Doctoral Thesis

Earthquake source study
joint inversion of broadband seismological data and SAR interferograms

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EARTHQUAKE SOURCE STUDY:
JOINT INVERSION OF BROADBAND SEISMOLOGICAL DATA
AND SAR INTERFEROGRAMS

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presented by

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Abstract

Earthquake source studies are generally implemented using different types of data: first, the local displacement field and accelerations (near-field data), and second, the wavefield generated by the earthquakes and recorded at teleseismic distances (far-field data). Many regions where earthquakes occur are poorly instrumented with no proper near field equipments like GPS or strong motions stations. The main topic of this work is the study of the rupture process of large earthquakes using 1) far field broadband seismological data acquired by the global seismological networks and 2) remote satellite measurements of the deformation field. The inversion of seismological data cannot always solve in a consistent manner for the source parameters of an event and is penalized by a trade off between rupture timing and location. On the other hand, geodetic data are sensitive to the spatial complexity of the faulting. More specifically InSAR data are able to map the coseismic ground deformation on large areas. In this thesis, we carry out a joint analysis of seismological and InSAR data and assess their resolving power with synthetic data. The joint inversion provides a reliable assessment of the spatio-temporal slip distribution of a rupture process.

We implement joint inversions on different large earthquakes: the Mw= 7.7, Nazca (Peru), November 12, 1996, subduction event; the Mw= 7.1, Hector Mine (California), October 16, 1999, strike-slip event; and the Mw= 7.5, Izmit (Turkey), August 17, 1999, strike-slip event. Improvements are further carried out by addition of other complementary data in the joint inversions: near-field strong motions, GPS data, complementary InSAR scenes and rupture offsets observations. From the different earthquakes source studies, we infer that the temporal and spatial complexities of a rupture can be resolved with better confidence when independent data sets are simultaneously inverted.
Résumé

La caractérisation des sources sismiques s’effectue en général à l’aide de différents types de données: premièrement, la mesure du champ de déformation cosismique du sol et du champ d’accélération local (données de champ proche), et deuxièmement, la mesure des ondes sismiques générées par les tremblements de Terre et enregistrées à distance télésismique (données de champ lointain). De nombreuses régions soumises à des séismes sont dépourvues de l’instrumentation appropriée telle que les stations GPS ou les accéléromètres pour les mouvements forts. Ce travail est principalement consacré à l’étude de la source sismique au moyen de 1) les données télésismiques large bande recueillies par les réseaux sismologiques mondiaux et 2) les données de déformation du sol fournies par les mesures radar satellitaires. 

La détermination des paramètres à la source au moyen de données sismiques ne permet pas toujours leur caractérisation précise. Il est de plus difficile d’ajuster à la fois l’estimation temporelle de la source et sa localisation spatiale précise uniquement au moyen de données sismiques. Les mesures géodésiques sont pour leur part sensibles à la géométrie spatiale de la rupture et, plus particulièrement, les données satellitaires InSAR qui enregistrent sur de larges surfaces la déformation cosismique liée à un tremblement de Terre. Nous avons développé dans cette thèse une méthode d’analyse combinée des données télésismiques et InSAR et nous évaluons leur résolution au moyen de données synthétiques. L’inversion jointe de ces données est un outil d’évaluation fiable de la distribution spatio-temporelle du glissement lors d’une rupture sismique.

Nous analysons différent grand tremblements de Terre au moyen de cette méthode: le séisme de subduction du 12 Novembre 1996, de Nazca (Pérou), Mw= 7.7; le séisme en décrochement du 16 Octobre 1999, à Hector Mine (Californie), Mw=7.1, le séisme en décrochement du 17 Août 1999 d’Izmit (Turquie), Mw=7.5. L’étude des sources est de plus améliorée grâce à l’utilisation de données complémentaires: les mesures de mouvement fort en champ proche, les données GPS, l’addition de mesures InSAR complémentaires et les observations de rupture en surface. L’étude de ces différents séismes nous a conduit à favoriser la description temporelle et spatiale d’un séisme au moyen de l’analyse combinée de données indépendantes afin d’obtenir une meilleure contrainte sur l’histoire spatio-temporelle d’une rupture sismique très complexe.
General introduction

1. The study of the large earthquakes

Earthquakes are one of the most dramatic hazards that are likely to cause heavy human losses and destroy entire cities on a scale of minutes. The more recent damaging events in Kobe (Japan, 1995), in Izmit and Duzce (Turkey, 1999) or ChiChi (Taiwan, 1999) recall that so far little is known about earthquakes physics that could prevent people from their deadly effects. Taking place in the earth’s interiors, a seismic event can be only detected or studied through its effects on the surface (Earth deformation and seismic waves). Decades of research involving numerous laboratories worldwide aim at investigating this large scale phenomena and trying to understand how it triggers, initiates, propagates, and stops.

2. Main goals of this thesis

This thesis focuses on the earthquake’s source study and more specifically on the space time history of the large events with a moment magnitude of 7 or more. For that purpose, different data types will be used. Most damaging earthquakes occur in areas that are not equipped with near-field instrumentation, we will then analyse them mainly using remote tools. Global seismic networks provide us with records of the far-field teleseismic data. The European Remote Sensing Satellites (ERS1/2) monitor the earthquake’s permanent ground deformation on large scale in the vicinity of the faulting. Though these data types provide detailed information on the rupture history, additional information coming from near-field observations (rupture offsets) or near-field instrumentation (GPS, strong-motions) are of major interest to enhance our knowledge of the source characteristics. The overall data sets allow the description of the geometry of faulting and the kinematic characteristics of the seismic sources, i.e the distribution in space and time of slip on the fault. Since an earthquake process is highly non linear, and the means to investigate it being somehow limited, different space time histories can be obtained. Our major goals are thus to restrain the possible solutions combining together independent informations available, to bring out the more insightful characteristics of the source process and to explore systematically the resolution capability of different data with concrete examples.
3. Content

In this thesis, we investigate the spatio-temporal characteristics of several recent earthquakes. Chapter 1 details the seismic source and its mathematical description for a kinematic approach. Some basics are given to solve the inverse problem, i.e. how to model a seismic process from observed data sets. Different inversion methods are reviewed and more details are given for the non linear search method we employed in the investigations. The data types are described and our investigation technique is explained (Fig.1). The following chapters are applications of the source investigation technique to different large earthquakes (Fig2). Chapter 2 describes the slip history of the 1996, Nazca (Peru), Mw=7.7, earthquake, a subduction event. It shows the potential of the joint inversion of Differential InSAR data and teleseismic data to infer precise spatio-temporal slip histories. This study validates the relevance and the potential of the joint inversion of seismic waves and dense remote geodetic measurements. Chapter 3 is the investigation of the 1999, Hector Mine (California) Mw=7.1, earthquake which occurred in the close vicinity of the 1992, Landers (California), Mw=7.2, earthquake. This strike-slip event is the best monitored seismic event to date. The very complex geometry of the faulting is inferred precisely from the InSAR technique that maps the complete ground deformation field. Further, we enhance the resolution power of the joint inversion method by combination of two InSAR scenes, GPS data, rupture offsets measurements and teleseismic data. Chapter 4 is dedicated to the study of the 1999, Mw=7.5 Izmit (Turkey) earthquake whose space-time history is inferred with InSAR, GPS, teleseismic, near field strong-motions and rupture offsets measurements. In this last part, we put the stress on the imperative requirement of complementary information provided by as many independent sources as possible to solve precisely the rupture history of large earthquakes in a joint inversion procedure.
INTRODUCTION

Figure 1: Schematic diagram of the different data types used in this thesis. Order of the spatial range of measurement is given for the geodetic data (GPS, rupture Offsets, InSAR), and spatial location of the seismic stations (Strong motions, and broadand teleseismic stations).

Figure 2: Earthquakes studied in this thesis. Left, surface projections of the fault models. Middle, earthquake centroid moment tensors (Harvard solutions). Right, different data types used for the investigation of the slip histories for each earthquakes.
Chapter 1

Imaging the source
1. The seismic source

An earthquake is a sudden rupture phenomena that takes place in the Earth’s crust or mantle caused by tectonic stress. The usual way to study these events is to get observational constraints provided mostly by the records of the seismic waves and near field observations (for shallow earthquakes). In this part, some theoretical background is presented in order to give insights into the earthquake process through the study of the ground deformation produced by a fault rupture.

1.1 The waves sources

1.1.1 An isotropic source

The majority of seismic sources originate from faulting. However underground explosions, generally human made also radiate seismic waves. They involve a complex process with the particularity that they generally occur with a spherical symmetry and their study provide a first approach of the seismic source.

Considering an explosion in an elastic infinite homogeneous medium, at distance r from the source, outside the very near source area where non linear processes occur, the infinitesimal strain theory is valid and the time-dependent pressure force, \( F(t) \), produced by the inelastically deformed interior, can be approximated by an impulse or a step, i.e. a transient or permanent effective pressure.

For an isotropic source, the seismic waves propagate outward with an equal amplitude in every directions on a spherical wavefront. The displacement field, \( u(r,t) \), is:

\[
\ddot{u}(r,t) = \left( \frac{1}{r^2} \right) \vec{F}(t - \frac{r}{\alpha}) + \left( \frac{1}{r\alpha} \right) \frac{\partial}{\partial t} \vec{F}(t - \frac{r}{\alpha})
\]

where \( r \) is the distance from the elastic radius \( r_e \), \( \alpha \) the P-wave velocity, and \( \tau = t-r/\alpha \) is the delay time (time required for the wavefront to reach the distance \( r \)).

The first term of equation (1) is called the near field term. It decays quickly with the distance \( (1/r^2) \) and if any step in the effective pressure occurs a static (permanent) deformation of the surrounding elastic media will result. The second term decays more slowly \( (1/r) \), dominating displacement at large distances, and corresponds to the far-field term which is proportional to the time derivative of the pressure force \( F(t) \). Those two terms are responsible for the observed
static deformations in the vicinity of a source and for the observed transient deformations at far distances.

### 1.1.2. The faulting source

A faulting event is approximated by a shear dislocation on a planar surface, which is not spherically symmetric but does have a low order symmetry. The amplitude and sense of the initial motion distributed over the P and S wavefronts in the vicinity of the source determine a radiation pattern that provides insights into the fault plane orientation, allowing the detailed study of the faulting mechanism.

The shearing motions on an area occur when the elastic strain accumulation overcomes the static frictional stress that resists motion. During an earthquake, sliding motion initiates at a point, the hypocenter, and a slip front expands outward over the fault, separating regions that are slipping from regions that have not yet slipped. The expansion of the rupture area is a function of space and time, $A(x,t)$, as is the corresponding slip function, $D(x,t)$. $D(x,t)$ gives the vector sliding motions on the fault (Fig. 1.1). For the long wavelength waves excited by the source motions, the rupture area and source volume are relatively small and can be approximated as a point source in space. Shorter wavelength waves are sensitive to finite extent and detailed variation of slip process on the fault, and they require a finite fault model.

![Figure 1.1: sketch of a rupture spreading from the hypocenter over a fault plane. The displacement field, $D(x,t)$, illustrates the heterogeneous behaviour of faulting (after Lay and Wallace, 1995).](image-url)
Planar faulting is defined as the slippage between two blocks of material in the plane connecting the two. The orientation of the fault plane in geographic coordinates is defined with two angular parameters: The *strike* of the fault, $\phi$, the azimuth of the fault projection onto the surface measured from north clockwise and the *dip* of the fault, $\delta$, the angle measured downward from the surface to the fault plane in the vertical plane perpendicular to the strike (Fig. 1.2).

![Figure 1.2: Standard definition of fault-plane and slip vector orientation parameters defined with the strike ($\phi$), the dip ($\delta$) and the rake ($\lambda$).](image)

The actual motion of the two blocks is defined by a slip vector. The direction of the slip vector is given by the angle of slip, or *rake*, $\lambda$, measured in the fault plane from the strike direction to the slip vector showing the motion of the hanging wall relative to the footwall. The magnitude of the slip is given by $D$, the total displacement of the two blocks. All parameters can vary greatly over finite fault surfaces and average values are used in the models. Three basic categories of fault are used to characterize motion on faults according to different values of dip and rake: thrust faulting, normal faulting, and strike slip faulting (Fig. 1.3).

![Figure 1.3: Examples of the basic styles of faulting for different slip vectors.](image)
1.2 Quantitative description of the seismic sources

1.2.1 Modeling of the earthquake process

Fault slip involves three main stages:

1. initiation of the fault sliding,
2. growth of the slip zone or rupture front expansion,
3. termination of source process.

The details of the fault slip impose a signature on the body waveforms, providing an empirical means for studying the faulting process using seismology. Two approaches of the earthquake phenomenology can be addressed:

The *dynamic simulation* that models the rupture propagation through initial conditions and physical laws (the friction laws, (1)). These models require an a priori knowledge of the fault properties at any points that are difficult to appraise before an earthquake.

*Kinematic modeling* is a relatively simple description of the faulting using a certain number of parameters (fault geometry and fault points slip histories (2)). Parameters are not derived directly from physical laws.

Kinematic modeling does not attempt to explain the earthquake process based on first physical laws. It provides an accurate description of the rupture history, which dynamic modeling cannot achieve. The kinematic process is often represented with a simplified force system, $f$, that produces equivalent seismic wave radiation. This approximation is valid for seismic waves with periods longer than the duration of rupture and for wavelengths that are large relative to the fault dimension: the complex faulting is replaced by a simple dislocation representation simulated by a *double couple force system*, applied within an elastic medium.

In this study, we aim at retrieving a detailed description of the faulting process. We will use a finite fault model that is represented with an equivalent distribution of double couples point sources that have the advantage to satisfy both near field and far field considerations of the seismic sources (Fig.1.4).

*Figure 1.4: Sketch of different fault models.*
*Up, the point-source model and its average parameters.*
*Down, finite fault model inferred from a distribution of point-sources (see text for details).*
1.2.2 Double couple point sources

The equation of a particle motion in an elastic medium is given by:

\[ \rho \frac{\partial^2 \mathbf{u}}{\partial t^2} = \mathbf{f} + (\lambda + 2\mu) \nabla(\nabla \cdot \mathbf{u}) - (\mu \nabla \times \nabla \times \mathbf{u}) \]  

(2)

where \( \mathbf{u} \) is the displacement of a particle in the medium, \( \lambda \) and \( \mu \) are the Lamé constants, \( \rho \) is the density and \( \mathbf{f} \) the force system.

1.2.2.1 The static deformation

The displacement field for a static deformation due to a double couple is obtained solving for the equation:

\[ \mathbf{f} + (\lambda + 2\mu) \nabla(\nabla \cdot \mathbf{u}(r)) - (\mu \nabla \times \nabla \times \mathbf{u}(r)) = 0 \]

(3)

Given a double couple force system, the static displacement field in polar coordinates, \( \mathbf{u}(r,\theta,\phi) \), in an infinite, isotropic, elastic, homogeneous medium with density \( \rho \), and elastic constant \( \lambda \), and \( \mu \) due to a force is: (equation 8.28, Lay and Wallace, 1995)

\[ u_r = \frac{M}{(4\pi\mu r^2)} \left( 1 + \frac{\Gamma}{2} \right) \sin^2 \theta \cdot \sin 2\phi \]

(4)

\[ u_\theta = \frac{M}{(4\pi\mu r^2)} \left( 1 - \frac{\Gamma}{2} \right) \sin 2\theta \cdot \sin 2\phi \]

(5)

\[ u_\phi = \frac{M}{(4\pi\mu r^2)} (1 - \Gamma) \sin \theta \cdot \cos 2\phi \]

(6)

with, for a Poisson solid

\[ \Gamma = \frac{(\lambda + \mu)}{\lambda + 2\mu} = \frac{2}{3} \]

(7)

where \( M \) is the finite moment of the double couple (\( M=\mu S \bar{D} \), with \( S \) a surface with an average displacement \( \bar{D} \)).
1.2.2.2 The transient deformation (far field term)

We consider equation (2) with a time dependent double couple force system.

$$\rho \frac{\partial^2 \mathbf{u}(r, t)}{\partial t^2} = \mathbf{f}(t) + (\lambda + 2\mu) \nabla \cdot (\mathbf{u}(r, t)) - (\mu \nabla \times (\nabla \times \mathbf{u}(r, t)))$$

(8)

The far-field radiation of each point source, i, located on a fault can be written as:

$$\mathbf{u}_i = \mathbf{u}_i^P + \mathbf{u}_i^S$$

(9)

where $\mathbf{u}_i^P$, $\mathbf{u}_i^S$, are the far-field terms for the P-waves and the S-waves respectively. These terms can be expressed, following Aki and Richards (1980), and Lay and Wallace (1995) (equations 8.65):

$$u^P(r, t) = \frac{1}{4\pi \rho r^3} R^P \frac{\partial}{\partial t} M \left( t - \frac{r}{\alpha} \right)$$

(10)

$$u^{SV}(r, t) = \frac{1}{4\pi \rho r^3} R^{SV} \frac{\partial}{\partial t} M \left( t - \frac{r}{\beta} \right)$$

(11)

$$u^{SH}(r, t) = \frac{1}{4\pi \rho r^3} R^{SH} \frac{\partial}{\partial t} M \left( t - \frac{r}{\beta} \right)$$

(12)

where $R^P$, $R^{SV}$, $R^{SH}$ are the radiation terms of the P, SV, and SH waves respectively. They are functions of the strike, dip, and rake of fault plane, and the seismic ray take-off angle. $\alpha$ is the velocity of the P-waves, $\beta$ is the velocity of the S waves, $\rho$ is the density of the medium. $\delta M/\delta t$ is the time derivative of the seismic moment function : $M(t) = \mu A(t)D(t)$ with $A(t)$ being the faulting surface, $D(t)$ the slip function, and $\mu$ the shear modulus.
1.2.3. The slip function $D(t)$

The slip function of an earthquake may have substantial irregularity, but it can often be approximated by a simple *ramp function*. The particle will need a finite time $t$ to achieve its final offset. The time derivative of slip at a point is called the *particle velocity*. The moment rate function ($\delta M/\delta t$), that arises from a ramp function is a *boxcar function* of length $t_r$, the *rise time* or the time required for a particle to achieve its final displacement (Fig.1.5). The moment rate gives the shape of the P- or S-waves energy radiated from the earthquake in the far field (equations 10, 11, 12). Hence, a fault approximated by a single point source (double couple), with a ramp function as slip history would have the P and S waves shaped like boxcars. The moment rate function of an earthquake that controls the amplitude of the body waves pulses is referred as the *source time function*. Determining the source time function of an earthquake allows the mapping of the spatio-temporal history of the fault slip.

![Figure 1.5: Relation between the displacement history of a particle on a fault and the far-field source time function (After Lay and Wallace, 1995).](image)

1.2.4. Kinematic finite source models

The study of earthquakes’slip history requires a more complex description than the simple point source approximation, namely a finite fault model in which the spatial extent of the rupture is taken into account. A finite fault can be discretized into series of point sources whose displacement histories are delayed in time as the rupture front expands. The summation of all point sources, accounting for the time delay, gives the complete slip function for the earthquake by linear superposition.
1.2.4.1. The one-dimensional Haskell source

This source model is represented by a rectangular fault where slip occurs simultaneously on the width of the fault (a “ribbon” fault) (Fig.1.6). The rupture starts at one end and terminates at the other with a finite velocity. The fault is divided into small segments and approximated as a series of point sources. The far-field displacement is the summation of the sub-event point source displacements:

for the P waves (equations are equivalent for the S waves):

\[ u^p(r, t) = \sum_{i=1}^{N} du_i^p \left( r_i, t - \frac{r_i}{\alpha} - \Delta t_i \right) \]  

where \( \Delta t_i \) is the lag between subevents, \( r_i/\alpha \) is the P-wave propagation time.

Figure 1.6: Sketch of one-dimensional fault of width W, length, L with individual fault segments of length, dx, for a rupture velocity \( V_r \). Rupture initiates at the edge of the fault (Star) Rupture initiates at the edge of the fault (Star)

The moment rate function of each segment, \( \mu A_i dD_i(t)/dt \), can be written as \( \mu w dx dD_i(t)/dt \) for a ribbon fault where \( w \) is the width of the segment. The far-field displacement for the P waves due to each elementary segment dx according to equations (10) is:

\[ du_i^p(r, t) = R_\alpha^p \frac{\mu}{4\pi \rho \alpha^3} r_0 d\chi D_0 \hat{F} \left( t - \frac{r(x)}{\alpha} - \frac{x}{v_r} \right) \]

with \( R_\alpha^p \), being the same for the subevent source points for the far-field approximation, \( x \) is the rupture length at time \( t \), \( v_r \) is the rupture velocity, \( D_0 \) the amplitude of the source time function, \( r_o \) and \( r \) are the distances between the slip initiation point and the currently slipping point to the point of observation respectively.
$F(t)$ is a ramp function with values equal to 0 if $t$ is inferior equal to 0, $t/r_r$ between 0 and $r_r$, the rise time, and 1 if $t$ is greater than $r_r$. The time derivative of this source function is a box car of length $r_r$, the rise time (Fig.1.5).

The entire far-field displacement is obtained integrating on the whole length of the fault:

$$u^j = \frac{R^j}{4\pi\rho c^3 r_0^2} \int_0^L F \left( t - \frac{r(x)}{c} - \frac{x}{v_r} \right) dx$$

(15)

$j$ being $P$, $c$ is $\alpha$, the P-waves velocity

$j$ being $S$, $c$ is $\beta$, the SH waves velocity

The integral (Eq. 2.14 to 2.24, Hernandez, 2000):

$$\int_0^L F \left( t - \frac{r(x)}{c} - \frac{x}{v_r} \right) dx$$

(16)

is computed with the far field approximation $r(x) \sim r_0 - x \cos \theta$, $\theta$ being the angle from which the observer sees the fault, giving:

$$\int_0^L F \left( t - \frac{r_0}{c} - \frac{x}{c} \left( \frac{c}{v_r} - \cos \theta \right) \right) dx$$

(17)

using the new variable:

$$\eta = t - r_0/c - \frac{x}{c} \left( \frac{c}{v_r} - \cos \theta \right)$$

(18)

$$d\eta = \frac{1}{c} \left( \frac{c}{v_r} - \cos \theta \right) dx$$

(19)

for $x=0$, $\eta = t-r_0/c$

for $x=L$, $\eta = t - r_0/c - L/c.(c/v_r-\cos \theta) = t-r_0/c - \tau_o$

$\tau_o = L/v_r - L\cos \theta/c$ is the apparent duration of the source that depends on the azimuth of observation. This azimuthal variability that involves a duration change of the source time function (or the seismic signal) without change in the seismic moment, is called directivity.
If $\theta$ is $\pi/2$ then $\tau_o$ is the actual duration of the earthquake:

$$\int_0^L \hat{f}(t-\frac{r_o}{c}-\frac{x(c/\nu_r-\cos \theta)}{\nu_r}) \, dx = -\frac{c}{\nu_r-\cos \theta} \int_{t-\frac{r_o}{c}}^{t-\frac{r_o}{c}-\tau_o} \hat{f}(\eta) \, d\eta = \frac{L}{\tau_o} \left[ F\left(t-\frac{r_o}{c}\right) - F\left(t-\frac{r_o}{c}-\tau_o\right) \right]$$  \hspace{1cm} (20)

the far-field displacement for the P and S-waves is then:

$$u^j = M_o R^j \frac{1}{4 \pi \rho c^3} \frac{1}{r_o} \frac{1}{\tau_o} \left[ F\left(t-\frac{r_o}{c}\right) - F\left(t-\frac{r_o}{c}-\tau_o\right) \right]$$  \hspace{1cm} (21)

with $M_o = \mu L w D_o$ the total seismic moment.

The far field displacement amplitude is thus proportional to the seismic moment, $M_o$, and inversely proportional to $\tau_o$.

Details of the faulting complexity are frequency and wavelength dependents. In order to give insights into those dependences, the slip function is studied in the frequency domain using the Fourier transform. In the frequency domain $\omega$ the Fourier transform of a ramp function $D(t)$ is:

$$TF(F(t)) = F(\omega) = \frac{1}{\tau \omega^2} \left(e^{-i\omega \tau} - 1\right)$$  \hspace{1cm} (22)

where $\tau$ is the rise time.

and:

$$TF\left(F\left(t-\frac{r_o}{c}\right)\right) = F(\omega) e^{-i\omega \frac{r_o}{c}}$$  \hspace{1cm} (23)

$$TF\left(F\left(t-\frac{r_o}{c}-\tau_o\right)\right) = F(\omega) e^{-i\omega \left(\frac{r_o}{c} + \tau_o\right)}$$  \hspace{1cm} (24)

$$TF\left(F\left(t-\frac{r_o}{c}\right) - F\left(t-\frac{r_o}{c}-\tau_o\right)\right) = F(\omega) e^{-i\omega \frac{r_o}{c}} \left(1 - e^{-i\omega \tau_o}\right) = \frac{1}{\tau \omega} \left(e^{-i\omega \tau} - 1\right) \left(1 - e^{-i\omega \tau_o}\right) e^{-i\omega \frac{r_o}{c}}$$  \hspace{1cm} (25)

The effect of the fault finiteness are accounted for by $\tau_o = L/c. (c/\nu_r-\cos \theta)$. 

\[\text{Chap. 1 Imaging the source}\]
The spectral density of the displacement field is then:

\[
|\tilde{u}(\omega)| = \frac{1}{4\pi \rho C^3 r_o} M_o R_c \left| \frac{\sin \left( \frac{\omega \tau_o}{2} \right)}{\omega \tau_o} \right| \left| \frac{\sin \left( \frac{\omega \tau}{2} \right)}{\omega \tau} \right|
\]  

(26)

where it appears that the displacement amplitude decreases with an increasing frequency \(\omega\).

The boxcar peak amplitude spectrum can be approximated as:

\[
\frac{\sin \left( \frac{\omega \tau}{2} \right)}{\frac{\omega \tau}{2}} \approx \begin{cases} 1 & \omega < \frac{\tau}{\tau_o} \\ \frac{\omega \tau}{2} & \omega > \frac{\tau}{\tau_o} \end{cases}
\]  

(27)

The spectrum of a boxcar (Eq.27; Fig.1.7) has a plateau at frequencies less than \(2/\tau\) and then decays in proportion to \(1/\omega\). The crossover frequency between the plateau and the \(1/\omega\) behavior defined by the intersection of the asymptotes to the low and high frequency spectra is called a corner frequency.

Assuming that the rise time is lower than the rupture duration (\(\tau < \tau_o\)), the convolution of two box cars (fault finiteness Eq.26) gives a peak amplitude spectrum (Eq.9.25, Thorne and Wallace, 1995) as:

\[
u(\omega) = \begin{cases} M_o & \omega < \frac{\tau}{\tau_o} \\ \frac{M_o}{\frac{\tau_o}{2}} & \frac{\tau}{\tau_o} < \omega < \frac{\tau}{\tau_o} \\ \frac{M_o}{\omega^2 \left( \frac{\tau \tau_o}{4} \right)} & \omega > \frac{\tau}{\tau_o} \end{cases}
\]  

(28)
The amplitude spectrum content of a seismic pulse should be flat at periods longer than the rupture time of the fault. At periods between the rise time and the rupture time, the spectra decays as $1/\omega$, and at high frequencies the spectra decays as $1/\omega^2$ ($\omega^2$ source model). The decay of far-field displacement spectra is a natural consequence of interferences between high frequency waves caused by the temporal and spatial finiteness of the source.

The study of the earthquakes’ spatial and temporal history is thus based on kinematic models that allow a representation of the seismic source for a certain range of frequencies where rupture history is characterized with discretized models and a restrained number of parameters used to describe the moment rate function. We describe below more sophisticated kinematic models of finite sources which allow coseismic slip to vary on the fault surface.

![Figure 1.7: Seismic source spectrum for a convolution of two box cars. The intersection of the asymptotes to the low-frequency and high-frequency portions define the two corner frequencies $\omega_{c1}$ and $\omega_{c2}$ (After Lay and Wallace, 1995).](image-url)
2. The inverse problem

In the first part, we addressed the question of the representation of the seismic source with a limited number of parameters. We now give some insights into the method to recover those parameters from the observed data. We recall the representation theorem (Eq. 3.2, Aki and Richards, 1980) that gives the basic approach of the slip on a fault:

\[ u_j(m, \tau) = \int_0^t \int G_{ij}(m, x, \tau, t)s^i(x, t)d\Sigma \]

where \( u_j(m,t) \) is the \( j \)th component of displacement at point \( m \) and time \( \tau \), \( G_{ij} \) is the impulsive response of the medium at the point \( m \) to a dislocation at the point \( x \) and time \( t \), \( s^i \) is the \( i \)th component of the slip on the fault and \( \Sigma \) is the fault surface.

With the knowledge of the displacement \( u_j \) at the point of observation and the impulsive response of the medium \( G_{ij} \), the source imagery has to estimate the slip \( s^i \) at the point \( x \) and time \( t \) on the fault.

2.1 Discretization of the kinematic problem:
the multi-time windows formulation

Different numerical approaches allow the estimation of the earthquake parameters either in the temporal domain (Olson and Apsel, 1982; Hartzell and Heaton, 1983; Das and Kostrov, 1990; Wald and Heaton, 1992) or in the frequency domain (Olson and Anderson, 1988; Cotton and Campillo, 1995, Ihmle, 1996).

Here, we use the multi-time windows approach of Olson and Apsel (1982) in the temporal domain. The fault is subdivided into planar surfaces approximated as point sources in which are defined two component slip vectors, \( s \), triggered at a time \( t \) (Fig.1.8). The slip within each subfault (Equation 2-2, Olson and Apsel, 1982) can be written for a discretized fault of \( N \) subfaults as:

\[ s^i(x, \tau) = \sum_{n=1}^{N} X_n(\hat{x}) \sum_{k=0}^{K} \delta_{nk} P_k(x, \tau) \]

with \( X_n \) being 1 if \( x \) is in the \( n \)th cell and 0 otherwise

with

\[ P_k(x, \tau) = F(\tau + k\partial t) \]
The two components (along strike and azimuth) vector $s_{nk}$ is the slip direction of the $n$th subfault at the $k$th time point. The function $P_k(x, \tau)$ is the time dependence of the $k$th slip. On each subfault, slip is allowed to occur in $K+1$ time windows at successive increments $\delta t$, slip varying according to a specified time function $F(t)$. $\tau$ is the timing of the rupture front that can be delayed by $\delta t$. Thus, the multi-time window formulation allows either a longer slip duration or a locally delayed rupture.

The space and time dependence being known, equation (31) gives:

$$u_j(\vec{m}, \tau) = \sum_{n=1}^{N} \sum_{k=0}^{K} s_{nk} g_{nj}(\vec{m}, \tau + k\delta t)$$

(35)

where the vector $g_{jn}(\vec{m}, \tau+k\delta t)$ is the green’s function for the $n$th subfault at position $\vec{m}$ in the $j$th component direction. In the work presented here, the synthetic green’s functions or the ground motion contributions for a unit deformation are computed for the dip-slip and strike-slip components on each subfault using ray theory (Nabelek, 1984), and stored before the inversion. If strong-motion data are used in the inversions, more complexity is added in the subfault parametrization: subfault contributions are computed by summing the responses of equally spaced point sources delayed in time according to the different source-to-station positions and to the local rupture front propagation (~75% of the shear wave velocity) (Legrand and Delouis, 1999). For the static inversion, the ground deformation is simply represented as a linear sum of $N$ subfaults contributions (see 1.2.2.1):

$$u_j(\vec{m}) = \sum_{n=1}^{N} s_{n} g_{nj}(\vec{m})$$

(36)

where $u_j(\vec{m})$ is the $j$th component of the static ground deformation at point $\vec{m}$, $s_{n}$ is the final slip vector of the $n$th subfault, and the vector $g_{nj}(\vec{m})$ is the Green’s function of the $j$th component for a unit deformation observed in $\vec{m}$ for the subfault $n$. In our study, the synthetic static ground deformations are computed from unit dislocations embedded in a elastic half-space (Savage, 1980) and stored before the inversion.
Figure 1.8: Schematic diagram illustrating the finite fault discretization with point-sources when using a), only teleseismic data or b), strong motions and teleseismic data. See text for details.
2.2 parametrization of the slip function

The inversion methodology of our study requires the estimation of the slip function at each subfault $j$ in order to model the data with regard to the equation (35) (Fig1.9):

- The rupture initiation time $\tau_n$
- The rake direction of the vector $s_{nk}$ (strike slip and dip slip components)
- The slip vector $s_{nk}$ amplitudes in each time windows, i.e. at each $k\delta t$

In our applications, the time slip function $F(t)$ is represented by a smooth ramp function within each time window, which is the integral of an isoscele triangle.

2.3 Another parametrization

Most of the source tomography are based on multi-time windows parametrizations. Another original representation of the slip function suggested by Cotton and Campillo (1995) and Hartzell et al (1996) is a modified cosine function (Fig.1.9):

$$\frac{d}{dt} F(t) = \frac{1 - \cos\left(\frac{2\pi t}{t_r}\right)}{t_r} \quad (37)$$

where $0 < t < t_r$, and $t_r$ is the rise time, and $t$ the time for the slip to occur.

The advantage of this formulation is the need of only one parameter to represent the rise time function but it restricts the time history of individual subfaults to a simple shape in comparison to the multiple time window representation (Ji et al, 2002).

Figure1.9: Two alternative functions of the source funtion used in our inversion. (top) $T_i$ is the rupture time, and $a_i$ are the triangular functions amplitudes. (Down) parametrization used in Cotton and Campillo (1995), where authors invert for the rupture time $T_i$, the slip and the rise time (different rise times are represented).
2.4 The inversion methods

The source parameters inferred with inversion algorithms are estimated by minimizing the difference between the observed and computed data. In the following, I give a short review of the different inversion methods used for the study of the source tomography.

- The linear inversion method (Hartzell and Heaton, 1983) requires that the rupture velocity and source risetime are chosen to calculate the synthetics data. As a consequence it reduces the solution space and this can be particularly critical for large events where those two parameters can vary greatly.

- In order to overcome this problem Hartzell (1989) and Hartzell and Langer (1993) used a non linear least square method. The problem is linearized and solved in an iterative manner. The rupture time and rise time are made free parameters. Nevertheless a maximum rupture velocity has to be fixed to set up the earliest possible time at which points of the fault fails. Thus, one has to perform different inversions with different velocity.

Solving for the rupture parameters is an ill-conditioned problem, i.e the parameters space is likely to present several local minima. The above mentioned standard least square inversions are starting from an initial model and are likely to converge toward a local minimum. As a consequence several starting models have to be tried to test the stability of the solutions.

A way to overcome the choice of initial guesses is to use non linear research algorithms that are not linked to any initial model and solve the forward problem a “certain” number of times:

- The Global grid search method that requires a systematic discretization of the entire parameter space. Each models is a combination of source parameters and is tested to select the one which best matches the observed data. Nevertheless, this method is generally performed in a restrained parameter space to strongly limit the number of tested solutions and thus the computer time.

- The Monte Carlo search method determines randomly chosen sets of parameters. Each model produces a set of synthetic data that are compared to the observed ones (the cost function). The minimum cost function found corresponds to the best model. This method requires to explore the whole parameter space. Thus, if the parameter space is large enough, the inversion can be instable and is likely to find solutions in local minima.

- The quasi-global search methods offer a compromise between the random search algorithms and the local inversions (least square inversions):
- The **genetic algorithms methods** start from a randomly chosen sets of model parameters that are crossed and mutated in order to generate iteratively new sets of models with lower cost function that tends to converge toward the absolute minima.

- The **simulated annealing method** is the method that is implemented in this thesis. This algorithm converges toward the absolute minimum thanks to a probability function, i.e the likelihood of a model to be accepted. The “**simulated annealing**” algorithm uses the **metropolis criteria** to select solutions.

Metropolis et al (1953) introduced an algorithm simulating a set of atoms in equilibrium at a given temperature. At each step of the algorithm a random displacement of an atom produces a variation of energy.

- If $\Delta E$ is negative, the new configuration is accepted and is taken as a reference for the next displacement.

- If $\Delta E$ is positive a probabilistic approach is used, and the probability for the new configuration to be accepted is:

$$ P(\Delta E) = \exp\left(\frac{-\Delta E}{Tk_b}\right) $$

Where $\Delta E$ is an energy change which corresponds to a random displacement for an atom, $k_b$ is the Boltzmann constant and $T$ is a given temperature.

The lower $\Delta E$, the higher the probability to accept the new configuration. A number between 0 and 1 is randomly chosen and compared to $P(\Delta E)$. If this number is inferior to $P(\Delta E)$, the new configuration is accepted, if not the previous configuration is kept. The iteration of this process simulates the displacements of atoms in contacts with an environment at temperature $T$.

If a cost function is used instead of $\Delta E$ and if the atoms distribution is replaced by a serial of parameters, the metropolis algorithm is an useful procedure to determine a serial of configuration in an optimization process. The ensemble of the accepted configurations samples the whole parameter space and concentrates more specifically around the minima of the cost function.

The “**simulated annealing**” algorithm is an implementation of the Metropolis criteria. The “temperature” $T$ decreases with iterations of the process. At the start of the procedure, the “temperature” is high, the system can get numerous configurations (it corresponds to atoms at high temperature) and escapes from secondary minima. Step by step the temperature is low-
ered and the space of possible configurations is shrank. This procedure a converges toward the absolute minima.

**Implementation of the method for the exploration for the source parameters:**

The procedure (Lundgren, 1999; Delouis et al, 1999) we are using in our inversion is derived from the 1-D simulated annealing which schematically works as follow:

1- Set up: The parameters space is set and represented by intervals for each of the variables. A number of iterations $N$ for the exploration of the entire space parameter is set, and a number $M$ of iterations is set for exploration within the interval of each parameter. A starting temperature $T$ is chosen.

2- The starting model $(X_1,X_2,....,X_l)$ is randomly chosen giving a primary cost function basically the Root Mean Square between the observed and computed data.

3- All the model parameters are kept fixed except the first one, $X_1$. The cost function of the models $(X,X_2,....,X_l)$ are computed for a set of randomly selected values of the first parameter within its domain and compared to the primary cost function:

   - If a better configuration is found, it is taken as the optimal configuration.
   - If a configuration presents a cost function greater than the primary one a probability function is made from the cost function $f(X)$:

     \[
     P(X) = \exp\left(-\frac{f(X)}{T}\right)
     \]  

     (37)

   where $T$ is the “temperature”.

   If the new configuration is kept according to the metropolis criteria, its cost function replaced the primary cost function.

   A new model configuration is found $(X’_1,X_2,....,X_l)$ after $M$ iterations

4- This operation is repeated for the following parameters $X_i$ and a new model $(X’_1,X’_2,....,X’_l)$ is taken as a new starting model after $M_1$ iterations.

5- The procedure of the points 2 and 3 is repeated $N$ times.

6- The temperature is decreased and the model space parameters is shortened by a chosen factor.

6- The process is stopped when the convergence criteria (precision on the parameters) is respected.
3. Data available for the study of the seismic source

Most damaging earthquakes occur in regions that are not equipped with adequate near-field instrumentation. One the main goal of this thesis is the detailed study of the fault kinematics using data accessible through remote instrumentation. Until recently the only remote source of information about an earthquake was the records of the teleseismic waveforms. Since the 1992 Landers Earthquake (Massonet et al, 1993), the coseismic ground deformation can be remotely accessed through the ERS1/2 satellites (Europe), or the J-ERS1 satellite (Japan). We give more details concerning the remote acquisition and processing of those data in the following paragraphs. In addition to these data, we complement the study of the large earthquake source parameters using data obtained from near field instrumentation or observations: the strong motion data, and the GPS data. Observed rupture offsets and aftershocks locations are also of interest to constrain the rupture models.

3.1. The Broadband teleseismic data

Quickly after a large event the digital broadband waveforms are available at stations worldwide through different organizations. The location of the events hypocenter, centroid, the focal mechanism, and the seismic moment are provided by the Harvard Centroid Moment Tensor catalog (HCMT) (http://www.seismology.harvard.edu/CMTsearch.html). The teleseismic broadband data collections for the different earthquakes studies are available through 2 major seismological organizations: The Incorporate Research Institutions for Seismology (IRIS)(http://www.iris.washington.edu) and the Observatories and Research Facilities for EUropean Seismology (ORFEUS) (http://orfeus.knmi.nl).

The data are collected at stations located between 30° and 90° of the epicenter in order to select waveforms that propagated in the mantle. We collect the vertical P waveforms and the transverse horizontal SH waveforms. The waveforms are selected according to their quality (noise to signal ratio, emergence of waves arrivals) and their azimuthal distribution.

The rough seismograms are first corrected from the background noise and bandpass filtered in general between 0.8Hz and 0.01Hz for the P-wave and between 0.4 and 0.01 for the Sh-waves. The waveforms are deconvolved from the instrument response and integrated to obtain the seismograms of the particle displacement at the station. The P and Sh theoretical travel times
are determined from the IASP91 travel times table and the observed arrivals are picked manually for each stations. We define afterwards time windows for each wave types in order to study the distincts seismic arrivals corresponding to failure of discrete asperities. Finally, the seismic signals are sampled to two counts per seconds for the inversion procedure.

3.2. The InSAR data

In this part, the Synthetic Aperture Radar differential interferometry technique is briefly described. The differential interferometry is based on the interference phenomenon, a paradoxal addition of waves. It requires that these waves are coherent with the same frequency and have parallel components. The interference phenomenon is illustrated by the Michelson interferometer (1887). The Synthetic Aperture Radar interferometry has been developed in the last two decades and its applications have soon revealed its major interest in the earth science fields. This technology designed for a satellite platform provided numerous and high quality results for the topography (Digital Elevation Model generation), the study of the vegetation, the oceanic currents, the glacial processes or the landslides (Massonet and Feigl, 1998; Rosen et al, 2000). One of the main applications is in the field of the crustal deformation research that includes the tectonic deformation, the volcanism and the subsurface fluid flow. Since the 1992 Landers earthquake (Massonet et al, 1993), the European Remote Sensing Satellites (ERS1/2) are providing continuous radar data that are used for the measurements of the earthquakes ground deformation potentially measured with a precision corresponding to millimeter-level displacements with a spatial resolution on the order of 100m on areas on the order of 100 km².

A differential interferogram is basically obtained from the combination of two SAR scenes (spanning the event in time) that are a record of radar waves amplitudes and phases on a target area. The interferograms are made from the phase informations. The phase change $\Delta \Phi$ between two satellites paths are proportional to the surface displacement (Fig. 1.10):

$$\Delta \Phi = \frac{4\pi}{\lambda} (\vec{D} \cdot \hat{\gamma}_1 - \vec{B} \cdot \hat{\gamma}_1)$$

(38)

where $\vec{B}$ the baseline vector (separation between the two radar antennas) estimated from the precise satellites orbits, $\vec{D}$ is the displacement. Both vectors being significantly less than the range distance. $\hat{\gamma}_1$ is the look vector of the satellite. $\lambda$ is the radar carrier wavelength. $\vec{D} \cdot \hat{l}_1$ is
the displacement in the range direction. In the final differential interferogram the phase changes are directly proportional to the displacement in the range direction:

\[
\Delta \Phi' = \frac{4\pi}{\lambda} \delta d
\]

where \( \delta d \) is the displacement in the satellite look direction.

Figure 1.10: Imaging geometry of side-looking radar (ERS) for InSAR applications, after Bürgmann et al (2000).

**B** is orbit separation baseline. \( l_1 \) is the look vector of the satellite \( (l_1-l_2 \text{ at distance}) \).

**B.l_1** is the contribution of the orbit separation baseline to measured change differences. The box in the bottom illustrates the range change caused by the surface displacement component in the range direction.
A displacement of $\lambda/2$ in the range direction gives one cycle of phase difference or one fringe. For the ERS microwave C-band radar this corresponds to a displacement of 28 mm (Fig1.11).

From the radar raw data to the geophysical measurement of the ground deformation, a long chain of processing is required in order to correlate the radar images and remove different phase contributions (orbits and topography) (table 1.1). The differential interferograms in our study are computed using the ROI_PAC software developed at the Jet Propulsion Laboratory and the California Institute of Technology that provides geodetic data directly exploitable for the earthquakes source study. The InSAR data are only a scalar measurement, i.e. the deformation is measured only in one direction. ERS satellites measurements are mostly sensitive to vertical deformations and insensitive to deformation parallel to the satellite track. This problem can be addressed by combining InSAR scenes from different azimuths. Another component of the displacement field can be obtained by computing the azimuth offset field between two amplitude radar images acquired before and after the earthquake. It provides the surface displacement component parallel to the satellite track on the ground but the error estimates are about 20 cm (Michel et al, 1999; Peltzer et al, 2001).
Table 1.1: Scheme of steps in processing SAR data for interferometric applications (Bürgmann et al, 2000)
3.4 Combination of independent data sets for the slip history of large earthquakes

In general when a data set is inverted alone for the recovery of earthquake’s source parameters, some trade-offs limit their resolving power for the slip history. Many earthquake source tomography based on the study of band-limited waveforms alone are penalized often by a trade-off between the rupture timing and the slip location. The InSAR data exhibit also a trade-off between the mechanism and the location of an earthquake, and between the location and the magnitude of slip (Lohman et al, 2002). One of the main goal of this thesis is to limit these trade-offs, specially the one that affects the slip location and timing. The joint inversions of geodetic and waveforms data present the advantage of improving the earthquake source characterization with a better control of the slip pattern over the fault independently from the rupture timing. The final slip resulting from the joint inversion is thus required to fit the final static slip providing an independent constraint on the rupture evolution (Wald and Heaton, 1994).

Further some a priori constraints lead to a better description of the earthquake process. Preliminary studies of the earthquake have thus to be performed in order to raise the ambiguities that could penalize the recovery of the slip history (Wald et al, 1996). The geometry and the location of the fault rupture have to be clearly estimated by different means. Our inversion strategy is to perform a preliminary inversion of the teleseismic bodywaves using the point source modeling code of Nabelek (1984) in order to estimate as a first insight the average characteristics of the earthquake. We thus resolve for:

- The focal mechanism and eventually its spatial variations
- The depth of the source and an estimation of its spatial extent
- The directions and average velocity of rupture propagation
- The single or multi events behaviour of the faulting
- The total seismic moment and source time function
Other data are also used such as aftershocks or reported surface offsets that help to constrain final shapes and locations of the rupture surfaces before inversion for the rupture history. We check finally for the relevance of an adequate crustal model for the source if available.

To close up this introduction on the far from trivial inverse problem for the earthquake slip history, we remind Olson and Apsel’s criteria (1982) for an acceptable solution:

1- The solution must explain the data

2- The solution must be physically reasonable, i.e. consistent with independent constraints

3- If more than one solution fits the data equally well, additional information must be supplied to uniquely define which solution is being obtained.

In addition, we deem that resolution analysis of the finite fault inversion should be carried systematically out for the data sets, using synthetic rupture models in order to address the spatio-temporal sensitivity of the results, and analyse the ability of the method to resolve trade-offs between the complex source phenomena.
Chapter 2

Joint inversion of Broadband teleseismic and InSAR data for the slip history of the Mw=7.7, Nazca ridge (Peru) earthquake of November 12, 1996

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Abstract

The slip distribution of the November 12, 1996, Mw=7.7, Peru earthquake is determined using broadband teleseismic waveforms, a differential SAR interferogram (InSAR) and a fault parametrization allowing slip and rupture velocity to vary along the rupture plane. Both data sets are inverted jointly to limit the trade-off between the space and time aspects of the rupture. The earthquake fault plane is located at the subduction interface, it strikes parallel to the trench and dips 30º North-East. By inverting synthetic data, we show how the InSAR and teleseismic data are complementary, and how the joint inversion produces a gain in the spatial and temporal resolution of the slip model, even with a SAR interferogram which covers only part of the coseismic deformation. The rupture of the 1996 Peru event initiated on the southern flank of the subducted Nazca ridge and propagated unilaterally toward the South-East (along strike) for more than 100km between 20 and 40km depth. The area of maximum slip (6 to 7m) is located 50km South-East of the hypocenter. The total seismic moment is $4.4 \times 10^{20}$ N.m (our joint inversion). The source time function is approximately 60s long and presents three major pulses of moment release. The dominant one which occurred between 30 and 45s does not correspond to the area of largest slip but to the rupture of a wide zone located about 100km away from the hypocenter where slip reaches only 2 to 3m. Computed coseismic coastal uplift correlates well with the location of raised marine terraces and with the topography of the coastal cordillera, suggesting that these features may be related to the repetition of 1996-type events at the interface between the Nazca ridge and the South-American plate.

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Introduction

In the past decades, considerable effort has been dedicated to the study of the rupture process of large earthquakes using seismic and geodetic data. One of the main outcomes of those studies is that the slip distribution associated with large earthquakes is generally heterogeneous. Understanding the space and time characteristics of large ruptures is particularly important for the study of earthquake dynamics, the earthquake cycle, and the assessment of seismic hazard. Traditionally, teleseismic P- and S-wave pulses have been used to image the rupture history in terms of total source time function, slip map and local slip-rate. However, the ability to resolve the details of the rupture depends strongly on type of data used, on the inversion algorithm and of course on the degree of complexity of the source itself. Broadband seismic data may not be sufficient for a unique and reliable mapping. A way to deal with this problem is to derive rupture models by matching different data sets which include multiple constraints from independent observations, such as strong motion, GPS, remote sensing and surface breaks (Wald and Heaton, 1994; Hernandez et al, 1999; Wright et al, 1999). We combine two independent data sets acquired at remote distances, far-field teleseismic P and SH waves and near-source SAR (Synthetic Aperture Radar) interferometric data. The geodetic InSAR data obtained from the ERS1/2 satellites can sample surface deformation over a large area around the earthquake, and provide constraint on the slip distribution that is totally independent from the rupture timing. In our joint inversion scheme, we fit simultaneously the waveforms and the geodetic data, in order to reduce the possible trade-off between rupture timing and slip location, and therefore to enhance the reliability of the resulting slip distribution. A similar approach has been successfully applied to the 1999, Izmit (Turkey), Mw=7.6 earthquake (Delouis et al, 2000, 2002), where broad-band, InSAR, GPS, strong-motion and surface break data have been combined in a single inversion scheme, allowing the exploration of the resolving power of the individual data sets.

The November 12, 1996, Mw=7.7, Peru-Nazca ridge earthquake is one of the first large subduction earthquakes whose surface deformation has been mapped by ERS1/2 satellites. This thrust event had a large extent, mostly offshore. The ERS1/2 satellites measurements monitored the entire on-shore coseismic ground deformation. Despite the large coverage of the event, the precise Digital Elevation Model needed to process the final differential interferogram could be obtained only for a limited area, restricting the geodetic control to the southern extent of the deformation field. Nevertheless, the inversions of the separate
and combined data sets are of interest to analyze how a partial control from remote sensing can complement the teleseismic data to map the source complexity in a remote location where no other data were available and where typically no surface breaks can be observed. Synthetic tests are carried out to assess the benefits of the joint inversion. They contribute to demonstrate the major improvements in the space and time resolution of slip that can be expected in the joint inversion of real teleseismic and InSAR data.

1. The 1996 Peru earthquake

The November 12, 1996, Mw=7.7, Peru earthquake occurred at the subduction interface between the Nazca ridge and the South American plate (Fig.1). It caused human losses and building damages, landslides within about 200 km of the epicenter, as well as coastal uplift (Chatelain et al., 1997). The mechanism corresponded to underthrusting, as indicated by the Harvard CMT solution (strike 312°, dip 33° and rake 55°). The epicenter has been relocated by Spence et al. (1999), offshore near the coast of Peru, at 14.99°S and 75.63°W. This event occurred in the vicinity of the former Mw=8.1, 24 August, 1942 epicenter and it has already been proposed that the 1996 and 1942 ruptures overlapped, at least partially, both events being located in front of the southern side of the subducting Nazca ridge (Spence et al., 1999; Swenson and Beck, 1999). Other recent large earthquakes took place further north of the Nazca ridge and the subduction interface between the rupture areas of the 1996, 1942 events and that of the 3 October, 1974 earthquake appears to be unbroken for several centuries (Kelleher, 1972; Dorbath et al., 1990; Swenson and Beck, 1999). The Nazca ridge is an aseismic and volcanic bathymetric high, an area of elevated crust about 200 km wide, located on the oceanic Nazca plate. Consumption of the Nazca ridge at the convergent boundary (7.6 cm/y) migrates southeastwardly along the coast due to its oblique orientation with respect to the trench (Hsu, 1992). The subduction of such a buoyant structure is still poorly understood and an accurate retrieval of the slip history would shed light on the relationship between the ridge and the South American plate in this area.
2. InSAR and teleseismic data

The raw differential interferogram was generated using two pairs of ERS1/2 scenes and the ESA precision orbits (PRC) (Table 1). We used a four-pass technique to process the data with the ROI-PAC software developed at the Jet Propulsion Laboratory and Caltech. The topography has been generated from a first pair of interferograms, the 1996-10-23 (ERS-1) and 1996-10-24 (ERS-2) raw tandem scenes. This has been flattened and unwrapped with the minimum cost flow algorithm of Costantini (1998). It was then unflattened (i.e. the earth curvature added back) and the differential interferogram formed from a second coregistered change pair interferograms (1992-07-13 and 1997-10-09) representing a time delay of four years before and one year after the event. Figure 2a shows the observed interferogram where displacements are represented with a 5 cm fringe cycle by interpolation on the original unwrapped interferogram with 2.8 cm fringe cycle. The northernmost fringes are unexpected and are possibly due to baseline errors which could produce a phase ramp, although the baselines were reestimated based on the amplitude offsets in the single look complex images. However the significant elevation change along track (0-5000m) limits our ability to accurately refine the baseline from the amplitude data alone. The phase ramp is largest in the azimuthal direction (along satellite track), and should be corrected for. Nonetheless, we can recognize the central area of subsidence associated with the thrust event (closed fringes in the lower half of the interferogram) and part of the uplift zone indicated by the high fringe rate observed near the coast, and confirmed by observations of coastal uplift (Ocola et al, 1997; Chatelain et al, 1997). The contribution of aftershocks to the final static ground displacement is not considered since the largest aftershocks are located mostly offshore and are several orders of magnitude smaller than the mainshock (Spence et al, 1999). We assume that the differential interferogram represents essentially the coseismic deformation. Nevertheless, in the absence of additional data we cannot rule out some postseismic contributions in the period from 12 November 1996 to 10 October 1997. For the inversions, we use a set of 1213 georeferenced points distributed over the whole interferogram, denser where the gradient in the ground displacements is larger (Fig.2b).

Seismological data have been collected through the IRIS and Geoscope networks. They consist of P and SH seismograms recorded at teleseismic distances. Seismic records have been deconvolved from the instrument response and integrated to obtain ground displacement. The seismograms have been bandpassed from 0.8Hz (P-waves) or 0.4Hz
(Sh-waves) to 0.01Hz. We model the first 90s of 13 P-waves signals and the first 120s of 11 SH-waves signals well distributed in azimuth around the source.

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3. Fault model and inversion procedure

Preliminary estimates of the fault strike, dip and rake, as well as the hypocentral depth were obtained from broadband teleseismic modeling using the method of Nabelek (1984). Values for those parameters were confirmed or adjusted by trial and error inversions performed with both the InSAR and teleseismic data. The rupture is parametrized by a fault plane aligned with the Peru trench, striking 307° and dipping 30° to the north-east. The center of the fault and the hypocenter are located at 33km along strike and at 28km depth respectively (Fig.3). The dimensions of the fault model are 180km along strike and 120km along dip. It is subdivided into 54 subfaults with dimensions of 20 by 20km². For each subfault strike and dip are held fixed. Their respective slip angle (rake) is allowed to vary independently from the neighbor subfaults within the range 50° +/- 15° (reverse-sinistral faulting). The inversion procedure follows the approach of Delouis et al (2000). The subfault slip rate functions are based on the parametrization of the source function from Nabelek (1984) and represented by a sequence of 4 isoscele triangular time windows of variable height, mutually overlapping, each having a duration of 5s and spaced by 2.5s. The maximum duration of slip on a subfault is thus 12.5s. A longer duration does not produce measurable changes in the modeling. The maximum allowed slip on each subfault is 7m and we verified that a higher bound for the maximum slip did not improve the modelling. Nevertheless, by limitating the maximum slip we favor smoother slip distributions. The rupture initiates at the hypocenter and propagation is represented by the
subfault rupture onset times that are allowed to vary within the range defined by two bounding rupture velocities, 1.8 and 3.5 km/s. We assume a simple half space crustal model with $V_p = 6.6$ km/s and $V_s = 3.8$ km/s. We solve for the following free parameters at each subfault: the rupture onset time, the amplitudes of the four elementary triangular time windows, and the rake.

Furthermore, we invert for one static offset to calibrate the InSAR data since the differential interferograms provide only a relative measurement of the deformation and the actual location of the “zero displacement fringe” is unknown. Due to the baseline errors accounting for a phase ramp along the track direction (azimuth -166°) shown by the additional fringes perpendicular to the satellite track in the northern part (Fig. 2a), two more free parameters have to be introduced in the inversion of InSAR data: the position along track of the ramp axis and the ramp slope. These are inverted using the InSAR data only and then kept fixed in the joint inversion (Fig. 2b). The displacement values are corrected (increased or decreased) depending on their location with respect to the ramp axis location.

The static near-source ground displacement is modeled using the dislocation formulation of Savage (1980). Each subfault is represented by a dislocation surface embedded in an elastic half-space.

The synthetic seismograms produced by simple shear dislocations (double couple) point sources at the center of each subfault, are computed using the ray theory for stations located at teleseismic distances (Nabelek, 1984).

The inversion is performed with a simulated annealing algorithm that allows for a quasi-global exploration of the model space. The primary cost function is the root mean square (RMS) error of the data fit normalized by the observed data. Both data sets have unit weight in the joint inversion. We consider that an appropriate weighting of the different data sets in the joint inversion should distribute the slight degradation of the data fit (with respect to the separate inversions) quite evenly on the different data sets. Since both the InSAR and the teleseismic data remain correctly fitted with equal weights in the joint inversion, we deem that the importance of both data sets is well balanced for unit weights.

An additional cost function is used in the InSAR and joint inversions to penalize solutions with seismic moments higher than a reference moment which is the Harvard Centroid Moment Tensor (HCMT) in the case of real data. This is particularly important since slip on subfaults located far from the InSAR data points may be easily overestimated.
4. Resolution tests with synthetic data

We investigate the space-time resolving power of the different data sets using synthetic slip maps, giving a special attention to the impact of an incomplete coverage of the displacement field from the InSAR data. The data processing and fault model are the same as in the inversions of the actual data. The number of free parameters, their bounding values, and the moment minimization are also kept identical to those that are used for the Peru event.

The synthetic data (Fig 4-5) are generated with a five patches model (Fig. 6a). A low level of random noise has been added to the data in order to include small deviations from the exact solutions: maximum +/-1 cm on the InSAR data (compared to the 2.8 cm of the ERS1/2 radar wavelength) and a modification of +/-10% on the amplitudes of the teleseismic data also randomly time-shifted by a maximum of +/-1 s. The synthetic slip map is composed by five asperities, each comprising four subfaults slipping 170 cm, plus 40 cm of slip at the hypocenter. Rake is 50° everywhere. The total scalar seismic moment is 5.57\times10^{20} \text{N.m}.

Rupture velocity is constant and equal to 2.7 km/s. All the elementary source time functions have the same amplitude. The rupture initiates at the same hypocenter location as for the actual earthquake.

We examine the resolving power for the teleseismic, InSAR, and combined data sets. Two different coverages of the synthetic InSAR data are considered (Fig. 4): one corresponds to the actual coverage (called “narrow”), and the other takes into account most of the inland ground displacement (called “wide”). The synthetic InSAR and joint inversions are performed for those two distributions.

Table 2 displays the RMS and scalar seismic moment resulting from the inversions. No moment minimization is used in the teleseismic inversion since the seismic moment remains in any case lower than the reference moment of the synthetic model. The resulting slip maps are presented on Figure 6.

When inverted alone, the teleseismic data tend to spread out slip over the entire fault plane (Fig. 6b). The asperities are nonetheless quite well resolved at the beginning of the rupture, in the vicinity of the hypocenter, but imaging deteriorates in the later SE part. This shows the difficulty to model the latest parts of the P and SH pulses when all the contributions from the source overlap in the teleseismic signals. Furthermore in the real case, the P and SH pulses are disturbed by the contribution of crustal complexities which are not taken into account in our tests.
The separate InSAR data inversions (Fig.6c-6d) give more accurate locations of the slip patches, but with the actual “narrow” coverage, the NW part of the rupture, around the hypocenter, is not properly retrieved, and slip at the NW bottom corner of the rupture is largely overestimated. This is related to the decreasing resolving power of geodetic data with distance to the subfaults. With the “wide” coverage of InSAR data, resolution improves, especially in the NW part. The joint inversions (Fig.6e-6f) combine the resolutions of the separate teleseismic and InSAR inversions, and retrieve the slip maps more completely and accurately.

The resolution of timing has also been assessed (Fig.7). Although all the inversions that incorporate the teleseismic data provide a good estimate of the rupture velocity, combining the two data sets (joint inversion) helps to retrieve the patches at the end of the rupture with the correct timing (hypocentral distance >100km, SE half of the fault plane). This confirms that the trade-off between rupture timing and slip location that affects seismological data inversions can be reduced through the addition of geodetic data.

5. Inversion results

The results of the separate and joint inversions of the real data sets are presented in Figures 8, 9, 10 (data modeling), and 11 (slip maps). The fit of the geodetic data (Fig.8) is good for both the InSAR and joint inversions, though it degrades slightly in the north with distance from the fault. Misfit does not exceed 5cm in both cases. Modeling of the teleseismic data in the separate and joint inversions are similar and there is no systematic misfit (Fig.9-10). As in the case of the synthetic tests, the teleseismic data tend to underestimate the seismic moment if compared to the HCMT moment (4.57 $10^{20}$N.m). Table 3 displays the RMS values for the inversions and shows that fit degrades only slightly when data are combined.

The slip distributions (Fig. 11) show common features along strike. The rupture propagated to the SE of the hypocenter. In preliminary inversions, we tested fault models extending more toward the NW but data do not require it.

In the teleseismic inversion, two areas of higher slip are found, one at the hypocenter and the other about 50km more to the SE along strike. The InSAR inversion does not retrieve any slip in the vicinity of the hypocenter, as expected from the resolution tests. The region of main slip (up to 680cm) is located at about 50km of the hypocenter to the SE and slightly downdip. Significant slip occurs also farther to the SE along strike.
The joint inversion combines the characteristics of the previous separate inversions: slip at the hypocenter, maximum slip 50km SE of it, and still large slip 50km further to the SE, i.e. 100km away from the hypocenter. A slip area previously found downdip (-40km, -30km on the fault plane) in the InSAR inversion is now translated updip to a neighbor subfault. The resolving power of the data is not sufficient to discriminate between these two neighbor locations for the peak slip. In all, the main slip area is about 100km long with most of slip restricted to a 60km wide strip along strike in the depth range 20 to 40km depth.

The time evolution of the rupture from the joint inversion is presented in Figure 12 with the corresponding overall source time function (STF). Rupture extends unilaterally along strike, i.e. from NW to SE, in about 60s. The STF displays three main pulses of energy: The first one, between 0 and 10s corresponds to the rupture of the hypocentral asperity. The second centered at 22s is associated with the breaking of the asperity that is found with the highest slip value. The second centered at 22s is associated with the breaking of the asperity that is found with the highest slip value. The third and major one, between 30 and 45s, involves a very large area of slip in the SE half of the rupture.

### Table 2. Misfit values and scalar seismic moment resulting from the inversions of the synthetic data

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Reference moment 5.67 10^20 N.m

### Table 3. Misfit values and scalar seismic moment resulting from the inversions of the actual data

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Harvard moment 4.57 10^20 N.m
6. Slip models and previous studies of the 1996 Peru earthquake

Previously, the rupture history of the 1996 Nazca ridge earthquake has been studied using teleseismic data by Swenson and Beck (1999), and by Spence et al (1999). In Figure 13, we compare the different inversion results. All studies show rupture propagation toward the SE but the results differ in the estimation of the seismic moment and in the details of the slip distribution. Comparison of Swenson and Beck (1999) and Spence et al (1999) shows that teleseismic data may be explained by seismic moments varying by a factor 4 or more (3.47 $10^{20}$ to 1.5 $10^{21}$N.m). These previous studies locate the maximum energy release (Swenson and Beck) or the main slip patch (Spence et al) about 100km to the SE of the hypocenter, where we also find the main contribution of slip and maximum moment release in the source time function (Fig.12). However, we find that the main asperity is located only 50km to the SE of the hypocenter. In addition to be controlled by the teleseismic data, this location is also constrained by the InSAR data in our study.

7. Relation between the 1996 earthquake and the coastal deformation

As mentioned before, the Mw=7.7, November 12, 1996, and the Mw=8.1, August 24, 1942, events damaged the same region (Chatelain et al, 1997) which displays a characteristic feature (Fig.14a) : a narrow uplifted coastal cordillera reaching 1000 m within the peruvian fore-arc, in front of the Nazca ridge (Hsu, 1992, Machare and Ortlieb, 1992). The probable repetition of such earthquakes at the subduction interface of the Nazca ridge, may have some tectonic implications and imaging precisely the distribution of slip for the last Peru event is essential to shed light on the contribution of those earthquakes to the coastal uplift of the fore-arc relief. A similar situation has also been proposed where an aseismic ridge is subducted under Costa-Rica (Marshall and Anderson, 1995).

We computed the vertical displacement at the surface due to our preferred earthquake model and compared it to the main geomorphological features of the coastal area (Machare and Ortlieb, 1992). The resulting vertical bulge is mostly located offshore and elongated about 200 km along strike (Fig.14a). It presents a maximum uplift (>50cm) above the area of maximum slip on the fault plane. Onshore, the uplift involves the South-Eastern part of the Pisco basin region where the vertical displacement occurs mainly along coast in the coastal cordillera region. The profile of coseismic uplift along the coast is very similar to
the topographic profile of the coastal cordillera, to the elevation of the uplifted marine terraces (Hsu, 1992, Machare and Ortlieb, 1992), and to the bathymetric cross section of the Nazca ridge (Fig.14b). We propose that the elevation profiles of the coastal cordillera and of the marine terraces reflect the cumulative effects of the repetition of earthquakes similar to the 1996 event as suggested by Swenson and Beck (1999). We note that the complexity of the earthquake source produces variable uplift along the coastal cordillera. Recurrent events may change the uplift pattern except if the source geometry is determined by the shape of the subducted Nazca ridge. It is therefore difficult to estimate a reliable coseismic uplift rate since we have access only to the slip map of a single event. Further, additional informations about the deformation field occurring in the interseismic period, which generally involves coastal subsidence, would be required to give an average return period for those earthquakes and to confirm their characteristic behaviour.

8. Conclusion

We observed that accurate slip map can be obtained from the joint inversion of remote access data which constrain both the timing of the rupture (teleseismic waveforms) and its spatial distribution (teleseismic+InSAR data). We show that the joint inversion of InSAR and teleseismic data resolve: 1) the location of the main asperities over the entire fault plane, despite the incomplete coverage of the surface displacement from the InSAR data and 2) the main characteristics of rupture timing. The south-eastern and central parts of the rupture plane are essentially resolved by the InSAR data, and the north-western part by the teleseismic data. The teleseismic data alone do not retrieve the same source model constrained with the addition of InSAR data. We show that synthetic tests are essential to understand the imaging power of individual data sets. The 1996 event (Mo=4.4 10^{20}N.m) ruptured the subduction interface on the southern flank of the Nazca ridge, as did the former Mw=8.2, 1942 earthquake. It propagated unilateraly toward the SE, the main asperity being located 50km away from the hypocenter. The slip area extended for more than 100km in total, and mostly between 20-40km depth. The similarity between coseismic uplift, topography of the coastal cordillera, and marine terraces suggests that coastal deformation may be related to the repetition of events similar to the 1996 earthquake.
Acknowledgments. This work was supported by ESA (cr. ERS-A03-194) and ETH zürich. Part of this work was carried out at the Jet Propulsion Laboratory, California Institute of Technology, under contract with the National Aeronautics and Space Administration (NASA). We express our gratitude to P. Briole, D. Wald, and H. Lyon-Caen for reviewing the manuscript. Figures have been generated using the Generic Mapping Tools (http://gmt.soest.hawaii.edu/). Topographic and bathymetric data were available through the web site: http://topex.ucsd.edu/marine_topo/mar_topo.html. This is a publication No. 1268 of the Institute of Geophysics, ETHZ.
Figure 1. Location of the $M_w=7.7$, November 12, 1996, Peru earthquake in front of the Nazca ridge. The position of the epicenter is indicated by the black dot. The dashed rectangle shows the surface projection of the fault plane model (this study) and the shaded area the extent of the rupture area obtained from the joint inversion of teleseismic and InSAR data. Also indicated is the Harvard CMT focal mechanism ($M_o=4.57 \times 10^{20}$ N.m), the rupture zone of the 1942, $M_w=8.2$, earthquake and the southern extent of the $M_w=8.1$, October 3, 1974, earthquake (Dorbath et al, 1990). The arrow indicates the relative velocity between the Nazca and the south american plates (7.6 cm/yr).
Figure 2. Location of the InSAR data and surface projection of the fault plane model. The epicenter is indicated by the black triangle while the small dots correspond to the centers of the subfaults. Each fringe (black-gray-white gradation) corresponds to 5 cm of displacement in the direction toward the satellite. (a) raw differential interferogram represented here with a 5 cm fringe cycle. (b) interferogram corrected with the phase ramp. White points represented on the interferogram are the data points used for the inversions.

Figure 3. Topographic cross section A-A’ of Fig.2, with the position the fault plane model (thick line).
Figure 4. Surface projection of the synthetic fault model with five slip patches (dark gray areas) used for the resolution tests together with the synthetic and inverted InSAR data. The light gray areas correspond to the “narrow” (left) and “wide” (right) coverages of InSAR data. The contour lines are drawn every 5 cm of ground displacement in the satellite line of sight. Modeled means from the joint inversions.

Figure 5. Focal mechanism and P waveforms modeling from the joint inversion with the synthetic 5 patches slip model and the “narrow” InSAR coverage.
Figure 6. Slip maps from the resolution tests. The centers of the subfaults are shown by the dots, the epicenter by a triangle. Rakes are shown by arrows whose length is proportional to slip and which indicate the underthrusting direction.
Figure 7. Timing of the rupture on the subfaults versus the distance to the hypocenter. Rupture timing of the subfaults from the reference model is represented by the empty squares. Black filled squares represent inversion results for subfaults whose location coincide with the synthetic slip patches and whose slip amplitude reaches at least half of the reference slip. Square size is proportional to the total slip of the subfault. Also shown are the maximum and minimum rupture time lines defined by the rupture velocities \(v_r\) 1.8 and 3.5 km/s respectively. The minimum slip duration corresponds to a single time window (5 s) and the maximum slip duration to 4 overlapping windows (12.5 s). The light dashed line represents the reference timing for a rupture velocity of 2.7 km/s.
Figure 8. Fit of the InSAR data for separate InSAR (left) and joint inversion (right). The observed displacement contour lines (corrected from the static offset) are shown in blue. The modeled InSAR data are shown in red. Three cross sections show how well the data are fit. The residual interferograms formed by the difference between observed and modeled data are displayed at the bottom.
Figure 9. Average focal mechanisms of the P and SH waves determined by the teleseismic inversion, and the waveform fitting.
Figure 10. Average focal mechanism of the P and SH wave determined by the joint inversion, and the waveform fitting.
Figure 11. Slip maps from the separate and joint inversions of the real data. Also shown is their corresponding average focal mechanism. The arrows whose size is proportionnal to slip indicate the underthrusting direction. Hypocenter is indicated by a triangle.
Figure 12. Source time function (right) and snapshots of the rupture history (left) for the 1996 Peru earthquake obtained with the joint inversion. Snapshots display cumulative slip in successive 5 s long time windows. Hypocenter is indicated by the triangle. (1), (2) and (3) indicate the three main pulses of moment release.
Figure 13. Solutions for the rupture of the 1996 Peru earthquake given by different studies. Swenson and Beck (1999) used an iterative, multi-station, pulse stripping method (Kikuchi and Kanamori, 1982; Kikuchi and Fukao, 1985) which locates the maxima of energy release on the fault plane. Spence et al (1999) performed a teleseismic inversion for the slip history based on a multi-time window method (Hartzell and Heaton, 1983, 1986; Hartzell and Langer, 1993). The dashed rectangle corresponds to the area of our fault model. Also shown is the respective focal mechanism for each inversion.
Figure 14. (a.) Tectonic sketch of the Pisco basin region and final vertical ground displacement (centimeters) inferred from the joint inversion results (thin contour lines). The dashed rectangle corresponds to the surface projection of the fault plane. (1) and (2) are two cross sections. (b.) Elevation profile across the coastal cordillera (1) its corresponding vertical coseismic uplift, and bathymetric profile (2) of the Nazca ridge. The finite deformation as figured by elevation of highest late Cenozoic marine surfaces is shown by the dotted curve (Machare and Ortlieb, 1992).
Chapter 3

Slip history of the 1999, October 16, Mw=7.1, Hector Mine earthquake (California) from the inversion of InSAR, GPS and teleseismic data

J. Salichon¹, P. Lundgren², B. Delouis¹*, D. Giardini¹

Submitted to BSSA
Abstract

The slip distribution of the Mw=7.1 Hector Mine earthquake, California, is investigated in space and time by jointly inverting broadband geodetic data and teleseismic data constrained by reported surface offsets. The geodetic data consist of a dense network of GPS data and synthetic aperture radar (SAR) interferograms both ascending and descending satellite tracks. According to the complexity of the earthquake rupture, we use a fault model of four partially overlapping segments discretized into 3x3km$^2$ patches. The source parameters on each sub-fault are estimated using a non-linear inversion scheme based on a simulated annealing method to explore the parameter space. We allow slip to vary in amplitude, direction, duration and rupture velocity. We first explore the space and time resolution of each data set and their combination using synthetics. The geodetic data map completely the coseismic deformation field of the Hector Mine earthquake and strongly constrain the spatial distribution of the final slip and the fault geometry. The spatial resolution is expected to be best for the upper ten kilometers over the entire fault model. The teleseismic data inversion exhibit a rather bad spatial resolution related to complexity of the rupture. The joint inversion of both geodetic and seismic data provides a robust slip history that fits simultaneously the independent data sets in space and time. The Hector Mine earthquake is a right lateral strike slip event that presents a heterogeneous distribution of slip at shallow depth (<12km). Most of the seismic moment is released in the vicinity of the hypocenter over two overlapping segments. The total seismic moment is 5.8x10$^{19}$ N.m with a peak displacement amplitude of about 6m. The velocity rupture is comprised between 2 and 3km/s for a total duration of about 15 seconds over a short extent (<50km).

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Introduction

The Eastern California shear zone was the area of two temporally and spatially close large events during the last decade. The 1999, Mw=7.1, Hector mine event (Fig.1) occurred about 20km north-east of the 1992, Mw=7.3, Landers earthquake and was recorded by regional and worldwide networks as well as remote sensing satellites. At least 40km of surface breaks were reported (USGS et al, 2000) with up to 5-6 meters of right lateral surface slip. The 1992, Landers event was the first case where the interferometric synthetic aperture radar (InSAR) technique was applied to a large earthquake (Massonet et al, 1993). The Hector Mine earthquake is a recent example for which coseismic ground deformation surrounding the rupture was almost entirely mapped by the ERS1/2 satellites in addition to numerous GPS stations. Accordingly, we expect to constrain the location of the major slip areas and robustly evaluate the entire final coseismic fault slip from those geodetic data.

Inverting teleseismic broadband data routinely provides spatial and temporal slip distribution of seismic sources, but the results may be biased by a trade-off between the timing of the rupture and the slip location. The earthquake source parameters are estimated here by inverting jointly InSAR, GPS data and teleseismic (P- and SH-) waveforms following the approach of Delouis et al (2000, 2002). By solving simultaneously for the main characteristics of the faulting in space and time with the condition that the final static slip fits the geodetic data, we reduce the teleseismic trades-off and better constrain the slip history. Different studies have been carried out to assess the spatial slip distribution of the 1999 Hector Mine earthquake with multiple data sets: GPS, InSAR, and InSAR azimuthal offset (Jónsson et al, 2002; Simons et al, 2002; Price et al, 2002). Temporal and spatial aspects were estimated combining regional strong motion, teleseismic broadband and geodetic data sets (Ji et al, 2002; Kaverina et al, 2002). Here, we perform a joint inversion of teleseismic broadband and geodetic data incorporating two SAR interferograms of the Hector Mine earthquake. We assume that the teleseismic waveforms are rich enough to constrain the whole slip history of the Hector Mine earthquake with the condition of being jointly inverted with geodetic data. Determining the complete rupture history of an earthquake with data provided only by remote instrumentation (broadband stations and ERS1/2 satellites) is important to address the study of earthquake kinematics worldwide. The two SAR scenes (ascending and descending satellite orbits) can be considered as independent geodetic data since each scene
represents the coseismic deformation projected onto the satellite line of sight with different azimuths. Using both the ascending and descending InSAR scenes allows a better description of the displacement field (Fujiwara et al., 2000; Fialko et al., 2001) and therefore better constrains the final slip distribution on the fault. The GPS measurements (Agnew et al., 2002) provide an additional constraint on the 3D coseismic displacements and potentially improve the final static slip distribution model. Trade-offs between source parameters inherent to the use of one single component of ground deformation can thus be limited (Wright et al., 2001). Finally, we consider the available fault surface offsets (USGS et al., 2000) to constrain coseismic displacement at the free surface.

1. DATA

Teleseismic data

The seismological data have been collected through the Global Seismological Network (IRIS/USGS, IRIS/IDA). They consist of P- and SH- waveforms recorded at teleseismic distances (30°<Δ<90°) (Fig.7). The seismograms have been deconvolved from the instrument response, integrated to obtain ground displacement and bandpassed from 0.8Hz to 0.005Hz (P-waves) and 0.4Hz to 0.01Hz (SH-waves). We model the first 50s of 17 P-wave signals and the first 70s of 13 SH-wave signals that are reasonably well distributed around the source, except for the southern quadrant where only a few stations were available. In the waveform modelling, we include an initial first weak P-arrival observed at several stations a few seconds before the most significant P-onset (e.g. stations CPUP and SJG on fig. 7).

SAR data

The differential SAR interferograms were computed from the SAR data provided by the European Space Agency’s ERS1/2 satellites. The data have been processed into differential interferogram using the ROI_PAC software at the Jet Propulsion Laboratory. Each differential interferograms represents the projection of the displacement vectors onto the radar line of sight. For the ERS satellites, the ground incidence angle at mid swath is approximately 23° from vertical. The sets of data consist of a descending scene (track 127), i.e. the satellite com-
ing from the northeast, and an ascending scene (track 077), i.e. the satellite coming from the southeast. The descending interferogram is computed from two SAR scenes acquired on 1999/09/15 and 1999/10/20 (spaning 35 days). The ascending interferogram is computed from the SAR data acquired on 1995/11/12 and 1999/11/21 (spaning ~4 years) (Table 1). In each case the perpendicular baselines for the interferograms were small, implying that the potential contribution of uncertainties in the USGS 30m, digital elevation model (DEM) much less than one fringe cycle (28 mm).

We assume that the pre-seismic displacement field for both interferograms is small compared to the coseismic displacement as indicated by the pre-seismic interferograms inspection (Fialko et al, 2001). The descending scene (t127) (Fig.2.a) samples essentially the coseismic displacement plus the immediate postseismic displacement in the four days following the earthquake. Accordingly, this InSAR scene describes mostly the coseismic deformation. The ascending scene (Fig.2.b) samples the coseismic and about one month of the postseismic deformation. Jacobs et al (2002) present a detailed analysis of the near field postseismic deformation from the ERS interferometry associated to the Hector Mine 1999 event. The inferred slip distribution reaches an estimated maximum post seismic slip of about 10cm near the surface (<3 km) that are small compared to the several meters of observed coseismic surface rupture on the same fault segment. Furthermore, since we use the offset rupture measurements to constrain the slip on the upper 3km of the fault model, we reduce even more the importance of the post seismic slip on the inversion. The descending scene t127 and the ascending scene t077 are sampled with 1642 and 1634 points respectively as shown in the Figure 2. We use a sampling algorithm to preferentially select points where the deformation gradient is higher.

GPS data

The coseismic data are from Agnew et al (2002) and we refer to this article for details relating to the processing of the GPS data. We invert for the three displacement components of 77 GPS stations (Figure 8) that are well distributed throughout the ruptured area.
Table 1: ESA precision Orbits

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</tr>
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</tr>
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<td>ERS2</td>
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<td>23979</td>
<td>19991121</td>
</tr>
</tbody>
</table>

2. FAULT MODEL

The reported surface breaks (USGS et al, 2000) and the InSAR interferograms indicate a complex segmentation of the rupture involving the Lavic Lake and the Bullion faults (Fig.1). We build a fault model based on those observations and on the aftershock locations (Hauksson, 2000; 2002). The north part of the rupture is characterized by a Y shape as primarily shown by the aftershock distribution. The northern and southern limits of the fault model are determined from the SAR interferograms. The closures of the north and south fringes (Fig.2) are assumed to characterize the ends of the main rupture, giving a total rupture length of about 50km. Thanks to the overall rich information, the complex geometry of the fault is assumed to be well constrained.

We use a fault parameterization allowing for a multiple-segment fault geometry, variable slip and variable slip velocity (Delouis et al, 2000; 2002). The fault is simulated by 4 segments (Fig. 3) with their respective upper limits coinciding with the surface breaks. Each segment is subdivided into subfaults with dimensions of 3km along strike and dip for a total of 144 subfaults. The respective strikes and dips are held fixed and the rake angle is allowed to vary within the range 175+/-20°. The individual source time functions are represented by 5 isosceles triangular functions of variable amplitude with 1 second duration, mutually overlapping for a maximum slip duration of 6 seconds. We set the maximum allowed slip on the subfault to 9 meters. The rupture propagation initiating at the hypocenter is represented by the subfault slip onset times that are allowed to vary within the interval defined by two rupture velocities, 1.8 and 3.5km/s. In total, we invert at each subfault for the rupture onset time, the amplitudes of
the triangular functions and the rake, i.e. 1008 free parameters. In addition, since the geodetic data provided by the two InSAR interferograms are only relative measurements (no reference points allowing for an absolute measurement), we invert for one static offset for each InSAR scene in order to calibrate the geodetic measurements.

The static near field displacements are computed with the dislocation formulation of Savage (1980) where each subfault is represented by a dislocation surface embedded in an elastic half-space. For the computation of the seismograms, we use ray theory (Nabelek, 1984). The synthetic seismograms for stations located at teleseismic distances are produced by simple shear dislocations (double-couple) at point sources located at the center of each subfault for a layered velocity model proposed by Jones and Helmberger (1998) (Table 2). The model space is explored in a quasi-global manner with a simulated annealing algorithm that is more likely to converge to the global minimum of the cost function than linearized inversion schemes (Imhle and Ruegg, 1997). The cost function is defined as the root mean square (RMS) misfit of the data normalized by the observed data. An additional cost function is used to minimize the total scalar moment (Mo). This allows the seismic moment to be close to that observed for the Harvard Centroid Moment Tensor. A reliable seismic moment estimated from long period waves (HCMT) is a strong constraint to obtain a physically reasonable solution (Das and Kostrov (1990)). In addition the minimization penalizes solutions that find slip at depth where the resolution power of the geodetic data is low. The moment minimization was required for the geodetic data inverted separately. No moment minimization is used for the teleseismic and joint inversions.

Furthermore, trial runs indicated that local slip complexities may appear in the slip model which are not constrained by the data. Therefore, inversions are stabilized by adding a smoothing constraint with respect to the inversion approach of Delouis et al (2002). We add a cost function that minimizes the standard deviation of the total slip between adjacent subfaults (8 at most), resulting in a smoothing of the slip distribution. Different runs for different smoothing coefficients were performed. The smoothing constraint is finally set to exhibit the most stable features in the slip distributions.
The different data sets are equally weighted in the joint inversion scheme after verifying that different weight combinations did not produce any important change in the overall data fit or in the slip distribution. The hypocenter position on the fault model has been set after several trial and error runs with inversions around the localization of Hauksson (2000, 2002). The hypocenter is basically controlled by the teleseismic body wave modelling and we locate it on the north-eastern segment (Fig. 3) at a depth of 7.5 km. In the inversions of the real data, slip values at the shallowest subfaults of the model are constrained to remain close (+/- 50 cm) to the maximum fault offset observed at the surface (indicated in Fig.3).

<table>
<thead>
<tr>
<th>Thickness (km)</th>
<th>P-wave velocity (km.s⁻¹)</th>
<th>S-wave velocity (km.s⁻¹)</th>
<th>Density (g.cm⁻³)</th>
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<tbody>
<tr>
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<td>half-space</td>
<td>7.85</td>
<td>4.40</td>
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</tr>
</tbody>
</table>

*Table 2: Layered velocity model (Jones and Helmberger, 1998)*
3. INVERSION RESULTS

Synthetic tests

The inversion algorithm is tested with synthetic InSAR, GPS and teleseismic data sets. These tests allow us to appraise the performance and limitations of the algorithm and to estimate its spatial and temporal resolution. The inversions are performed with the same parametrization, data configuration and data processing used for the actual data inversions, but without surface offset constraints. Noise has been added to the synthetic data sets: +/- 5cm of random displacement for the InSAR data, +/- 3cm and +/- 6cm of random displacement on the horizontal and vertical GPS measurements respectively, +/- 10% of signal amplitude plus a random time-shift of +/- 1s for the teleseismic data. The synthetic slip model is characterized by three slip patches of regular shape composed of 12 subfaults and distributed on the fault plane as shown in Figure 4.a. We set the subfault displacements of two of the slip patches to 600cm and the third to 300cm. The rake angle is uniform and set to 175°. The rupture velocity is constant and equal to 2.7km/s. The elementary triangular source time functions have the same amplitudes. The rupture initiates at the same hypocenter location as for the inversion of the actual data. We perform different inversions in order to study the improvement provided by each of the data sets, separately and then jointly (Fig. 4.b-h).

The spatial resolutions of the two separate InSAR scenes (Fig. 4.b-c) are similar and fairly good at shallow depths (down to 12km) in spite of sparse spurious slip on the edges of the fault segments. The joint inversion of the two interferometric scenes (Fig. 4.d) gives improved results, constraining slip patches at their proper locations down to 12km. The inversion of the GPS measurements (Fig. 4.e) provides a good appraisal of the original slip location similar to the inversion of the InSAR data.

The spatial resolution of the teleseismic data is low (Fig. 4.f) compared to that of the InSAR and GPS data. The proper location the slip patches is not retrieved. The slip distribution is smeared over a large area. Patches are shifted without an accurate control over their azimuth and depth positions because of the trade-off between rupture timing and slip location. More specifically, location of slip on overlapping segments 1 and 3 is not constrained.
The final static slip distributions (Fig. 4.g-h) retrieved by the joint inversions (SAR+Broadband, SAR+GPS+Broadband) are mostly constrained by the geodetic data at shallow depth. The resolution of both InSAR and GPS data sets being similar (down to 12km) over the entire fault plane, no major improvements is observed when GPS data are added in the joint inversion. Beneath 12km, the decreasing spatial resolution of the different data is shown by the downdip spreading of the slip distribution.

From these resolution tests, we infer that the geodetic data well constrain the entire spatial slip distribution of the Hector Mine earthquake to depths down to 12km. Given the good spatial resolution of the geodetic data, the main contribution of the teleseismic records in the joint inversions is to constrain the temporal evolution of the rupture process. In Figure 5, we compare the temporal evolution of the synthetic rupture with that resulting from the joint inversion. The main features of the temporal evolution of the rupture are well retrieved though rupture velocity estimates for significant displacements oscillate around the correct velocity.

**Inversion for the Hector Mine earthquake**

We present in Figures 6, 7 and 8, the data fits resulting from the GPS, InSAR and teleseismic joint inversion. The final slip distributions obtained from the separate and the joint inversions with constraints on the surface offset are presented in the Figure 9. No large misfit is observed and the RMS values resulting from the joint inversion are only slightly degraded with respect to the separate inversions (Table 3). All data types require the rupture to extend mainly over an area south of the hypocenter (segment 3), plus a slip patch located just north-west of it (segment 2). The maximum extent of the rupture is about 30km toward the south and 15km toward the north, starting from the hypocenter. The combined inversion of the InSAR scenes (Fig.9.a) retrieves slip at shallow depth between 0-12km where we expect a good spatial resolution as shown by the resolution tests. The inversion of the GPS data presents a similar but rougher slip distribution (Fig. 9.b) probably because of the sparse location of the GPS stations compared to the evenly distributed InSAR measurements.
The teleseismic inversion (Fig. 9.c) locates significant slip patches on the segments 2 and 3 in agreement with the geodetic inversions though slip is located at 10-15km depth and disconnected from the surface offsets. Slip retrieved on the bottom northern edge of the segment 3 is more likely mislocated as shown by the synthetic tests. Small slip is found on segment 4 but it is still not well constrained by the teleseismic data. Low slip at the hypocenter is associated with the smooth start of the bodywaves (Fig. 7).

The slip distribution resulting from the joint inversions (Fig. 9.d-e) of InSAR+Broadband data and InSAR+Broadband+GPS data are similar, close to the final slip distribution obtained by inversion of the geodetic data alone, without significant additional misfit of the teleseismic data with respect of the single teleseismic data inversion. The contribution of the GPS data to the joint inversion is a noticeable increase in the overall slip amplitude.

The slip distribution obtained by the joint inversion of the geodetic and broadband data exhibits three main slip areas, or main patches (a1,a2, and a3, Fig.9e), all located at shallow depths (0-12km). According to the resolution tests for the joint inversion (Fig.6.h), we interpret slip at the bottom of segment 2 as a poorly resolved deep slip feature. Displacement at the hypocenter is small and in agreement with the low energy initial pulse of the P-waveforms as discussed before.

Surface offsets and slip at depth are in agreement and maximum slip is reached on the main slip patch a1 with a final slip of 593cm. The overall seismic moment over segments 1 and 2 represents about 42% of the total seismic energy release (2.46 $10^{19}$ N.m). Segment 3 (slip patch a2) represents about 39% of the total moment. The last 19% of seismic moment is located on segment 4 that includes the last slip patch a3.

The time evolution of the rupture is represented with slip distribution snapshots within 2 seconds windows (Fig. 10). The overall time evolution exhibits a bilateral rupture with a total duration of about 15 seconds. No significant slip is retrieved during the first 2 seconds of the rupture. Slip occurred simultaneously over many parts of the fault, especially between 6 and 12 seconds after the rupture initiation. Rupture velocity remains close to 2km.s$^{-1}$ in the shallow part of the fault where most of the slip occurred, but it increases on segment 4 to reach about 3km.s$^{-1}$. Rupture propagation on this segment appears to be different from that of the other segments as earlier noted by Ji et al (2002). Rupture slows down at the southern edge of seg-
ment 3 while slip is simultaneously triggered on segment 4 (snapshots between 4 and 8s, fig.10). The surface breaks mapped in the corresponding area (USGS et al, 2000, Fig.1) indicate a rather complex segmentation implying a degree of complexity in the rupture pattern that we do not solve for in our study.

The source time function is shown in Figure 10. The seismic energy is mostly released within an interval of 12 seconds. The short duration is explained by the bilateral rupture. Its shape is simple and quite symmetric presenting a maximum around 8 seconds after rupture initiation when slip occurred simultaneously on several fault segments, and two secondary maxima of energy release.

<table>
<thead>
<tr>
<th></th>
<th>Teleseismic data</th>
<th>Combined InSAR data</th>
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<th>Teleseismic and InSAR data</th>
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<td></td>
<td></td>
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<td>5.8</td>
<td>5.9</td>
<td>5.7</td>
<td>5.8</td>
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</tbody>
</table>

Harvard seismic moment: $5.98 \times 10^{19}$ N.m
Conclusion

The overall rupture duration of the Hector Mine earthquake is about 15 seconds, and the total seismic moment release is $5.8 \times 10^{19}$ N.m close to the Harvard seismic moment ($5.98 \times 10^{19}$ N.m). Three main slip areas are found for a maximum peak slip of about 600 cm. Rupture velocities vary from one segment to another. Rupture propagates first on segments 1, 2 and 3 where most of the slip occurred (about 80% of the total seismic moment), with a rupture velocity of about 2 km/s, and ends on segment 4 (about 20% of the total moment) with a rupture velocity of about 3 km/s. The rupture history that we inferred from the joint inversion of teleseismic, GPS, and InSAR data succeed in recovering the details of the spatio-temporal slip distribution of the Hector Mine earthquake. The teleseismic data alone cannot resolve robustly the slip distribution of such a complex fault rupture where we observed that about 80% of the total seismic moment was released within a small 20 km long area. Only an approximate location of slip on the segments could be recovered. The same poor resolution of the teleseismic data was observed for the 1992, Landers earthquake (Wald and Heaton, 1994). The static final slip distribution being strictly constrained by the geodetic data in the joint inversion of InSAR, GPS and teleseismic data, the inferred rupture history is comparable to the ones derived from the near field seismic data (Fig. 11) (Ji et al, 2002; Kaverina et al, 2002). The different parameterizations, data types, bandwidth of the waves used, and the inversion procedures contribute to produce differences in the results. Timing differences in the bodywaves arrival picks, various hypocenter locations, and the low resolution of the teleseismic data (Figure 9.c) add uncertainties in the estimation of the event’s duration. Furthermore, the local seismic data used in the previous studies with a higher frequency content are more sensitive to the dynamics of the earthquake. In our study, discrepancies in the final slip distributions inferred from the inversion of the geodetic data and from the inversion of the teleseismic data alone point out the need to systematically provide a combined analysis of multiple independent data sets. In addition we observed that having access to the complete static coseismic deformation field is a major improvement to enhance the reliability of the inferred spatio-temporal history of a large event usually studied using only far-field teleseismic broadband stations.
Figure 1. Hector Mine and Landers earthquakes locations. The position of the Hector Mine epicenter (used in our inversion) is indicated with the grey filled circle. The surface projection of the segments of the fault model are indicated by the rectangles. The Hector Mine aftershocks locations (October 1999-December 1999) from Hauksson (2002) are indicated with black dots. Half arrows indicates the sense of the strike-slip motion. Harvard Centroid Moment Tensor are indicated for both earthquakes. Compressional quadrants are filled in grey. The fault segments are indicated by the numbers between brackets.
Figure 2. Observed unwrapped differential interferograms with 2 different line of sights: (a) Descending track (T127) and (b) ascending track (T077). Black arrows indicate the azimuth and direction of the ERS satellites. Ground to satellite displacements are represented here by white-to-black fringes and each corresponding to 10cm of displacement. Grey dots correspond to the InSAR data points used in this study. Rectangles are the surface projection of the segments of the fault segments and their inner black dots are the surface projection of the center of the 144 subfaults of the model.
Figure 3. Side view, from the east, of the fault model used in this study with respect to the surface rupture location. The grid corresponds to the subfault positions. The triangle indicates the position of the hypocenter (our inversion). Thick black rectangles delimit the fault segments with their respective strike (str.) and dip. Surface offsets in centimeters used to constrain (within +/- 50 cm) slip on the uppermost subfaults of the model are indicated at the top of segments. The heavy grey line with an irregular shape drawn in between the segments represents the surface trace of the rupture. Dashed lines connect the fault segments to the corresponding surface trace. At the bottom of the Figure, the subfault parametrization and its moment rate elementary function (see text for details) are illustrated.
Figure 4. Slip maps from the resolution tests. Arrows indicate the direction of slip of the hanging wall, their length being proportional to slip magnitude. Dots indicate the centers of the subfaults. From top to bottom: (a) synthetic model, (b) and (c) InSAR descending and ascending inversion, (d) combined InSAR data inversion, (e) GPS data inversion, (f) broadband teleseismic inversion, and (g) joint inversion of InSAR and Broadband teleseismic data, (h) InSAR, GPS and broadband teleseismic data joint inversion. Dashed circles indicate the position of the patches of the synthetic model. Triangle indicates the position of the hypocenter.
Figure 5. Time evolution for the synthetic model (left) and for slip model resulting from the joint inversion of teleseismic + InSAR+ GPS data (right) given at intervals of 2 seconds (cumulative slip maps). Dashed circles correspond to rupture velocities of 2 and 3 km.s$^{-1}$, solid lines correspond to the rupture velocity of the synthetic model (2.7 km.s$^{-1}$) (left) and that of the joint inversion (right).
**Figure 6.** Time evolution for the synthetic model (left) and for slip model resulting from the joint inversion of teleseismic + InSAR+ GPS data (right) given at intervals of 2 seconds (cumulative slip maps). Dashed circles correspond to rupture velocities of 2 and 3 km/s, solid lines correspond to the rupture velocity of the synthetic model (2.7 km/s) (left) and that of the joint inversion (right).
Figure 7. Waveform fitting of the teleseismic broadband from the joint inversion of the teleseismic, InSAR and GPS data. Observed (solid lines) and computed (dashed lines) waveforms are displayed for the P-waves (top) and the Sh-waves (bottom). For each signal, the stations azimuth (az) is indicated. Right: Azimuthal distribution of the P- and Sh- signals used for the inversions. The focal mechanism is the one at the hypocenter. Compressional quadrants are filled in grey.
Figure 8. Coseismic GPS horizontal component (Fialko et al., 2002) shown with the black arrows and computed horizontal component inferred from the joint inversion of teleseismic GPS and InSAR data (White arrows). Surface breaks are indicated with dotted lines. Surface projection of the fault model is indicated by the rectangles filled in grey.
Figure 9. Slip maps for the inversion of the real data with surface offset constraints (view from the east): (a) for the combined InSAR data inversion, (b) for the GPS data inversion, (c) for the teleseismic inversion, (d) for the joint inversion of the InSAR and teleseismic data, (e) for the joint inversion of InSAR, GPS and teleseismic data. Arrows indicate the direction of slip of the hanging wall, their length being proportional to slip magnitude. Fault segment boundaries are indicated by vertical dashed lines. The corresponding location of the fault model with respect to the surface rupture is indicated at the bottom of the Figure (e). Seismic scalar moments (Mo) are indicated for each inversion.
Figure 10. (left) Time evolution (2s cumulative slip maps) of the rupture from the joint inversion of InSAR, GPS, and teleseismic data with constraint on surface offsets. At the bottom is displayed the final static slip map. Dotted lines correspond to rupture velocities of 2 and 3 km.s\(^{-1}\) presented as references and thick lines to the rupture front in the joint inversion model. (right) Overall source time function (evolution of moment rate with time) of the Hector Mine earthquake from the joint inversion of GPS, InSAR and teleseismic data.
Figure 11. Comparison between the final coseismic slip map resulting from the joint inversion of InSAR, GPS, and teleseismic data with constraint on the surface offsets (this study, left) with that of Ji et al (2002) obtained from the joint inversion of teleseismic, strong-motion, GPS data, and measured surface offset (right). Superimposed dashed rectangles and contours indicate the position of the Ji et al model and their inferred peak slip areas respectively. Respective position of the fault models (surface projections) and of the hypocenters (triangles) used in the inversions are displayed at the top of the Figure with the reported surface rupture. Mo indicates the total seismic moment found for each joint inversions. Arrows indicate the direction of slip with respect to the hanging wall and are proportional to slip. Also indicated are the respective percentages of seismic moment releases on the fault segments (Seismic moment values are between brackets).
Chapter 4

Joint Inversion of InSAR, GPS, Teleseismic and Strong Motion Data for the Spatial and Temporal Distribution of Earthquake Slip: Application to the 1999 Izmit Mainshock

B. Delouis¹, D. Giardini¹, P. Lundgren² and J. Salichon¹
Abstract

The space-time distribution of slip of the 17 August 1999 Izmit earthquake is investigated by inverting synthetic aperture radar (SAR) interferometry and Global Positioning System (GPS) data together with teleseismic broadband and near-field strong-motion records. Surface offsets are used as an added constraint. A special emphasis is given to the analysis of the resolution of the different data sets. We use a four-segment finite fault model and a nonlinear inversion scheme, allowing slip to vary in amplitude, direction and duration, as well as variable rupture velocity. From the inversion of synthetic data, we find that the best spatial resolution can be expected in the upper half of the fault model (above 12 km) where coverage of the interferometric SAR data is good (western half of the rupture), and near the GPS and strong-motion stations. Teleseismic data is found to have a lower resolution that is more evenly distributed over the fault model. The joint inversion of all the data sets has an increased resolving power compared to the separate inversions, and gives a more robust description of the space and time distribution of slip. Our study shows the importance of resolution tests in evaluating the reliability of earthquake kinematic models and it confirms that an excellent fit of a single kind of data does not necessarily imply a good retrieval of the kinematic properties of an earthquake. The Izmit rupture, which is almost pure right-lateral strike-slip faulting, is dominated by the bilateral breaking of a central asperity located between 29.7E (about 10 km west of the city of Gölcük) and 30.4E (eastern margin of Sapanka Lake), with slip reaching 6 to 8 m is in the depth range 6 to 12 km. The western termination of the rupture is found near the city of Yalova but large slip ends around 29.7E (about 10 km east of Hersek Delta). A second area of large slip is required by all the data sets further East toward the city of Düzce, between 30.7E and 31.1E (Karadere and Düzce faults). This eastern slip zone, which is separated from the main central asperity by an area of very reduced slip, is less well constrained by the data. However, a strong motion station near the city of Düzce helps to locate a high slip patch near 31.1E in the depth range 6 to 12 km. The total seismic moment resulting from the joint inversion is $2.4 \times 10^{27}$ dyne-cm. Most of the energy release occurred in a short time, less than 15 sec, corresponding to the bilateral breaking of the central asperity. Rupture propagation is relatively uniform and fast toward the West, with a rupture velocity close to 3.5 km/sec. Propaga-
tion of large slip toward the East is initially slower but it accelerates during a short time interval about 10 sec after rupture nucleation. Eastward progression then slows down to less than 2 km/sec after 15 sec and rupture almost vanishes in amplitude ~20 sec after initiation. Rupture propagation then proceeds on the easternmost Karadere and Düzce fault segments, east of 30.7E, from 22 to ~50 sec. Supershear rupture propagation is not required to model the waveforms considered in this study. The hypocenter of the Düzce earthquake, which occurred three months later (12 November 1999, Mw=7.2), is located in the immediate vicinity of the easternmost slip patch of the Izmit earthquake.

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Nota Bene: This study with multiple data sets follows the previous study combining teleseismic broadband data and InSAR interferograms:

Introduction

The Izmit earthquake of 17 August 1999 (Mw=7.5 to 7.6) is the last and westernmost event of a sequence of large earthquakes which migrated from East to West along the North Anatolian Fault during the last century (Barka, 1999; Toksöz et al., 1999; Stein et al., 1997). The earthquake produced a large amount of destruction and a large number of casualties in several cities of western Turkey, in particular Gölcük, Izmit, Yalova, Adapazari and Gölyaka (MSK intensities X, General Directorate of Disaster Affairs, Ankara), which are located in the vicinity of the causative fault segments delineated by approximately 120 km of surface faulting (Barka, 1999; Reilinger et al., 2000a; Fig. 1). The objective of this work is to determine the space-time distribution of slip for the Izmit mainshock. A robust assessment of the slip history is a fundamental requirement for further studies aimed at relating the source properties to damage, computing accurate stress transfers, characterizing the seismic cycle and hazard related to the North-Anatolian fault, and deriving dynamic rupture models. By determining the space distribution of slip we intend also to address the open questions of the nature of the western (offshore) and eastern termination of the rupture, of the location of the main asperities along the 1999 Izmit rupture, and of the relation between surface and deep slip. Following Kanamori (1981), we associate asperities to areas on the fault which accumulated large stress before the earthquake and which underwent large slip during the main shock.

Although source time functions (STF) are reliably and routinely obtained from teleseismic broadband data, recent studies aimed at deriving kinematic models for large earthquakes have shown that slip maps for a given event may vary significantly depending on the kind of data used (e.g. Cohee and Beroza, 1994a; Wald and Heaton, 1994; Cotton and Campillo, 1995, for the Landers earthquake). The same is true for the 1999 Izmit earthquake. Although bilateral rupture propagation is present or implicit in all models, the dominant slipping zone, or major seismic moment release, has been located either west of the hypocenter (Yagi and Kikushi, 2000; Bréger et al., 2000) or east of it (Tibi et al., 2001; Bouchon et al., 2000). By inverting GPS data, Reilinger et al. (2000b) found three main patches of slip, one west of the hypocenter and two east of it, all confined to the upper 10 km of the crust. In other models large slip patches reach a depth of 15 to 20 km (Yagi and Kikushi, 2000; Bouchon et al., 2000). We will show that a large part of these discrepancies have their origin in the limited power of resolution of the different data sets when used separately. According to Bouchon et al. (2000) part of the
rupture propagated toward the West at supershear velocity, and we will consider this possibility.

It has been shown that slip inversions using seismological data only are likely to be affected by the trade-off between rupture timing and slip location (e.g. Cohee and Beroza, 1994b). A way to solve this problem is to combine seismological and geodetic data. Geodetic data, like InSAR or GPS, provide an independent constraint on the space distribution of slip. InSAR data have been used in a number of earthquake studies (e.g. Massonnet et al., 1993; Feigl et al., 1995; Meyer et al., 1996; Ozawa et al., 1997; Wright et al., 1999; Peltzer et al., 1999; Wright et al., 2001; Feigl et al., 2001). Hernandez et al. (1999) adopted a two-step approach in order to limit the space-time trade-off. They first constrained the slip distribution of the 1992 Landers earthquake with the geodetic data (InSAR + GPS) to recover in a second step the temporal details of rupture propagation by inverting the strong motion records. The implicit assumption with this procedure is that the geodetic data are rich enough to constrain by themselves the slip distribution. In most cases, this is probably too optimistic an assumption since measuring points are restricted to the earth surface and generally offer an incomplete coverage of the deformed area. Our approach is to invert jointly geodetic and seismological data, including InSAR, to retrieve simultaneously the temporal and spatial characteristics of the rupture. In the present paper, we give special emphasis to the analysis of the spatial resolution and we show how each independent data set, geodetic, teleseismic and strong motion, help to control different parts of the fault model. Finally, we show how the joint inversion of the combined data sets gives the most robust picture of the rupture, and how separate inversion of individual data sets may provide only a partial and often poorly resolved image of the source process. In a previous paper we performed the first simultaneous inversion of InSAR and teleseismic data (Delouis et al., 2000). Here, and for the first time, a simultaneous inversion of InSAR, GPS, teleseismic, and strong motion data with constraint from surface breaks is carried out.
1. Data

Access to most of the geodetic, teleseismic, strong-motion and surface offset data was provided very quickly after the Izmit earthquake, and this opened an almost unique possibility to combine a large amount of complementary data.

**Geodetic data: InSAR and GPS**

The differential SAR interferogram, shown on Figure 12, was computed using InSAR data from the European Research Satellites (ERS-1 and 2) acquired on August 13, 1999 and September 17, 1999 by the European Space Agency (ESA). The raw data (ERS-2, track 157, frames 815 and 801) were processed using the JPL-Caltech developed ROI_PAC software. Flattening, the removal of fringes in the initial interferogram caused by Earth's curvature and orbits of the two satellite passes, and computation of the final differential interferogram is calculated in two steps. First a differential interferogram is computed through the removal of a simulated interferogram based on precise orbit information from the European Space Agency and a precise digital elevation model (DTED 90 m posting digital elevation model for NW Turkey from the U.S. National Imagery and Mapping Agency [NIMA]). This differential interferogram is then unwrapped, using a branch-cut algorithm with bridges used to connect regions separated by low coherence. Then the baseline parameters are estimated using a SVD least squares inversion and a new simulation (topography and Earth curvature effects) is removed from the initial interferogram to more precisely flatten the interferogram.

This ERS pair has the shortest coseismic pair possible for this earthquake (35 days) with the first image acquired four days before the earthquake. The perpendicular baseline (separation) of the two ERS satellite orbits was small (~30 m), minimizing sensitivity to topographic uncertainties in the digital elevation model.

As can be observed on Figure 12, the coverage of the InSAR data is good for the western part of the rupture but InSAR data are lacking near the fault in the eastern part. In the SAR inversions, we use a subset of 1300 points from the interferogram in order to limit the number of data points and to sample more densely the deformation were it is larger (near the fault). Resampling was carried out in two steps. First we kept only every fifth point from the original interferogram. Then we defined different intervals of distance to the fault and within each interval we kept one point every 7, 50, 120, or 200, depending on the distance to the fault. The
actual InSAR data points used in the inversions are displayed on Figure 4. We also processed the available ERS-1 pair (990812-990916) with precise orbits for coseismic deformation and found that the difference with the ERS-2 differential interferogram was small.

Any postseismic deformation taking place within the SAR frame area during the month following the İzmit earthquake is included in our SAR interferogram. In their joint inversion of InSAR, GPS and Spot measurements to estimate the slip distribution of the İzmit mainshock, Feigl et al. (2001) corrected the ERS-1 SAR interferogram from the postseismic deformation. However, GPS measurements in the source area show that the postseismic deformation of the ground surface over a 75 day period following the mainshock did not exceed about 5 cm (Burgmann et al., 2000; Reilinger et al., 2000b). Therefore it was small with respect to the meter level satellite measured coseismic displacements (see observed profiles on Fig. 12). This postseismic deformation has been attributed to aseismic fault slip of as much as 0.43 m on and below the coseismic rupture (Reilinger et al., 2000b), a small amount compared to the 6 to 8 m of maximum coseismic slip, and our interferogram covers only 40% of that 75 day period.

Another source of errors which can affect the results of InSAR inversions for the İzmit earthquake is related to tropospheric artifacts (Reilinger et al., 2000b; Feigl et al., 2001). Reilinger et al. (2000b) discarded the possibility of finding a coseismic slip model explaining both the differential GPS and InSAR data, noting that an error of a few tens of mm in SAR range displacement at 50 km from the fault may alter the estimate of slip on the fault by several meters. However, we found that this problem can be avoided by down-weighting or down-sampling the InSAR data far from the fault. In addition, the use of teleseismic data makes us less dependent on far fringes to constrain deep slip.

Coseismic GPS data (Fig. 3) are from Reilinger et al. (2000b), and we refer to that paper for details concerning the processing of the GPS data.
Telesismic data

Broadband seismograms recorded at teleseismic distances by the IRIS and GEOSCOPE networks were deconvolved from the instrument response, integrated to obtain displacements, and equalized to a common magnification and epicentral distance. The data were bandpassed from 0.8 Hz (P waves) or 0.4 Hz (SH waves) to 0.01 Hz. We model the first 60 sec of 13 P wave signals and the first 80 sec of 12 SH signals, well distributed in azimuth around the source (Fig. 2).

Strong motion data

We use six strong-motion accelerograms from the Earthquake Research Department of the General Directorate of Disaster Affairs, Ankara. Most of the records are from standard Kinematics SMA-1 instruments, except station SKR which is equipped with digital Geosys GSR-16 instrument with a force balance sensor. The stations are located within 50 km from the earthquake fault (Fig. 1). The accelerograms were integrated twice to get the ground displacements. The SMA-1 data are bandpassed from 0.1 Hz to 0.5 Hz. The low-cut frequency was chosen in order to limit the influence of long period noise enhanced by the double integration. The 0.5 Hz high-cut frequency aims at reducing high frequencies which we cannot intend to model with a simple crustal model. In order to avoid large differences in amplitude among the stations, station SKR was also band-pass filtered, between 0.05 and 0.5 Hz, though the lowest frequencies may be retrieved down to DC at that station, as shown at the end of this paper. Station SKR which has absolute timing was aligned in absolute time. For those stations without absolute timing, we performed a series of single station inversions in order to determine the minimum time shift required to obtain a good fit between the synthetic and observed waveforms. The resulting time shifts were used to align those stations and we verified a posteriori that they were adequate in the multi-station and multi-data set inversions. The vertical components of the strong motion records, which are more difficult to model with a simple crustal model, are excluded from the inversion, with the exception of station SKR which is located very close to the fault. The NS component of station SKR did not work properly and cannot be used.
Surface offset data

The average offset values observed along the coseismic rupture were released very quickly to the scientific community (IPGP, ITU, USGS and SCEC, 1999, Web sites) and were confirmed afterwards with the publication of more detailed observations (e.g. Honkura et al., 2000; Barka et al., 2000; Awata et al., 2000). Surface breaks extend for about 120 km, from the southern bank of the Izmit Bay near Gölcük up to the longitude of the city of Düzce in the East (Fig. 1). The earthquake was almost pure right-lateral strike-slip with maximum offsets reaching 4 to 5 m west of Gölcük and East of Sapanka Lake (Fig. 3). Localized normal faulting connecting two strike slip segments has also been described east of Gölcük (Barka, 1999).

2. Fault model and inversion procedure

We use a uniform parameterization allowing for a multi-segment finite fault geometry, variable slip and variable rupture velocity. Our approach combines the multiple time-windows formulation (Olson and Aspel, 1982; Hartzell and Heaton, 1983) with a nonlinear inversion scheme incorporating rupture onset times as free parameters. A quasi-global exploration of the model space is carried out with a simulated annealing algorithm. Examples of use of simulated annealing in variable-slip inversions can be found in Ihmlé (1998) and Lundgren et al. (1999). Simulated annealing, which allows us to fully solve nonlinear inverse problems without dependency on a starting model, requires the a priori definition of bounding values for the free parameters. Convergence of the simulated annealing procedure is based here on the simultaneous minimization of the RMS misfit and of the total seismic moment. If different data sets are used, the RMS error is a weighted average of the normalized RMS errors of the individual data sets. In the joint inversion, InSAR, GPS, teleseismic and strong-motion data have weights 1.0, 1.0, 1.2, and 1.0 respectively, the 20% higher weight for the teleseismic data being motivated by the higher normalized misfit resulting from the teleseismic inversion. We do not use a specific smoothing constraint but the minimization constraint on the total seismic moment results in relatively smooth models and strongly reduce spurious slip. The weight of the moment minimization with respect to the RMS minimization is not the same in all the inversions. To estimate an optimum (maximum) value for that weight for each data set, i.e. to find...
solutions minimizing the seismic moment, we carried out several preliminary runs increasing the weight, until the degradation of the data fit became easily perceptible by inspection.

The earthquake rupture is simulated by four segments whose strike and intersection with the surface coincide with the main surface breaks (Fig. 1 and Table 1). The epicenter is located near the city of Izmit (40.76N, 29.97E, Kandilli Observatory), but the source depth is not well constrained (Toksöz et al., 1999). The scalar seismic moment from the Harvard CMT catalogue was initially $2.1 \times 10^{27}$ dyne-cm (Mw=7.5) but it was later revised to $2.88 \times 10^{27}$ dyne-cm (Mw=7.6). Estimates of fault dip and hypocentral depth were obtained from preliminary broadband teleseismic modeling using the method of Nabelek (1984). Values for those parameters, as well as for the length of the westernmost and easternmost segments of the rupture, were confirmed or adjusted after a series of trial and error inversions including both the geodetic and seismological data. To discretize the rupture, each fault segment is subdivided into rectangular subfaults with dimensions 7.5 km along strike and 4.5 km along dip, resulting in a total of 115 subfaults. Strikes, dips and segment dimensions are held fixed (Table 1) but the rake at each subfault is allowed to vary within the range $180° \pm 20°$ (right-lateral slip). The evolution of slip amplitude with time is assumed to be uniform over each subfault. Subfault slip rate functions are represented by a sequence of six isosceles triangular time windows of variable height. Time windows are 2 sec long and mutually overlapping. With six time windows the maximum duration of slip at any single point of the fault model is 7 sec. Bounding values on individual time window heights are positive and such that final slip may reach a maximum of 8 m on each subfault. We verified that a longer allowed duration of slip or a larger allowed amplitude of slip does not improve the modeling. The rupture initiates at the hypocenter and overall propagation is accounted for by the rupture onset times of the subfaults. Rupture onset times are allowed to vary within a certain range defined by two bounding rupture velocities, 1.8 and 3.5 km/s. This latter maximum bounding rupture velocity corresponds to the shear wave velocity of our crustal model (see below). In summary, we estimate the following free source parameters, at each subfault: the rupture onset time, the amplitudes of the six triangular time windows, and the rake. In addition, we invert for two static offsets required to calibrate the InSAR data north and south of the fault. We do not include a parameter to allow for possible orbital tilt across the image.
Subfault strong-motion and teleseismic contributions are computed by summing the responses of 15 equally spaced point sources delayed in time according to the varying source-to-station position and to the propagation of a local rupture front on each subfault. This local propagation occurs in the model at 75\% of the shear wave velocity. We verified that the results are not sensitive to the choice of this locally fixed rupture velocity. In any case, at a broader scale, the inversion of the rupture onset times allows for a variable rupture velocity. Ground motion contributions for both dip-slip and strike-slip components on each subfault are computed and stored before inversion. During the inversion, they are combined to obtain motions corresponding to any rake.

In some inversions, slip values at the shallowest subfaults of the model were constrained to remain close to the maximum offset observed at the surface shown on Figure 3. In the simulated annealing algorithm, this is carried out by using bounding values close to the observed surface offset for slip on the shallowest fault segments. Small scale complexities in surface breaks are deemed to be a surface effect and are ignored.

Synthetic seismograms produced by simple shear dislocation (double-couple) point sources are computed using ray theory for stations located at teleseismic distances (Nabelek, 1984). For near-field stations, we use the exact analytical expressions including both the far-field and near field waves given by Johnson (1974) for an elastic half space. More details are given in Legrand and Delouis (1999). Near-field static displacements (for InSAR and GPS data) are computed using the formulation of Savage (1980), each subfault being represented by a dislocation surface embedded in an elastic half space. We assume a simple half space crustal model, with \( V_p = 6.0 \) km/s, \( V_s = 3.47 \) km/s, and the Lamé elastic constants \( \lambda = \mu = 3.3 \times 10^{10} \) Pa. With this simple velocity structure, we limit ourselves to relatively low-frequency waveform modeling (see previous section) and to near-source strong motion stations (less than 50 km away from the fault).
3. Resolution tests with synthetic data

We present first the application of the inversion scheme to synthetic data which allows us to assess the resolving power of the different data sets. The fault geometry, the number of free parameters and their bounding values, as well as the data configuration and processing are the same for these synthetic simulations and for the actual data inversions. The synthetic SAR interferogram and the synthetic GPS data are displayed in Figure 4 and examples of observed and synthetic waveform data are presented in Figure 5. Synthetic data are generated using the synthetic slip model presented in Figure 6-top (slip distribution) and 7 (time evolution). A low level of noise was added to the synthetic data before inversion, which varies between +/- 3 and +/- 5 cm for the InSAR and GPS data respectively, and +/- 10% of the amplitude of the signals for the teleseismic and strong-motion data. Moreover, synthetic seismograms were randomly time-shifted by up to +/- 1 sec. The synthetic slip model is characterized by five asperities of regular shape distributed on the four fault segments (numbered a1 to a5, Fig. 6-top). The slip angle (rake) varies slightly around a central rake value of 180 degrees from one asperity to the other but since the real data do not require strong variations in slip direction we do not emphasize resolving this parameter. Heterogeneity in the rupture propagation has been incorporated in the following way. From the hypocenter, at asperity a1, the rupture velocity is 3.4 km/s toward the East and 2.1 km/s toward the West. The rupture velocity used to compute rupture times for asperities a2 and a3 are 2.6 km/s and 2.1 km/s respectively. All rupture times on segments 3 and 4 (easternmost segments) correspond to an average rupture velocity of 2.6 km/s with an added delay of 2 sec. Figure 7 displays the time evolution in the synthetic model. A discretization into 2.5 sec time intervals is used to limit the number of snapshots, although this relatively coarse discretization introduces some imprecision in the appearance of the rupture velocity.

Figure 6 presents the slip distribution resulting from the separate inversions of the different data sets as well as that resulting from their simultaneous joint inversion. No constraint on the surface offset is used. The InSAR inversion locates relatively well the position of the asperities, with the exception of the easternmost one (a5), but their shape is approximate and the resolution clearly degrades with depth. The deep root of asperities a2, a3 and a4 are not resolved. Asperity a5 is translated into some mislocated slip in the deep part of segment 3. This latter effect can be attributed to the lack of coverage for the InSAR data in the eastern part of the
model. The GPS inversion locates all the asperities, including a5, as a result of the relatively good distribution of GPS station around the rupture. However, as in the case of the InSAR inversion, the shape of the asperities is only approximately retrieved and there is almost no resolution in the lower part of the model. Slip above and below the hypocenter tends to be underestimated by the GPS inversion. The spatial resolution of the teleseismic inversion appears to be low. Localized asperities are transformed into spread out slip, and the actual shape of the asperities cannot be retrieved. The strong-motion inversion does a relatively good job in retrieving the position and shape of asperities in the upper part of the model close to stations SKR, IZT and DZC, which are located near the fault. However, the presence of station IZT just above the hypocenter is not sufficient to provide a good resolution in the western part of the hypocentral asperity a1. The strong-motion data provide almost no resolution in the deep part of the model. The spatial resolution clearly improves with the joint inversion. The deep root of asperity a1 is better resolved though the deep parts of the other four asperities are imprecisely defined. Figure 5 shows how the waveform misfit remains small even in the case of the joint inversion. Misfits in Figure 5 are representative of the way data are matched at all the stations. Misfit is even smaller for the separate inversions. Since slip distributions resulting from separate inversions give an incomplete picture of the actual slip distribution (Fig. 6), we argue that an excellent fit of a single kind of data does not necessarily imply a good retrieval of the kinematic properties of an earthquake rupture. This is true in the case of the Izmit earthquake, even though the area covered by the InSAR and GPS data is quite wide, the azimuth distribution of teleseismic stations is good, and the distribution of strong-motion stations is quite favorable with six stations well distributed within less than 50 km of the fault.

Summarizing the above observations, we expect the best spatial resolving power in the upper part of the model where coverage of the InSAR data is good, and near the strong-motion and GPS stations. A low spatial resolution but more evenly distributed over the whole model is expected from the teleseismic data. Combining all the data sets improves spatial resolution, though it is not possible to retrieve all the details of the asperities. Seismic moment from the InSAR, GPS, teleseismic, strong-motion and joint inversions are respectively 66%, 76%, 77%, 76%, and 90% of that of the input synthetic model. The underestimation of the seismic moment may result from the minimization constraint that we apply on the seismic moment in the inversion process but it results also from the partial spatial coverage of the data, especially
in the case of the InSAR data. The joint inversion combines the coverage of the different data sets and gives the best estimates of the seismic moment. It is clear that individual data sets provide only a very partial imaging of the slip distribution, and that the joint use of multiple data is required.

We restrict our analysis of the temporal evolution of the rupture to the result of the joint inversion. Only minor differences are observed when comparing Figures 7 (synthetic timing) and 8 (inverted timing). Westward propagation is slow until \( t = 12.5 \) s, then becomes faster, ending finally at about \( t = 25 \) s. Eastward propagation is fast until \( t = 7.5 \) s, then slow between 15 and 27.5 s. The rupture starts on segment 3 at \( t = 25 \) s and reaches the eastern edge of the model at about \( t = 42.5 \) to 45 s. The main characteristics of rupture timing are well retrieved by the joint inversion.

We also included variations of the slip duration in the synthetic model. Figure 9 displays how the slip time histories (slip curves) at individual subfaults are matched by the joint inversion. Slip curves were obtained by integrating the slip rate time functions. We limit our analysis of the resolution of the slip curves to those subfaults where slip exceeds 4 m in both the synthetic and inverted models. Though some large slip discrepancies are observed (Fig. 9), the duration of slip is well retrieved in most cases.
4. Slip history of the Izmit earthquake

The slip distributions resulting from the separate and joint inversions of the real data are presented in Figure 10 and 11, respectively without and with the surface offset constraint. They share some common features: 1) the absence of slip at depth in the western part of the model, 2) high slip values (4 to 8 m) in the central portion of the model from about 10 km west of Gölcük up to the area immediately east of Sapanka Lake, and 3) significant slip on the eastern-most segments. Without the constraint on the surface offset, slip at the surface is overestimated at several places, for all four data sets. The InSAR and GPS models are characterized by shallow slip mostly restricted to the upper half of the model from Lake Sapanka to the West. In the East the InSAR models exhibit deep slip but we can infer that it is not resolved due to the poor coverage of InSAR data in this area, as confirmed by the synthetic tests. The GPS models locate an asperity in the vicinity of stations DKMN and KDER which are the stations closest to the eastern segments. The teleseismic models exhibit large slip values at greater depth, in particular in the hypocentral region. Although the synthetic tests indicated the tendency for the teleseismic inversion to spread out the distribution of slip, the teleseismic models from the inversion of the real data are dominated by a concentrated central pattern of slip, located on segments 1 and 2. This is taken as an indication that the principal asperity of the Izmit earthquake is indeed quite concentrated in space. The strong-motion models exhibit the most complex slip pictures, but from the synthetic tests we can infer that deep slip is poorly resolved and that the best resolution is to be expected near the strong-motion stations SKR, IZT and DZC, and more precisely east of the hypocenter in the upper part of segment 2. The effect of adding the constraint on the surface offset is essentially to restrict the highest slip values (6 to 8m) to the depth range 6 to 12 km.

The joint models combine the best resolved parts of the parameter sets estimated from individual data sets. The western part of the rupture is controlled mostly by the InSAR, GPS and teleseismic data, the central part by the teleseismic and InSAR, while control in the eastern part comes essentially from the strong motion and GPS data. Data modeling for the joint inversion with constraint on the surface offset is presented on Figures 12 to 15. Misfit for the InSAR data does not exceed 10 to 15 cm in the near fault area (see profiles on Figure 12). The overall match of the GPS coseismic vectors is good (Figure 13), though we note small angular misfits at the two largest displacement stations SISL and SMAS, as well as some underestima-
tion of the displacement amplitude at KDER. Noticeable misfit of the teleseismic amplitudes is essentially limited to the CHTO and TATO stations for the P-waves and to the ATD, COLA and MAJO stations for the SH-waves (Fig. 14). The misfit of the strong-motion signals is remarkably small (Fig. 15). Table 2 lists the normalized RMS values and seismic moments for the separate and joint inversions. Normalized RMS (NRMS) is dimensionless. A NRMS value of 0.0 would mean perfect matching. A NRMS value of 1.0 would mean bad matching, corresponding for instance to cases where the computed data would be zero everywhere or would be two times larger than the observed ones. NRMS values larger than 1.0 would imply either anticorrelation or computed data more than two times larger than the observed ones. As expected, the addition of new data sets in the inversion results in a slight increase of the RMS values, i.e. of the misfit. However, as can be observed in Figures 12 to 15, the overall fit of all the data sets remains excellent in the joint inversion. Seismic moment from the different inversions varies between 2.0 and 2.6 x 10\(^{27}\) dyne-cm, 2.4 x 10\(^{27}\) dyne-cm being the most robust estimates resulting from the joint inversion. Our estimates of the seismic moment is hence 17% lower than the Harvard CMT scalar moment (2.88 x 10\(^{27}\) dyne-cm).

<table>
<thead>
<tr>
<th>Data set</th>
<th>RMS_SAR</th>
<th>RMS_GPS</th>
<th>RMS_TELE</th>
<th>RMS_SM</th>
<th>Mo (dyne cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SAR</td>
<td>0.07</td>
<td></td>
<td></td>
<td></td>
<td>2.0 x 10(^{27})</td>
</tr>
<tr>
<td>SAR + SOC</td>
<td>0.08</td>
<td></td>
<td></td>
<td></td>
<td>2.0 x 10(^{27})</td>
</tr>
<tr>
<td>GPS</td>
<td></td>
<td>0.07</td>
<td></td>
<td></td>
<td>1.7 x 10(^{27})</td>
</tr>
<tr>
<td>GPS + SOC</td>
<td></td>
<td>0.10</td>
<td></td>
<td></td>
<td>1.6 x 10(^{27})</td>
</tr>
<tr>
<td>TELE</td>
<td></td>
<td></td>
<td>0.41</td>
<td></td>
<td>2.3 x 10(^{27})</td>
</tr>
<tr>
<td>TELE + SOC</td>
<td></td>
<td></td>
<td>0.41</td>
<td></td>
<td>2.3 x 10(^{27})</td>
</tr>
<tr>
<td>SM</td>
<td></td>
<td></td>
<td></td>
<td>0.25</td>
<td>2.4 x 10(^{27})</td>
</tr>
<tr>
<td>SM + SOC</td>
<td></td>
<td></td>
<td></td>
<td>0.27</td>
<td>2.6 x 10(^{27})</td>
</tr>
<tr>
<td>JOINT</td>
<td>0.12</td>
<td>0.13</td>
<td>0.51</td>
<td>0.36</td>
<td>2.2 x 10(^{27})</td>
</tr>
<tr>
<td>JOINT + SOC</td>
<td>0.12</td>
<td>0.16</td>
<td>0.51</td>
<td>0.37</td>
<td>2.4 x 10(^{27})</td>
</tr>
</tbody>
</table>

A notable feature of the strong-motion models is the absence of slip in the hypocentral area, in contrast with the teleseismic and joint models (Fig. 10 and 11). The small effect on the strong-motion seismograms of large slip in the hypocentral area is illustrated by Figure 16 and is at the origin of the low resolution of the strong-motion data in this area of the model. We note, however, the relatively large and constructive effect of hypocentral slip at station IZT. Teleseismic broadband signals are on the average more sensitive to slip in the hypocenter area, as illustrated in Figure 16 for two representative stations. As shown by Figure 15, the strong-
motion data are equally well fitted with large slip at the hypocenter. The slip model resulting from the joint inversion with constraint on the surface offset (Fig. 11-bottom) we consider the most robust to describe the source process. The western termination of the rupture is located near the city of Yalova, at about 29.3E, but large slip ends near 29.7E. In the East, each type of data require the presence of consequent slip on the two easternmost segments (Karadere and Düzce faults, Fig. 1), but clearly it is the strong-motion station DZC situated near the city of Düzce which provides the most accurate constraint on the easternmost part of the rupture. If station DZC is removed from the data sets, slip stops at about 31.0E. On the other hand, we verified that with station DZC included, a joint inversion with a fault model incorporating only the first three segments would produce an accumulation of slip at the eastern edge of segment 3 (Karadere fault), confirming that slip should be located more to the East on segment 4 (Düzce fault). In order to match the waveforms at DZC, our model with surface offset constraint requires 5 to 6 m of slip between 6 and 12 km depth at about 31.1E (Figure 11-bottom). The rupture of the Mw=7.2, 12 November 1999 Düzce earthquake (Tibi et al., 2001; Ayhan et al., 2001) suggests a lower dip for the Düzce fault segment (segment 4). With a fault dipping 55 to 65 degrees to the North (instead of 85 degrees), the asperity at 31.1E is still found by the joint inversion, but with lower slip values (3 to 4 m), a decrease expected since the fault plane becomes closer to station DZC. Since this asperity depends primarily on station DZC, it is important to stress that the dominant period of the near field displacement waveforms that we are matching are too long (> 5 - 10 s) for a basin effect to bias our modeling of station DZC. An area of very reduced slip is found between 30.4E and 30.7E, forming a noticeable gap in the slip distribution cross section.

From Gölcük to Sapanca Lake, there is a good overall correspondence between surface offsets and slip at depth (Fig. 11-bottom) though in detail surface slip does not mimic slip at depth. On the other hand, deep slip associated with the easternmost asperity has almost no expression at the surface.
The time evolution of the rupture for our preferred model is shown in Figure 17. Rupture propagation toward the West is relatively homogeneous and occurs at a velocity close to 3.5 km/s. Note that with the snapshot representation used here the estimation of the rupture velocity is imprecise at the beginning of the rupture, especially within the first 5 sec. Propagation toward the East appears to be initially slower than toward the West. This is reflected by the fact that before t=7.5 sec large slip amplitudes are almost restricted to the area at, and west of, the hypocenter. Around t=10 sec, large slip progresses much more rapidly toward the East, producing a strong directivity effect contributing to the high amplitudes observed at the strong motion station SKR. However, this fast progression occurs well within the supershear velocity boundary of 3.5 km/s when considering the total rupture history starting at the hypocenter. To further test the need for a supershear rupture, we carried out a specific joint inversion where the maximum bounding rupture velocity was 5 km/s instead of 3.5 km/s. The effect of relaxing the constraint on the rupture velocity was to produce minor patches of slip propagating faster than 3.5 km/s but without improvement of the waveform fit. We conclude that supershear rupture is not required. From 10 to 15 sec, the rupture still propagates relatively fast toward the East, at 3 km/s or more. Then, between 15 and 20 sec, it slows down to less than 2 km/s. Between 20 and 22.5 sec, the rupture almost vanishes in amplitude. This is corroborated by the absence of significant slip in the easternmost part of segment 2 as observed in the final distribution of slip (Fig. 17-bottom). Rupture starts to propagate on segment 3 at about t=22.5 sec. Rupture velocity across segments 3 and 4 appears to be less well constrained than across the two other segments. After t=35 sec, propagation stops but slip continues in some places, especially at the easternmost asperity.

The overall source time function (STF) for the complete rupture is shown in Figure 18. The peculiarity of the STF is that most of the energy is released in the first 15 sec, a short duration for an earthquake of that size. As shown by Figure 17, this is explained by the bilateral rupture of the central asperity located on segments 1 and 2 (approximately between 29.7E and 30.4E, Fig. 11-bottom), which is almost completed in the first 15 sec. The main characteristics of the overall STF are found also from the individual teleseismic and strong-motion inversions, indicating that even though slip maps from individual data sets may differ significantly the STF tends to be more robust.

Slip histories at individual subfaults indicate in most cases a short duration of slip and the 7
sec available for slip are fully used at one subfault only (Figure 19). At several places, slip of 7 to 8 m occurs in less than 4 sec. The waveforms modeled in this study are dominated by periods longer than the individual slip durations. However, as shown by Legrand and Delouis (1999), the long periods which dominate in signals produced by finite faults in the near-field are built by constructive and destructive interference of seismograms produced by small portions of the fault, which have themselves a higher frequency content. A proper estimation of the slip duration (rise time) at individual subfaults is required in order to reproduce those low frequencies. Moreover, we have seen that the synthetic tests carried out in this paper (Fig. 9) indicate that slip curves may be retrieved with some degree of confidence, at least where slip is large. Though individual slip histories may not be strictly constrained, the overall short duration of dislocation time histories in the model is well established.

We test whether relevant information may have been lost in the process of filtering the lowest frequencies of the strong-motion seismograms. In order to address this question, we corrected the E-W and vertical components of SKR, the only digital station, for baseline shift in acceleration using a modified version of the approach proposed by Iwan (1985), then proceeded with the double integration to get the displacement signals. In Figure 20 we compare the unfiltered observed signals with the unfiltered synthetics generated with the model resulting from the joint inversion of InSAR, GPS, teleseismic and strong-motion data with constraint on the surface offset. The match is excellent, in particular for the spectacular E-W component displaying a fast rise and a static offset of about 170 cm. We note that though the displacement was very large (170 cm) at that station the peak ground acceleration was only 0.4 g.
Conclusions

The fast availability of the geodetic, seismological and surface offset data from the Izmit earthquake provides a unique occasion to constrain the rupture process of a very large earthquake with independent and complementary sources of information. From our resolution tests with synthetic data and from the inversion of the real data we make the following inferences:

1) An excellent fit of a single kind of data does not necessarily imply a good retrieval of the kinematic properties of an earthquake. The overall source time function of the rupture may be easily retrieved but great caution should be taken when considering slip maps from individual data sets.

2) Resolution and sensitivity tests are highly recommended in kinematic rupture studies, especially when a single kind of data is used. In the case of the Izmit earthquake, we show that the location of slip in the easternmost part of the rupture cannot be constrained properly by the InSAR and teleseismic data, while strong-motion, and to some extent GPS data, are unable to resolve slip in the hypocentral area.

3) In the case of a near vertical strike-slip fault, the best spatial resolution can be expected in the upper part of the fault model where InSAR coverage is good and near strong-motion and GPS stations. The resolution of InSAR, GPS, and strong-motion data degrades with depth.

4) The resolution of teleseismic data tends to be lower overall but more evenly distributed within the model. This means that teleseismic data help control slip in the deep portion of the model when combined with the other data sets.

5) Without a constraint on the surface offsets, slip near the surface can easily be overestimated.

6) InSAR, GPS, teleseismic and strong-motion data can be inverted simultaneously to obtain a more robust image of the rupture process of an earthquake. The spatial resolution of the joint inversion is better than for the separate inversions of the individual data sets. For that reason, the temporal complexities of the rupture can be retrieved with more confidence in the simultaneous inversion.
The Izmit rupture is dominated by the bilateral breaking of a central asperity located approximately between 29.7E (10 km or so west of Gölcük) and 30.4E (east margin of Sapanka Lake), with right-lateral strike-slip reaching 6 to 8 m in the depth range 6 to 12 km. The western termination of the rupture is found near the city of Yalova, at about 29.3E, but large slip ends about 25 km more to the East. A second area of large slip is required by all the data sets further East toward the city of Düzce, between 30.7E and 31.1E (Karadere and Düzce faults). This eastern slip zone, which is separated from the main central asperity by an area of very reduced slip, is less well constrained by the data. However, strong motion station DZC near the city of Düzce helps to locate a high slip patch in the depth range 6 to 12 km near 31.1E. The hypocenter of the Düzce earthquake, which occurred three months later (12 November 1999, Mw=7.2), is located in the immediate vicinity of this easternmost slip patch of the Izmit earthquake.

The correlation between deep and surface slip is relatively good for the main slip zone from Gölcük to Sapanka Lake, though in detail surface slip does not mimic slip at depth. In the case of the easternmost asperity, there is no correlation since large slip did not reach the surface. Caution must be taken not to consider that small scale structural complexities and slip offsets observed at the surface can be simply projected downdip along the fault.

The seismic moment resulting from the joint inversion is $2.4 \times 10^{27}$ dyne-cm. Slip lasts 45 to 50 sec but most of the energy release occurs in a short time, less than 15 sec, corresponding to the bilateral breaking of the central asperity. Rupture propagation is relatively uniform toward the West, with a rupture velocity close to 3.5 km/sec. In our inversions, we found no need for a rupture front travelling from the hypocenter faster than the shear wave velocity. Propagation of large slip toward the East is initially slower but it accelerates during a short time interval situated about 10 sec after rupture nucleation. Eastward progression slows down to less than 2 km/sec after 15 sec and rupture almost vanishes in amplitude about 22 sec after initiation. Although slip dies out between 30.4E and 30.7E, the rupture propagates more to the East across the Karadere and Düzce fault segments. From 25 sec on, rupture propagation is limited to these two easternmost segments. Although the slip time histories at individual subfaults may not be strictly constrained, the overall short duration of dislocation time histories in the model is well established.
The Izmit earthquake is another example of a large event displaying a heterogeneous slip distribution. In the light of the present results, the relation between the Izmit and Düzce earthquakes must be carefully explored.

Acknowledgements
This work was supported by ESA (cr. ERS-A03-194) and ETH. Part of this work was carried out at the Jet Propulsion Laboratory, California Institute of Technology, under contract with the National Aeronautics and Space Administration (NASA). We express our gratitude to R. Burgmann and K. Feigl for reviewing the manuscript. We thank the Earthquake Research Department of the General Directorate of Disaster Affairs, Ankara, for providing an easy access to the strong-motion data of the Izmit earthquake through its web page, and R. Reilinger for sending us the GPS data. This is publication No. 1213 of the Institute of Geophysics, ETHZ.
Figure 1. Location of the Izmit earthquake. The position of the epicenter (Kandilli Observatory) is indicated near the city of Izmit (open triangle) and the heavy black line displays the location of the main surface breaks (Barka, 1999; IPGP, ITU, USGS and SCEC, 1999, Web sites). The surface projection of the fault model used in this study is also shown. S1 to S4 are the four fault segments of the model, S3 and S4 corresponding respectively to the Karadere and Düzce faults.

Figure 2. Azimuthal distribution of the P and SH teleseismic signals used in this study, draw around their respective focal mechanism for the central fault segment bearing the hypocenter (segment S2 on Fig. 1 and Table 1). The compressional and dilational quadrants are indicated by the + and - symbols respectively.
Figure 3. Surface offsets, in meters, used to constrain shallow slip in our fault model. After the web sites of the IPGP, ITU, USGS and SCEC (1999). The four segments of our fault model are indicated (segments 1 to 4).
Figure 4. (Top) Synthetic SAR differential (coseismic) interferogram. One gray cycle or "fringe" (black-gray-white) represents here 5 cm of ground-to-satellite displacement. Open circles show the location of the InSAR data points used in the inversions of synthetic and real data. (Bottom) Synthetic GPS coseismic data. The interferogram and GPS vectors were computed from the synthetic model shown on Figure 6-top. The thick black lines show the position of the fault segments used in this study.
Figure 5. Example of observed (left) and synthetic (right) waveform data. The vertical component of the P-waves at two teleseismic broadband stations BGCA and BINY is represented in the top frame. The lower frame shows the East and North component at two strong-motion stations, SKR and DZC respectively. Synthetic signals, to which a low level of noise has been added (solid line, right column) are computed using the synthetic model shown in Figures 6-top (slip map) and 7 (time evolution). Waveform resulting from the joint inversion of InSAR, GPS, teleseismic and strong-motion synthetic data (Fig. 6-bottom and 8) are drawn in dashed lines.
Figure 6. Slip maps from the resolution tests. From top to bottom are displayed the slip maps for the synthetic model, and then for the InSAR, GPS, teleseismic, strong-motion, and joint inversions. The joint inversion combines the four different synthetic data sets. Positions of the strong-motion stations and of the GPS stations located less than 40 km from the fault are indicated, projected into the fault line, above the corresponding models. Dots indicate midpoints of the subfaults.
Figure 7. Time evolution of the rupture for the synthetic model given at intervals of 2.5 sec. Lines corresponding to rupture velocities of 2, 3, 4 and 5 km/sec are represented for reference. The local propagation on individual subfaults is taken into account, as well as the slight EW overlap of segments 2 and 3.
Figure 8. Time evolution of the rupture from the joint inversion of the InSAR, GPS, teleseismic and strong-motion synthetic data. See also caption of Fig. 7.
Figure 9. Slip time functions for individual subfaults for the synthetic model (solid curves) and for the model resulting from the joint inversion of the InSAR, GPS, teleseismic and strong-motion synthetic data (dashed curves). Slip time functions are drawn around the slip map from the joint inversion for those subfaults where slip exceeds 4 m in both the synthetic and inverted models. The maximum amplitude and duration of slip allowed in the model are 8 m and 7 sec respectively. In order to be able to compare the shape and duration of individual slip functions, inverted slip curves were time shifted, when necessary, to be put in coincidence with the synthetic curves. Slip functions are labeled according to the subfault numbers in the model. There is a general agreement between the synthetic and inverted slip curves but large misfits are observed at subfaults 26, 31, 17, 78 and 79.
Figure 10. Slip maps from the inversion of the real data, without constraint on the surface offset. From top to bottom are displayed the slip maps for the InSAR, GPS, teleseismic, strong-motion, and joint inversion of the four data sets. See also caption of Figure 6.
Figure 11. Same as Fig. 10, but with constraint on the surface offset. The slip map at the bottom (joint inversion) is our preferred model. The surface offset values in meters used to constrain slip at the shallowest subfaults (see also Fig. 3), as well as the longitude, are indicated above the joint model.
Figure 12. Top left. Observed unwrapped SAR interferogram. One gray cycle or "fringe" (black-grey-white) represents here 5 cm of ground-to-satellite displacement. Bottom left. Computed SAR interferogram. Right. Four fitting profiles in the direction sub-perpendicular to the fault. Black and white lines on the profiles show the observed and computed ground-to-satellite displacements respectively. Computed means from the model resulting from the joint inversion of real data with constraint on the surface offset (Fig. 11-bottom).
Figure 13. Coseismic GPS vectors fitting. Observed and computed horizontal displacements are represented by open and black arrows respectively. Computed means from the model resulting from the joint inversion of real data with constraint on the surface offset (Fig. 11-bottom)
Figure 14. Waveform fitting of teleseismic broadband data. Observed (solid lines) and computed (dashed lines) are shown for the P (top) and SH (bottom) waves. Computed means from the model resulting from the joint inversion of real data with constraint on the surface offset (Fig. 11-bottom and 17). For each signal, the station azimuth (az) is indicated.
Figure 15. Waveform fitting of strong-motion data. Observed (solid lines) and computed (dashed lines) are shown for the components used in this study. Computed means from the model resulting from the joint inversion of real data with constraint on the surface offset (Fig. 11-bottom and 17). The map in the top-left corner reminds the station locations.
Figure 16. Effect of a single asperity (a) located in the hypocentral area on the strong-motion seismograms (b). The almost negligible effect at most of the stations, especially at SKR and DZC which exhibit the largest observed amplitudes, explains the low resolution of the strong-motion data in this area of the fault model. In (c) is shown the effect of the same asperity on the teleseismic P-wave seismograms at two representative stations BGCA and BINY. The effect is clearly larger on the teleseismic data.
Figure 17. Time evolution of the rupture for the joint inversion of InSAR, GPS, teleseismic and strong-motion data with constraint on the surface offset, given at intervals of 2.5 sec. Lines corresponding to rupture velocities of 2, 3, 4 and 5 km/sec are represented for reference. The local propagation on individual subfaults is taken into account, as well as the slight EW overlap of segments 2 and 3.
Figure 18. Source time function (STF) for the whole rupture. From the joint inversion of InSAR, GPS, teleseismic and strong-motion data with constraint on the surface offset. The STF depicts the evolution of moment rate with time.
Figure 19. Slip time functions for individual subfaults for the model resulting from the joint inversion of the InSAR, GPS, teleseismic and strong-motion data with constraint on the surface offset. Slip time functions are drawn around the slip map from the joint inversion for those subfaults where slip exceeds 4 m only. The maximum amplitude and duration of slip allowed in the model are 8 m and 7 sec respectively. Slip functions are labeled according to subfault numbers in the model. The 7 sec available for slip are fully used at subfault 36 only.
Figure 20. Waveform fitting of the unfiltered strong-motion displacements at station SKR. Observed seismograms are drawn in solid lines while those computed with the model resulting from the joint inversion of InSAR, GPS, teleseismic and strong-motion data with constraint on the surface offset are shown in dashed lines.
Chapter 5

General Conclusions
1. Summary of the results

We developed an inversion method allowing us to characterize the earthquake source process with a limited number of parameters (Delouis et al., 2000). The fault models are discretized, divided into subfaults where the rupture propagation time, the slip amplitude, direction and duration are investigated according to the formulation of Hartzell and Heaton (1983). Source models are derived from multiple observation data sets available for the study of a particular earthquake.

In this thesis, we particularly focus on the systematic joint inversion of teleseismic broadband and InSAR data. These geodetic data are potentially able to map completely the 3-D deformation field due to an earthquake with a spatial resolution on the order of 100m on areas on the order of $10^4\,\text{km}^2$. Both data sets are potentially available worldwide and in a nearly real time. The geometry of the faulting being strongly constrained by the geodetic data, the robustness and the uniqueness of the spatio-temporal history of large events are noticeably enhanced. Moreover, an earthquake’s source parameters can then be investigated in detail in any place they occur. In addition, the method easily allows for the use of other data sets to improve or complement the detailed description of the space and time distribution of the slip: GPS measurements, rupture offsets, or local strong motions.

The simulated annealing scheme used here to explore the parameter space of a non linear process is an efficient search algorithm that likely converges toward the absolute minimum and provides stable solutions without need to test any a priori slip distributions. This semi-global search method only requires an a priori definition of bounding values for the free parameters.

The different studies of the earthquake source parameters have confirmed the spatial and temporal heterogeneities of the slip history at many scales. The interpretations of earthquake processes in terms of subevents or a propagating rupture, i.e addressing the problem of the spatial pattern of the moment release (Ihmle, 1998; Das and Kostrov, 1994) are potentially better constrained by the combination of geodetic and teleseismic data than with the single use of seismic data.

We showed the importance to systematically perform resolution tests with synthetic data. Evaluating the ability of the different data sets used separately and jointly to recover properly the source parameters is an important issue to test the relevance and uniqueness of a solution.

We observed that the seismological data generally have a poor resolving power in terms of spa-
tial location of slip areas (trade off between rupture time and location). Different additional constraints such as smoothing, or moment minimization (targeting the Harvard Centroid Moment Tensor solution) help to stabilize the inversions (Das ans Kostrov, 1990). Nevertheless, stabilizing solutions too much leads to a poor resolution (Backus and Gilbert, 1968). Further, seismic data exhibit very different resolving powers depending on their frequency content and on the complexity of the earthquakes: the spatial extent, the rupture mechanism including the nature of the rupture propagation (unilateral or bilateral) and the segmentation of the faulting. As a consequence, we deemed that seismic data should be systematically used with complementary data, and more specifically with InSAR data that map the deformation field with an high spatial resolution in order to improve the kinematic modeling.

We observed that InSAR data have a limited resolution at depth, resolving better the shallower deformed part of the rupture even if the entire deformation field is measured (Chap3. Hector Mine earthquake; Chap.4. The Izmit earthquake). Nonetheless, we observed that even when only part of the deformation field was accessible with spatial measurements, it could contribute to enhance the slip history models (Chap.2, The Nazca ridge earthquake, Chap.4 The Izmit earthquake). Locating precisely the gross slip features of a rupture improves noticeably the temporal resolution and we observed that the overall rupture time history is well retrieved in the joint inversion scheme. On the other hand, obtaining robust particle histories would require a near field instrumentation in the vicinity of faulting (Chap4 The Turkey earthquake). According to the types of data sets, the geometry of the faulting and its spatial location, we can expect to get into different levels of details in the description of a rupture process with a kinematic approach. Further, the more independent data sets that we have available, the more we are able to robustly calculate the complexity of earthquakes.
2. Limitation of the data and possible enhancements

2.1 The teleseismic data

Despite all the improvements that can be carried out by the joint inversion of multiple data sets, the propagation effects in the vicinity of a rupture are still difficult to be deconvolved from the seismograms: the crustal complexities that are not taken into account in our test models produce a noticeable perturbation of the seismic pulses and should be corrected for, especially for subduction zone earthquakes (water layer and complexity of the medium). We used only simple velocity structures limiting our investigations to low frequency waveform modeling (<1 hz). Furthermore, we used preferentially seismic data at teleseismic distances: the bodywaves have simple structural interactions and propagate into the lower mantle where the velocity structure is smooth. Some improvements in the determination of the velocity structures, like 3-D models in the vicinity of major fault zones could enhance the resolving power of near-field seismic data (Graves and Wald, 2001) or may allow broader frequencies in inversions. Another way to overcome this problem for the teleseismic data would be the consideration of small earthquakes located near the larger event of interest. If the small event has a short, simple (impulsive) source time function, its recordings approximate the Earth’s Green function including attenuation, propagation, instrument and radiation pattern effects. Nevertheless the conditions for such use of an empirical Green function is that both the small and large events have the same source depth, focal mechanism and at least two orders of magnitude difference in the seismic moments.

The study of the earthquakes’ high frequency radiations (1 to 15 Hz) provides information about the dynamics of the rupture. However, the difficulties encountered in modeling waveforms at high frequencies prevent them from being used in kinematic approaches. Nonetheless, the estimation of high frequency radiation areas (Hartzell et al, 1996) could help to further constrain the slip history of an earthquake by locating the areas of initiation and stop of the rupture process as a priori constraints in the kinematic modeling.
2.2 The InSAR data

The InSAR data have provided a tremendous advancement in the field of the measurement of earth deformation. For one decade, continuous records of radar data have improved the possibility for earth scientists to have access to large topographic and deformation measurements all over the continental earth. Nevertheless, their interpretation and exploitation have to deal with specific problems inherent to the InSAR technique in order to avoid misinterpretation of the data. Bürgmann et al (2000) sum up (Table 5.1) the main errors that could lead to noise or artifacts. These perturbations or additional physical contributions out of the field of the earthquake geophysics fall into three categories: *Phase noise* introduced by a variety of sources associated with the radar and wave propagation and scattering physics (Chap 4, Turkey earthquake); *Height noise* introduced in the mapping process by uncertainties in the dimensions of the interferometer; *Geometric distortion* associated with radar imaging as illustrated by the differential interferometric scene of the Peru earthquake (Chap 2). Nonetheless most of these problems can be overcome when error sources are properly identified.

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<td>Phase errors</td>
<td>Random broad-band, characterized by interferometric correlation</td>
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<td>Generally systematic, can be random if platform motion is severe and uncompensated</td>
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<td>Data gaps</td>
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*Table 5.1: Errors in InSAR measurements (Bürgmann et al, 2000)*
Another limitation of the InSAR data is the temporal sampling (Figure 5.1): The above mentioned limitations constrain the acquisition of radar images to form a differential interferogram. Thus, the time delay between two suitable radar scenes in our field of interest can vary greatly from one event to another: 4 years for the Peru earthquake to 35 days for the Hector Mine and Turkey earthquakes though earthquakes are a minute scaled phenomenon. The deformation field is therefore a contribution of the coseismic but also pre-seismic and post-seismic deformations. Hopefully, the multiplication of the radar scenes lead to hypotheses that take into account all the observed phenomena in order to remove or study them (Jacobs et al, 2002). Another way to overcome this problem is to deploy a galaxy of satellites covering the whole earth that would be able to acquire real time events.

Figure 5.1: Sketch of the potential spatio-temporal sampling of different geodetic measurements used to determine ground displacement.
At last, a limitation of the InSAR data more specific to the study of the ground deformation is that measurements of InSAR data provide only one component of the ground motion. Since more and more radar scenes are available to compute interferograms, the study of earthquakes using InSAR data will be noticeably improved by complementary SAR scenes from different orbits (ascending and descending) but also by the measurement of the amplitude offsets that provide a less accurate but potentially useful third displacement component in the azimuth direction of the satellite (Michel et al, 1999). Nonetheless, if only one component is available, the InSAR data can be complemented by the use of the GPS stations, even sparsely located that have the advantage to provide an absolute 3-D measurement. Differential interferometry is then still a potentially improving tool to study the earth deformation: Pre-, co-, and post-seismic deformation on faults are likely to be systematically monitored and will provide growing information that will contribute to earthquake source studies, hazard assessment, stress transfer, or seismic cycle at different time scales comprising eventually some studies of ultraslow earthquakes.
3. Perspectives

3.1. Improvement of the inversion method

The semi-global search methods are now often used to explore some finite dimension parameters space and provide reliable solutions for non-linear problems. Linearized least squares inversions, despite their instability and their a priori input models, have the advantage to provide a solution with an error estimate. An equivalent estimate on parameters for the non-linear methods could be implemented to image for instance the trade-offs between the parameters and to provide an estimation on the confidence we could expect from an inversion. For instance, Sambridge (1999) has implemented a new approach for the geophysical inversions, the neighborhood algorithm. It uses the information given by different models collected by a search process and allow us to obtain an appraisal of the resolution and the trade-off in the model parameters or any combinations of them without the need to further solve the forward problem.

3.2. Improving the static displacements modeling

Most of the geodetic modeling of earthquakes are computed with the assumption of a half-space. Savage (1987) has shown that this approximation for static calculations may be inadequate and slip distributions inferred from them may contain artifacts. More recently, Wald and Graves (2001) discuss the importance of 3D, or 1D layered models compared to an half-space approximation. In their study, they have shown that the static displacements are sensitive to the 3-D structures, but less than the waveforms. They point out that geodetic inversions are limited by a trade-off between fault depth and slip magnitude and recommend that the fault plane geometry be fixed in space. Further, using half-space models can result in overestimation of the slip determination. Nevertheless, with this approximation geodetic data can still provide additional resolution when jointly inverted with seismic data. Further, Cattin et al (1999) noticed that the effects of superficial layers have a larger influence on the horizontal displacements than on the vertical ones. Consequently, the half-space approximation is still adequate when InSAR measurements are used since they map principally this component. Until now, few seismic source studies have implemented the use of layered or 3-D models (Hernandez et al, 1999; Ji et al, 2002).
The growing number of accurate geodetic measurements (GPS, InSAR) that will provide the entire 3-D deformation fields of the earthquakes will require a more precise modeling that will include 3-D crustal model, limiting the half space approximation to areas where no detailed crustal models are available.

### 3.3. The detailed slip distribution as an input for other studies

Detailed slip histories are of interest for many forward modeling such as the modeling of site effects, the stress transfer study with possible improvements of the coulomb stress modeling, or the relationship between large events: for instance interactions between the 1999, Landers and Hector Mine (California) earthquakes (Price et al, 2002), or between the 1999, Izmit and Duzce (Turkey) earthquakes (Figure 5.2). It could also have some implications in the study of the spatio-temporal interactions of small and large earthquakes (cf. ETH project, Jochen Wössner PhD project) with improvements in the field of hazard assessment. Finally, detailed slip maps provided by kinematic modelings can be used as a constraint in the dynamic modeling of the rupture process (Olsen and Madariaga, 1999; Peyrat et al, 2001).
Figure 5.2: Slip maps of the 1999, Izmit and Duzce earthquakes inferred from joint inversions (Delouis et al, in prep.)
Appendix

Joint inversion of InSAR and teleseismic data for the slip history of the 1999 Izmit (Turkey) earthquake

B. Delouis (1), P. Lundgren (2), J. Salichon (1) and D. Giardini (1)
Abstract

The slip history of the Izmit earthquake is investigated by jointly inverting SAR and teleseismic data with a multi-segment, variable-slip, finite fault model. Surface offsets are used as an added constraint. The highest slip values (5 to 7 m of right-lateral displacement) are associated with a strong central asperity surrounding the hypocenter. The bilateral breaking of this main asperity, which lasted about 10 sec, dominates the source time function. From the Gölcük-Izmit area to Lake Sapanca, a relatively good correspondence is observed between slip at depth and at the surface, but farther to the West and to the East slip tends to be confined to the near surface. Slip in the easternmost part of the rupture (Düzce segment) is not well resolved. Our analysis finds the InSAR and teleseismic data to be compatible and complementary. This study shows the potential of the joint inversion of both data sets for a rapid determination of the spatio-temporal history of large earthquakes.

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Appendix

Introduction

The Izmit earthquake of 17 August 1999 (Mw=7.3 to 7.6) was one of the most destructive earthquakes in recent decades, causing several thousand casualties. It occurred in the westernmost part of the North Anatolian Fault which experienced a sequence of large earthquakes migrating from East to West during the period 1939-1967 [Barka, 1996]. At least 110 km of surface breaks were reported (Figure 1) with up to 5 meters of right-lateral slip (IPGP, ITU, USGS and SCEC, 1999, Web sites). The epicenter (40.76N, 29.97E, Kandilli observatory) was near the city of Izmit (Figure 1). The purpose of this work is to study how slip was distributed in space and time along the causative fault by modelling interferometric synthetic aperture radar (InSAR) and teleseismic data, taking into account the observed surface offsets. InSAR and seismological data have so far very rarely been combined [Hernandez et al., 1999; Wright, et al., 1999] and we aim at inverting them both independently and jointly for the slip history of the Izmit earthquake. The advantage of the joint inversion of geodetic and seismological data is that it limits the trade-off between the rupture timing and the slip location [Wald and Heaton, 1994].

Data

The differential SAR interferogram (Figure 2, top) was computed using SAR data from the European Research Satellites (ERS-1 and 2) acquired on August 13, 1999 and September 17, 1999. The raw data (ERS-2, track 157, frames 815 and 801) were processed using the JPL-Caltech developed ROI_PAC software. We used the "2-pass" method [Zebker and Goldstein, 1986] to generate the differential interferogram. This is done by removing a synthetic interferogram based on precise orbit information from the European Space Agency and the effects of topography using the DTED 90 m posting digital elevation model (DEM) for NW Turkey from the U.S. National Imagery and Mapping Agency. This ERS pair has the shortest coseismic time separation possible for this earthquake (35 days) with the first image acquired four days before the earthquake. The perpendicular baseline (separation) of the two ERS satellite orbits was small (~30 m), minimising sensitivity to topographic uncertainties in the DEM. The interferogram was unwrapped using a branch-cut algorithm, with bridges used to connect regions separated by low coherence. As shown by Figure 2, the coverage of the SAR data is good for the western part of the rupture but SAR data are lacking near the fault in the eastern part due to
low coherence. The interferogram was resampled in order to reduce the number of points in the inversion and to obtain a lower density of points far from the fault. We also processed the available ERS-1 pair (990812 – 990916) with precise orbits for coseismic deformation and found that the difference with the ERS-2 differential interferogram was small.

Seismological data consist of P and SH seismograms recorded at teleseismic distances by the IRIS and GEOSCOPE networks. Records were deconvolved from the instrument response and integrated to obtain displacements. We model the first 60 sec of the P wave trains and the first 80 sec of the SH signals.

We use also data on the average offsets observed at the surface along the coseismic rupture (IPGP, ITU, USGS and SCEC, 1999, Web sites).

Fault model and inversion procedure

The earthquake rupture is simulated by three fault segments whose upper limits coincide with the main surface breaks at the surface (Figure 1). Each segment is itself subdivided into rectangular subfaults with dimensions 10 km along strike and 5 km along dip, resulting in 75 total subfaults. Strikes and dips are held fixed (Table 1) but the slip angle is allowed to vary within the range 180+/-20 degrees (right-lateral slip). Slip at each subfault is represented by a moment rate source time function. Individual source time functions are themselves represented by two elementary triangular functions mutually overlapping, each having a duration of 4 sec. The rupture initiates at the hypocenter and propagation is accounted for by the rupture onset times of the subfaults. Rupture onset times are allowed to vary within a certain range defined by two bounding rupture velocities, 1.8 and 3.5 km/s. In summary, the parameters to be inverted for are, at each subfault: the rupture onset time, the amplitudes of the elementary triangular functions, and the slip angle. Larger ranges of slip angle and rupture velocity, as well as larger numbers of elementary triangular functions were tested, without improvement nor significant changes in the solutions. In agreement with the teleseismic data, a hypocentral depth of 12.5 km is used. For the computation of synthetic teleseismic seismograms, each subfault is approximated by a point source placed at its center and the ray theory is used [Nabelek, 1984]. Near-field static synthetic displacements (for InSAR data) are computed using the dislocation formulation of Savage (1980). The inversion itself is performed by using a simulated annealing procedure, the primary cost function being the RMS error of the data fit normalised.
with the observed data. An additional cost function is used in order to minimize the total seis-
mic moment. Simulated annealing, which allows for a quasi-global exploration of model
space, has been recently used in variable-slip fault inversions [e.g. Ihmlé, 1998; Lundgren et
al., 1999]. We include among the parameters to be inverted for two static offsets, required to
calibrate the InSAR data south and north of the fault. In two specific separate inversions, as
well as in the joint inversion, slip values at the shallowest subfaults are constrained to remain
close to the average offset observed at the surface. From the data processing and inversion we
found that a correction for orbital tilt across the image was not required. In the joint inversion,
the individual cost functions (normalised RMS errors) of the InSAR and teleseismic data are
equally weighted.

<table>
<thead>
<tr>
<th>Table 1. Geometry and dimensions of the fault model</th>
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<tr>
<td>Strike (deg)</td>
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<tr>
<td>Segment 1</td>
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<td>Segment 2</td>
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<td>Segment 3</td>
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**Results and resolution tests**

The slip distributions corresponding to the different inversions carried out are presented in Fig-
ure 3. The modelling of the data corresponding to the joint inversion can be seen in Figures 2
(SAR data) and 4 (waveforms). The relatively simple shapes of the P and SH waves reflect the
smooth and continuous character of the rupture. This characteristic is also indicated by the
smoothly varying pattern of sub-parallel fringes on the SAR interferogram (Figure 2, top).
Seismic moments resulting from the inversions range between 1.9 and 2.2 x 10^{27} dyne.cm.

We carried out some resolution tests with the InSAR data in order to evaluate how the spatial
distribution of slip is constrained. Synthetic InSAR data were generated using two different
fault models (Figure 5) which we inverted with the same parametrisation as done with the real
data. A random noise was added to the synthetic data so as to have a final misfit error similar to
that of the real inversions. The synthetic tests show that InSAR inversions resolve well the
shallow part of the models (above ~12 km depth) for segment 1 and for the western part of seg-
ment 2, with only partial recovering of features at mid-depth. Features in the deepest part of
the model are almost not recovered at all.
Appendix

Slip maps for the real data inversions (Figure 3) display a clear overall similarity, except for segment 3 (Düzce segment). However, a close examination of the results shows that each data set has some specific contributions. In agreement with the synthetic tests, slip from the InSAR inversions is essentially restricted to the upper half of the model for segments 1 and 2. Teleseismic data require deeper slip near the hypocenter. In contrast to teleseismic data, InSAR data require added shallow slip in the western part of segment 1 and constrain relatively well the western termination of the rupture. Without constraint on the surface offset, both the teleseismic and SAR inversions tend to overestimate slip values near the surface.

In contrast to the teleseismic and joint models, InSAR models display very high slip values (up to 8 m) in the deep part of segment 3 (Figure 3). Continuous GPS measurements in the source area indicate some postseismic deformation of the ground surface, but much smaller than the meter level coseismic ground displacement that we model here (Burgmann et al., 2000). It is hence likely that the high slip values on segment 3 are mostly related to the poor coverage of the InSAR data in this area (see the eastern part of Figure 2). This is confirmed by the synthetic tests (Figure 5) which show that this part of the fault model is not well resolved. Additional types of data controlling the eastern part of the rupture are needed to improve resolution there, since slip along this segment may have strong implications for the triggering of the 12 November 1999 Düzce earthquake.

The slip model resulting from the joint inversion with constraint on the surface offset (Figure 3f and g) we consider the best to describe the source process since it provides an acceptable fit to all the data. The highest slip values, with 5 to 7 m of right-lateral displacement, observed in the hypocentral area, are likely to correspond to a strong (resistant) asperity within the fault zone. The cities of Izmit and Gölcük (Figure 1), which were highly damaged, are located approximately above this high slip zone where it reaches the surface. From the Gölcük-Izmit area to Lake Sapanca, a relatively good correspondence is observed between surface offsets and slip at depth. However, further to the East and to the West, slip tends to be confined to the near surface. The time evolution of the rupture (Figure 3g) displays an essentially bilateral rupture propagation with a relatively smooth rupture front crossing the main slip zone and the junction between segments 1 and 2 at a velocity varying between 2.8 and 3.1 km/sec. The overall source time function (STF, Figure 4) shows two episodes of seismic moment release for a total duration of about 50 sec but it is dominated by a first pulse lasting 10 sec which corresponds to the bilateral breaking of the main asperity. The second episode corresponds to the easternmost part of the rupture.
Implications

This study shows the potential of the joint inversion of SAR and teleseismic data for the investigation of the spatio-temporal history of large earthquakes. Both data sets have an almost real-time availability and with such an approach slip-histories of large events may be obtained soon after their occurrence. This kind of rapid determination may be particularly useful for the fast modelling of potential damage patterns and stress transfers on neighbouring faults.

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Figure 1. Location of the Izmit earthquake. The epicenter, the HCMT focal mechanism, the main surface breaks (IPGP, ITU, USGS and SCEC, 1999, Web sites), and the surface projection of the fault model are indicated.
Figure 2. Top. Observed unwrapped SAR interferogram. One gray cycle or "fringe" (black-grey-white) represents 5 cm of ground-to-satellite displacement. Center. Computed SAR interferogram. Bottom. 4 fitting profiles in the direction subperpendicular to the fault. Black and white lines on the profiles show the observed and computed ground-to-satellite displacements respectively. Computed means from the joint inversion.
Figure 3. (a) Average surface offsets used to constrain the slip of the shallowest subfaults (IPGP, ITU, USGS and SCEC, 1999, Web sites). (b) to (e) Slip maps from the 4 separate inversions carried out with or without constraint on the surface offset. Black dots indicate the center of the subfaults. (f) Slip map, and (g) time evolution of the rupture for the joint inversion. The isochrons of the rupture front are labelled in sec.
Figure 4. Waveform modelling from the joint inversion. Observed (solid) and computed (dashed) seismograms are shown for the P and SH waves around their respective focal mechanism at the hypocenter, together with the overall moment rate source time function (STF).
Figure 5. Synthetic models (S1 and S2) and the corresponding inverted models (I1 and I2) used in the resolution test of the SAR inversion.
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