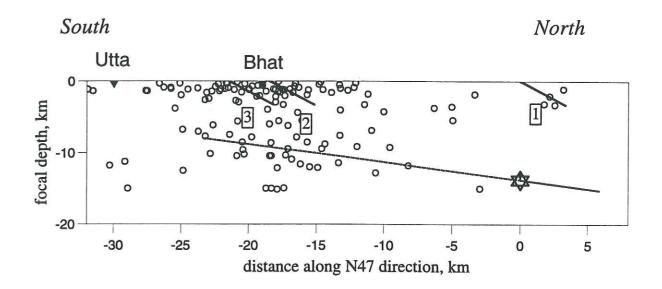


Figure 10. Cross section across the Uttarkashi region showing the fault plane (dashed line) and the epicenter (star) used in this study and the main thrusts mapped on the field (1. Vaikrita thrust, 2. Jutogh thrust, 3. Chail thrust). Aftershocks of the Uttarkashi earthquake are also plotted alongside the cross section (circles).

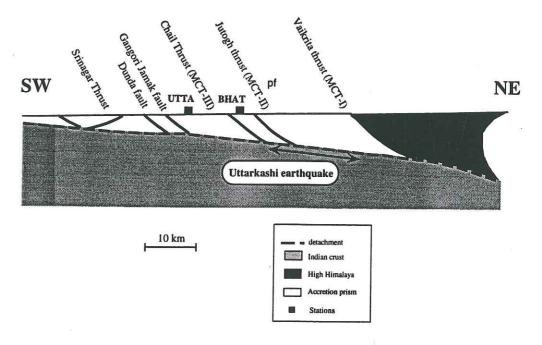


Figure 11. Interpretative cross section showing our tectonic interpretation of the Uttarkashi region and the localization of the earthquake in this tectonic scheme.

#### Acknowledgments

We thank R. Madariaga and two anonymous reviewer for their comments and suggestions that have inproved this version. We have also benefited from useful discussions with P. Lefort. This project has been supported by project Mecalith (INSU/CNRS). Strong motion data was readily provided by A.R. Chandrasekaran. The work was started during the visit of B.K. Rastogi to Grenoble in 1993 under European Council Department of Science and Technology Programme.

# Chapitre II-4

Histoire de la rupture du séisme de Northridge: résultats et limites d'une inversion de mouvement fort

Article en préparation

# 4. Histoire du glissement du séisme de Northridge: résultats et limites d'une inversion de mouvement fort

Slip history of the 1994 Northridge earthquake: results and limitations of strong motion inversion.

(preliminary version)

#### **Fabrice Cotton and Michel Campillo**

#### A. Abstract

Strong motion records of the 1994 Northridge earthquake are used to determine a model of the rupture history. Regarding the fault geometry, we consider two different models. One is a 40° dipping fault given by the global distribution of aftershocks. The second model has a larger dip (53°) that is suggested by the precise location of aftershocks 15 km West of the epicenter where the largest moment occured. This dip is also consistent with the Harvard Centroïd Moment Tensor determination of Dziewonski et al. (1994). We invert the bandpassed (0.1-1.0Hz) ground velocity to determine the distribution of slip, rupture time and rise-times on the fault plane. Two different inversion procedures are used. One is a linearized frequency domain inversion technique which operates through an iterative perturbation of a starting model using locally determined gradients of the misfit surface. The second one is a genetic inversion method which samples the model space more globally. We show that it is difficult to model the remote stations and the high frequency part of the data (more than 0.6 Hz). This study also shows that whatever the inversion method and the chosen fault plane slip, maxima are around 2 meters and preferably located west of the hypocenter. The inversions performed with the two different fault planes give almost the same fit to the data. Because of the non-uniqueness of the solution and the lack of resolution of strong motion inversion at depth, small wavelength features of the slip are not constrained by the data.

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#### B. Introduction

On 17 January 1994 at 12:30 (UT) a magnitude 6.7 (Mw) earthquake occurred in the north east region of the Los Angeles area. The Northridge earthquake resulted from thrust faulting on a south dipping ramp located beneath the San Fernando Valley.

Considerable previous works (that aid this study) has been done on the Northridge earthquake (Wald et al., 1995; Zhao and Kanamori, 1995; Dreger, 1994) but some questions remain unresolved.

One of the main question about this earthquake deals with the spatial distribution of coseismic slip on the fault plane. One interesting feature of previous studies is the variation of the degree of complexity in the models obtained with different type of data. Slip distributions obtained from geodetic inversions are smooth (Wald et al., 1995). The slip is confined to a single, central asperity. On the opposite, slip distributions obtained with teleseismic, regional or strong motions data are more complex and show several subevents (Dreger, 1994; Wald et al., 1995, Zeng et al., 1995). This variation in the degree of complexity of slip distributions found with different type of data has not been observed in studies of previous earthquake like the Loma Prieta or the Landers earthquake. For example, several sub-events were found in the case of the Landers earthquake, whatever the dataset used. In the case of the Northridge earthquake the ruptured area and the source duration are relatively small (teleseismic time function suggests a total rupture duration of 7 sec). Since we try to resolve shorter wavelength features compared to the Landers earthquake, the geodetic inversion has probably not the capacity (particularly at depth) to resolve these short wavelength variations of the slip distribution. On the opposite, is the resolution of strong motion inversions sufficient to resolve these short wavelength variations? The problem is that using standard inverse methods of seismic waveforms, it is not difficult to obtain heterogeneous solutions that fit the data acceptably well. This is illustrated by the important discrepancies between rupture models of the same earthquake published by different authors using same data sets. Past studies have concentrated only on deriving a model that fits the strong motion without emphasizing on assessing the solution uniqueness and the spatial resolution. To explain the differences between different solutions most of the authors suggest that differences in the data, various data weighting or fault parametrizations produce the discrepancies. Another explanation is that generally the solution is non unique and many solutions, perhaps associated with different local minima, will produce an acceptable fit to the data. Recently several authors in different context have used semiglobal inversion methods like simulated annealing or genetic inversion methods to test the non-uniqueness of the solution by finding large numbers of acceptable solutions (Sambridge and Drijkoningen, 1992; Jin and Madariaga, 1993; Lomax and Snieder, 1994...). To constrain the existence and the localization of the Northridge earthquakes subevents found through the analysis of strong motion data we use such a genetic inversion method and try to find a population of models which fit the data to see if all these "good" models have common characteristics. These genetic inversion results are also compared to a more classical local inversion method.

#### projection strike north 120 degrees

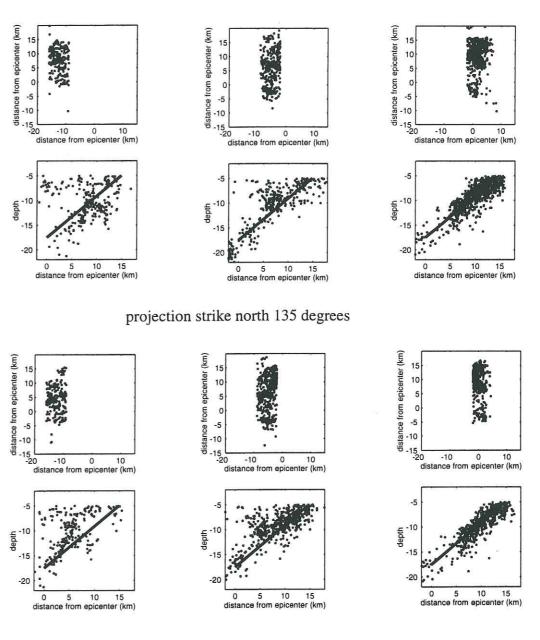


Figure 1. Map views (top) and cross-sections (bottom) of the aftershocks distribution for the Northridge earthquake. These aftershocks have been relocated in a 3D model by Zhao and Kanamori (1995)

Another important question about this earthquake is the actual geometry of the fault. No surface faulting was observed and this event is not obviously associated with any surficial geologic feature. The modeling of teleseismic surface-waves (Dziewonski et al., 1994). indicates a strike near 130° and a dip of 53°. First motion mechanisms (USGS, 1994) indicate a strike of 100° and a smaller dip of 40°. These observations show that there is an ambiguity in assuming the location and the geometry of the fault at depth. Do strong motion data constrain the fault plane to that geometry? To answer to this question, we model the strong motion data with different fault locations given by cross-sections of relocated aftershocks.

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## C. Geometry of the fault.

Using the aftershocks relocated by Zhao and Kanamori (1995) we have tried to give an interpretation of the fault surface at depth. In Figure 1 we show N120° and N153° strike-normal cross-sections. For each strike, 3 cross sections extend from the south-east (near the epicenter) to the northeast. The mainshock rupture as determined from previous waveform data (Dreger, 1994; Zeng et al., 1995, Wald et al., 1995) started at the southeastern corner of this zone and extended about 20 km to the northwest. All these waveform studies have used a fault plane dipping 40° on which most of the moment occurred between 10 and 15 km northwest of the epicenter. This fault plane is plotted as a thick black line on Figure 1. Along the east side, the aftershocks form a dense cluster and in this part of the fault the cross-section show that this dip of 40 degrees agrees with closely with the distribution of the aftershocks. However, the spatial distribution of aftershocks does not show the orientation of this 40° dipping plane in the west, where most of moment release was found by previous waveform inversions. In this part of the fault, the aftershocks distribution shows a more dipping plane close to the dip of 53° found by the Centroid Moment solution. To evaluate the possible variation of the fit between data and synthetic due to a change in the geometry of the mainshock fault, we consider two fault plane locations. The first fault plane agrees closely with the Centroïd Moment Tensor (Dziewonski et al. ,1994) solution and the western part of the aftershocks distribution and is a 53° degrees dipping and N135° striking fault plane. The second one agrees with the western part of the aftershocks distribution and the fault planes chosen in previous studies and is a 40° dipping and N120° striking fault plane. A more realistic model might incorporate variations in dip both in depth and along strike. But such a model will not serve the purpose of this study which is to evaluate the possible variation of the fit between data and synthetic due to a change in the position of the mainshock fault within the limits of knowledge of the actual geometry.

# D. Strong motion modeling

#### Data selection and processing

The strong motion data used in this study come from three main sources: The California Division of Mines and Geology network, the U. S. Geological Survey, and the University of

Table 1. Velocity structure model used to compute the strong-motion Green's functions. The "rock" model is taken from Kanamori and Hadley (1975).

#### Rock model.

Thickness (m)	Vp (m/s)	Vs (m/s)	Density (kg/m³)	Qp	Qs
2000.	4100.	2300.	2500.	300.	300.
2000.	5500.	3200.	2800.	500.	500.
22000.	6300.	3650.	2900.	500.	500.
6000.	6800.	3900.	3100.	500.	500.
0.	8200.	4700.	3200.	500.	500.

#### Soil model.

Thickness (m)	Vp (m/s)	Vs (m/s)	Density (kg/m³)	Qp	Qs
500.	3000.	1000.	2300.	300.	300.
1500	4100.	2300.	2500.	300.	300.
2000.	5500.	3200.	2800.	500.	500.
22000.	6300.	3650.	2900.	500.	500.
6000.	6800.	3900.	3100.	500.	500.
0.	8200.	4700.	3200.	500.	500.

Southern California. To minimize path effects, which could be difficult to distinguish from source effects, only stations within 30 km of the epicenter have been used. These stations are located in a region of extremely complex and varied geology. For example, the accelerograms recorded in the region South East of the epicenter contain large long period wavetrains which are not seen at the other stations. These arrivals probably correspond to the exicated surface waves in the basin (Olsen and Archuleta, 1995). Since we are not able to take into account such effects in our modeling, these types of records has been eliminated in our analysis. The strong motion stations finally selected and used in this study are listed in Table 1. Figure 2 shows that these strong motion data are well distributed around the source.

In practice an important element for the inversion is the duration of the record. Possible slip contribution from the entire width and length of the fault must be accounted for. However, the part of the signal we try to invert must be terminated before the arrival of significant reverberations or scattered energy resulting from a complicated structure which details are unknown. In other words, we do not want to map propagation path effects into features of the source. In practice, a time window of 15s was judged to be most suitable and applied from to the beginning of each signal.

Each of the stations we have selected recorded 3 components of ground acceleration. Because of the difficulty in modeling high frequencies, velocity records are used in the inversion. The accelerations records are then integrated to velocity and filtered from 0.1 to 1.0 Hz. This filtering is done to remove frequencies outside the bandwidth for which Green's function could be conveniently calculated.

#### ♦ Parametrization of strong motion modeling

The fault is parametrized as follows. A rectangular plane is assumed with fixed strike and dip. In the case of a dip of 53° and a strike of 135° the model fault plane extends from a depth of 4km to 20 km giving a fault width of 25 km. In the case of a dip of 40 degrees and a strike of 120 degrees the fault plane extend from 4 km to 24 km giving a fault with the same width of 25 km. The fault length of the fault is 20 km.

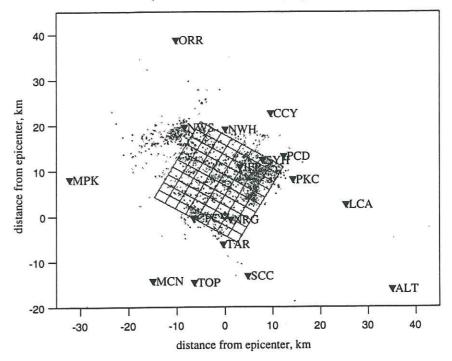
The fault is then subdivided into 80 subfaults of equal size (2.5 by 2.5 km). An appropriate number of 81 point sources is uniformly distributed over each subfault so that failure appears smooth for a rupture front crossing it in the frequency range of the inversion (0.1-1.0 Hz). The hypocenter is fixed at a depth of 18.6 km in both models and is indicated by a star in Figure 2.

With this parameterization, the ground motion V at a given station i and a given angular frequency  $\omega$  can be represented (Cotton and Campillo, 1995) as a linear sum of n subfault contributions, each appropriately delayed in time to account for front propagation:

$$V_{i}(\omega) = \sum_{k=1}^{n} \operatorname{slip}(k) \exp^{-i\omega t(k)} U_{ki}(\omega) S_{k}(R_{(k)}, \omega)$$
(1)

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## fault plane: strike N120°, dip 40°



## fault plane: strike N135, dip 53°

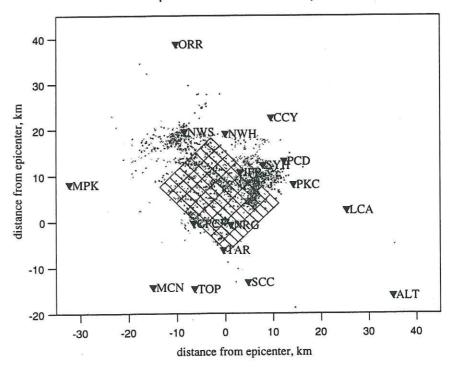


Figure 2. Selected strong motion stations (full triangles) and aftershocks (small open circles) The gray lines depicts the surface projection of the model fault plane used in the inversions. This fault plane is divided 80 subfaults (top, strike N120° and dip 40°, bottom: strike N135° and dip 53°).

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 $Slip_k$ ,  $t_k$ , and  $S_k$  are the final slip, the rupture time, and the slip time function of the k th subfault. The slip time function depend on a single variable: the rise time  $R_k$ ,  $u_{ki}$  represents the ground motion for a unit constant slip on the subfault k with a given source mechanism.

#### ♦ Green function computation

Given the above geometry for the fault, an appropriate velocity structure is required for the computation of the crustal response. The point source Green functions are calculated for a layered velocity model using the discrete wavenumber integration method (Bouchon, 1981) associated with the reflection transmission matrix method (Kennett, 1983). After several tests, we decided to use different crustal models for the computation of Green's function at rock and soil sites. The velocity model used to compute strong-motion Green's function at rock site is taken from Kanamori and Hadley (1975). To better evaluate elastic properties beneath soil stations a slower layer of 500m was added to this model and used to compute Green's function at soil sites. The width of this layer and the elastic properties of this layer were evaluated with the 3-D geological model of the Los Angeles area developed by Magistrale et al. (1994). These models are described Table 2.

#### ♦ Data weighting

Preliminary test of the inversion have suggested that the modeling of vertical components is difficult. This suggests more contamination from path and site effects in the vertical data. For this reason we have chosen to give a higher weight (by a factor of 2) to the horizontal components. Each component is weighed proportionally to the maximum of the energy of the spectra, so the small amplitude components have a small weight in the inversion. In previous inversions with that method (Cotton and Campillo, 1995a) we had chosen to equally normalize the observed and synthetic spectra at each station. Preliminary tests of the inversion of Northridge data have shown that with this type of normalization the weight given to remote stations in the back azimuth like ALT was too high and the fit to the other stations near the epicenter was unsatisfactory. Finally each station was normalized and then weighted by log(E)+6.0 where E is the total energy of the signal spectra at each station.

## E. Gradient method inversion

#### ♦ Procedure

The inversion for fault slip, rupture time and rise time use the method described in details in Cotton and Campillo (1995a) and already used in Cotton and Campillo (1995b) and Cotton et al. (1995c). This method is a linearized frequency domain inversion technique. We introduce the local slip rupture time, and rise time on each subfault as parameters to be directly and simultaneously evaluated without constraints through the inversion of the records. Assuming an initial starting model, we get the iterative solution using an inversion algorithm based on the work

Table 2 strong motion stations

Code	A 1				
Code	Adress	Lat	Lon	distance from	Codes
				epicenter	
ALT	Alhambra Freemont School	34.070	118.150	39	BRT
ORR	Castaic. Old Ridge Route	34.564	118.642	40	CRT
MCN	Monte Nido Fire Station	34.087	118.693	21	GN
JFP	Granada Hills Jenson Filtr. Plant	34.312	118.496	12	G S
MPK	Moorpark	34.288	118.881	33	CST
NWH	Newhall LA County Fire Dept	34.387	118.530	20	C S
PCD	Pacoima Dam Downstream	34.334	118.396	19	CN
PKC	Pacoima Kagel Canyon Fire	34.288	118.375	17	CRT
SCC	Sepulvada Canyon Contr Facility	34.097	118.478	15	G R
SYH	Sylmar 6-story County	34.326	118.444	15	CST
TOP	Topanfa. Fire Station	34.084	118.599	16	GR
NRG	Canoga Park 17645 Saticoy	34.209	118.517	1	US
LCA	La Crescenta	34.238	118.254	26	US
NWS	Newhall 26385	34.391	118.622	21	UR
TAR	Tarzana	34.160	118.534	6.0	CST

Codes: C= C.D.M.G., G=U.S.G.S., U=USC, S and R indicate use of soil or rock velocity structure for synthetic ground motions. T indicate the stations with absolute time.

by Tarantola and Valette (1982). This method operates through an iterative perturbation of the solution using locally determined gradients of the misfit surface.

# ♦ Configuration of the inversion

As a starting model we impose a value of 1m on a part of the fault as illustrated in Figure 3. The subfaults located at the edge of the fault have no slip. The seismic moment of this starting model is 0.13 10²⁷ dyn/cm. This moment is equal to the moment chosen by the Centroid Moment Tensor solution of Dziewonski et al. (1994) .The initial rise time for each subfault is 1 sec and is the same on all the fault plane. The a priori variance for the slip, the rupture time and the rise-time were chosen to be respectively 160, 160 and 80. These values give the possibility to find heterogeneous slip distribution but also variations in the rupture front velocity and rise time values. The quality of the solution is evaluated through an error analysis and a resolution study. First, the fit to the data is evaluated using the variance reduction between theoretical and observed seismograms (a perfect fit between data and synthetic gives a variance reduction of 100%). This variance reduction was proposed by Cohee and Beroza (1994) and already used in Cotton and Campillo (1995a). The variance reduction is used for all the data but also for each frequency and station component. In a second step we calculate the resolution matrix of the inversion which illustrates how well the inverse problem can be solved with our data. This resolution analysis is described in Cotton and Campillo (1995a) and allows to compare the relative resolution of different parts of the fault.

# Determination of absolute times

For strong motions inversion the time of the observed and synthetic records has to be related to the same origin time. This is naturally achived when the trigger time of the instrument is known. On the other hand Ellsworth and Beroza (1994) noted that the rupture of the Northridge earthquake began with a small foreshock followed by the main rupture episode. Based on these observations, we chose not to take into account the foreshock and to begin the modeling of the rupture 0.5 sec after the origin time. The origin time of the data and synthetic presented in this study is then 12:30:55.7 GMT. Of the 17 stations which have been selected 5 have readable trigger times. A correct estimation of the trigger time of the other station is important because it is clear that timing differences can shift the location of the patches of rupture. The trigger time of the 12 remaining stations are estimated in the following manner. A preliminary inversion of the data is done with only the stations with absolute time (stations ORR, MPK, SYH, TAR and ALT). The observed and synthetic records predicted by this inversion results are then compared to find the trigger time of the stations without absolute time. Figure 4a and Figure 4b show the results of this inversion performed with the fault plane dipping 53° and the fault plane which is dipping 40°. The trigger time found by this study and used in the following part of this work are listed in Table 1. These first results show that the 12 stations which are not used in the inversion are for most of them not too badly predicted by the model found with only 5 stations and this conclusion is important to correctly determinate the trigger times of those stations. Clearly this rupture model

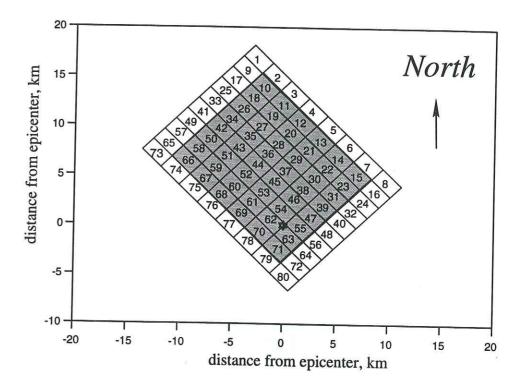


Figure 3. Subfaults discretization of the fault plane. The starting model of the gradient inversion has a slip of 1 m in the gray area and a slip of 0 m in the white one. In the genetic inversion only the subfaults in the gray area were used. This choice was done to limit the number of parameters.

obtained with 5 stations with absolute time and a good azimuthal distribution is sufficient to produce reasonable estimations of the phase and amplitude of the other stations.

#### ♦ Results

The inversion is then performed with all the 17 stations. The same weight is given to all these stations and the results are shown Figure 5a for the fault plane dipping 53° and Figure 5b for the fault plane dipping 40°. When the station coverage is increased from 5 to 17 stations we see that the same general characteristics of the slip distribution remains. This suggests that the solution is relatively stable with regard to this set of stations: different set of data give the same inversion results. The second point is that the models found with the two differents fault planes have some similarities: the slip maximum is near 2 meters for both models and located 10 kilometers west of the hypocenter, the hypocenter is a region of significant slip (around 1 meter) and low slip values (around 0.5 meters) are found at the top of the fault. The rise time values are between 0.7 and 1.2 seconds in both models. Rupture velocities varies between 1.0 km/s to 4km/sec. In region of high moment release the rupture velocity is close to 3km/sec. The variance reduction and the seismic moment of the model found with the fault plane dipping  $53^{\circ}$  are 37% and 0.17.  $10^{27}$ dyne/cm. These values are equal to 36% and 0.16. 10²⁷ dyne/cm for the fault plane dipping 40°. The difference in the waveform fit between these two models are not significant. The seismic moment found by the Centroïd Moment Tensor analysis of Dziewonski et al. (1994) is 0.16. 10²¹ dyn/cm and the seismic moment evaluated through geodetic analysis is 1.4+/- 0.2 10²⁷ dyn/cm (Wald et al, 1995). The seismic moments of both models are coherent with the moment found in these studies. To conclude, it is not possible to say which of these models is closer to the actual fault plane. In other words, when the geometry is unknown heterogeneous slip distributions located on different fault planes can equally fit the data, even with a good station coverage.

# ♦ Influence of the choice of the velocity model.

To test the influence of the choice of a different velocity model for the soil and the rock stations, we have performed an inversion with the same velocity model for all the stations. This velocity model is the velocity model of Hadley and Kanamori (1975). The fault and inversion parameterization are the same than those used to produce the model of Figure 5b. The results are shown Figure 6. The slip distribution pattern of Figure 5b and Figure 6 are the same but the slip found with this new inversion are higher (the seismic moment is equal to 0.17.  $10^{27}$  dyn/cm). The fit of this model is lower (31%) and some stations located on soil are underpredicted by synthetics (for example, station Sylmar). The presence of soft materials that amplify the ground motion cannot be neglected.

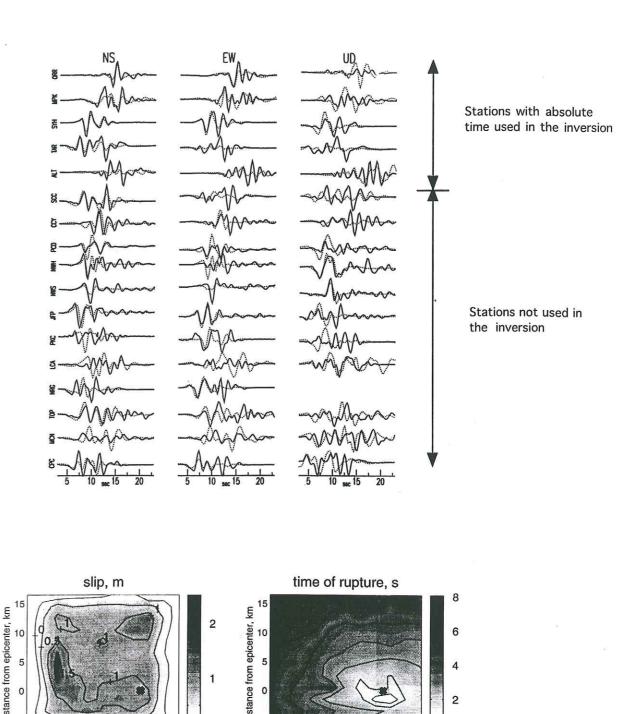


Figure 4a. Results of gradient inversion performed only with the station with absolute time (ORR, MPK, SYH, TAR and ALT) and the fault plane dipping 53°. The top of the figure shows a comparison between observed (solid) and synthetic (dashed) ground velocities. The bottom of the figure shows the slip distribution obtained in the inversion. The contour interval of the slip distribution (left of the figure) is 0.5 meter. The scale bar shows the slip shading in meters. Concentric lines show the advance of the rupture front at 1 sec intervals (right of the figure).

-15

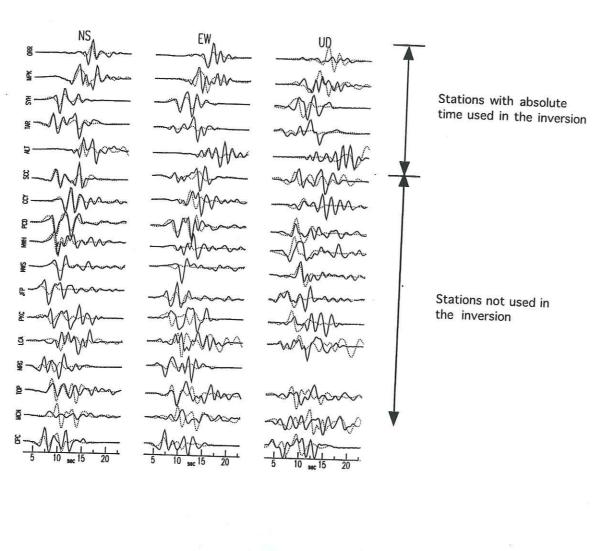
-10

distance from epicenter, km

-10

-5

distance from epicenter, km



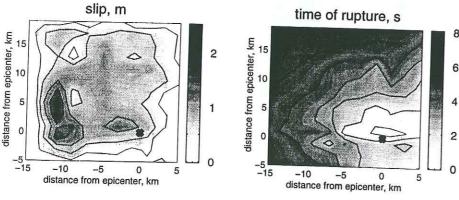
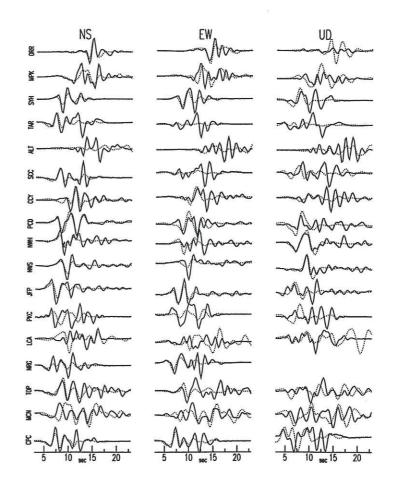


Figure 4b. Results of the gradient inversion performed only with the stations with absolute time (ORR, MPK, SYH, TAR and ALT) and the fault plane dipping 40°. Otherwise same as Figure 4a.



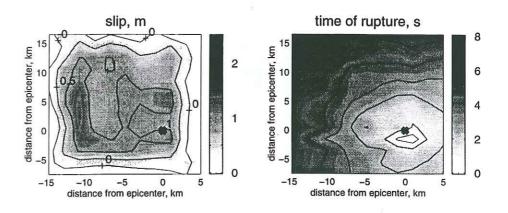
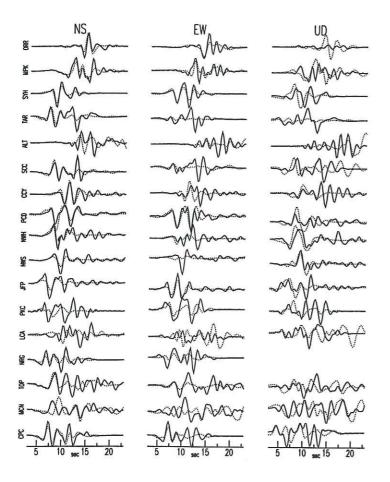


Figure 5a. Results of the gradient inversion performed with all the stations and the fault plane dipping 53°. Otherwise same as Figure 4a.



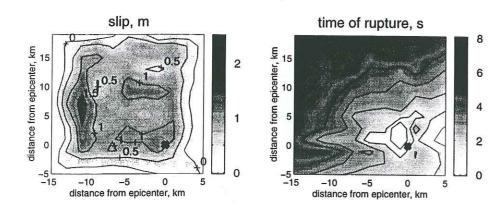
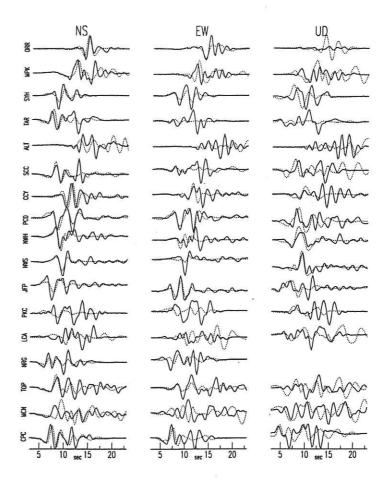


Figure 5b. Results of the gradient inversion performed only with all the stations with absolute time and the fault plane dipping 40°. Otherwise same as Figure 4a.



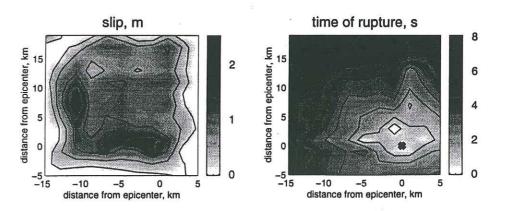
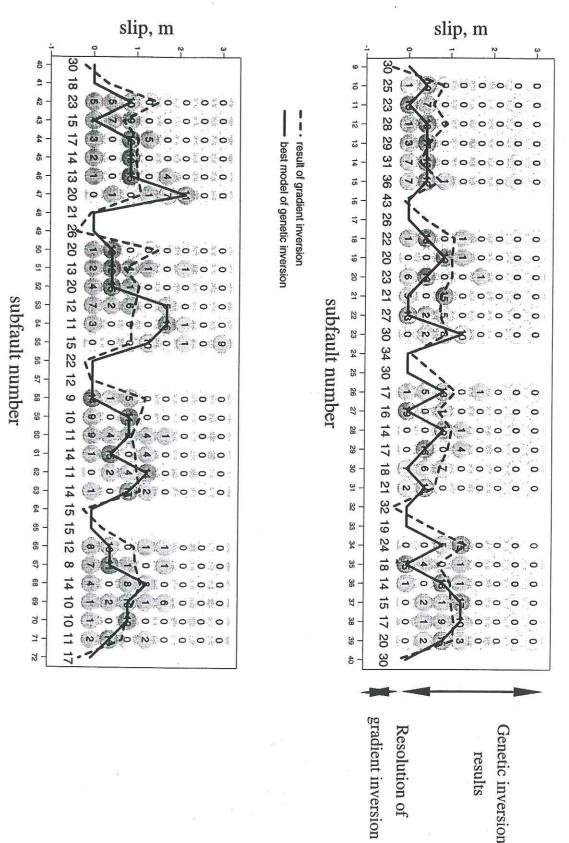


Figure 6. Results of the gradient inversion performed with all the stations and the fault plane dipping 53°. In that case, the velocity model is the same for all the stations (velocity model of Hadley and Kanamori, 1975). Otherwise same as Figure 4a.

## F. Hybrid Inversion

#### Procedure

Traditional approaches where the problem is linearize as in the previous inversion procedure are dependent on the starting model and the inversion can converge to a local minimum. On the other hand, grid searches are of practical interest only for simple problems involving a few model parameters. Semiglobal inversion methods like simulated annealing or genetic inversion methods provides an attractive alternative. Simulated annealing uses an analogy with physical annealing in thermodynamics systems. This method has its origins in the work of Metropolis et al. (1953) and was developed by Kirkpatrick et al. (1983). Recently this method has been applied in the field of geophysics by Courboulex et al. (1995) for the deconvolution of earthquake records by empirical Green's function and Hartzell and Liu (1995) for the inversion of teleseismic body waves. Genetic algorithms (GA) operate an analogy between biological system and function optimization. Genetic algorithms were first developed by Holland (1975) and were investigated recently by several authors (Stoffa and Sen, 1991; Sambridge and Drijkoningen, 1992; Zeng et al., 1995 and Lomax and Snieder, 1995). Beginning with a random populations of solutions, succeeding populations are created by 1) selection (saving solutions with smaller misfit, 2) crossover- (combining part of two solutions to form new trial solutions), and 3) mutation (changing the values of some of the parameter of a solution). The GA produces a large set of solutions which sample the solution-space globally but which can rapidly converge to a minimum. In our case the number of parameters (240) is much higher than the number which has been solved in previous genetic inversions studies and to solve this problem we have to limit the number of models in the model space. For this reason, the number of subfaults used in the inversion is limited and only the subfaults in the gray area of Figure 3 are used in the inversion. A slip of 0 meter is given to the other subfaults. Only the slip and the rupture time are inverted and a value of 1s is given to the rise-time of all the subfaults. In genetic inversions, the continuous range of each parameter is replaced by a set of discrete values of parameters. We first define the search width of each of the model parameters. For the slip the bounds chosen were 0 and 3m with a discretisation interval of 0.43 m. For the rupture time, the bounds were between tr+1.0 sec and tr-1.0 sec where tr is the rupture time given by a rupture velocity of 3km/sec. For the rupture time the discretisation interval is 0.66 sec. With this rough discretisation of the parameters, it is clear than the models found by the genetic inversion are not the optimum ones. The goal of the genetic inversion is not to find directly the global minimum but to explore globally the model space to find an appropriate model that is not far from this global minimum. The best model found by the genetic inversion is then chosen as starting model in the gradient inversion method to move from this model to the global minimum. This method is called hybrid inversion method in the following and is applied to the inversion with the fault plane dipping 53°.



selected this value. The figure shows the slip distribution of the results of the gradient method (same as Figure 4a) and the results of the hybrid method (same as Figure 7). The resolution of

take 8 possible values between 0 and 3 meters.

7. This figure

shows the result of the genetic inversion. For each subfault, the slip can

A numbered circle indicates for each slip value,

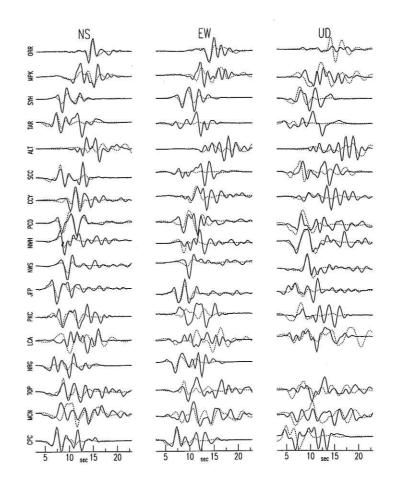
#### ♦ Genetic inversion configuration

Here we configure the GA to find sets of acceptable solutions representing all regions of the solution space that give a misfit below some acceptable level. The GA we use has been developed by Lomax and Snieder (1994). This algorithm is similar to that described by Sambridge and Drijkoningen (1992) but is configured to find many acceptable solutions and to avoid rapid convergence in secondary minima. We set the rates of cross-over and mutation to 0.7 and 0.02. In addition, the best solution of each generation is not explicitly saved. As already mentioned by Lomax and Snieder (1994), these adjustments tend to produce a smaller but more stable and diverse set of acceptable models relative to a GA configured for rapid convergence. The number of possible models is 10⁹¹. At each generation 60 models are tested and 150 generations are computed. These values were chosen after several trials. To find the starting model with the genetic inversion method, we prefer smooth model over oscillatory models and for this reason we have used a roughness penalty function (Nolte and Frazer, 1994).

#### ♦ Results

We show in Figure 7 the distribution of slip values corresponding to the best 20 models found by the genetic inversion for each subfault. The variance reduction achieved by the genetic algorithms for these 20 models are between 25% and 22%. It is interesting to note that even though each parameter is discritized in only a few values (8 for the slip and 4 for the rupture time), the genetic algorithm has allow to find a large set of diverse models with a significant fit value. On the other hand, it is constructive to look for similarities between all the « good » models found by the genetic inversion. These similarities represent the more robust and better resolved parts of the solution. One interesting feature is that the slip values found in the different « good » models are distributed on a set of 2 or 3 adjacent values on most of the subfaults. For example, the slip on the subfault number 42 is equal to 0 m in 5 models, 0.41m in 5 models and 0.82 m in 10 models. For this subfault and at that level of fit between data and synthetic, models with slip values egal to 0.42m +/- 0.42 m are selected by the inversion procedure. For some subfaults, the results show a different feature: for the subfault number 55 (near the hypocenter), 5 models have a slipvalue of 3 m and 10 a value of 1.2 meter: in this case, the scatter of the solution is high and the inversion is finding different classes of models with very different slip patterns. With the data we use, it is not possible to chose between these different classes of models: the scatter of the results described by these acceptable solutions indicates regions of model space that are not well constrained by this inversion. According to these results, 3 features found by the genetic inversion are however robust: (1) the inversion requires low slip at shallow depths to fit the synthetics (2) most of the subfault have a maxima lower than 2 meters, (3) the scatter of the solution is higher at depth.

To performe the hybrid inversion, the best model of the genetic inversion is then used as a starting model in the linearized inversion. The results of this inversion is shown Figure 8. The fit



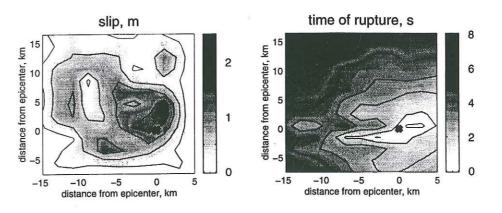


Figure 8. Results of the hybrid inversion performed with all the stations and the fault plane dipping 53°. Otherwise same as Figure 4a.

of this model is equal to 38% so almost the same value than the result found by the gradient method alone. The moment of this model is equal to  $0.14 \cdot 10^{27} \text{dyne/cm}$ .

Finally, we also compare in Figure 7 all the results found by the different types of inversion (gradient alone and hybrid method). For each subfault the solution found by the linearized gradient method (model described in Figure 4a) and the result of the hybrid inversion (model described in Figure 8) are shown. The important point is that the model found with the gradient method alone has a fit which is almost the same than the fit obtained with the hybrid inversion and the main features of the two models are the same. It means that the solution found with the gradient method alone is not far from the global minimum.

We also show on the figure the resolution of each slip value found by the gradient inversion study alone. The striking feature of the resolution studies of the linearized inversion is that the resolution at depth is less than the resolution at the top of the fault. The genetic inversion shows the same feature: near the surface all the models who show a good fit have slip which are in a narrow range between 0 and 0.84 m. At depth, the values of the slip are distributed in a larger range between 0.0 and 1.6 meters.

### G. Discussion and conclusion

In this study, we show that when the geometry is unknown heterogeneous slip distributions located on different fault planes can equally fit the data. Finding details of the spatial distribution of coseismic slip on the fault plane with strong motion studies, can not be performed if the geometry of the fault is not well-known. For this reason, relocation of aftershocks and precise studies of the geometry of the fault (Zhao and Kanamori, 1995; Hauksson, 1995) must be performed before the source history study.

The variance reduction gives us the possibility to compare the quality of the inversion in the present case to the inversion performed in the study of other earthquakes like the Landers earthquake. In the case of the Northridge earthquake the variance reduction is in the best case around 40%. In the case of the Landers earthquake the variance reduction reached 70% (Cotton and Campillo, 1995b). One possible explanation is the frequency range used in these two inversions. In the case of the Landers earthquake, displacement in the frequency band between 0.05 and 0.5 Hz were inverted. In the case of the Northridge earthquake the source duration is relatively short and we try to resolve shorter wavelength features compared to the Landers earthquake. For this reason we use a larger frequency range between 0.2 and 1.0 Hz. Figure 9 shows that at frequencies lower than 0.5 Hz the inversion of Northridge data is also able to find good variance reductions higher than 50%. On the other hand, at frequencies more than 0.5 Hz the results are worse because the crustal model we use is too simple to account for actual propagation.

The genetic algorithm is an interesting method of data fitting without the need for the derivative information or a good starting model close to the global minimum. The main interest of semi-global inversions method is that they give information about the uniqueness of the solution. In the case of the Northridge inversion, we have seen that details of the slip distribution solution obtained by these inversions should not be considered as robust specially at depth where the

resolution is low and several models with different slip distribution can equally fit the data. However, the hybrid method does not lead to results which are much better than solutions found with the gradient method alone. It means that, in this case, the solution found with the gradient method alone is not far from the global minimum. The problem of uniqueness of the solution must be studied for each earthquake and satisfactory approach involves a careful balance between random global search and retention of multiple acceptable solutions. From our experience of genetic inversion, three practical difficulties must however be highlighted. The first problem is when a model is generated which is not particularly close to the global solution but relatively good compared to the rest of the population. It then makes multiple copies of itself and begins to dominate the population early in the life of the algorithm. The result of this is a lack of diversity in the final population of good model. An other aspect of the genetic algorithm which is somewhat unsatisfactory is that it is possible to improve the efficiency of the algorithm by some exploration of the parameters controlling the algorithm (size of the population, cross over and mutation probabilities...). A number of preliminary trials runs are then necessary in order to achieve an efficient implementation of the algorithm. The third difficulty is that strong motion inversion studies use a large numbers of parameters to image the rupture history on the fault plane. For this reason to keep reasonable time calculations, only rough discretisation of the parameters can be used with this method.

Finally let's summarize the robust features of the Northridge rupture history found by this analysis. Whatever the geometry considered in the inversion, 2 maxima are found in all the models. One is located 10 km west of the epicenter and his location is almost the same in all the models. The other one is located up-dip North west of the epicenter and this localization vary from one model to another. All the models show significant slip (1m) near the hypocenter. Another robust feature is that the inversion requires low slip at shallow depths to fit the synthetics. This corresponds well with the idea that the rupture does not break the free surface and probably did not propagate to depth shallower than 5 km. The models slip maxima are less than 2.0 meters and the seismic moment of the models are between 1.4 and 1.7  $10^{27}$  dyn/cm.

#### Akckowledgments

This work was initiated during a visit of the two authors to the Institute for Crustal Studies, University of California, Santa Barbara. We thank Ralph Archuleta for helpful suggestions. Discussions with J. Steidl, A. Tumarkin, C. Nicholson, K. Olsen and F. Bonilla are greatly aknowledged. A. Lomax provided us with the genetic inversion code used in this study.

Chapitre II-5

Conclusion méthodologique

## 5. Conclusion méthodologique

Les chapitres précédents ont illustré la mise au point et l'utilisation d'une méthode d'imagerie de la rupture sismique à l'aide de données de champ proche. Ce travail a comporté 3 aspects: le traitement des données accélérométriques, le développement de la méthode d'imagerie et enfin l'interprétation des résultats en terme de physique de la rupture. Nous essayons ici de faire le point sur les méthodes d'analyse que nous avons mis au point et nous discutons ensuite de la façon de les améliorer.

## A. Caractéristiques et originalité de la méthode

#### Utilisation des mouvements forts

Pour imager la rupture sismique et déterminer les variations spatio-temporelles du glissement sur le plan de faille 3 types de données (données géodésiques, données télésismiques et données de mouvement fort) sont classiquement utilisées. On peut penser que les données de mouvements forts sont intrinsèquement mieux adaptées à l'imagerie de la rupture. En effet, pour les données télésismiques, la variation des fonctions de Green provenant de différentes parties de la faille est simplement une différence de phase. Les données géodésiques ne permettent quant à elles qu'une image de la répartition spatiale du glissement et ne donnent aucune information sur la cinématique de la rupture.

#### ♦ Paramétrisation du problème

La paramétrisation que nous avons utilisé permet de décrire la rupture avec un nombre faible de paramètres par sous-faille, ceci tout en ne limitant pas *a priori* les modèles possibles. L'inversion s'effectue dans le domaine des fréquences ce qui stabilise l'inversion: les longues périodes du signal sont résolues indépendamment des courtes périodes.

#### ♦ Contrôle de la qualité de l'inversion

Nous avons essayé au cours de cette étude de donner au lecteur la possibilité de contrôler la qualité de l'image de l'inversion obtenue et notamment de répondre aux questions suivantes: Quelle est la partie des données que le modèle explique vraiment ? Quelle méthode d'inversion donne les meilleurs résultats ? Plusieurs classes de modèles permettent-elles d'expliquer les données ? Quelles sont les paramètres vraiment résolus par l'inversion ? Pour répondre à ces questions plusieurs outils ont été développés.

Constatant que la mesure visuelle de l'accord entre les données et synthétiques ne permet pas de comparer diverses inversions entre elles, nous avons utilisé pour chaque station un indicateur quantitatif de la qualité de l'accord (réduction de variance définie par Cohee et Beroza (1994)).

Une mesure de l'accord entre données et synthétiques a aussi été effectuée fréquence par fréquence pour montrer la bande spectrale où nos résultats sont valides.

Dans les méthodes par multifenêtrage où le nombre de paramètres est important une contrainte de positivité est introduite pour stabiliser l'inversion. La façon dont cette contrainte de positivité est introduite ne permet pas d'analyser la résolution de l'inversion (Hartzell, 1989). Au chapitre II-1, nous avons vu qu'avec la paramétrisation utilisée dans notre méthode on peut trouver sans contrainte de positivité un modèle réaliste avec peu de vitesses de glissement négatives. Le calcul de la matrice de résolution est alors possible et ce résultat nous a par exemple permis de montrer que la résolution est plus faible en profondeur qu'en surface. Au chapitre II-3, dans l'étude du séisme d'Uttarkashi, nous montrons qu'analyse de résolution et contrainte de positivité ne sont pas contradictoires. Il suffit d'effectuer un changement de variable et inverser non pas le glissement mais le logarithme du glissement (Tarantola, 1987).

L'inversion du glissement et de la vitesse de glissement est un problème non linéaire. La méthode d'inversion par gradient que nous avons décrite au chapitre II-1 est une méthode locale. Les paramètres non contraints par les données dépendent entièrement de leur valeur dans le modèle de départ. Plusieurs fois au cours des chapitres précédents nous avons exploité cette propriété et effectué l'inversion avec plusieurs modèles de départ. Ceci nous a ainsi permis de voir les zones de notre modèle de Landers où la vitesse de rupture est contrainte par les données.

# ♦ Utilisation de méthodes d'inversion semi-globale

Un problème important des méthodes locales d'inversion non linéaires consiste à choisir *a priori* un "bon" modèle de départ. De plus, lors de ce type d'étude comment être sûr que le minimum de la fonction coût n'est pas un minimum secondaire? Nous avons vu au chapitre II-4 que l'utilisation de méthodes d'inversion semi-globales telles que l'inversion génétique peuvent aider à résoudre ce type de problème et permettent d'explorer plus complètement l'espace des paramètres. Ces méthodes permettent de trouver une population de "bons" modèles. Les paramètres communs à tous ces modèles sont bien contraints par les données. Il faudrait maintenant tester d'autres types de méthodes semi-globales (recuit-simulé, métropolis) afin de bien analyser les avantages et les défauts de chacune de ces méthodes.

# B. Conditions nécessaires à son application

# Avoir de bonnes données

Notre méthode nécessite des données de champ proche n'ayant pas saturées lors du choc principal. Il est important de connaître ou de retrouver les temps absolus de ces enregistrements. Les temps absolus des accélérogrammes sont malheureusement bien souvent absents car la plupart des utilisations classiques des données de mouvements forts en risque sismique ne nécessitent pas la connaissance précise du temps. Par exemple, l'étude du séisme de Northridge a montré que dans une région bien instrumentée (les environs de Los Angeles) seul un tiers des stations avaient un temps absolu. Retrouver ces temps absolus avec une bonne précision n'est par ailleurs pas toujours

facile car le déclenchement peut se faire sur la première P mais aussi sur des phases S plus énergétiques.

Nous avons vu qu'une bonne répartition azimutale est importante et ceci est un problème certain pour l'analyse de la source des séismes en subduction ou des séismes Himalayens (peu de stations à l'intérieur de la chaîne). L'analyse du séisme de Northridge a montré qu'avec une bonne couverture azimutale et 5 stations on peut trouver un modèle qui prédit de façon acceptable les formes d'ondes des autres stations.

L'inversion déterministe que nous avons mise au point utilise la partie basse fréquence des enregistrements des mouvements forts. L'étude du séisme de Northridge a montré qu'il était très dur de modéliser les fréquences supérieures à 0.5 Hz. Pour cette raison il est difficile de résoudre les variations spatiales du glissement de petite longueur d'onde (<5 km). La méthode est donc plus spécifiquement adapté à l'étude des "gros" séismes (Magnitude 7). Pour des séismes plus petits d'autres méthodes utilisant par exemple des fonctions de Green empiriques ont été testées (Courboulex, 1995).

Le traitement des données comprend plusieurs étapes: la digitalisation (pour les données analogiques), la correction instrumentale et l'intégration numérique éventuelle. Ce traitement limite la bande de fréquence "utile". Pour des données de SMA-1, la limite basse fréquence est égale à 0.1 Hz. Pour des données de type « TERRAscope » la dynamique est plus grande et on peut descendre la bande utile jusqu'à 0.05 Hz.

#### ♦ Connaître la géométrie de la faille

La géométrie de la faille sur laquelle le processus de rupture est analysé est choisie *a priori*. Nous avons vu lors de l'étude du séisme de Northridge l'influence du choix de cette géométrie sur la distribution de glissement finalement trouvée par l'inversion. Dans le cas d'un séisme où le plan de faille est mal connu, il faut attendre que la relocalisation des répliques ait été effectuées pour mieux estimer la position du plan de faille.

#### Connaître le modèle de vitesse

Pour retrouver la complexité de la source, il est indispensable de bien savoir représenter l'effet de la propagation de l'onde de la source à la station. Au cours de cette étude nous avons calculé les fonctions de Green dans un milieu à couches planes. Pour que le résultat de l'inversion soit bon il importe donc que nous connaissions un modèle de vitesse plan représentatif de la réalité.

# C. Comment améliorer l'inversion ?

# Mieux estimer les fonctions de Green.

Le modèle plan que nous utilisons pour représenter l'effet de la propagation de l'onde de la source à la station ne prend pas en compte les effets possibles des structures hétérogènes autour de la source et les effets de site. Un des dangers possibles d'une étude de la source à l'aide de fonction de Green non représentatives de la propagation consiste à reporter dans la fonction source des complexités non prise en compte par la fonction de Green calculée. Pour atténuer ce type d'erreur plusieurs précautions peuvent être prises. Au chapitre II-3 et II-4 nous avons ainsi décidé de limiter la durée du signal utilisée dans l'inversion. Ceci permet de ne pas inverser les arrivées tardives du signal. Nous avons aussi lors de l'étude de Northridge décidé de calculer les fonctions de Green dans des milieux à couches planes différents selon les stations. Nous avons contraint notre analyse à des fréquences inférieures à 1Hz pour limiter l'influence des hétérogénéités de petite longueur d'onde. Pour aller plus loin et mieux estimer les fonctions de Green dans des milieux hétérogènes, deux possibilités sont envisageables. On peut tout d'abord espérer dans les prochaines années pouvoir calculer les fonctions de Green numériques dans des milieux tridimensionnels. Tout le problème consistera à caractériser le milieu de propagation (études géotechniques de subsurface, technique pétrolière de sismique réflexion, tomographie). L'autre voie possible consiste à utiliser l'enregistrement des répliques d'un séisme comme fonctions de Green (fonctions de Green empiriques, FGE). Cette méthode est prometteuse mais son utilisation nécessite un certain nombre de conditions qui peuvent être contraignantes lors de l'étude des grands séismes:

- (1) Les mécanismes au foyer des FGE et du séisme étudié doivent être identiques. (2) Dans le cas où le séisme étudié est une source de grande dimension, il faut obtenir une couverture du plan de faille par les petits séismes car il est difficile d'utiliser une seule réplique en considérant que le milieu dans la région de la source est homogène.
- (3) Il faut que les stations ne saturent pas lors du séisme principal et déclenchent lors des répliques. Dans le cas d'un fort séisme seuls les accélérographes ne saturent pas. En revanche, ils ne déclenchent pas lors des répliques. Il faut donc pour enregistrer les répliques déployer après le séisme principal des vélocimètres à côté des accélérographes.

Un tel déploiement a été effectué en 1994 à la suite du séisme de Northridge par l'Institut des Etudes Crustales de Santa Barbara en Californie. Leur expérience montre que les conditions 1 et 2 restent difficiles à réunir et nécessitent de très bonnes localisations et donc un très bon réseau.

# D. Modifier la paramétrisation

Pour améliorer l'imagerie on pourrait aussi modifier la paramétrisation utilisée. Dans cette méthode, la forme de la fonction source est fixée *a priori* et le choix de cette fonction source n'est pas facile *a priori* (chapitre II-2) Lorsque le nombre de données est important on pourrait utiliser une fonction source dont la forme est paramétrisée par quelques paramètres. Le contrôle de ces paramètres par les données permettrait de contraindre la forme de la fonction source.

Une deuxième modification possible consiste à supprimer la discrétisation en sous-faille. Chaque sous-faille est composée de point-sources, regroupés artificiellement en sous-faille dont les dimensions sont choisies *a priori*. Le choix de la largeur de chaque sous-faille est déterminé (Chapitre II-1) en effectuant quelques tests pour voir quelle taille donne le meilleur compromis résolution-accord avec les données. On peut envisager à l'avenir une inversion sans sous-faille avec une longueur de corrélation corrélant les points source proches et l'inversion pour chaque point source du glissement, du début du glissement et du temps de montée. Ce type d'inversion a déjà été utilisé en tomographie (Nercessian et al., 1984) et permet d'obtenir directement des images « lisses». Son application au problème qui nous intéresse ne pose pas de problème théorique. Le seul problème est pratique car avec cette formulation il faut manipuler et inverser des matrices beaucoup plus grandes.

# Supprimer le couplage entre paramètres en incorporant d'autres types de données

Un des problèmes inhérents à la méthode d'inversion est du au couplage entre les paramètres. Nous l'avons vu au chapitre B1, distribution spatiale du glissement et vitesse du front de rupture ne sont pas résolus indépendamment. Le couplage est aussi important entre le temps de montée et l'amplitude du glissement sur chaque sous-faille (chapitres II-1 et II-2). Pour supprimer ce couplage il faut envisager d'incorporer d'autres types de contraintes dans l'inversion.

Certaines équipes inversent déjà simultanément les données géodésiques et de mouvement forts (Wald et Heaton, 1994, Wald et al., 1995). Les données géodésiques permettent d'avoir une image statique du glissement. Avec la densification des réseaux de mesure GPS (Global Positionning System), le nombre de données géodésiques collectées lors d'un grand séisme est de plus en plus grand. Tout le problème consiste à savoir s'il n'y a pas eu de glissement asismique dans les jours ou même les minutes qui ont suivis le séisme et si l'inversion des mouvements forts et des données géodésiques imagent bien le même phénomène: la répartition spatiale du glissement pendant le séisme. De plus, les données géodésiques sont surtout sensibles aux variations du glissement situées près de la surface et la résolution des inversions effectuées avec ce type de données est limitée en profondeur. En revanche, l'utilisation des images radars et notamment l'interférométrie permettent maintenant de mesurer la projection des déformations du sol selon une direction particulière en tout point autour de la faille (Massonet et al., 1993) le nombre de points de mesures compense sans doute le manque de résolution de chaque mesure prise individuellement. Un challenge intéressant consiste à incorporer ce type de données dans les inversions et à les coupler avec les données de mouvements forts.

Une autre solution à tester consiste à trouver indépendamment la vitesse du front de rupture, par exemple en utilisant la polarisation des ondes hautes fréquence (Bouin et Bernard, 1994), et à inverser ensuite linéairement la répartition spatiale du glissement sur le plan de faille.

#### Imager les hautes fréquences

Les inversions que nous avons effectuées utilisent la partie basse-fréquence des mouvements forts (<1.0 Hz). Pour les hautes fréquences, les milieux simples que nous utilisons pour calculer les fonctions de Green sont insatisfaisants (effet des petites hétérogénéités et de 1' anisotropie). Considérer les basses fréquences simplifie le problème, permet une inversion déterministe telle que nous l'avons effectuée mais induit une perte d'information que nous aimerions pourtant bien exploiter. A haute fréquence, la rupture est probablement incohérente (Bernard et Herrero, 1994). Pour décrire l'histoire du glissement à de petites longueur d'ondes il faudrait utiliser un nombre si important de paramètres que l'inversion n'a aucune chance de réussir. Ainsi une approche réaliste de l'utilisation de la partie haute fréquence ne doit pas considérer les variations détaillées de la phase mais seulement l'amplitude du spectre ou l'enveloppe du signal temporel haute fréquence (Cocco and Boatwright, 1993). Le paramètre à résoudre ne doit pas être l'histoire détaillée de la rupture mais plutôt un paramètre décrivant l'énergie haute fréquence radiée en chaque point de la faille. Localiser les endroits où sont rayonnées les hautes fréquences par rapport au front de rupture basse fréquence, analyser le contenu de ce rayonnement haute fréquence et notamment comprendre s'il peut permettre de comprendre quels sont les mouvement normaux au plan de faille (Brune et al., 1990) constituent des voies de recherches intéressantes.

#### ♦ Passer d'une image cinématique à une image dynamique

Les techniques d'imagerie de la rupture telles que nous les avons décrites permettent d'améliorer notre compréhension de la cinématique de la rupture. Néanmoins, ces résultats cinématiques restent insatisfaisants pour comprendre pleinement la physique de la rupture. Ces solutions décrivent en effet la rupture sous la forme de simples dislocations. Elles nous apportent très peu en ce qui concerne les niveaux de contraintes atteints sur la faille ou les lois constitutives des roches de la zone de faille. Pour franchir ce pas, Quin (1990) utilise la description cinématique du séisme d'Impérial Valley obtenue par Archuleta (1984) pour déterminer les chutes de contrainte dynamique en chaque point de la faille. Quin utilise un programme de simulation dynamique de la rupture (Das, 1980): les temps de rupture en chaque point sont donnés par l'image cinématique. Par essai-erreur (plus de 200 modèles ont été testés) Quin détermine la distribution spatiale des chutes de contrainte dynamique permettant d'ajuster la durée de la rupture et le déplacement final du modèle dynamique aux résultats donnés par le modèle cinématique d' Archuleta. Plus récemment d'autres tentatives ont été effectuées par Miyatake (1992) et Fukuyama et Mikumo (1993) pour convertir des modèles cinématiques obtenus par inversion des formes d'onde en modèles dynamiques. Tous ces travaux suivent le même principe et consistent à trouver des modèles dynamiques de la rupture cohérents (glissement et temps de ruptures identiques) avec les modèles cinématiques issus de l'inversion des formes d'onde. D'autres approches sont possibles. Imaginons que le glissement soit connu sur la faille grâce à l'inversion cinématique. Grâce au théorème de représentation, on peut alors calculer le déplacement en tout point du milieu en dehors des points sources. On peut ensuite calculer les déformations, puis les contraintes en utilisant la loi de Hooke. On obtient ainsi les variations de contraintes en chaque

point situé hors du plan de faille. Pour aller plus loin, on peut chercher à calculer par une intégration des contributions locales l'histoire des chutes de contrainte sur la faille. Ceci permettrait de calculer de façon exacte les changements de contrainte dynamique en tout point du milieu correspondant à l'image cinématique de la rupture obtenue par inversion des formes d'onde.

Troisième partie Séisme et contrainte

### Chapitre III-1

Stabilité du glissement pendant le séisme de Landers. L'indication d'une chute de contrainte partielle ?

> Article publié dans Geophysical Research Letters VOL 22, pages 1921-1924, 1995

#### III. Séisme et contrainte

1. Stabilité du glissement pendant le séisme de Landers. L'indication d'une chute de contrainte partielle?

#### A. Résumé

Nous utilisons les données de haute qualité du séisme de Landers pour étudier en détail la capacité des inversions de mouvements forts à trouver les variations spatiales de la direction de glissement sur le plan de faille. Dans une première étape, cette direction de glissement est choisie constante sur le plan de faille. Plusieurs inversions avec différentes valeurs de la direction de glissement montrent qu'un décrochement pur donne le meilleur accord entre données et synthétiques. Dans une seconde étape, la direction de glissement n'est plus constante sur le plan de faille et peut varier spatialement. Les variations spatiales de la direction de glissement sont bien résolues par les données. Ces variations restent faibles: le séisme est un décrochement pur avec pratiquement aucune variation de la direction de glissement sur le plan de faille. En chaque point de la faille, la direction de glissement est colinéaire à la direction de la contrainte appliquée. La contrainte appliquée est elle même la somme de la contrainte existante avant le séisme et de la variation de contrainte dynamique due au processus de rupture. La stabilité spatiale de la direction de glissement sur une faille où le processus de rupture est clairement hétérogène suggère que la contrainte existante avant le séisme est homogène et qu'elle est supérieure à la variation de contrainte dynamique.

#### Stability of the rake during the 1992, Landers earthquake. An indication for a small stress release?

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Abstract. We use the high quality data set of the Landers Data and a priori parameterization earthquake to study in detail the ability of strong motion inversions to resolve the rake during the rupture process. Initially, we constrain the rake to be constant on the fault. Several inversions with different rake values show that a pure strike slip gives the best fit to the data. In a second step, we have allowed the rake to vary spatially. Spatial rake variations are well resolved by the data and remain small; the data strongly constrain the earthquake to be a pure strike-slip event with almost no variation of the rake. We assume that the direction of the slip vector is always close to the direction of the applied stress, which itself is the vector sum of the preexisting stress and the dynamic stress changes. The stability of the rake on a fault with a dynamical rupture process which is clearly heterogeneous suggests that the pre-existing stress level is quite homogeneous and that it dominates over the dynamic stress changes.

#### Introduction

In their study of the 1989 Loma Prieta earthquake several authors (Beroza [1991], Hartzell et al. [1991], Wald et al. [1991] and Steidl et al. [1991]) allow the rake to vary. All find strong variations in the slip direction over short distances (about 15 km). But the rake distributions in these studies are significantly different from one another and the question of the resolution of the rake in the case of Loma Prieta has been addressed by Beroza [1991] and by Cocco et al. [1994].

The 28 June 1992 Landers earthquake provides one of the most complete sets of near-source strong motion records. The good azimuthal distribution of strong motion instruments in the vicinity of the Landers event allows the characteristics of dynamic rupture propagation to be well resolved. Several studies (Cohee and Beroza, 1994; Wald and Heaton, 1994; Cotton and Campillo, 1995) have used these data to infer the detailed history of the rupture by waveform matching. The slip in all these models has been satisfactorily modeled with a constant rake of 180°. The difficult question is whether rake values are really constant and equal to 180° over the fault, or whether the data do not constrain these rake values so that other values of rake could equally well explain the strong motions. Using this unique set of records, we have concentrated our analysis on the rake variations: we find that this absence of rake variation is actually required by the data.

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The strong motion station distribution is shown in Figure 1. Each of the 11 stations used in this study recorded three components of ground acceleration. The accelerograms are band-pass filtered between 20 s and 2.0 s, then doubly integrated in the frequency domain to obtain displacements.

The crustal model used in this study is that described in a forward modeling study of the Landers earthquake rupture by Campillo and Archuleta [1993], which was only slightly modified from Kanamori and Hadley [1975]. We assume that the surface offset and the aftershocks distribution give an image of the actual fault at depth. These observations lead us to consider a model which consists of three vertical distinct segments (30 km, 20 km, and 30 km long) oriented with different strikes (Figure 1).

#### Strong motion modeling and inversion procedure

The fault parameterization and modeling procedure is that of Cotton and Campillo [1995] except that we now allow the rake to vary by solving for both strike-slip and dip-slip components. We divide the fault plane into 48 subfaults. Each subfault is allowed to slip once. The ground motion  $\nu$  at a given station i and a given frequency  $\omega$  can be represented as a linear sum of n subfault contributions:

$$v_{i}(\omega) = \sum_{k=1}^{n} (\Delta s_{k} u_{k}(\omega) + \Delta d_{k} u_{k}(\omega)) \exp^{-i\omega t_{k}} S_{k}(R_{k}, \omega)$$

where  $u_{_{\!\scriptscriptstyle H}}$  and  $u_{_{\scriptscriptstyle H}}^{'}$  represent the ground motion for a unit constant strike-slip (rake equal to 0°) and dip-slip (rake equal to 90°) on the subfault k.  $\Delta s$ ,  $\Delta d$ , t, and S are the strike slip, the dip slip,

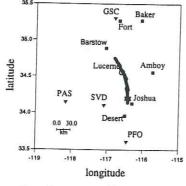


Figure 1. Location map displaying CDMG (squares). TERRAscope (triangles), and Southern California Edison (octagon) strong-motions stations used in this study. Heavy solid lines represent the surface rupture of the earthquake.

the beginning rupture time and the source function  $(S(t)=0.5(1+tanh(t+R/2.0)^2)$  of the k-th subfault depending on a single variable: the rise time R. In practice, each subfault is represented by an array of point sources separated by 500 meters. The point-source Green's functions are computed using a discrete wavenumber integration method associated with a reflection transmission matrix method.

The ground motion  $v(\omega)$  can be considered as a function of the following parameters:  $\Delta s_{t}$ ,  $\Delta d_{t}$ ,  $t_{t}$  and  $R_{t}$ . The elements of d consists of real and imaginary part of displacement spectra from all stations and all components (in the following d is represented by 3960 elements). The parameter vector p and data vector d are related by the function model vector f as d=f(p). If we assume an initial parameter vector  $p_p$ , we can iterate to the solution  $p_{\nu}$  by linearization of f around  $p_{\rho}$  at the first iteration and around  $p_{k-1}$  at each subsequent iteration k. Using the observed data vector  $d_0$  and an inversion algorithm Figure 3. Strong motion seismograms of the Landers based on Tarantola and Valette [1982], p is given by:

$$p_{k+1} = p_k + b(A_k^i C_{\cdot i}^{-i} A_k + C_{\cdot i}^{-i})^{-i} (A_k^i C_{\cdot i}^{-i} (d_o - f(p_k)) + C_{\cdot i}^{-i} (p_o - p_k))$$

Here  $A_k$  is the Jacobian matrix of  $f(p_k)$ , b is a damping constant between 0 and 1 used to prevent divergence, and  $C_n$  and  $C_d$  are the covariance matrices for p and d. We assume that offdiagonal elements of  $C_n$  and  $C_d$  are equal to 0. The diagonal elements of  $C_d$  and  $C_n$  are given by the data and parameter variances. In this inversion, we give a greater weight to the TERRAscope stations, which have an acquisition system with higher dynamic range by giving smaller variances ( $C_d$ =0.25) to these stations relative to the others  $(C_d=1.0)$ .  $C_p$  is equal to 81.0 for all the parameters. These diagonal values were found after several test inversions to find the best convergence. The observed and synthetic amplitude spectra at each station are normalized by the maximum observed amplitude of the three components.

#### Results

As a first step, we have tried several inversions with constant rake on all the subfaults. In those inversions the ratio between the strike slip and the dip slip is held constant and the inversion gives only the rise time, the time of rupture and the slip amplitude (144 unknows). The fit to the data is evaluated using the variance reduction between theoretical and observed seismograms as defined by Cohee and Beroza [1994]. A variance reduction of 100% means that the data are perfectly matched by our synthetics. Figure 2 gives the final variance reduction for each chosen value of rake. The rake which gives

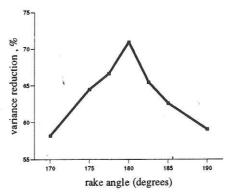
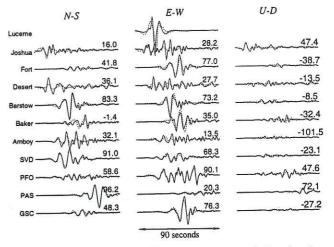


Figure 2. Variance reductions given by inversions performed with different rake directions. In those inversions the rake direction is the same on the all fault plane.



earthquake (solid line) compared with synthetic seismograms (dotted line). Each pair of data and theoretical seismograms is plotted at the same amplitude scale with the variance reduction shown to the right of each pair.

the best fit (variance reduction of 71%) is equal to 180° (pure strike slip). A variation in rake of 10° around this pure strike slip decreases the fit to the data significantly: with a rake of  $190^{\circ}$  and  $170^{\circ}$  the variance reduction are, respectively, 58% and 59%. These results also suggest that this type of study cannot be performed with only a visual evaluation of the fit. Indeed, models with a variance reduction of 58% give fits to the data which, visually, seem acceptable.

A s a second step, the rake is allowed to vary spatially in the inversion (192 unknowns). The variance reduction between the data and synthetics with a starting model for the rake of 180° is 71.5 %. Figure 3 shows the fit between data and synthetics. This value is very close to the variance reduction found in the inversion with constant rake of 180°, described above and presented in Cotton and Campillo [1995]. In other words, we do not improve the fit to the data by permitting the rake to vary during the procedure from an initial value of 180°. The difficult point is to know is whether spatial rake variations really are small or whether the inversion simply cannot determine these

Since the inversion is not purely linear, the final values for the parameters which are not well constrained by the data are mostly dependent on the a priori value chosen in the starting model. If we choose different starting models, we identify the unresolved portion of the solution by the parameters which do not converge to a common value. We have performed our inversion procedure with 3 initial values of rake on the subfaults of the starting model po: 170°, 180° and 190°. On each subfaults, the inversion procedure leads to final values of the rake which give a better fit to the data. For each a priori rake value chosen as a starting model, the numbers of subfaults with a given rake value in the final model are presented in an histogram Figure 4. Regardless of the initial values of the rake, most of the final values are between 175° and 185°. Even, if the initial a priori values of the rake is equal to 190° or 170°, the inversion changes these initial values to new values close to 180°. This shows that the final rake values do not depend on our a-priori choices. It is therefore clear from these tests that the data require the rupture to be almost pure strike-slip (when significant moment release occurred).

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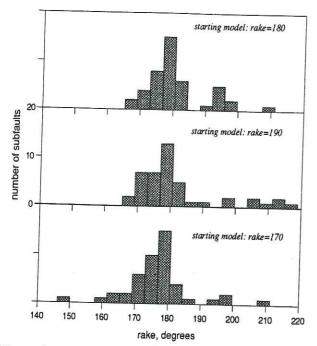


Figure 4. Distribution of the final rakes values found by our inversion with different a priori starting model (top: starting Landers region. These data show that prior to the Landers value of the rake equal to 170°, middle: 180°, bottom: 190°)

To illustrate this point, we have calculated for each subfault the mean and standard deviation of the rake values found by the inversions with three different starting models. Figure 5 shows for each subfault, the amplitude and direction of this mean slip vector of the west wall of the fault. Considering only the subfaults where the amplitude of slip is large (more than 1.5 m) we notice that most of the rakes are close to 180° (Figure 6). Most of the deviations of the rake with respect to  $180^{\circ}$  occur only on the Emerson and Camp Rock faults, farther than 40 km from the epicenter. One can also notice negative values of slip on the edge of the actual rupture zone. In spite of the lack of positivity constraints, these nonphysical negative values remains rare. This absence of positivity constraints allows to reduce the influence of the a priori choices on the final result of the inversion.

#### Discussion and conclusion

Nearly all the waveform inversion studies which have been done over the past 10 years to determine the slip histories of earthquakes have shown that the slip amplitudes, rupture velocities and rise times have a heterogeneous distribution on the fault plane. This is also the case for the Landers earthquake. Our results which prove the stability of the value of the rake, therefore seems surprising. How could the rake be spatially constant in a rupture process which has a very heterogeneous behavior?

The total stress at a point on the fault plane is given by the frictional stress which is the result of an interaction between the pre-existing stress and the dynamic stress changes. If we assume that a highly heterogeneous rupture process results in spatial variability of dynamic stresses and that the sliding friction stress is collinear with the slip velocity, the stability Figure 6. Rake values function of the distance from the

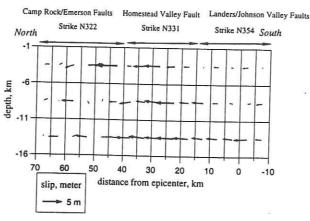
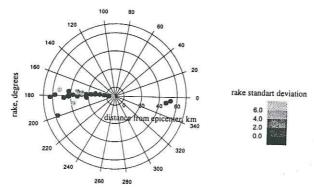


Figure 5. Vector plot of the rake as a function of position on all the subfaults.

leads to the 2 following propositions. 1): the spatial variations of the pre-existing stress are small and 2): this preexisting stress probably dominates dynamic stress variations (partial stress release).

Hauksson [1994] used focal mechanisms of the background seismicity to determine the state of stress before 1992 in the earthquake spatial stress variations were small in agreement with our proposition 1). Hauksson [1994] also studied the aftershocks of the 1992 event and showed that the direction of the deviatoric stress tensor did not change significantly from the pre-event state in the southern and central part of the fault. This is where about 70% of the moment release occured (Cohee and Beroza, 1994; Wald and Heaton, 1994; Cotton and Campillo, 1995), and is consistent with our proposition 2), at least for the southern part of the fault. According to Hauksson [1994], since the strike of the rupture changes as it propagates to the north, the angle between the maximum principal stress and the orientation of fault ranged from  $37^\circ$  (near the epicenter) to  $58^{\circ}$  (in the north). This means that the ratio between the dynamic shear stress release and the preexisting shear stress becomes greater as the rupture propagates to the north which could explain the greater variability of rake found on the Camp Rock and Emerson faults. Hauksson [1994] also found a greater variability of the aftershocks in the northern section, and interpreted this to mean that much of the uniform component of the shear stress had been removed on the Emerson and Camp Rock faults. However, all the inversions performed with the data of the Landers earthquake (Cohee and Beroza, 1994; Wald



of the rake on a fault which has a heterogeneous rupture process epicenter when significant slip (>1.5m) was found.

and Heaton 1994; Cotton and Campillo, 1995) have shown for this section of the fault, although the slip was very high (>5m), it was located only near the surface. Therefore coseismic slip effect did not involve the entire brittle crust in the northern segment, and could not have relieved all the preexisting stress there.

It is difficult to give a quantitative evaluation of the ratio of initial stress to stress drop. Comparing our inversion results with published spontaneous rupture models is not easy because all of these models have been run for high ratio of initial stress to stress drop or under the assumption of a constant rake. For example, in Das [1981] the ratio of initial stress to stress drop is 5. These values can be considered as upper bounds of the values required to ensure a rupture with a constant rake, assuming an homogeneous initial stress. The development of 3D dynamic codes of spontaneous rupture in stratified media is needed to study quantitatively the variability of dynamic stresses and rake in heterogeneous rupture processes. An original approach is proposed by Spudich [1992] who discusses the conditions under which it is possible to determine the absolute stress level at a point on a fault from observation of temporal rake rotations at that point. He shows that if the rake rotates, there is a unique absolute stress consistent with this motion. We cannot use these results because the inversion of temporal rake variations strongly increases the number of parameters and therefore the rake at one point of the fault is constrained to be constant in our inversion.

Finally, how can we explain the strong rake variations found for the Loma Prieta earthquake but small variations for the Landers earthquake? One possible explanation is a bias in the rake estimates due to a lack of resolution in the data set of the Loma Prieta earthquake (Beroza [1991]). In fact, Cocco et Tarantola, A. and B. Valette, Generalized nonlinear inverse problem al. [1994] has shown that strong motion inversions are able to find many physically acceptable models which fit these data equally well. The partition between the constrained and unconstrained parts of the solution depends on the distributions of stations, fault geometry and mechanism. It must be analysed for each particular case. A second possible explanation of the contrast of amplitude of rake variations between the Loma Prieta and Landers earthquakes may be due to the near total stress drop and near-frictionless heterogeneous faulting during the 1989 Loma Prieta, California, earthquake as Zoback, M.D. and G.C. Beroza, Evidence for near-frictionless faulting suggested by Zoback and Beroza [1993] who analysed the variation of focal mechanism of aftershocks. On the contrary, Landers earthquake released only partially the stress on the Johnson Valley and Homestead Valley faults.

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## Chapitre III-2

Variations de contraintes dynamiques associées à la rupture sismique

Article soumis à Geophysical Journal International

# 2. Variations des contraintes dynamiques due à la rupture sismique dans un milieu à couches planes

#### A. Résumé

Les équations permettant de calculer les variations de contraintes dynamiques dans un milieu plan sont présentées. Nous utilisons une méthode de réflectivité et de nombre d'onde discret pour calculer le champ des contraintes radié par une source sismique. Les expressions développées dans cette partie représentent un outil intéressant pour l'analyse des phénomènes associés aux variations de contraintes dynamiques engendrées par un séisme.

## Dynamic stress variations due to shear faults in a plane layered medium.

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#### **Abstract**

A complete set of expressions is presented for the computation of elastic dynamic stress in plane layered media. We use a discrete-wavenumber reflectivity method to compute the stress field radiated by any moment-tensor source. The expressions derived here represent an interesting tool both for the observational and theoretical analysis of dynamic stress changes associated with earthquake phenomena.

Key words: dynamic stress, triggered seismicity, rupture complexity, wavenumber, reflectivity

#### B. Introduction

Earthquake rupture complexity can be partly due to triggered slip phenomena. In this process, slip on one fault can be triggered by an earthquake on another fault, or by a distant segment rupturing on the same fault. If the triggered slip comes soon and close after faulting, it will be usually considered as part of a complex dynamic event. Such heterogeneous slip distributions have been revealed by numerous waveform inversions in the last decade (e.g. Olson and Apsel, 1982; Hartzell and Heaton, 1983; Fukuyama and Irikura 1986; Beroza and Spudich, 1988). All these inversions are based on kinematic fault models. Their results show slip values as a function of space and time but they do not describe the associated stress changes on and around the fault. Nevertheless, it is of great interest to quantify the radiated transient stress waves. These stress waves can probably reach a high level and trigger events on other sections of the fault which were not themselves ready to self-nucleate.

In the other case, if the slip triggering is delayed compared to the main event rupture time, it is most likely related to changes in the static-stress field. Since it was first pointed out by Das and Scholz (1981), it has been recognized that the increase in static-stress produced by an earthquake can trigger, with a certain time delay, earthquakes on nearby faults. Using analytical expressions to quantify static-stress changes (Chinnery, 1963, Okada 1985, Okada, 1992), numerous papers in the last ten years have shown that aftershocks occur in preferred locations where static-stresses due to the main shock have

increased (e.g. Reasenberg and Simpson, 1992; King et al., 1994)

At large distances (more than tens of kilometers), the computed static-stress changes appear to be too low to cause triggering (Hill et al., 1993). The M 7.4 Landers earthquake however triggered widespread remote seismicity in the western United States (Hill et al., 1993; Anderson et al., 1994). These distant triggered slips could be related to the dynamic-stress field propagating from the main rupture event. For this reason, several studies have tried to quantify dynamic transient-stress by estimating peak dynamic strain from surface ground-motion records (Hill, 1993). Since frequency dependence and variability of stress with depth and geometric structure could be important to the understanding of dynamic triggering of earthquakes, there have been also attempts to quantify the full dynamic stress or strain tensor (Gomberg and Bodin, 1994; Anderson et al., 1994, Spudich et al. 1995). Spudich et al. (1995) have used the displacements recorded by the UPSAR (U.S.G.S. Parkfield Small Aperture Array) array to determine the surface strain and stress tensor as a function of time. Gomberg and Bodin (1994) have developed an approach to quantify the theoretical dynamic strain tensor at remote distances from the main event. Synthetic seismograms including the fundamental and first higher modes of surface waves are computed. Differentiation of the particle displacement expressions then yields the dynamic strains. Closer to the fault however, particle motion is not dominated by surface waves and the quantification of theoretical dynamic stress tensor changes is not possible with this method.

Both the phenomenon of distant triggered seismicity and the complex multievent nature of earthquakes suggest that the quantification of transient stresses is important. This article describes a study that attempts to give theoretical expressions of stress tensor changes as function of time associated with shear rupture on a fault plane. We present a complete set of expressions for the calculation of dynamic stress variations resulting from a shear point source in a plane layered elastic medium. These expressions are valid for all source-station distances. The main goals of this paper are, (1) to give the expressions of the stress tensor field radiated by a shear rupture in a plane layered medium, (2) to show numerical results of dynamical stress changes due to a strike-slip rupture.

#### C. Method of computation

We compute the stress field radiated by the six elements of the moment tensor using the reflectivity method of Kennett and Kerry (1979) with the discrete wavenumber decomposition of the Green's functions given by Bouchon (1981) for an axisymetric medium. These two methods are well known and have been extensively used to compute displacement fields in plane layered media. Computing stress field implies one more differentiation step from displacement to stress field, once displacements have been derived from potentials. We used as a starting point an extensively used and tested code (Coutant 1990, unpublished), which computes the displacements radiated by moment

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sources. In this paper, we present the analytic formulas giving the stress Green's functions for six independent moment sources (3 dipoles and 3 double couples) in a homogeneous space, using the discrete wavenumber representation with cylindrical coordinates. These formulas have been arranged to be included in a reflectivity propagation method. The resulting program allows one to compute stress at several receivers, due to several sources, in any geometry except when source and receiver are at the same depth.

Let us briefly describe the method of computation. We use the formalism given by Muller, (1985), where seismic wavefields are expressed with the three scalar potentials  $(\phi, \psi, \chi)$  representing respectively P, SV and SH waves in the cylindrical coordinates  $(r, \theta, z)$  system. Displacement field is given by:

$$\mathbf{u}(r,\theta,z) = \nabla \phi + \nabla \times (\mathbf{e}_z \psi) + \nabla \times \nabla \times (\mathbf{e}_z \chi) \tag{1}$$

We distinguish three major steps in the computation:

(a) From the expressions giving the potentials radiated by single forces (see e.g. Muller, 1985, Bouchon, 1981), we derived the upward and downward potentials in the source layer(s) radiated by six independant moment sources. The radiated fields can be expressed as the linear combination of the six elementary potential sources given by (see appendix A for detail):

(b) The potential fields for these six elementary sources are propagated through the layers. In the receiver layer(s) each of them yields upgoing and downgoing potentials, including possible conversions. This step corresponds to the reflectivity propagation and is not detailed in the paper. We obtain the following resulting potential terms in the receiver layer(s):

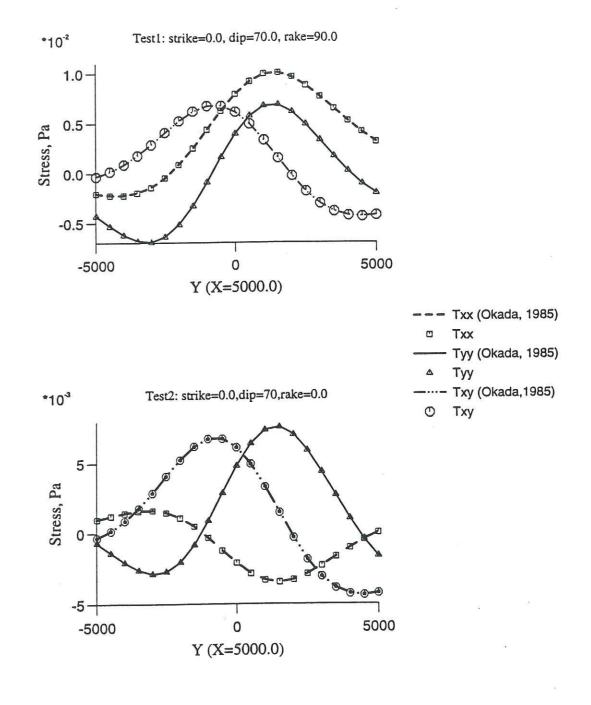


Figure 1. Comparison of stress tensor static changes computed by our method with those derived from the analytical expressions of Okada (1985) These comparison are done for two different point sources and 21 stations located at the surface of a half space (Txz, Tzz and Tyz are null because of the boundary conditions at the free surface).

(c) Previous potentials are differentiated to obtain displacements using Equation 1, and then stress is computed using relations derived from Hooke's law:

$$\begin{split} &\tau_{rr} = (\lambda + 2\mu)u_{r,r} + \frac{\lambda}{r}u_{\theta,\theta} + \lambda u_{z,z} + \frac{\lambda}{r}u_{r} \\ &\tau_{\theta\theta} = \lambda u_{r,r} + \frac{\lambda + 2\mu}{r}u_{\theta,\theta} + \lambda u_{z,z} + \frac{\lambda + 2\mu}{r}u_{r} \\ &\tau_{zz} = \lambda u_{r,r} + \frac{\lambda}{r}u_{\theta,\theta} + (\lambda + 2\mu)u_{z,z} + \frac{\lambda}{r}u_{r} \\ &\tau_{zr} = \mu \Big(u_{z,r} + u_{r,z}\Big) \\ &\tau_{z\theta} = \mu \bigg(u_{\theta,z} + \frac{1}{r}u_{z,\theta}\Big) \\ &\tau_{r\theta} = \mu \bigg(r \bigg(\frac{u_{\theta}}{r}\bigg)_{,r} + \frac{1}{r}u_{r,\theta}\bigg) \end{split}$$

Details of stress and displacement expressions as functions of the source potentials given by Equation 3 may be found in appendix B.

Using discrete wavenumber decomposition with cylindrical coordinates is known to produce small artifacts. A plane wave propagating vertically is sometime observed in all expressions including the term:  $J_0(k_r,r)$ , where  $J_0$  is the zero order Bessel function. To reduce this artifact, we use the solution proposed by Herrmann and Mandal (1986).

#### D. Tests

To check the validity of our expressions, we have performed two tests: (a) comparison with static stresses obtained analytically, (b) comparison with dynamic stress obtained using a finite difference approximation of the derivative operator from computed displacements.

#### (a) comparison with static results

In the last decade, several authors have proposed analytical expressions for static displacement, strain and stress due to a point source in an elastic medium (Chinnery, 1963,

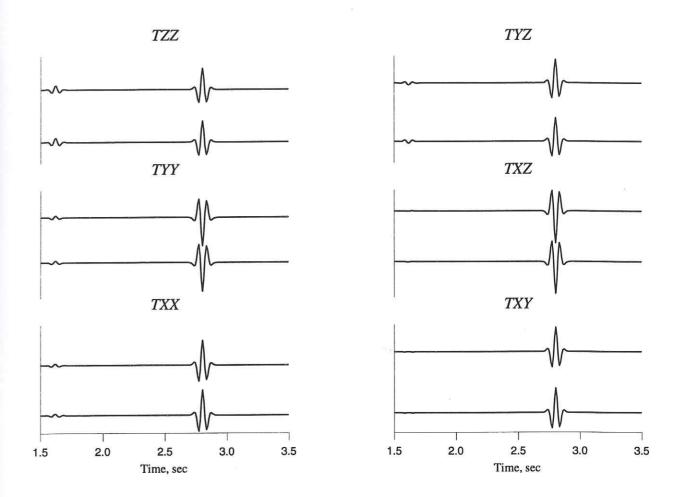


Figure 2. Comparison of dynamic stress variations obtained with our stress Green's function (top) and with finite difference approximation of spacial derivative (bottom).

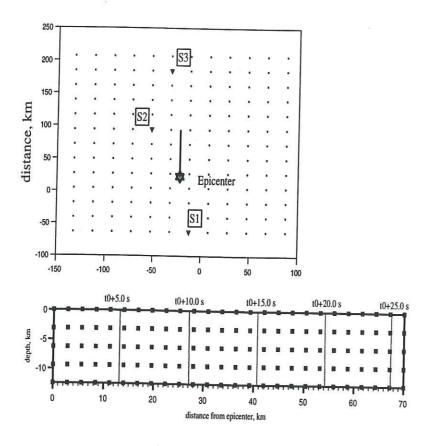


Figure 3. Top: map view of the fault (solid line) and position of the stations (circles) used in the simulations of Figures 4,5,6. Triangles give the location of the 3 observation points S1, S2 and S3. Bottom: the fault plane is 12.5 km high and discretized into 130 point sources (black squares). The rupture front has a velocity of 2.7 km/sec and is propagating from the epicenter to the north. Its position is represented every 5 seconds by black lines on the figure.

Okada 1985, Okada, 1992...). The expressions described in this paper allow the computation both of the dynamic stress associated with propagating elastic waves and the static stress variations. The purpose of this test is to check the static values obtained with our computation against the values obtained from the analytical expressions given by Okada (1985). In these tests (Fig. 1), static stress changes are compared for two different point sources and 21 stations in an elastic halfspace ( $v_p$ = 4000 m/s ,  $v_s$ = 2700 m/s and  $\rho$ = 2500 kg/m³). The coordinates of sources are x=0, y=0 and z= 5000 m and the stations are at the free-surface (x=5000, y between -5000 and 5000 m). Static changes obtained from our calculations match perfectly the values calculated with the analytic expressions of Okada (1985).

#### (b) comparison with finite difference results

We now check the accuracy of the expressions for stress Green's functions by comparing our results with those obtained by computing numerically stress from displacement. We first compute the ground displacement at four closely located stations whose coordinates are given in meters by: S1: (1000, 4000, 5000), S2: (1000.05, 4000, 5000), S3: (1000, 4000.05, 5000) and S4: (1000, 4000, 5000.05). The source is a point source with a moment of 1N.m, a strike of 10°, a dip of 20° and a rake of 30°. The characteristics of the medium are  $v_p$ = 4000 m/s ,  $v_s$ = 2390 m/s and  $\rho$ = 2500 kg/m³. Spatial derivatives of the displacement are approximated using first order finite differences. From them, we calculate the strain variations and, using Hooke's law, the stress variations near the station 1. These stress variations compare very well (Fig. 2) with the results obtained directly with the expressions presented above.

#### E. Example of computation for a finite fault

We present results of dynamical stress changes due to a shear rupture on a finite fault and try to illustrate the different effects that control transient stress variations. In these examples, time histories of the stress tensor computed with the expressions described above are presented. In-plane  $\tau_{\beta}$  and normal stress  $\sigma_{\beta}$  (extension positive) changes on observation planes arbitrarily oriented are also calculated. To characterize the conditions under which failure occurs in rocks, one of the more widely used criteria is the Coulomb failure criterion:  $\sigma_c = \tau_{\beta} + \mu_c \ \sigma_{\beta}$  (where  $\mu_c$  is the friction coefficient). Assuming a given coefficient of friction  $\mu$ , we compute the spatio-temporal variations of this failure criterion. In these simulations, a vertical fault (70 km long and 12.5 km large) is assumed to be located at the center of a grid containing 156 stations (Fig. 3). The fault is represented by 130 point sources, whose focal mechanism is pure right-lateral strike-slip (strike=0). The tapered slip on the fault is set to 4.2 m. These parameters represent an earthquake with a moment of 6.2  $10^{19}$  N.m. The rupture initiates at the south of the fault and the rupture front velocity is 2.7 km/s. The source function at each point source is a

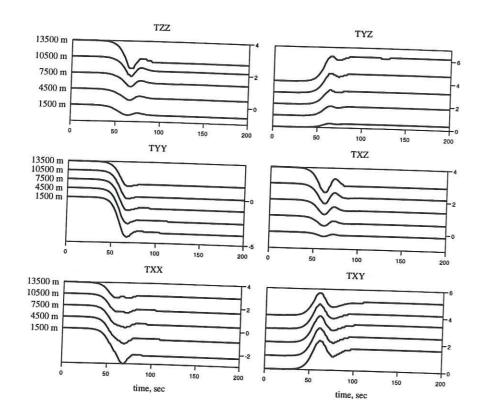


Figure 4. Stress tensor in bars observed at observation point S2 located in meters at (x,y)=(94311.6, -52497.6) for different depths varying between 1500 and 13500 m. Traces are shifted by 1.0 bar.

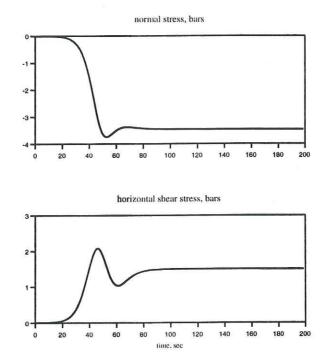


Figure 5. Horizontal in-plane stress and normal stress changes on a vertical plane striking N0 situated at S2 (depth 5500m)

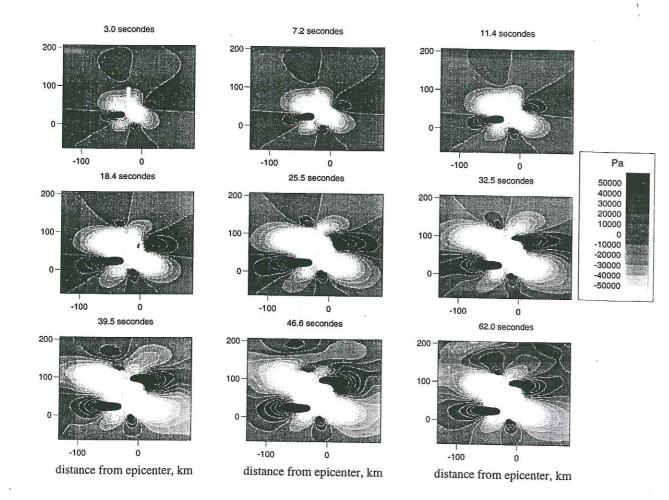


Figure 6. Coulomb failure spatio-temporal changes. These maps are obtained from the Coulomb failure criterion histories calculated at each of the 156 stations shown in Figure 3, along vertical planes striking North and with a friction coefficient equal to 0.4. Stress is sampled at a depth of 5500 m. The maximum and minimum stress changes exceed the plotted gray bar range.

ramp with a rise-time of 10 seconds. We compute for each station 360 seconds of stress time history with a frequency band of 0-0.18 Hz. This means that the distance between 2 point sources is less than one sixth of the minimal wavelength.

Variation of transient stress with depth

To illustrate the dependence of transient stress variations on depth, stress tensor histories are calculated beneath an observation point S2 at different depths between 1500 m and 13500 m. These results (Fig. 4) show that dynamic stress changes at depth are not correctly represented by estimations from surface seismograms and Figure 4 clearly shows that the xz, yz and zz components of stress are increasing with depth. This dependency has also been noted by Gomberg and Bodin (1994) and Spudich et al. (1995). These numerical tests confirm the analysis obtained by Spudich et al. (1995) in their study of dynamical stress variations detected at Parkfield after the Landers strike-slip earthquake: in the upper crust, transient stresses associated with horizontally oriented tractions on vertical planes (xx,yy and xy components) dominate. As already mentioned by Spudich et al. (1995), this means that the transient stresses due to a major strike-slip event tend to favor faulting across vertical planes in the upper crust, and other orientations below. Should these dynamic stresses trigger seismicity by direct working and weakening of material, triggered seismicity would exhibit different mechanism with depth, which has not yet been observed.

#### Distance and directivity effects

To illustrate distance and directivity effects, normal and horizontal in-plane stresses are computed on vertical fault planes striking North at a depth of 5500 m for each of our 156 stations. Figure 5 shows the results of these calculations at the observation point S2. The Coulomb failure criterion (CFC) is calculated with a friction coefficient of 0.4 and Figure 6 shows the results obtained by interpolating the CFC computed at each station. Figure 7 shows the CFC time histories at observation points S1, S2 and S3 located at different distances from the fault. One can notice that static stresses rapidly decrease with increasing distance and are significant only in the neighborhood of the fault (static stress changes observed at locations S1 and S3 are null). On the other hand, dynamic stress associated with propagating waves and specifically surface waves, have a falloff of less than R⁻² (Hill et al., 1993; Anderson et al., 1994) and are still important at observation points S1 and S3. The difference in amplitude between S1 and S3 is striking because these two points are at the same distance from the fault. This is clearly a directivity effect: the fault rupture propagates northward and waves have focused toward the north. Directivity also explains the high amplitude stress pulse observed at S2.

#### F. Conclusions

We have developed a method to compute numerically theoretical stressgrams in order to estimate the dynamical stress tensor associated with fault rupture. This method represents

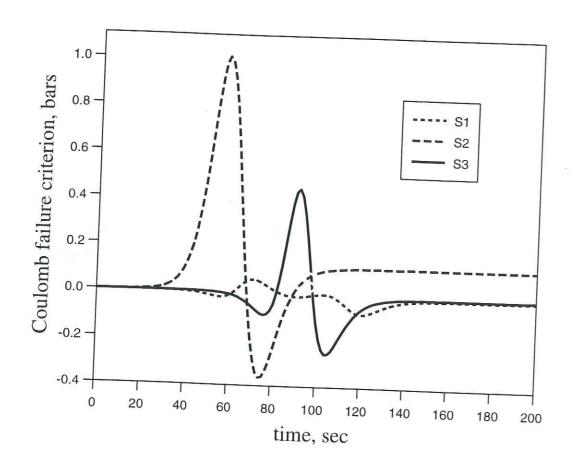


Figure 7. Coulomb failure time histories at the observation point S1, S2 and S3

a powerful tool both for the observational and theoretical analysis of dynamic stress changes associated with earthquake phenomena. For example, quantification of dynamic stress variations is important in order to understand the dynamic triggering of earthquakes, and our numerical results illustrate directivity effects and the depth dependence of dynamic stress tensor variations. The authors are ready to provide the source code to any interested researchers.

The following notations are used in the next expressions:

 $J_n = J_n(k_r,r)$ , where  $J_n()$  is the Bessel's function of order n.

 $k_r = \frac{2\pi n}{L}$  is the radial wavenumber and  $r = ||\vec{r} - \vec{r_0}||$  is the distance between source and receiver.

 $k_{\alpha} = \frac{\omega}{\alpha}, k_{\beta} = \frac{\omega}{\beta}$  are the P, S-wave wavenumbers,

 $\gamma, \nu = \sqrt{k_{\alpha}^2 - k_r^2}, \sqrt{k_{\beta}^2 - k_r^2}$  are the P, S-wave vertical wavenumbers with  $Im(\gamma) < 0$  and  $Im(\nu) < 0$ .

We also define the following variables:

$$C_f$$
:  $\frac{1}{2\rho\omega^2L}$ 

$$K_0$$
:  $k_r J_0$ 

$$K_1: k_r J_0 - 2J_1/r$$

$$K_2$$
:  $\frac{J_1}{r}$ 

$$K_3 = \frac{\partial K_1}{\partial r} : \quad -\frac{2}{r} K_1 - k_r^2 J_1$$

$$K_4 = \frac{\partial K_2}{\partial r}$$
:  $\frac{K_1}{r}$ 

$$K_5 = \frac{\partial J_1}{\partial r}$$
:  $k_r J_0 - J_1/r$ 

$$K_6 = \frac{\partial K_3}{\partial r}$$
:  $-k_r^2 K_5 + \frac{2}{r^2} K_1 - \frac{2}{r} K_3 = -k_r^2 (K_1 + K_2) - 2K_7$ 

$$K_7 = \frac{\partial K_4}{\partial r}$$
:  $-\frac{3K_1}{r^2} - k_r^2 K_2 = \frac{K_3 - K_4}{r}$ 

$$K_8 = \frac{\partial K_5}{\partial r}$$
:  $K_3 + \frac{K_1}{r}$ 

$$K_9 =: \frac{K_3 - 4K_4}{r}$$

$$K_{10} =: \qquad \frac{4K_4 - K_3}{r} + K_6 = K_6 - K_9$$

$$K_{11} =: \frac{K_3}{r}$$

$$K_{12} =: \frac{K}{\tau}$$

$$K_{13} =: \frac{K_8}{r}$$

$$cs2 = \frac{k_r^2}{i\nu};$$
  $cs3 = \frac{k_\beta^2}{i\gamma};$   $cs9 = \frac{k_\beta^2 - 2k_r^2}{i\gamma}$ 

#### 0.1 Potential Green's functions for a homogeneous space

Using previous definitions, they are:

0.1.1 Single couples  $M_{xy}$  and  $M_{yx}$   $(\frac{\partial F_x}{\partial y_0}$  et  $\frac{\partial F_y}{\partial x_0})$ 

 $M_{xy}$ :

$$\begin{split} \phi(r,z,\theta) &= C_f F_x \cos \theta \sin \theta \, \, \frac{k_r^2}{i\nu} \, \left( k J_0 - \frac{2J_1}{r} \right) \, e^{-i\nu |z-z_0|} \\ \psi(r,z,\theta) &= C_f F_x \cos \theta \sin \theta \, \, sign(z-z_0) \, \left( k J_0 - \frac{2J_1}{r} \right) \, e^{-i\gamma |z-z_0|} \\ \chi(r,z,\theta) &= C_f F_x \, \left( \frac{k_\beta^2}{i\gamma} \right) \, \left( \sin^2 \theta \, \left( k J_0 - \frac{2J_1}{r} \right) + \frac{J_1}{r} \right) \, e^{-i\gamma |z-z_0|} \end{split}$$

 $M_{yx}$ :

$$\begin{split} \phi(r,z,\theta) &= C_f F_y \cos\theta \sin\theta \ \frac{k_r^2}{i\nu} \left(kJ_0 - \frac{2J_1}{r}\right) \, e^{-i\nu|z-z_0|} \\ \psi(r,z,\theta) &= C_f F_y \cos\theta \sin\theta \ sign(z-z_0) \left(kJ_0 - \frac{2J_1}{r}\right) \, e^{-i\gamma|z-z_0|} \\ \chi(r,z,\theta) &= C_f F_y \ \left(-\frac{k_\beta^2}{i\gamma}\right) \ \left(\cos^2\theta \ (kJ_0 - \frac{2J_1}{r}) + \frac{J_1}{r}\right) \, e^{-i\gamma|z-z_0|} \end{split}$$

0.1.2 Single couples  $M_{xz}$  and  $M_{zx}$ 

 $M_{xz}$ :

 $M_{zx}$ :

$$\begin{split} \phi(r,z,\theta) &= C_f F_x \cos\theta \ sign(z-z_0) \ (-k_r^2) J_1 \ e^{-i\nu|z-z_0|} \\ \psi(r,z,\theta) &= C_f F_x \cos\theta \ (-i\gamma) J_1 \ e^{-i\gamma|z-z_0|} \\ \chi(r,z,\theta) &= C_f F_x \sin\theta \ sign(z-z_0) \ (-k_\beta^2) J_1 \ e^{-i\gamma|z-z_0|} \\ \phi(r,z,\theta) &= C_f F_z \cos\theta \ sign(z-z_0) \ (-k_r^2) J_1 \ e^{-i\nu|z-z_0|} \\ \psi(r,z,\theta) &= C_f F_z \cos\theta \ sign(z-z_0) \ (-k_r^2) J_1 \ e^{-i\nu|z-z_0|} \end{split}$$

 $\chi(r,z,\theta)=0$ 

0.1.3 Single couples  $M_{yz}$  and  $M_{zy}$ 

 $M_{yz}$ :

 $M_{zy}$ :

$$\begin{split} \phi(r,z,\theta) &= C_f F_y \sin\theta \ sign(z-z_0) \ (-k_r^2) J_1 \ e^{-i\nu|z-z_0|} \\ \psi(r,z,\theta) &= C_f F_y \sin\theta \ (-i\gamma) J_1 \ e^{-i\gamma|z-z_0|} \\ \chi(r,z,\theta) &= C_f F_y \cos\theta \ sign(z-z_0) \ (k_\beta^2) J_1 \ e^{-i\gamma|z-z_0|} \\ \phi(r,z,\theta) &= C_f F_z \sin\theta \ sign(z-z_0) \ (-k_r^2) J_1 \ e^{-i\nu|z-z_0|} \\ \psi(r,z,\theta) &= C_f F_z \sin\theta \ sign(z-z_0) \ (-k_r^2) J_1 \ e^{-i\gamma|z-z_0|} \end{split}$$

 $\chi(r, z, \theta) = 0$ 

**0.1.4** Double couple  $M_{XY} = M_{xy} + M_{yx}$   $(F = F_x = F_y)$ 

Using previous definitions  $(K_1, K_2)$  to simplify the expressions we get:

$$\begin{split} \phi(r,z,\theta) &= C_f F \sin 2\theta \ \left(\frac{k_r^2}{i\nu}\right) \ K_1 \ e^{-i\nu|z-z_0|} \\ \psi(r,z,\theta) &= C_f F \sin 2\theta \ sign(z-z_0) \ K_1 \ e^{-i\gamma|z-z_0|} \\ \chi(r,z,\theta) &= C_f F \ \cos 2\theta \ \left(-\frac{k_\beta^2}{i\gamma}\right) \ K_1 \ e^{-i\gamma|z-z_0|} \end{split}$$

0.1.5 Double couple  $M_{XZ}$   $(F = F_x = F_z)$ 

$$\phi(r, z, \theta) = C_f F \ 2 \cos \theta \ sign(z - z_0) \ (-k_r^2) \ J_1 \ e^{-i\nu|z - z_0|}$$

$$\psi(r, z, \theta) = C_f F \cos \theta \ \left(\frac{k_\beta^2 - 2k_r^2}{i\gamma}\right) \ J_1 \ e^{-i\gamma|z - z_0|}$$

$$\chi(r, z, \theta) = C_f F \sin \theta \ sign(z - z_0) \ (-k_\beta^2) \ J_1 \ e^{-i\gamma|z - z_0|}$$

**0.1.6** Double couple  $M_{YZ}$   $(F = F_y = F_z)$ 

$$\phi(r, z, \theta) = C_f F \, 2 \sin \theta \, sign(z - z_0) \, (-k_r^2) \, J_1 \, e^{-i\nu|z - z_0|}$$

$$\psi(r, z, \theta) = C_f F \sin \theta \, \left(\frac{k_\beta^2 - 2k_r^2}{i\gamma}\right) \, J_1 \, e^{-i\gamma|z - z_0|}$$

$$\chi(r, z, \theta) = C_f F \cos \theta \, sign(z - z_0) \, (k_\beta^2) \, J_1 \, e^{-i\gamma|z - z_0|}$$

0.1.7 Dipole  $M_{XX}$ 

$$\phi(r,z,\theta) = C_f F_x \left(\frac{k_r^2}{i\nu}\right) \left(\cos^2\theta \ K_1 + K_2\right) e^{-i\nu|z-z_0|}$$

$$\psi(r,z,\theta) = C_f F_x \operatorname{sign}(z-z_0) \left(\cos^2\theta \ K_1 + K_2\right) e^{-i\gamma|z-z_0|}$$

$$\chi(r,z,\theta) = C_f F_x \cos\theta \sin\theta \left(\frac{k_\beta^2}{i\gamma}\right) K_1 e^{-i\gamma|z-z_0|}$$

0.1.8 Dipole  $M_{YY}$ 

$$\phi(r, z, \theta) = C_f F_y \left(\frac{k_r^2}{i\nu}\right) \left(\sin^2 \theta \ K_1 + K_2\right) e^{-i\nu|z-z_0|}$$

$$\psi(r, z, \theta) = C_f F_y \ sign(z - z_0) \left(\sin^2 \theta \ K_1 + K_2\right) e^{-i\gamma|z-z_0|}$$

$$\chi(r, z, \theta) = C_f F_y \cos \theta \sin \theta \left(-\frac{k_\beta^2}{i\gamma}\right) K_1 e^{-i\gamma|z-z_0|}$$

#### 0.1.9 Dipole $M_{ZZ}$

$$\begin{split} \phi(r,z,\theta) &= C_f F_z \ (-i\nu k_r) \ J_0 \ e^{-i\nu|z-z_0|} \\ \psi(r,z,\theta) &= -C_f F_z \ sign(z-z_0) \ J_0 \ e^{-i\gamma|z-z_0|} \\ \chi(r,z,\theta) &= 0 \end{split}$$

## 0.2 Potential Green's functions for a layered medium, expressions in the source layer

Replacing the elementary sources defined by Equation 2 in the text, in the previous expressions gives us:

#### **0.2.1** Double couple $M_{XY}$

$$\phi(r, z, \theta) = C_f F \sin 2\theta \cos 2 K_1 S_1$$
  
$$\psi(r, z, \theta) = C_f F \sin 2\theta K_1 S_4$$
  
$$\chi(r, z, \theta) = -C_f F \cos 2\theta \cos 3 K_1 S_5$$

#### **0.2.2** Double couple $M_{XZ}$

$$\phi(r, z, \theta) = C_f F \ 2 \cos \theta \ (-k_r^2) J_1 \ S_2$$
  
$$\psi(r, z, \theta) = C_f F \cos \theta \ cs9 \ J_1 \ S_3$$
  
$$\chi(r, z, \theta) = C_f F \sin \theta \ (-k_{\beta}^2) J_1 \ S_6$$

#### **0.2.3** Double couple $M_{YZ}$

$$\phi(r, z, \theta) = C_f F 2 \sin \theta \ (-k_r^2) J_1 S_2$$
  
$$\psi(r, z, \theta) = C_f F \sin \theta \cos \theta J_1 S_3$$
  
$$\chi(r, z, \theta) = C_f F \cos \theta k_\beta^2 J_1 S_6$$

#### 0.2.4 Dipole $M_{XX}$

$$\phi(r, z, \theta) = C_f F_x \cos 2 (\cos^2 \theta \ K_1 + K_2) \ S_1$$
  
$$\psi(r, z, \theta) = C_f F_x (\cos^2 \theta \ K_1 + K_2) \ S_4$$
  
$$\chi(r, z, \theta) = C_f F_x \cos \theta \sin \theta \cos 3 \ K_1 \ S_5$$

#### 0.2.5 Dipole $M_{YY}$

$$\begin{split} \phi(r,z,\theta) &= C_f F_y \ cs2 \ (\sin^2 \theta \ K_1 + K_2) \ S_1 \\ \psi(r,z,\theta) &= C_f F_y \ (\sin^2 \theta \ K_1 + K_2) \ S_4 \\ \chi(r,z,\theta) &= -C_f F_y \cos \theta \sin \theta \ cs3 \ K_1 \ S_5 \end{split}$$

#### 0.2.6 Dipole $M_{ZZ}$

$$\phi(r, z, \theta) = C_f F_z (-i\nu k_r) J_0 S_1$$
  
$$\psi(r, z, \theta) = -C_f F_z J_0 S_4$$
  
$$\chi(r, z, \theta) = 0$$

#### 0.3 Displacement Green's functions in receiver layer

We evaluate here the displacements in the  $i^{th}$  layer at depth z where  $z_i < z < z_i + h_i$  ( $h_i = i^{th}$  layer thickness). When starting to propagate the potentials with the reflectivity algorithm, we set the upgoing/downgoing source potentials in the source(s) layer(s). At the end, we know the upgoing/downgoing potentials in any layer and we deduce the corresponding displacements by derivating:

$$\begin{split} u_r(r,\theta,z) &= \frac{\partial \phi}{\partial r} + \frac{\partial^2 \psi}{\partial r \partial z} + \frac{1}{r} \frac{\partial \chi}{\partial \theta} \\ u_\theta(r,\theta,z) &= \frac{1}{r} \frac{\partial \phi}{\partial \theta} + \frac{1}{r} \frac{\partial^2 \psi}{\partial \theta \partial z} - \frac{\partial \chi}{\partial r} \\ u_z(r,\theta,z) &= \frac{\partial \phi}{\partial z} - \frac{\partial^2 \psi}{\partial r^2} - \frac{1}{r} \frac{\partial \psi}{\partial r} - \frac{1}{r^2} \frac{\partial^2 \psi}{\partial \theta^2} \end{split}$$

The last  $u_z$  expression can be rewritten using the equation of propagation for  $\psi$ :

$$u_z(r,\theta,z) = \frac{\partial \phi}{\partial z} + k_r^2 \psi$$

#### 0.4 Propagating the six elementary sources

Each of the six elementary sources located in a  $s^{th}$  source layer radiate in any layer the following  $\phi$ ,  $\psi$  or  $\chi$  potentials with an upgoing and a downgoing components:

$$S_1 oup S_1^{\phi} \ ^{up}, \ S_1^{\phi} \ ^{down}, \ S_1^{\psi} \ ^{up}, \ S_1^{\psi} \ ^{down}$$
 $S_2 oup S_2^{\phi} \ ^{up}, \ S_2^{\phi} \ ^{down}, \ S_2^{\psi} \ ^{up}, \ S_2^{\psi} \ ^{down}$ 
 $S_3 oup S_3^{\phi} \ ^{up}, \ S_3^{\phi} \ ^{down}, \ S_3^{\psi} \ ^{up}, \ S_3^{\psi} \ ^{down}$ 
 $S_4 oup S_4^{\phi} \ ^{up}, \ S_4^{\phi} \ ^{down}, \ S_4^{\psi} \ ^{up}, \ S_4^{\psi} \ ^{down}$ 
 $S_5 oup S_5^{\chi} \ ^{up}, \ S_5^{\chi} \ ^{down}$ 
 $S_6 oup S_6^{\chi} \ ^{up}, \ S_6^{\chi} \ ^{down}$ 

The  $\rightarrow$  symbol represents all the reflection/transmission computed with the reflectivity algorithm. We have then:

$$S_n^{\phi \ up} = P_n^{\phi \ up} e^{i\nu(z-z_i)}$$

$$S_n^{\phi \ down} = P_n^{\phi \ down} e^{-i\nu(z-z_i)}$$

$$S_n^{\psi \ up} = P_n^{\psi \ up} e^{i\gamma(z-z_i)}$$

where  $P_n^{??}$  are the reflectivity coefficients.

#### 0.5 Propagating a general source

A general seismic source radiating  $\phi$  and/or  $\psi$  and/or  $\chi$  potential(s) in the source layer is a linear combination of the six elementary sources with r-dependant coefficients. Let us take a simple example where the seismic source is expressed as A(r)  $S_1(z_0)$  in the source layer. The radial displacement for instance observed at z depth in any layer is equal to:

$$u_r(r,\theta,z) = \frac{\partial}{\partial r} A(r) \left(S_1^{\phi up}(z) + S_1^{\phi down}(z)\right) + \frac{\partial^2}{\partial r \partial z} A(r) \left(S_1^{\psi up}(z) + S_1^{\psi down}(z)\right)$$

The upgoing and downgoing derivatives only differ by the sign due to the  $\frac{\partial}{\partial z}$  derivative. We have thus:

$$\frac{\partial}{\partial z}A(r)(S_1^{\psi\ up}(z)+S_1^{\psi\ down}(z))=A(r)(-i\gamma S_1^{\psi\ down}+i\gamma S_1^{\psi\ up})$$

Using the following notation:

$$S_1^{\psi}(z) = S_1^{\psi \ up}(z) + S_1^{\psi \ down}(z)$$

and

$$\epsilon i \gamma S_1^\phi(z) = \epsilon i \gamma S_1^{\psi\ up}(z) + \epsilon i \gamma S_1^{\psi\ down}(z)$$

where  $\epsilon=1$  for the upgoing wave and  $\epsilon=-1$  for the downgoing. We obtain a more simple expression:

$$u_r(r, \theta, z) = S_1^{\phi}(z)A'(r) + \epsilon i\gamma S_1^{\psi}A'(r)$$

## 0.6 Expressions of displacement for double couples, dipoles, forces

Using the previous notations, we obtain for the displacements:

	Contribution from potentials	$\phi$	₽.	×
$M_{XY}$	er er			
	$u_r(r, heta,z) = \tilde{z}$	$C_f F \sin 2\theta \cos K_3 \left( S_1^{\phi} + \epsilon i \gamma S_1^{\psi} \right)$	+ $C_f F \sin 2\theta K_3 \left( S_A^{\phi} + \epsilon i \gamma S_A^{\psi} \right)$	$+ C_f F 2 \sin 2\theta \ cs 3K_4 S_x^{\chi}$
	$u_{\theta}(r,\theta,z) =$	$C_f F 2 \cos 2\theta \cos 2K_4 \left(S_1^{\phi} + \epsilon i \gamma S_1^{\psi}\right)$	+ $C_f F 2 \cos 2\theta K_4 \left( S_4^{\phi} + \epsilon i \gamma S_4^{\psi} \right)$	
	$u_z(r, heta,z) =$	$C_f F \sin 2\theta \cos 2 K_1 \left( \epsilon i \nu S_1^{\phi} + k_r^2 S_1^{\psi} \right)$	+ $C_f F \sin 2\theta K_1 \left(\epsilon i \nu S_4^{\phi} + k_r^2 S_4^{\psi}\right)$	
$M_{XZ}$				
	$u_r(r,\theta,z) =$		+ $C_f F \cos \theta \cos \theta K_5 \left(S_3^{\phi} + \epsilon i \gamma S_3^{\psi}\right)$	$-C_f F \cos \theta (k_\beta^2) K_2 S_6^{\chi}$
	$u_{\theta}(r,\theta,z) =$	$-C_f F 2 \sin \theta \ (-k_r^2) K_2 \ (S_2^\phi + \epsilon i \gamma S_2^\psi)$	$- C_f F \sin \theta \cos W_2 \left( S_3^{\phi} + \epsilon i \gamma S_3^{\psi} \right)$	$+ C_f F \sin \theta (k_\beta^2) K_5 S_6^{\chi}$
	$u_z(r, heta,z) =$	$C_f F 2 \cos \theta \ (-k_r^2) J_1 \ (\epsilon i \nu S_2^{\phi} + k_r^2 S_2^{\psi})$	+ $C_f F \cos \theta \cos \theta \int_1 (\epsilon i \nu S_3^{\phi} + k_r^2 S_3^{\psi})$	
$M_{YZ}$				
	$u_r(r, heta,z) =$	$C_f F 2 \sin \theta \ (-k_r^2) K_5 \ (S_2^{\phi} + \epsilon i \gamma S_2^{\psi})$	+ $C_f F \sin \theta \cos \theta K_5 \left( S_3^{\phi} + \epsilon i \gamma S_3^{\psi} \right)$	$- C_f F \sin \theta \ (k_{\beta}^2) K_2 S_{6}^{\chi}$
	$u_{\theta}(r,\theta,z) =$	$-C_f F \ 2\cos\theta \ (-k_r^2) K_2 \ (S_2^{\phi} + \epsilon i \gamma S_2^{\psi})$	$- C_f F \cos\theta \cos\theta K_2 \left( S_3^{\phi} + \epsilon i \gamma S_3^{\psi} \right)$	$-C_f F \cos \theta (k_{\beta}^2) K_5 S_{\kappa}^{\chi}$
	$u_z(r, heta,z) =$		+ $C_f F \sin \theta \cos \theta J_1 \left( \epsilon i \nu S_3^{\phi} + k_r^2 S_3^{\psi} \right)$	
$M_{XX}$				
	$u_{r}(r,\theta,z) =$	$C_f F (K_3 \cos^2 \theta + K_4) cs2 (S_1^{\phi} + \epsilon i \gamma S_1^{\psi})$	+ $C_f F \left( K_3 \cos^2 \theta + K_4 \right) \left( S_4^{\phi} + \epsilon i \gamma S_4^{\psi} \right)$	+ $C_f F \cos 2\theta \cos 3K_4 S_5^{\chi}$
	$u_{\theta}(r,\theta,z) =$	$-C_f F \sin 2\theta \cos K_4 \left(S_1^{\phi} + \epsilon i \gamma S_1^{\psi}\right)$	$- C_f F \sin 2\theta K_4 \left( S_4^{\phi} + \epsilon i \gamma S_4^{\psi} \right)$	$-C_f F \cos \theta \sin \theta \ cs3K_3 S_5^{\chi}$
	$u_z(r, heta,z) =$	$C_f F (K_1 \cos^2 \theta + K_2) \cos (\epsilon i \nu S_1^{\phi} + k_r^2 S_1^{\psi})$	+ $C_f F \left( K_1 \cos^2 \theta + K_2 \right) \left( \epsilon i \nu S_4^{\phi} + k_r^2 S_4^{\psi} \right)$	
$M_{YY}$				
	$u_r(r, \theta, z) =$	$\binom{\psi}{1}$	+ $C_f F \left( K_3 \sin^2 \theta + K_4 \right) \left( S_4^{\phi} + \epsilon i \gamma S_4^{\psi} \right)$	$-C_fF\cos 2\theta \ cs3K_4S_5^{\chi}$
	$u_{\theta}(r,\theta,z) =$		+ $C_f F \sin 2\theta K_4 \left( S_4^{\phi} + \epsilon i \gamma S_4^{\psi} \right)$	$+ C_f F \cos \theta \sin \theta \ cs3 K_3 S_5^{\chi}$
	$u_z(r, heta,z) =$	$C_f F \left( K_1 \sin^2 \theta + K_2 \right) cs2 \left( \epsilon i \nu S_1^{\phi} + k_r^2 S_1^{\psi} \right)$	+ $C_f F (K_1 \sin^2 \theta + K_2) (\epsilon i \nu S_4^{\phi} + k_r^2 S_4^{\psi})$	8
MZZ				
	$u_r(r, heta,z) =$	$C_f F k_r^2 i \nu J_1 \left( S_1^\phi + \epsilon i \gamma S_1^\psi \right) +$	$+ \qquad C_f F \ k_r^2 \ J_1(S_4^\phi + \epsilon i \gamma S_4^\psi)$	
	$u_{ heta}(r, heta,z) =$	0		
	$u_z(r, \theta, z) =$	$-C_f F k_r i \nu J_0(\epsilon i \nu S_1^\phi + k_r^2 S_1^\psi)$	$- C_{f}F k_{r} J_{0}(\epsilon i \nu S_{4}^{\phi} + k_{r}^{2} S_{4}^{\psi})$	

Using program conventions, the previous expressions can be rearrange into:

```
\mathbf{F}_{\mathbf{x}}:
 u_r(r,\theta,z) = C_f F_x \cos\theta \left(K_5 \left[cs2(S_1^{\phi} + \epsilon i\gamma S_1^{\psi}) + (S_4^{\phi} + \epsilon i\gamma S_4^{\psi}) \right]
 cs3 K_2 S_5^{\chi})
 u_{\theta}(r,\theta,z) = -C_f F_x \sin \theta \left(K_2 \left[cs2(S_1^{\phi} + \epsilon i \gamma S_1^{\psi}) + (S_4^{\phi} + \epsilon i \gamma S_4^{\tilde{\psi}}) \right] \right)
 cs3 K_5 S_5^{\chi})
 u_{z}(r,\theta,z) = C_{f}F_{x}\cos\theta \ J_{1}\left[cs2(\epsilon i\nu S_{1}^{\phi} + k_{r}^{2}S_{1}^{\psi}) + (\epsilon i\nu S_{4}^{\phi} + k_{r}^{2}S_{4}^{\psi})\right]
 \mathbf{F}_{\mathbf{v}}:
 u_r(r,\theta,z) = C_f F_y \sin\theta \left(K_5 \left[cs2(S_1^{\phi} + \epsilon i\gamma S_1^{\psi}) + (S_4^{\phi} + \epsilon i\gamma S_4^{\psi}) \right] \right)
 cs3 K_2 S_5^{\chi}
 u_{\theta}(r,\theta,z) = C_f F_y \cos \theta \left(K_2 \left[cs2(S_1^{\phi} + \epsilon i \gamma S_1^{\psi}) + (S_4^{\phi} + \epsilon i \gamma S_4^{\psi}) \right]
 cs3 K_5 S_5^{\chi})
 u_{z}(r,\theta,z) = C_{f}F_{y}\sin\theta J_{1} \left[cs^{2} \left(\epsilon i\nu S_{1}^{\phi} + k_{r}^{2}S_{1}^{\psi} \right) + \left(\epsilon i\nu S_{4}^{\phi} + k_{r}^{2}S_{4}^{\psi} \right) \right]
 \mathbf{F}_{\mathbf{z}}:
 u_r(r,\theta,z) = -C_f F_z J_1 \left[k_r^2 \left(S_2^{\phi} + \epsilon i \gamma S_2^{\psi} \right) + \frac{k_r^2}{i \gamma} \left(S_3^{\phi} + \epsilon i \gamma S_3^{\psi} \right) \right]
 u_{\theta}(r,\theta,z) = 0
 u_z(r,\theta,z) = C_f F_z J_0 \left[k_r \left(\epsilon i \nu S_2^{\phi} + k_r^2 S_2^{\psi} \right) + \frac{k_r}{i \gamma} \left(\epsilon i \nu S_3^{\phi} + k_r^2 S_3^{\psi} \right) \right]
 M_{XY}:
 u_r(r,\theta,z) = C_f F \sin 2\theta \left(K_3 \left[cs2(S_1^{\phi} + \epsilon i\gamma S_1^{\psi}) + (S_4^{\phi} + \epsilon i\gamma S_4^{\psi}) \right] \right)
 2 cs3 K_4 S_5^{\chi}
 u_{\theta}(r,\theta,z) = C_f F \cos 2\theta \left(2K_4 \left[cs2(S_1^{\phi} + \epsilon i \gamma S_1^{\psi}) + (S_4^{\phi} + \epsilon i \gamma S_4^{\psi}) \right] \right)
 cs3 K_3 S_5^{\chi}
 u_z(r,\theta,z) = C_f F \sin 2\theta \ K_1 \left[cs2(\epsilon i\nu S_1^{\phi} + k_r^2 S_1^{\psi}) + (\epsilon i\nu S_4^{\phi} + k_r^2 S_4^{\psi}) \right]
 M_{XZ}:
 u_r(r,\theta,z) = C_f F \cos \theta \ (K_5 \left[-2k_r^2 (S_2^{\phi} + \epsilon i \gamma S_2^{\psi}) + cs9(S_3^{\phi} + \epsilon i \gamma S_3^{\psi}) \right]
 k_{\beta}^2 K_2 S_6^{\chi}
 u_{\theta}(r,\theta,z) = C_f F \sin \theta \left(K_2 \left[2k_r^2 (S_2^{\phi} + \epsilon i\gamma S_2^{\psi}) - cs9(S_3^{\phi} + \epsilon i\gamma S_3^{\psi}) \right] \right)
 k_{\beta}^2 K_5 S_6^{\chi}
 u_{z}(r,\theta,z) = C_{f}F\cos\theta J_{1} \left[-2k_{r}^{2}(\epsilon i\nu S_{2}^{\phi} + k_{r}^{2}S_{2}^{\psi}) + cs9(\epsilon i\nu S_{3}^{\phi} + k_{r}^{2}S_{3}^{\psi}) \right]
 M_{YZ}:
 u_r(r,\theta,z) = C_f F \sin \theta \left(K_5 \left[-2k_r^2 (S_2^{\phi} + \epsilon i \gamma S_2^{\psi}) + cs9(S_3^{\phi} + \epsilon i \gamma S_3^{\psi}) \right] \right)
 k_{\beta}^2 K_2 S_{\beta}^{\chi}
 u_{\theta}(r,\theta,z) = C_f F \cos \theta \left(K_2 \left[-2k_r^2 (S_2^{\phi} + \epsilon i \gamma S_2^{\psi}) + cs9(S_3^{\phi} + \epsilon i \gamma S_3^{\psi}) \right] \right)
 k_{\beta}^2 K_5 S_6^{\chi}
 u_{z}(r,\theta,z) = C_{f}F\sin\theta \ J_{1} \left[-2k_{r}^{2}(\epsilon i\nu S_{2}^{\phi} + k_{r}^{2}S_{2}^{\psi}) + cs9(\epsilon i\nu S_{3}^{\phi} + k_{r}^{2}S_{3}^{\psi}) \right]
 M_{XX}:
 u_r(r,\theta,z) = C_f F \left((K_3 \cos^2 \theta + K_4) \left[cs2(S_1^{\phi} + \epsilon i \gamma S_1^{\psi}) + (S_4^{\phi} + \epsilon i \gamma S_4^{\psi}) \right]
 + \cos 2\theta \cos K_4 S_5^{\chi}
 u_{\theta}(r,\theta,z) = -C_f F \sin 2\theta \left(K_4 \left[cs2(S_1^{\phi} + \epsilon i\gamma S_1^{\psi}) + (S_4^{\phi} + \epsilon i\gamma S_4^{\psi}) \right] \right)
 \frac{cs3}{2}K_3S_5^{\chi}
 u_{z}(r,\theta,z) = C_{f}F\left(K_{1}\cos^{2}\theta + K_{2}\right)\left[cs2(\epsilon i\nu S_{1}^{\phi} + k_{r}^{2}S_{1}^{\psi}) + (\epsilon i\nu S_{4}^{\phi} + k_{r}^{2}S_{4}^{\psi})\right]
 M_{YY}:
 u_r(r,\theta,z) = C_f F \left((K_3 \sin^2 \theta + K_4) \left[cs2(S_1^{\phi} + \epsilon i \gamma S_1^{\psi}) + (S_4^{\phi} + \epsilon i \gamma S_4^{\psi}) \right] \right)
 -\cos 2\theta \cos K_4 S_5^{\chi}
 u_{\theta}(r,\theta,z) = C_f F \sin 2\theta \left(K_4 \left[cs2(S_1^{\phi} + \epsilon i\gamma S_1^{\psi}) + (S_4^{\phi} + \epsilon i\gamma S_4^{\psi}) \right] \right)
 \frac{cs3}{2}K_3S_5^{\chi}
u_{z}(r,\theta,z) = C_{f}F(K_{1}\sin^{2}\theta + K_{2})\left[cs2(\epsilon i\nu S_{1}^{\phi} + k_{r}^{2}S_{1}^{\psi}) + (\epsilon i\nu S_{4}^{\phi} + k_{r}^{2}S_{4}^{\psi})\right]
 M_{ZZ}:
u_r(r,\theta,z) = C_f F J_1 k_r^2 \left[i\nu (S_1^{\phi} + \epsilon i\gamma S_1^{\psi}) + (S_4^{\phi} + \epsilon i\gamma S_4^{\psi}) \right]
u_{\theta}(r,\theta,z) = 0
u_z(r,\theta,z) = -C_f F J_0 k_r \left[i\nu (\epsilon i\nu S_1^{\phi} + k_r^2 S_1^{\psi}) + (\epsilon i\nu S_4^{\phi} + k_r^2 S_4^{\psi}) \right]
```

#### 1 Stress Green's functions

#### 1.1 Definitions

Let e and  $\sigma$  be the linearized strain and stress tensors. Using the Hookes's law we obtain:  $\tau_{ij} = \lambda \theta \delta_{ij} + 2\mu e_{ij}$ 

```
We write:
 \tau_{rr} = \lambda(e_{rr} + e_{\theta\theta} + e_{zz}) + 2\mu e_{rr}
 \tau_{\theta\theta} = \lambda(e_{rr} + e_{\theta\theta} + e_{zz}) + 2\mu e_{\theta\theta}
 \tau_{zz} = \lambda(e_{rr} + e_{\theta\theta} + e_{zz}) + 2\mu e_{zz}
 \tau_{zr} = 2\mu(e_{zr})
 \tau_{z\theta} = 2\mu(e_{z\theta})
 \tau_{r\theta} = 2\mu(e_{r\theta})
 where strain components e_{ij} in cylindrical coordinates (r, \theta, z) are given by:
 e_{\theta\theta} = \frac{u_{\theta,\theta}}{r} + \frac{u_r}{r}
 e_{zz} = u_{z,z}
 e_{r\theta} = \frac{1}{2}(u_{\theta,r} - \frac{u_{\theta}}{r} + \frac{u_{r,\theta}}{r})
 e_{rz} = e_{zr} = \frac{1}{2}(u_{z,r} + u_{r,z})
 e_{\theta z} = e_{z\theta} = \frac{1}{2}(u_{\theta,z} + \frac{1}{x}u_{z,\theta})
 we finally find:
 \tau_{rr} = (\lambda + 2\mu)u_{r,r} + \frac{\lambda}{r}u_{\theta,\theta} + \lambda u_{z,z} + \frac{\lambda}{r}u_{r}
 \tau_{\theta\theta} = \lambda u_{r,r} + \frac{\lambda + 2\mu}{r} u_{\theta,\theta} + \lambda u_{z,z} + \frac{\lambda + 2\mu}{r} u_{r,z}
\tau_{zz} = \lambda u_{r,r} + \frac{\lambda}{r} u_{\theta,\theta} + \frac{\lambda + 2\mu}{r} u_{z,z} + \frac{\lambda}{r} u_r
 \tau_{zr} = \mu(u_{z,r} + u_{r,z})
 \tau_{z\theta} = \mu u_{\theta,z} + \frac{\mu}{r} u_{z,\theta}
 \tau_{r\theta} = \mu(r(\frac{u_{\theta}}{r})_{r} + \frac{1}{r}u_{r,\theta})
 We note
A_j = S_j^{\phi} + \epsilon \iota \gamma S_j^{\psi}
B_i = \epsilon \iota \nu S_i^{\ \phi} + k_r^2 S_i^{\ \psi}
C_i = S_i^{\chi}
 and their derivatives with respect to z
 A_{j}' = \epsilon \iota \nu S_{j}^{\phi} - \gamma^{2} S_{j}^{\psi}
B_i' = -\nu^2 S_i^{\phi} + \epsilon \iota \gamma k_r^2 S_i^{\psi}
C_i' = \epsilon \iota \gamma S_i^{\chi}
```

#### 1.2 Expressions of stresses for double couples, dipoles and forces

Using the previous notation and the expression of the displacements found in paragraph 5.3, we obtain for the stress tensor:

```
M_{XY}
 T_{rr}(r,\theta,z) =
 \sin 2\theta
 [\kappa K_6 + \lambda K_9](A_1 cs2 + A_4) +
 \lambda K_1(B_1'cs2 + B_4') + 2K_7(\kappa - \lambda) cs3C_5
 T_{tt}(r,\theta,z) =
 \sin 2\theta
 [\lambda K_6 + \kappa K_9](A_1 cs2 + A_4) +
 \lambda K_1(B_1^{\dagger}cs2 + B_4^{\dagger}) + 2K_7(\lambda - \kappa) cs3C_5
 T_{zz}(r,\theta,z) =
 \sin 2\theta
 [\lambda K_6 + \lambda K_9](A_1cs2 + A_4) +
 \kappa K_1(B_1'cs2 + B_4') +
 T_{zr}(r,\theta,z) = \mu \sin 2\theta
 K_3(A_1'cs2 + A_4') +
 K_3(B_1cs2 + B_4) +
 2K_4 cs3C_5'
 T_{zt}(r,\theta,z) = \mu \cos 2\theta
 2K_4(A_1'cs2 + A_4') +
 2K_4(B_1cs2 + B_4)
 K_3 cs3C_5'
 T_{rt}(r,\theta,z) = \mu \cos 2\theta
 4K_7(A_1cs2 + A_4)
 (K_{10}) \ cs3C_5
M_{XZ}
 T_{rr}(r,\theta,z) =
 [\kappa K_8 + \lambda K_4](-2k_r^2A_2 + A_3cs9) +
 \lambda J_1(-2k_r^2B_2' + B_3'cs9) -
 2\mu K_4 k_B^2 C_6
 T_{tt}(r,\theta,z) =
 [\lambda K_8 + \kappa K_4](-2k_r^2 A_2 + A_3 cs9)
 \lambda J_1(-2k_r^2B_2'+B_3'cs9) +
 2\mu K_4 k_\beta^2 C_6
 T_{zz}(r,\theta,z) =
 \cos \theta(
 [\lambda K_8 + \lambda K_4](-2k_r^2 A_2 + A_3 cs9)
 \kappa J_1(-2k_r^2B_2'+B_3'cs9)
 T_{zr}(r,\theta,z) =
 \mu \cos \theta
 K_5(-2k_r^2A_2'+A_3'cs9)
 K_5(-2k_r^2B_2 + B_3cs9)
 K_2 k_{\beta}^2 C_6'
 T_{zt}(r,\theta,z) =
 \mu \sin \theta
 -K_2(-2k_r^2A_2'+A_3'cs9)
 K_2(-2k_r^2B_2 + B_3cs9) +
 K_5 k_{\beta}^2 C_6'
 T_{rt}(r,\theta,z) =
 -2K_4(-2k_r^2A_2 + A_3cs9)
 \mu \sin \theta
 K_3 k_{\beta}^2 C_6
M_{YZ}
 T_{rr}(r,\theta,z) =
 [\kappa K_8 + \lambda K_4](-2k_r^2 A_2 + A_3 cs9)
 + \lambda J_1(-2k_r^2B_2' + B_3'cs9)
 2\mu K_4 k_{\beta}^2 C_6
 T_{tt}(r,\theta,z) =
 [\lambda K_8 + \kappa K_4](-2k_r^2 A_2 + A_3 cs9)
 \lambda J_1(-2k_r^2B_2' + B_3'cs9) +
 2\mu K_4 k_\beta^2 C_6
 T_{zz}(r,\theta,z) =
 [\lambda K_8 + \lambda K_4](-2k_r^2 A_2 + A_3 cs9)
 \sin \theta(
 \kappa J_1(-2k_r^2B_2'+B_3'cs9)
 T_{zr}(r,\theta,z) =
 \mu \sin \theta
 K_5(-2k_r^2A_2'+A_3'cs9)
 K_5(-2k_r^2B_2 + B_3cs9)
 K_2 k_{\beta}^2 C_6
 T_{zt}(r,\theta,z) =
 \mu \cos \theta
 K_2(-2k_r^2A_2'+A_3'cs9)
 K_2(-2k_r^2B_2 + B_3cs9)
 K_5 k_{\beta}^{2} C_6'
 T_{rt}(r,\theta,z) =
 \mu \cos \theta
 2K_4(-2k_r^2A_2 + A_3cs9)
 K_3 k_{\beta}^2 C_6
```

```
M_{XX}
[(\lambda K_7 + \kappa K_{12}) - \cos 2\theta (2\kappa K_{12}) + \cos^2 \theta (\lambda K_6 + \kappa K_{11})](A_1 cs2 + A_4) + (\lambda K_2 + \cos^2 \theta \lambda K_1)(B_1' cs2 + B_4') - 2\mu K_7 \cos 2\theta \ cs3C_5
T_{tt}(r, \theta, z) =
 [(\lambda K_7 + \lambda K_{12}) - \cos 2\theta(2\lambda K_{12}) + \cos^2 \theta(\lambda K_6 + \lambda K_{11})](A_1 cs2 + A_4) + (\kappa K_2 + \cos^2 \theta \kappa K_1)(B_1' cs2 + B_4')
T_{zz}(r,\theta,z) =
T_{zr}(r,\theta,z) =
 (\mu K_4 + \cos^2 \theta \mu K_3)[(A_1' cs2 + A_4') + (B_1 cs2 + B_4)] + \cos 2\theta K_4 \mu \ cs3C_5'
 \mu \sin 2\theta (-K_4(A_1'cs2 + A_4') - K_4(B_1cs2 + B_4) - \frac{K_3}{2} cs3C_5')
T_{zt}(r,\theta,z) =
T_{rt}(r,\theta,z) =
 \mu \sin 2\theta (-2K_7(A_1cs2 + A_4) - \frac{K_{10}}{2} cs3C_5)
 M_{YY}
 [(\kappa K_7 + \lambda K_{12}) + \cos 2\theta(2\lambda K_{12}) + \sin^2 \theta(\kappa K_6 + \lambda K_{11})](A_1 cs2 + A_4) + (\lambda K_2 + \sin^2 \theta \lambda K_1)(B_1' cs2 + B_4') - 2\mu K_7 \cos 2\theta cs3C_5
T_{rr}(r,\theta,z) =
 [(\lambda K_7 + \kappa K_{12}) + \cos 2\theta (2\kappa K_{12}) + \sin^2 \theta (\lambda K_6 + \kappa K_{11})](A_1 cs2 + A_4) + (\lambda K_2 + \sin^2 \theta \lambda K_1)(B_1' cs2 + B_4') + 2\mu K_7 \cos 2\theta \ cs3C_5
T_{tt}(r,\theta,z) =
T_{zz}(r,\theta,z) =
 [(\lambda K_7 + \lambda K_{12}) + \cos 2\theta(2\lambda K_{12}) + \sin^2 \theta(\lambda K_6 + \lambda K_{11})](A_1cs2 + A_4) + (\kappa K_2 + \sin^2 \theta \kappa K_1)(B_1'cs2 + B_4')
 (\mu K_4 + \sin^2 \theta \mu K_3)[(A_1' cs2 + A_4') + (B_1 cs2 + B_4)] - \cos 2\theta K_4 \mu \ cs3C_5'
T_{zr}(r,\theta,z) =
T_{zt}(r,\theta,z) =
 \mu \sin 2\theta (K_4(A_1'cs2 + A_4') + K_4(B_1cs2 + B_4) + \frac{K_3}{2} cs3C_5')
 \mu \sin 2\theta (2K_7(A_1cs2 + A_4) + \frac{K_{10}}{2} cs3C_5)
T_{rt}(r,\theta,z) =
```

```
2\mu K_4 \ cs3C_5
2\mu K_4 \ cs3C_5
 2\mu K_4 cs3C<sub>5</sub>
 K_2
K_5
K_3
 K_2
K_5
K_5
 + 1 1
 + 1 + + + +
 \lambda J_1(B_1'cs2 + B_4') \ \lambda J_1(B_1'cs2 + B_4') \ \kappa J_1(B_1'cs2 + B_4') \ K_5(B_1cs2 + B_4) \ K_5(B_1cs2 + B_4) \ K_2(B_1cs2 + B_4)
 \lambda J_1(B_1'cs2 + B_4') \\ \lambda J_1(B_1'cs2 + B_4') \\ \kappa J_1(B_1'cs2 + B_4') \\ K_5(B_1cs2 + B_4) \\ \frac{J_1}{r}(B_1cs2 + B_4)
 \lambda J_0 k_r (B_2' + rac{B_3'}{i\gamma}) \ \lambda J_0 k_r (B_2' + rac{B_3'}{i\gamma}) \ \kappa J_0 k_r (B_2' + rac{B_3'}{i\gamma}) \ J_1 \mu k_r^2 (B_2 + rac{B_3}{i\gamma})
 + + + 1
 ++++1
 +++++
 [\kappa K_8 + \lambda K_4](A_1cs2 + A_4)
[\lambda K_8 + \kappa K_4](A_1cs2 + A_4)
[\lambda K_8 + \lambda K_4](A_1cs2 + A_4)
K_5(A_1'cs2 + A_4')
-K_2(A_1'cs2 + A_4')
-2K_4(A_1cs2 + A_4')
 [-\kappa K_5 - \lambda K_2)]k_\tau^2 (A_2 + \frac{A_3}{\iota\gamma})
 [-\lambda K_5 - \kappa K_2]k_\tau^2 (A_2 + \frac{A_3}{\iota\gamma})
 [-\lambda K_5 - \lambda K_2]k_\tau^2 (A_2 + \frac{A_3}{\iota\gamma})
 -J_1\mu k_\tau^2 (A_2' + \frac{A_3'}{\iota\gamma})
 [\kappa K_8 + \lambda K_4](A_1cs2 + A_4)
[\lambda K_8 + \kappa K_4](A_1cs2 + A_4)
[\lambda K_8 + \lambda K_4](A_1cs2 + A_4)
K_5(A_1'cs2 + A_4')
K_2(A_1'cs2 + A_4')
K_2(A_1'cs2 + A_4')
2K_4(A_1'cs2 + A_4')
 [-\kappa K_5
T_{rr}(r, \theta, z) = T_{tt}(r, \theta, z) = T_{tt}(r, \theta, z) = T_{zz}(r, \theta, z) = T_{zr}(r, \theta, z) = T_{zt}(r, \theta, z) = T_{rt}(r,
 \sin \theta(
\sin \theta(
\sin \theta(
\mu \sin \theta(
\mu \cos \theta(
\mu \cos \theta(
 \cos \theta (
\cos \theta (
\cos \theta (
\mu \cos \theta (
\mu \sin \theta (
\mu \sin \theta (
 T_{rt}(r, \theta, z) = T_{tt}(r, \theta, z) = T_{zz}(r, \theta, z) = T_{zr}(r, \theta, z) = T_{zt}(r, \theta, z) = T_{rt}(r,
 T_{tt}(r, \theta, z) = T_{tt}(r, \theta, z) = T_{zz}(r, \theta, z) = T_{zr}(r, \theta, z) = T_{zt}(r, \theta, z) = T_{zt}(r, \theta, z) = T_{rt}(r,
 T_{rr}(r, \theta, z) =
T_{tt}(r, \theta, z) =
T_{zz}(r, \theta, z) =
T_{zr}(r, \theta, z) =
T_{zt}(r, \theta, z) =
T_{zt}(r, \theta, z) =
T_{rt}(r, \theta, z) =
 \begin{aligned} [\kappa K_5 + \lambda K_2] k_r^2 (A_4 + i\nu A_1) \\ [\lambda K_5 + \kappa K_2] k_r^2 (A_4 + i\nu A_1) \\ [\lambda K_5 + \lambda K_2] k_r^2 (A_4 + i\nu A_1) \\ [\lambda K_5 + \lambda K_2] k_r^2 (A_4' + i\nu A_1') \end{aligned}
 F_Z
 F_Y
```

+ 1 1 1

 $\lambda J_0 k_r (B'_4 + i\nu B'_1)$   $\lambda J_0 k_r (B'_4 + i\nu B'_1)$   $\kappa J_0 k_r (B'_4 + i\nu B'_1)$   $J_1 \mu k_r^2 (B_4 + i\nu B_1)$ 

 $\sim$ 

 $cs3C_{\rm s}'$   $cs3C_{\rm s}'$  $cs3C_5$   $\sim$ 

 $cs3C_5$ 

 $cs3C_5'$  $cs3C_5'$  $cs3C_5$ 

## Chapitre III-3

Variations de contraintes dynamiques associées au séisme de Landers

## 3. Variations de contrainte dynamique associées au séisme de Landers

#### A. Introduction

Pour mieux estimer les variations de contraintes engendrées par un séisme, nous avons appliqué la méthode décrite dans les paragraphes précédents au séisme particulier de Landers où nous avons la chance de bien connaître la source sismique et le milieu. La faille est modélisée de la même façon qu'aux chapitres II-1 et III-1 par une source étendue comprenant 3 segments non plans d'orientation différente. Cette source est représentée par 3768 point- sources. Le milieu de propagation est un modèle stratifié plan légèrement modifié par rapport au modèle de vitesse proposé par Hadley et Kanamori (1975) pour la Californie du Sud. Ce modèle a déja été décrit aux chapitres II-1 et III-1. La cinématique de la rupture (vitesse du front de rupture et variation spaciotemporelle du glissement) est celle du modèle A trouvée au chapitre II-1 (Cotton et Campillo 1995a) par inversion des formes d'onde des enregistrements en champ proche. La modélisation est effectué de 0 à 0.3 Hz dans le domaine de fréquence où les déplacements du sol synthétiques prédits par notre modèle de rupture donnent un bon accord avec les données.

Cette modélisation permet de montrer les variations de contraintes engendrées par un séisme sur un exemple réaliste. Les variations de contraintes seront tout d'abord modélisées sur les plans de faille de deux séismes (Big Bear et de Little Skull Mountain) ayant apparamment été déclenchés par l'évènement de Landers (Figure 1). Les résultats de nos calculs seront comparés aux résultats de calculs statiques pour le séisme de Big Bear (King et al., 1994 et Harris et al. 1995) et aux calculs dynamiques de Gomberg et Bodin (1994) pour le séisme de Little Skull Mountain (1995).

#### B. Modélisations des variations de contrainte a Big Bear.

Le séisme de Big-Bear a eu lieu 3 heures après le séisme de Landers. D'une magnitude égale à 6.2 il est présenté comme la réplique la plus importante du séisme de Landers (Hauksson et al., 1993). Du fait d'un séisme précurseur le mécanisme et la profondeur de ce séisme sont mal contraints. Pour effectuer nos modélisations, le mécanisme proposé par Hauksson et al. (1993) a été retenu (strike=55°, dip=85° et rake=-10°) et nous avons choisi une profondeur de 6500 mètres.

Les résultats de la modélisation (variation de contrainte normale, variation de la contrainte tangentielle et la variation du critère de Coulomb) sont présentés Figure 2. La perturbation dure environ 40 secondes. La variation du critère de rupture de Coulomb atteint un pic de 4 bars et la valeur statique de cette variation est égale à 1.3 bars. Ces calculs confirment qualitativement les résultats statiques de Harris et al. (1992) et King et al. (1994): le séisme de Landers a créé une

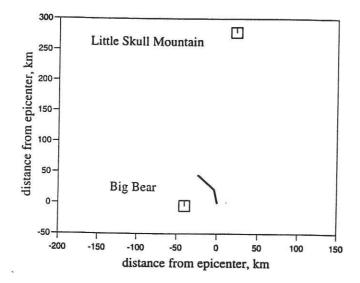


Figure 1. Localisation des séismes de Little Skull Mountain et Big Bear par rapport à la faille de Landers. Le séisme de Big Bear a eu lieu 3 heures après le séisme de Landers. Le séisme de Little Skull Mountain a eu lieu 20 heures après.

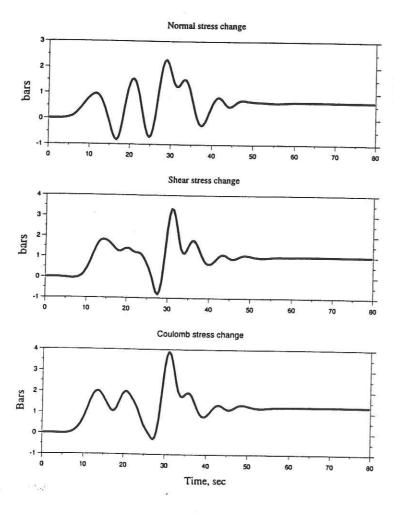


Figure 2. Variations de contrainte normale (haut) et tangentielle (milieu) sur le plan de faille du séisme de Big Bear. Le mécanisme retenu (direction 55°, pendage 85° et plongement -10°) est celui déterminé par Hauksson (1993). Au bas de la figure est représentée la variation temporelle du critère de rupture de Coulomb calculée avec un coefficient de friction égal à 0.4.

augmentation du critère de rupture de Coulomb sur la faille de de Big Bear. La revue des valeurs des variations statiques du coefficient de Coulomb trouvée par les différentes équipes donne un ordre de grandeur de l'erreur de ce type de calcul. Harris et al. (1992) et King et al. (1994) utilisent les mêmes formules analytiques développées par Okada (1992), mais font des choix différents en ce qui concerne le coefficient de friction, les propriétés du milieu élastique et la géométrie du plan du séisme de Big Bear. En fonction de ces choix les valeurs trouvées vont de 0.7 (Harris et al., 1992) à 2.9 (valeur maximale proposée par King et al., 1994). La valeur de la variation statique de 1.3 bar trouvée par nos calculs est donc intermédiaire entre ces deux résultats.

## C. Modélisation des variations de contrainte à Little Skull Mountain.

Le séisme de Little Skull Mountain (LSM) a eu lieu 22 heures après le séisme de Landers à plus de 200 kilomètres au Nord de celui ci. Il est considéré, avec une magnitude Ms=5.4, comme le séisme lointain déclenché le plus important (Bodin et Gomberg, 1995). La Figure 3 présente les variations de contraintes normales et tangentielles calculées sur le plan de faille le plus probable (d'après un mécanisme au foyer et une analyse de Harmsen, 1994). Sur la Figure 4 sont présentées les variations de contraintes normales et tangentielles calculées sur un plan représentatif de l'orientation moyenne des failles dans cette région. Au bas de chacune de ces figures sont présentées les variations temporelles du critère de rupture de Coulomb calculées avec un coefficient de friction de 0.4. On peut retenir de ces calculs que la variation maximum du critère de Coulomb sur le plan de faille du séisme de LSK a été entre 0.5 et 1.0 bar selon la géométrie retenue. La perturbation majeure est due à l'arrivée des ondes de surface. Sa durée est de 30 secondes et la période du signal est comprise entre 10 et 15 secondes. A cette distance, le niveau de contrainte statique est nul. Les calculs de Gomberg et Bodin (1994) donnent les mêmes ordre de grandeur. Par contre le signal obtenu par ces auteurs est beaucoup plus haute fréquence car leur modélisation s'effectue dans un milieu où l'atténuation n'est pas prise en compte.

## D. Variations relatives des contraintes dynamiques et statiques en fonction de la distance à la faille.

Les études de sismicité déclenchée distinguent classiquement la sismicité déclenchée lointaine (apparamment déclenchée par les variations de contraintes dynamiques) et la sismicité proche (apparamment déclenchée par les variations de contraintes statiques). Cette division entre répliques lointaines et répliques proches est artificielle. Pour mieux apprécier les variations relatives des contraintes dynamiques et statiques en fonction des distances à la faille, nous avons ainsi modélisé les variations de contrainte à différentes distances de Landers (Figure 5). Les variations du critère de rupture de Coulomb ont été effectuées sur des décrochements sénestres de direction nord-sud. Les conclusions qui suivent restent vraies si on change le mécanisme de

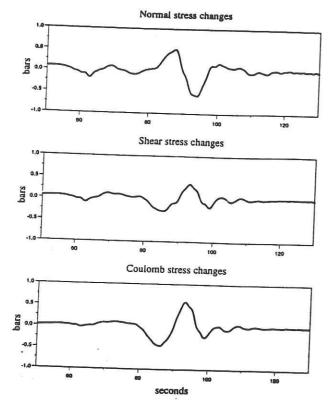


Figure 3. Variation de contrainte normale (haut) et tangentielle (milieu) sur le plan de faille du séisme de Little Skull Mountain. Le mécanisme retenu (direction 55°, pendage 56° et plongement 0°) est celui déterminé par Harmsen (1994). Au bas de la figure est représentée la variation temporelle du critère de rupture de Coulomb calculée avec un coefficient de friction égal à 0.4.

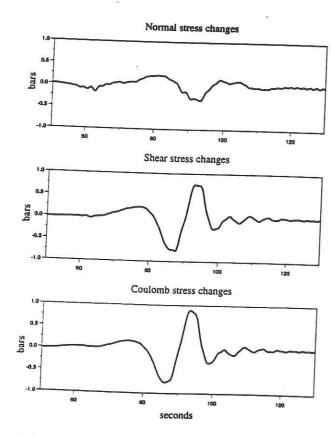


Figure 4. Variation de contrainte normale (haut) et tangentielle (milieu) sur le plan de faille du séisme de Little Skull Mountain. Le mécanisme retenu (direction 55°, pendage 56° et plongement 0°) est donné par l'orientation des failles du critère de rupture de Coulomb calculée avec un coefficient de friction égal à 0.4.

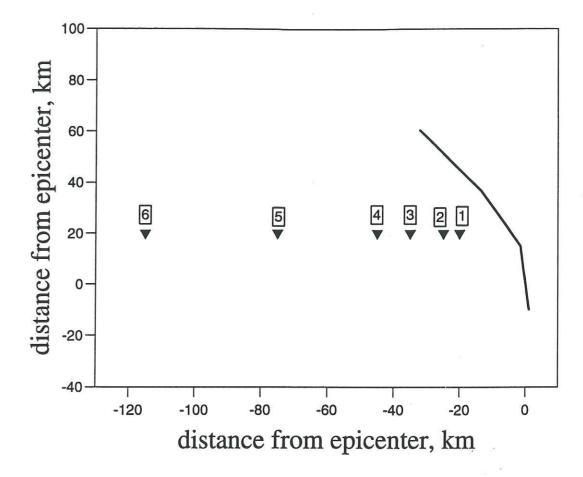


Figure 5. Les triangles représentent les localisations où les variations de contraintes des Figures 6 et 7 sont calculées. Le trait plein représente la faille du séisme de Landers (1992). L'histoire cinématique du glissement est celle que nous avons trouvé par inversion des mouvements forts au chapitre II-1.

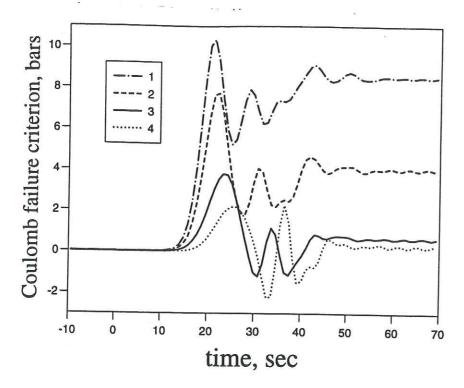


Figure 6. Variations temporelles du critère de rupture de Coulomb en fonction de la distance à la faille. Les calculs ont été effectués pour des décrochements senestres de direction Nord-Sud. Le coefficient de friction utilisé est égal à 0.4.

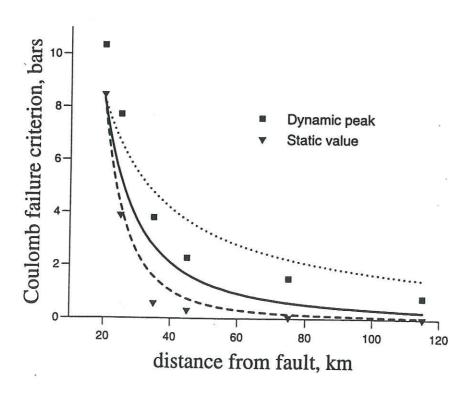


Figure 7. Pics des variations de contrainte dynamiques (carrés) et statiques (triangles) calculés à différentes distances r de la faille (voir Figure 6). Ces variations sont comparées à des lois de décroissance en r⁻¹ (pointillés), r⁻² (trait plein), r⁻³ (tirets).

ces failles. La Figure 6 montre les variations temporelles des variations de critère de Coulomb pour les différentes localisations choisies. On remarque que quelle que soit la distance à la faille, les variations de contraintes dynamiques sont très supérieures aux variations de contraintes statiques. Nous comparons ensuite (Figure 7) les pics atteints par les variations de contraintes dynamiques aux valeurs statiques. Ces variations sont comparées à des lois de décroissance en r⁻¹ r⁻² et r⁻³. On voit que les variations de contraintes statiques diminuent en r⁻³, décroissance théorique prédite par les formules analytiques de Okada (1992) à partir de 60 kilomètres de la faille les variations statiques deviennent négligeables. Les variations des pics de contrainte dynamique varient en r⁻² dans les premiers 50 kimomètres à r⁻¹ plus loin de la faille. Le rapport pic de la variation dynamique, valeur de la variation statique augmente rapidement et est au moins supérieur à 2 dans la région où les variations de contraintes statiques ne sont pas négligeables.

#### E. Interprétation des résultats et discussion

Les calculs précédents montrent que quelle que soit la distance à la faille la variation du critère de Coulomb dynamique est toujours supérieure à la variation de contrainte statique. Pourquoi les variations dynamiques ne déclenchent-elles pas la rupture alors que la variation de contrainte statique (apparamment) réussit à déclencher le séisme ? Comment expliquer le délai entre ce pic du critère de rupture de Coulomb du au passage des ondes et le moment où la rupture intervient (3 heures par exemple dans le cas du séisme de Big Bear)?

Pour interpréter cette observation plusieurs hypothèses peuvent être considérées. La première a été discutée par Spudich et al. (1995). Dans cette hypothèse, la rupture a lieu lorsque la résistance de la roche est dépassée. Avec cette approche classique, le déclenchement intervient lorsque le critère de Coulomb atteint un certain seuil. Ce seuil de rupture est situé au delà du pic atteint lors du passage des ondes. Pour expliquer le fait que ce seuil soit atteint 3 heures plus tard, on peut supposer que la pression fluide a augmenté peu à peu, diminuant la contrainte normale et ainsi augmentant le critère de rupture de Coulomb. Celui ci dépasse la valeur qui avait été atteinte lors du passage des ondes, franchit le seuil de résistance de la faille: il y a rupture. Nos calculs montrent que dans le cas de Big Bear la différence entre variations du critère de Coulomb statiques et dynamiques est de 3 bars, avec  $\mu$ =0.4, il faut que la pression fluide ait augmenté d'au moins 7.5 bars pour provoquer la rupture. Cette valeur est équivalente à une remontée de la nappe phréatique de 75 mètres, variation qui n'a pas été observée.

Un second type d'approche consiste à considérer un séisme comme une instabilité (Sholz, 1990). Cette instabilité n'a pas lieu lorsque la contrainte atteint son maximum, mais au delà du maximum lorsque la résistance de la faille (la friction sur le plan de faille) diminue plus vite que la charge qu'elle subit (Figure 8). Pour que l'affaiblissement de la résistance ait lieu, il faut donc que dans la zone de nucléation du séisme un certain glissement ait eu lieu (affaiblissement de la friction du au déplacement). Ce glissement s'effectue de manière quasi-statique et requiert un certain temps pour avoir lieu. Il n'est donc pas sensible aux variations rapides du champ de

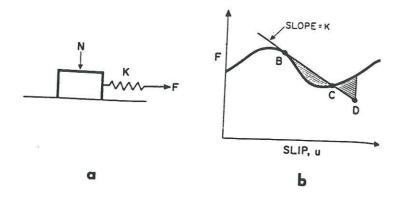
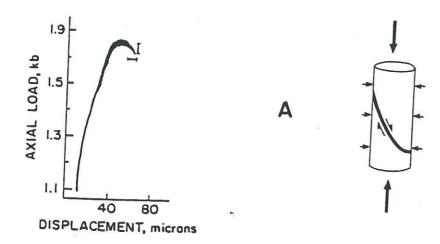


Figure 8. Schéma illustrant l'origine de l'instabilité de friction. (a) modèle de bloc-ressort. (b) diagramme force-déplacement montrant un cas hypothétique dans lequel la force de friction qui "retient" le bloc (courbe noire épaisse) diminue lorsqu'il y a déplacement du bloc plus vite que la force de rappel (trait fin) du ressort: il y a instabilité. d'après Sholz, 1990)



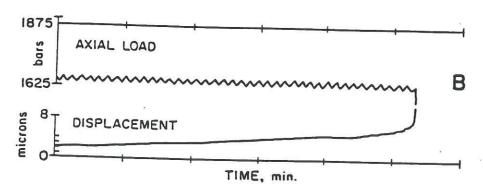


Figure 9. Résultats d'expérience de laboratoire par Johnson (1981). Le glissement entre les 2 lèvres d'un cylindre de granite coupé en biseau est mesuré dans un essai de compression triaxiale. Dans cet expérience, à un raccourcissement graduel se surimpose une variation rapide en dent de scie. La figure du haut montre la relation entre la contrainte axiale imposée par le piston et le déplacement du au raccourcissement. La figure du bas montre les variations de la contrainte axiale et dû glissement les 5 dernières minutes avant une instabilité du glissement.

contrainte et ce type de système peut toujours supporter des charges plus hautes de courte durée alors que des charges moins hautes mais plus longues vont créer l'instabilité. Ce type de réponse a été observé en laboratoire (Johnson, 1981). Un cylindre de granite est coupé en biseau et est soumis à une augmentation de contrainte axiale dans un essai triaxial (Figure 9). Ce chargement s'effectue de façon progressive par raccourcissement du cylindre. A ce raccourcissement graduel à vitesse constance se surimpose une variation rapide "en dent de scie". On mesure à la fois le raccourcissement du cylindre et la charge axiale qui lui est imposée. Le raccourcissement progressif conduit à un pic de la charge axiale de 1670 bars. du fait du glissement entre les 2 lèvres, cette charge diminue ensuite. Le glissement entre les 2 lèvres se fait tout d'abord de façon continue avant qu'une instabilité se produise. La Figure 9b permet d'observer des variations de la charge et du déplacement 5 minutes avant l'instabilité. Pendant ces 5 dernières minutes la contrainte oscille régulièrement autour d'une valeur moyenne. Le glissement accélère peu à peu et l'instabilité se produit alors que le chargement n'est pas à sa valeur maximale. Cette expérience montre que le déclenchement de l'instabilité n'est pas influencé par les variations rapides ("dynamique") du chargement Par contre la valeur moyenne élevée ("statique") de ce chargement produit un glissement "asismique" entre les deux lèvres de la discontinuité. Ce glissement diminue la résistance de la faille qui ne peut plus répondre de façon asismique: il y a instabilité ("séisme")

Avec ce modèle, la corrélation entre variations de contrainte statique et distribution spaciale des répliques ainsi que le délai entre application de la contrainte et le déclenchement s'expliquent très bien. Par contre, ce mécanisme n'explique pas le déclenchement à de grandes distances là où la variation de contrainte statique peut être tenue comme négligeable. De nombreux auteurs considèrent ainsi que les ondes de surface produites par le séisme ont déclenché à grande distance des phénomènes annexes. Une première classe de mécanisme fait intervenir des magmas ou des fluides sous pression (montée de bulle (Linde et al.,1994), liquéfaction d'un magma partiellement cristallisé par les variations basse fréquence des contraintes ce qui modifie l'état de contrainte ambiant et déclenche les séismes (Hill et al., 1994)). Un type de mécanisme ne faisant pas intervenir de magmas ou des variations de contraintes dynamique est aussi proposé par Bodin et Gomberg, 1994. Dans leurs modèles numériques, ces auteurs supposent la connexion des failles à travers la "Eastern California Shear Zone" située au Nord du séisme de Landers. Les variations de contraintes statiques calculées dans un tel milieu où de nombreuses failles sont connectées en réseau et peuvent glisser librement en réponse aux variations de contraintes sont d'un ordre de grandeur supérieur aux variations statiques dans un demi-espace élastique telles qu'elles ont été calculées par exemple par King et al. (1994). Pour conclure, on peut retenir que pour choisir entre ces différentes interprétations nous manquons d'observations fiables.

Dans les paragraphes précédents nous avons calculé les variations de contrainte statiques et dynamiques dans un milieu élastique. Cette approche est intéressante pour expliquer les relations entre un séisme déclencheur et un séisme déclenché mais il convient de rappeler les limites de ce type de calcul. Tout d'abord, l' estimation des variations du critère de rupture de Coulomb est difficile car pour effectuer ces calculs le nombre de choix a priori est important et ces choix sont mal contraints. Le premier choix concerne la valeur du coefficient de friction. On peut estimer

que ce coefficient est compris entre 0.7 (valeurs trouvées dans les expériences de friction en laboratoire (Sholz, 1990)) et 0.2 (valeur proposée par certains sismologues (Heaton, 1990). Or, les résultats et notamment les cartes de variations statiques du critère de rupture de Coulomb, sont très dépendants du choix de ce coefficient (Figure 4 de King et al., 1994). Les résultats varient aussi en fonction de la distribution de glissement choisie sur le plan de faille du séisme déclencheur et il est parfois difficile d'estimer le glissement en profondeur lorsqu'aucune inversion en champ proche n'a été effectuée. La géométrie du plan de faille choisie pour le séisme déclenché et les vitesses du milieu élastique interviennent aussi dans les résultats. Le second problème concerne les limites du modèle élastique. Ainsi, ces calculs ne prennent pas en compte les glissements asismiques qui peuvent avoir lieu après le séisme. De tels glissements (les calculs de Bodin et Gomberg (1994) le montrent bien) redistribuent les contraintes de façon importante. Plus près de la faille, la déformation n'est pas localisée sur un plan. Les études géologiques montrent que la bande de cisaillement est beaucoup plus large (200 mètres lors du séisme de Landers, Johnson et al., 1994), la modélisation élastique est clairement insuffisante et on connait mal la façon dont les contraintes sont redistribuées près de la faille (microséismes pendant ou après le séisme principal, déformations asismiques). De plus, l'aspect temporel de la réponse du milieu aux variations de contrainte statique n'a pas encore été bien exploré. Les premières études n'ont pas tenu compte de ce type de problème: Stein et al. (1994) montrent ainsi que le séisme de Long Beach (1933) et de Kern County (1952) ont augmenté le critère de Coulomb de 0.1 bars au site du séisme de San Fernando (1971). Cette variation de contrainte de 0.1 bar créée par ces deux séismes a-t'elle vraiment influencé le déclenchement d'un séisme plusieurs dizaines d'années plus tard ? Dans une étude récente, Harris et al. (1995) essayent ainsi d'estimer le temps pendant lequel l'effet des variations de contraintes statiques engendrées par un séisme persistent et influencent la sismicité. Ils étudient ainsi la sismicité des séismes californiens de magnitude supérieure à 5 entre 1968 et 1995. Ces calculs montrent que pendant une année et demie après un événement, les séismes sont situés préférentiellement à des endroits où les variations du critère de Coulomb engendrées par le premier séisme sont positives. Après ce délai, la corrélation disparait et les séismes suivants sont statistiquement aussi nombreux dans des zones où les variations sont négatives que positives.

# Quatrième partie Synthèse des résultats et conclusions

#### IV. Synthèse des résultats obtenus et conclusions

Au travers de cette étude nous avons examiné trois séismes particuliers: le séisme de Landers, le séisme d'Uttarkashi et le séisme de Northridge. Nous ne rappellerons pas ici les résultats concernant la sismotectonique et la description de la rupture relative à chaque séisme. De la même façon nous ne répéterons pas dans ce chapitre les conclusions concernant les méthodes utilisées. Ce chapitre regroupe plutôt une synthèse des résultats et des réflexions pouvant aider à comprendre les questions concernant la physique de la rupture que nous avions posées au chapitre I.

#### 1. Le problème du temps de montée.

Le problème de la mesure du temps de montée et ses implications théoriques ont été plus spécifiquement abordés au chapitre II-2. Plusieurs points peuvent être retenus.

Tout d'abord, les premières estimations des temps de montée à l'aide des enregistrements de champ proche (Heaton, 1990) ont été effectuée grâce aux méthodes d'inversion par multifenêtrage. Ces inversions ont utilisé peu de fenêtres (en général 3) et cette paramétrisation ne permettait pas de trouver des temps de montée importants. Plus récemment, le nombre de triangles et donc la durée possible de la rupture dans les modèles a été augmentée et des temps de montée plus grands ont été obtenus (Mendoza et al., 1994; Hartzell and Langer, 1993).

Il existe un couplage entre les temps de montée et le moment sismique des modèles trouvés par inversion. Un temps de montée du modèle trop faible par rapport à la réalité peut permettre de trouver des sismogrammes synthétiques qui donnent un bon accord avec les données. En revanche, le moment de ce modèle sera inférieur au moment sismique trouvé par d'autres techniques (géodésie, études télésismiques longues périodes).

L'étude de Landers a montré qu'aux endroits où le glissement a été significatif (glissement supérieur à 3 mètres) le temps de montée est compris entre 3 et 4 secondes. Cette durée est faible par rapport à la durée totale de la rupture qui est d'environ 20 secondes. L'imagerie de la rupture a par ailleurs montré que la distribution du glissement est hétérogène et est constitué de trois sous-évènements qui ont une largeur de 15 kilomètres. Cette distance est proche de la distance sur laquelle la rupture se développe à un moment donné (de 9 à 12 km si la vitesse de rupture est égale à 3 kilomètres par seconde).

Une compilation des résultats publiés depuis 1990 montre que de nombreuses études ont trouvé des séismes composés de plusieurs sous-évènements dont le temps de montée est de l'ordre de la durée requise pour casser chaque sous-évènement. Ces observations sont compatibles avec des mécanismes de type fracture où la radiation d'un point de la faille se termine à l'arrivée d'une phase d'arrêt ("healing phase") à partir des bords de chaque sous-évènement.

Pour conclure, deux hypothèses permettent d'expliquer des temps de montée courts par rapport à la durée totale de la rupture. Le premier est décrit par Heaton (1990): l'arrêt du glissement en un point s'opère automatiquement après le passage du front de rupture. Cette autocicatrisation (« self-healing ») est due à une loi de friction où la friction diminue avec la vitesse de glissement (« velocity weakening »). L'autre hypothèse est plus classique: le plan de faille est hétérogène et la rupture est composée de plusieurs fractures. Les observations ne permettent pas de trancher définitivement entre ces deux hypothèses.

#### 2. Contrôle de la rupture

#### A. Synthèse des résultats obtenus à Landers

Les 3 séismes étudiés dans les chapitres précédents ont tous été modélisés par des modèles complexes où le glissement et la vitesse de rupture sont hétérogènes sur le plan de faille. La géométrie des failles des séismes d'Uttarkashi et de Northridge n'est pas connue de façon précise. Pour cette raison les résultats obtenus pour ces deux séismes ne permettent pas une discussion approfondie des relations entre les propriétés de la rupture (distribution spatiale du glissement par exemple) et des paramètres qui pourraient contrôler l'hétérogénéité de cette rupture (sauts de faille ou variations lithologiques). Dans le cas du séisme de Landers cette discussion est possible car la rupture a atteint la surface et la géométrie est bien contrainte par les relevés géologiques de terrain et la distribution spatiale des répliques.

La Figure 1 montre une carte récente du tracé des failles (Sieh, 1995) ayant joué pendant le séisme de Landers. Les ruptures principales sont disposées en échelon. Les terminaisons de plus de la moitié des ruptures du séisme de 1992 correspondent à des extrémités de faille (cercles pleins sur la Figure 1). La terminaison de la rupture sur la partie sud de la faille de la Johnson Vallée apparaît par exemple comme l'extrémité de cette faille. Ceci est aussi le cas pour les extrémités nords de la faille de la Homestead Vallée et de la faille Emerson. En 4 endroits, (cercles ouverts sur la Figure 1) la rupture s'est au contraire arrêtée au milieu d'une faille plutôt qu'à son extrémité.

Comparons maintenant les variations de la géométrie et les modèles de rupture que nous avons obtenus. La rupture du séisme de Landers est caractérisée par deux sauts de faille (flèches sur la Figure 2). Un premier saut a lieu entre la faille de la Johnson Vallée et la faille Homestead. Plus loin la rupture saute de la faille Homestead à la faille Emerson. La Figure 2 montre que les sauts de faille correspondent à des endroits où d'après notre modèle le glissement est plus faible. Cette corrélation suggère ainsi un contrôle des propriétés de la rupture par la géométrie. Les inversions proposées par Cohee and Beroza (1994) et Wald et Heaton (1994) donnent des résultats qui varient dans le détail (Figure 3) mais qui globalement confirment la distribution de la rupture en 3 sous-évènements dont les limites correspondent aux jonctions entre les différentes failles. Par contre, la rupture se termine au milieu de la faille Camp Rock à un endroit où il n'y a pas (du moins en surface) de discontinuité géométrique du plan de faille.

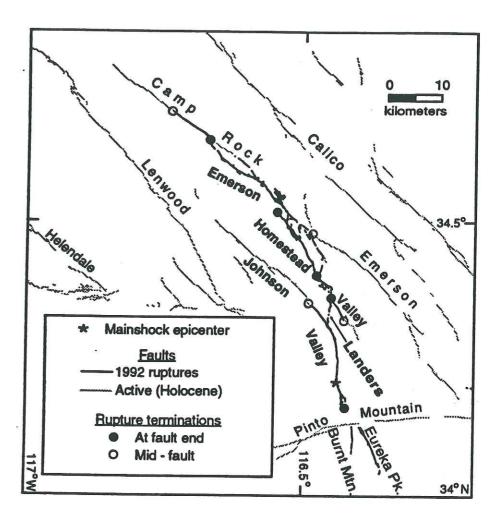


Figure 1. Géométrie des failles ayant joué pendant le séisme de Landers (1992). D'après Sieh (1995).

Depuis l'étude du séisme de Landers décrite au chapitre II-1 nous avons essayé de compléter les informations relatives à ce séisme. Nous avons ainsi comparé précisément les images de la rupture aux résultats de la tomographie 3D obtenue par J. Lee et C. Nicholson (1993, 1995) par inversion des temps d'arrivée des ondes P et des données gravimétriques. Les résultats de la tomographie (variation de la vitesse des ondes P par rapport à un modèle de référence) et la distribution du glissement sur le plan de faille sont présentés sur les mêmes plan de coupe et à la même échelle. La comparaison de ces deux images ne montrent pas de corrélation claire entre les variations de la vitesse dans le milieu et les variations du glissement sur la faille. On peut, au risque de "surinterpréter", remarquer que la rupture s'arrête au milieu de la faille Camp Rock là où la vitesse dans le milieu change et que ces deux limites (arrêt du glissement et variation de la vitesse) présentent la même forme. A cet endroit la faille entre par ailleurs dans un craton granitique (Nicholson, communication personnelle).

Récemment de nombreuses études de paléosismicité ont aussi permis d'accroître notre connaissance du séisme de Landers. Le dernier séisme à avoir eu lieu sur les failles de la , Homestead Vallée et la faille Emerson a eu lieu il y a 6000-9000 ans. Les études paléosismiques ont mis en évidence des ruptures plus récentes sur la faille Camp Rock et la partie sud de la faille Emerson. Ces observations suggèrent que ces failles ont joué il y a peu longtemps et qu'elles ont pu constituer des barrières de relaxation Sholz (1990). Cette hypothèse ne se vérifie pas pour la partie nord de la faille de la Johnson Vallée sur laquelle la rupture ne s'est pas propagée en 1992 et sur laquelle on n'a pas mis en évidence de rupture récente.

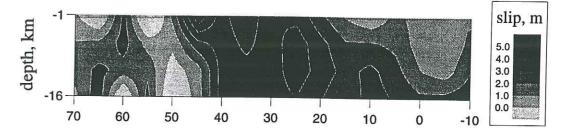
Pour conclure, la répartition spatiale du glissement pendant le séisme de Landers peut être expliquée avec des arguments d'ordre géométrique. La rupture a eu lieu le long d'un système de faille en échelon. Chaque échelon correspond à un endroit où le glissement est plus faible. Dans la majorité des cas, la rupture s'arrête lorsqu'elle rencontre la fin d'un segment de faille ou une zone qui avait déjà joué récemment.

#### B. Contrôle de la rupture: synthèse des derniers résultats

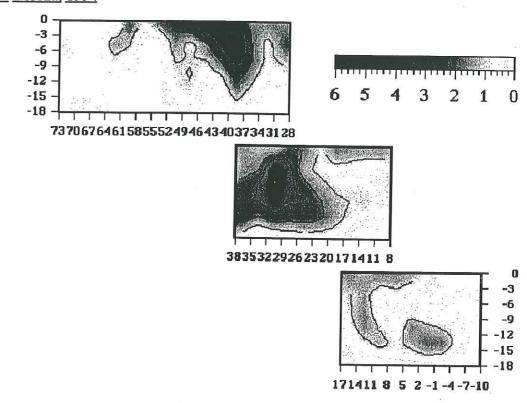
Ainsi, les résultats de notre étude de Landers et les études récentes de paléosismicité sur ce séisme confortent plutôt l'idée classique que la complexité de la rupture n'est que le reflet d'une hétérogénéité sous-jacente de la zone sismogénique.

Que penser de l'idée opposée selon laquelle la complexité de la rupture peut être considérée comme la conséquence de la dynamique non-linéaire du système ? Récemment Rice et Ben Zion (1995) apportent des éléments nouveaux à cette discussion. Tout d'abord, la plupart des modèles ayant montré que la complexité peut apparaître sur des faille "lisses" sont des modèles simplifiés et surtout discrets (Carlson et Langer, 1989; Bak et Tang, 1989). Ben Zion et Rice (1995) suggèrent que ces résultats dépendent de la discrétisation des modèles et par exemple, dans le cas de réseaux de blocks reliés entre eux par des ressorts (modélisation utilisée par Carlson et Langer, 1989), de la longueur du ressort. Les modèles considérés ici ne sont donc pas représentatifs de modèles lisses mais sont au contraire fondamentalement hétérogènes. Campillo et al. (1995) montrent par ailleurs des exemples clairs de la limitation des analogies utilisant des modèles de type bloc-ressort

#### Cotton and Campillo, 1995



#### Cohee and Beroza, 1994



#### Wald and Heaton 1994

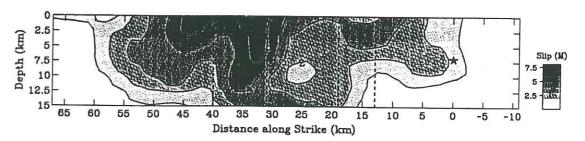


Figure 3. Distributions spatiales du glissement pendant le séisme de Landers déterminées par Wald et Heaton (1994), Cohee et Beroza (1994) et l'inversion décrite au chapitre II-1. Ces résultats ont tous été obtenus par inversion des données de mouvement forts.

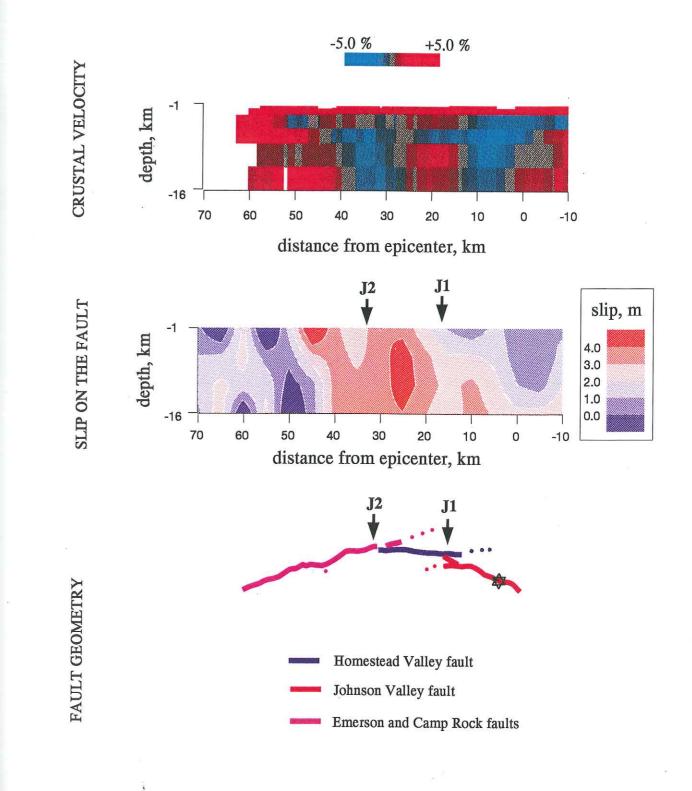


Figure 2. Comparaison entre les variations de vitesse dans le milieu (d'après les résultats de Lee, 1995), la distribution spatiale du glissement et la géométrie de la faille

pour représenter le glissement entre deux blocs élastiques. D'autres types de modélisation ont été effectuées avec des modèles fondés sur la mécanique des milieux continus par modélisation du glissement entre deux blocs élastiques. Les résultats les plus récents ont été développés par Rice et Ben-Zion (1995). Modélisant des failles lisses, ces deux auteurs ne retrouvent pas une distribution des séismes dont la répartition fréquence-taille est de type Gutenberg-Richter. Ces modèles théoriques de faille lisse où la dicrétisation est petite par rapport à la zone de nucléation produisent en effet des séismes réguliers à la fois dans le temps et dans l'espace. Rice et Ben Zion (1995) déduisent donc que la loi de Gutenberg-Richter observée dans la réalité reflète la répartition des tailles de faille dans la région ou le spectre des hétérogénéités sur le plan de faille. Une étude récente paléosismique et instrumentale de la sismicité des failles majeures californiennes (Wenousky, 1995) montre par ailleurs que les failles subissent de gros séismes dont la taille est sous-estimée par la loi de Gutenberg-Richter établie avec les séismes plus petits. En d'autres termes, tout comme dans les modèles de Rice et Ben Zion (1995) la sismicité sur un segment de faille donné est rythmée par des gros séismes de tailles identiques contrôlés par la géométrie.

Pour bien comprendre les limites de ces résultats théoriques il convient de mentionner que dans ce type de calcul les résultats dépendent de la loi constitutive choisie. Rice et Ben Zion (1995) utilisent des lois de friction tirées d'expériences de laboratoire effectuées à des vitesses de glissement situées entre  $10^{-7}$  et 1mm/sec. Cochard (1995) montre ainsi qu'en utilisant une loi de friction où le frottement décroît lorsque la vitesse de glissement augmente, il est possible d'obtenir un état de contrainte hétérogène à la fin de la rupture sur une faille "lisse" avant la rupture. En étudiant des séismes successifs sur la même faille il obtient des séismes de tailles diverses et la contrainte sur le plan de faille se maintient dans un état hétérogène. En revanche, la répartition taille-fréquence de ces séismes successifs n'obéit pas à une loi puissance de type Gutenberg-Richter.

Quelles sont les lois constitutives à des vitesses de glissements ou à des glissements importants ? Que donne la résolution de ce type de problèmes en trois dimensions ? Voici quelques problèmes à résoudre pour améliorer ce type de modélisation.

#### C. Implications en terme de risque sismique

En risque sismique il est important de savoir si les failles actives sont segmentées en différentes parties où se produisent des séismes de façon indépendante. En d'autres termes, connaissant la géométrie de la faille, peut on prédire le séisme maximum qui peut avoir lieu sur cette faille ?

Il y a une dizaine d'années est apparu le principe du séisme caractéristique (Schwartz et Coppersmith, 1984) selon lequel une faille joue de manière sismique par des séismes successifs de nature identique (même longueur et même magnitude). Cette notion est née d'études paléosismiques sur les failles de San Andréas et Wasatch . Les exemples de rupture répétée sur un même segment de faille sont rares. Sélectionnant rigoureusement les quelques données californiennes disponibles (Sieh, 1995) défend l'idée que des segments "lisses" sans hétérogénéité géométrique évidente en surface, glissent de façon "caractéristique": aux endroits où les glissements ont été décimétriques, les événements précédents étaient eux aussi d'ordre

décimétriques. De la même façon, là où le glissement a été métrique la paléosismicité montre que les glissements précédents ont eux aussi été métriques. A son avis, les modélisations théoriques où d'un événement à l'autre se produisent des glissements très irréguliers sur un même segment ne reflètent pas la réalité et la sismicité sur un segment de faille donné est au contraire rythmée par des gros séismes de tailles identiques ce qui correspond aux derniers modèles de Rice et Ben Zion (1995) et aux observations de sismicité instrumentale et de paléosismicité de Wesnousky (1995). Le séisme de Landers montre par ailleurs qu'un séisme peut passer d'un segment à un autre ou d'une faille à une autre et qu'il n'est pas possible d'anticiper la longueur d'un futur événement sismique en ayant analysé la segmentation des failles en surface.

La synthèse de toutes ces idées nous conduit à envisager un modèle de séisme composé de plusieurs sous-évènements. La largeur de chaque sous-évènement est contrôlée géométriquement par la segmentation des failles. D'un séisme à l'autre sur une même faille, le nombre de segments glissant pendant le séisme peut varier et en fonction du nombre de sous-évènements la longueur du séisme varie. En revanche le glissement moyen sur chaque segment est toujours le même quelle que soit la longueur du séisme. Ce modèle est il réaliste ? Plusieurs questions se posent immédiatement: ainsi quel est le facteur contrôlant l'amplitude du glissement sur chaque sousévènement ? Si le glissement est le même d'un séisme à l'autre sur le même segment alors le glissement doit être contrôlé par une propriété locale intrinsèque au segment. Quelle est cette propriété ? La dimension du segment ? Lors de Landers, l'inversion montre que d'un sousévènement à l'autre le glissement est très variable et n'est pas du tout corrélé avec la largeur de la zone ayant glissé (glissement de 6 mètres sur la faille Camp Rock pour une rupture qui est restée superficielle; glissement de 4 mètres sur la faille Homestead pour une rupture qui a cassé l'ensemble de la couche fragile). L'amplitude du glissement est elle contrôlée par les propriétés locales de la lithologie ? Qu'est ce qui définit précisément un segment ? Quels segments vont casser normalement indépendamment de leurs voisins ? Quels segments vont casser dans un séisme avec plusieurs sous-évènements ?

#### 3. Cinématique de la rupture et contrainte absolue

Donner une évaluation du rapport entre la chute de contrainte et la contrainte absolue initiale sur le plan de faille est un problème difficile. Nous avons étudié précisément les variations de la direction du glissement pendant le séisme de Landers (chapitre III-1). Notre étude montre que le glissement a gardé une direction constante pendant la rupture, cela alors que la répartition spatiale du glissement et les vitesses du front de rupture sont hétérogènes. En un point de la faille, la direction de la vitesse de glissement est colinéaire à la contrainte agissante. Cette contrainte agissante, est la somme de la contrainte préexistante (contrainte tectonique) et de la variation de contrainte dynamique résultant du processus de rupture. Si on suppose qu'un processus de rupture très hétérogène engendre de grandes variations spatiales des variations de contraintes dynamiques, on en déduit que la stabilité de la direction du vecteur glissement sur le plan de faille indique que la contrainte tectonique préexistante a été plus grande que les variations dynamiques. En d'autres termes, la chute de contrainte associée à la rupture a été partielle pendant le séisme de Landers. Ainsi l'étude de la cinématique de la rupture peut peut-être permettre d'avoir des

informations sur le niveau de contrainte absolu agissant sur la faille. Dans le cas de Landers, les variations de la direction de glissement sont faibles. Cette faille n'a pas joué depuis plusieurs milliers d'années, la contrainte tectonique tangentielle est élevée, la chute de contrainte est partielle et la rupture s'effectue avec une friction importante (faille « forte »). Dans le cas de Loma Prieta, la récurrence des séismes est plus élevée, les contraintes tectoniques tangentielles sont plus faibles, les chutes de contraintes totales et la rupture s'effectue avec une friction faible (failles « faibles »). Pour ces séismes et lorsque la rupture est hétérogène la direction du glissement est plus variable (observation faite à Loma Prieta par plusieurs équipes). Pour aller plus loin et confirmer ces hypothèses, des modélisations numériques en 3 dimensions où la direction de glissement n'est pas contrainte *a priori* sont nécessaires.

## 4. Variations de contrainte dynamique et sismicité déclenchée

Notre travail combiné avec les résultats d'autres équipes montre que la résistance des failles est une fonction dépendante du temps. De faibles variations des contraintes agissant sur des périodes longues (heures ou semaines) peuvent être corrélées avec le déclenchement des séismes alors que des variations plus importantes de durée faible ne semblent avoir aucun effet immédiat. Les estimations fiables des variations du critère de rupture de Coulomb statiques et dynamiques sont difficiles car ce type de calcul dépend de nombreux paramètres mal controlés (coefficient de friction, orientation des failles...). L'étude de l'évolution dans le temps des variations de contrainte statique crées par un séisme est un sujet de recherche intéressant. En effet, à des échelles de temps de l'ordre de l'année ou de la dizaine d'années on ne sait pas comment ces variations statiques sont redistribuées.

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Titre de l'ouvrage

Imagerie de la source sismique par inversion des mouvements forts et étude des variations de contraintes associées aux séismes

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Nom de l'établissement: Université Joseph Fourier- Grenoble I Résumé

Une méthode d'inversion des accélerogrammes de champs proche dans le domaine fréquences est tout d'abord décrite. Cette méthode permet de trouver le glissement, la vitesse du front de rupture et la durée de la rupture en chaque point de la faille. méthode a été appliquée à trois séismes: le séisme californien de Landers (1992), le séisme d'Uttarkashi (1991) situé au nord de l'Inde et le séisme de Northridge en Californie (1994). La répartition spatiale du glissement montre la présence de plusieurs sous-évènements. largeur semble controlée par la segmentation de la faille. La durée de la rupture en chaque point est compatible avec un mécanisme de type fracture où la radiation d'un point de la faille se termine à l'arrivée d'une phase d'arrêt émise à partir des bords de chaque sous-évènement.

Dans une deuxième partie, nous nous sommes variations de interessés aux contraintes associées aux séismes. Notre étude de la cinématique du séisme de Landers montre que le glissement a gardé une direction constante pendant la rupture alors que celle-ci hétérogène. Nous interprétons ce résultat par une chute de contrainte partielle pendant le séisme. Une méthode de calcul du champ de contrainte dynamique est ensuite décrite. Cette méthode est appliquée au séisme de Landers. Il est montré que la résistance des faille est une fonction dépendante du temps. De faibles variations "statiques" des contraintes peuvent être corrélées avec le déclenchement séismes alors que des variations "dynamiques" plus importantes mais de faibles durées n'ont eu aucun effet immédiat.

#### Mots clés

Sourcesismique - Mouvements forts - Inversion - Imagerie - Inversion génétique - Landers - Uttarkashi - Northridge - Sismicité déclenchée - Variations de contrainte dynamique.