Comparisons of Geological Models to GPS Observations in Southern California

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Abstract

Although the surface geology of southern California has been extensively studied, debate continues over the subsurface geometry and style of deformation. The two end-member models in the Los Angeles basin are thin-skinned and thick-skinned deformation. This thesis presents two studies which analyze some of the implications of these two models.

A consequence of thin-skinned tectonics is fault propagation folding (FPF) associated with buried thrust faults. The analysis of deformation due to FPF presented here indicates that deformation on axial planes of folds would affect surface displacements. In Chapter 2 a model of the FPF associated with the 1994 Northridge, California, earthquake is tested to determine whether folding along axial planes inferred from geological models accounts in part for the coseismic surface displacements measured using the Global Positioning System (GPS). GPS observations are inverted to determine the distribution of slip on planes of dislocations coincident with the main rupture plane, as determined from the aftershocks and main shock, as well as axial planes predicted by geological models. Although the inversion for FPF deformation is better in a normalized root mean square (NRMS) sense than inversion for deformation on the main rupture plane alone, the sense and magnitude of axial plane displacements do not agree with geological predictions.

In Chapter 3 a kinematic block model is presented for evaluating geological observations of fault slip rates in southern California. The blocks are rigid over geological time scales and conform to the path integral constraint: deformation accumulated between two points is path-independent. Fault slip rates are inverted to determine block motions. Coseismic deformation on bounding faults of blocks is determined by modification of the standard elastic method. Interseismic velocities at GPS sites are predicted by subtracting the coseismic displacements, divided by the earthquake recurrence time, from the geological rates. The thick-skinned and thin-skinned block models are tested. The existing geological fault slip data is integrated through these two kinematically consistent models. The inversion for block motions in the thin-skinned model is marginally better statistically than the thick-skinned model. However, the thick-skinned model predicts interseismic velocities that correspond best to geodetic observations.
This dissertation is dedicated to my mother,
Kay Souter.
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# Table of Contents

Abstract ......................................................................................................................... 2  
Dedication .................................................................................................................... 3  
Acknowledgments ........................................................................................................ 4  
Table of Contents ......................................................................................................... 5  
Chapter 1: Introduction ............................................................................................... 6  
Chapter 2: Fault propagation folding ......................................................................... 9  
Chapter 3: Three-dimensional block models ............................................................ 39  
Chapter 4: Conclusions .............................................................................................. 80  
Appendix I ..................................................................................................................... 82
Chapter 1
Introduction

In this dissertation I present two studies of geologically based models of geodetic observations in the Los Angeles region. The philosophy I have followed throughout this work is that more is learned about deformational styles and characteristics from simple models of crustal deformation than from overly complex models. I have avoided models in which there are too many degrees of freedom to evaluate the success of the models. Toward this end I have generally used previously determined geological and geophysical evidence, such as fault geometry and slip rates determined from geological studies, to constrain my models. This results in fewer parameters to solve for and allows concentration on key questions.

The underlying question that I address is what style of deformation prevails in the Los Angeles region. Two geological models exist for deformation in the Los Angeles basin: thick-skinned and thin-skinned. In thin-skinned tectonics, thrust faults merge at depths of 12-20 km with subhorizontal detachment faults. In thick-skinned tectonics, faults continue at depth into the ductile lower crust. Many ancient exhumed fold and thrust belts such as the southern Appalachians display thin-skinned characteristics [Harris and Bayer, 1979]. Standard methods have been devised for inferring subsurface thin-skinned fault geometry from surface geology (Suppe, 1985). However, it has recently been suggested that the deformation in the Los Angeles basin is thick-skinned in style.

It is generally difficult to distinguish between the two styles from the surface expression of deformation alone. But, like any hypothesis, each deformational style has implications that can be tested. In this dissertation I explore some of the consequences of thin-skinned tectonics and thick-skinned tectonics in the Los Angeles region. The existence of thin-skinned tectonics implies fault propagation folding (Chapter 2), and the existence of potentially seismogenic detachment faults and higher far-field displacements for a given slip rate on a thrust fault (Chapter 3). Deformation through thick-skinned tectonics does not require fault propagation folding or underlying subhorizontal detachment faults, and predicts lower far field displacements for a given rate of deformation on a thrust fault.

The first study arose from the hypothesis that fault propagation folding occurring during or immediately after an earthquake would affect surface displacements. Geodetic observations of surface displacements after the Northridge earthquake do not fit a standard elastic half-space model of distributed slip on a plane of dislocations that is constrained to lie in the aftershock plane. In
Chapter 2, I develop a linear forward model of the relationship between fault propagation folding and surface displacements, and then invert the coseismic GPS observations for deformation on the fault plane and fault propagation fold axial planes.

In the second study geological observations of fault geometry and displacement rates are systematically inverted to determine block motions in a block model of regional deformation in the Los Angeles region (see Chapter 3). Externally obtained geological information about fault slip rates and geometry is critical to setting up and constraining block models of the Los Angeles region. The effects of the thin-skinned model on far field displacements are incorporated into a kinematically consistent model. The results are compared to a thick-skinned model with different relationships between fault slip rates and far field deformation. The predictions of the thin-skinned and thick-skinned models are then evaluated by comparison with geological observations.

The plausibility of seismogenic detachment faults is also addressed in Chapter 3. Fault slip rates are not affected by the existence of detachment faults. However interseismic surface displacement rates are affected by detachment faults. Predictions of surface displacements from the thin-skinned model and from the thick-skinned model are compared to independent geodetic observations.

In the final chapter I discuss the merit of each model and summarize the results. Possibilities for future work are explored in detail.
References


Chapter 2
Fault propagation fold growth during the 1994 Northridge, California, earthquake?

Abstract

Geological models of buried thrust faults indicate that fault propagation folds (FPF) form and grow with a geometry that depends on that of the fault [Suppe, 1985; Suppe and Medwedeff, 1990]. The displacement gradient fields for faults and kink folds are very similar, and both can be modeled using dislocations. In this chapter a geological model of the FPF associated with the January 17, 1994 Northridge, California, earthquake is tested to determine whether folding along axial planes inferred from geological models accounts in part for the coseismic surface displacements measured with Global Positioning System (GPS). Coseismic deformation on both the main rupture plane and active axial planes of related folds is tested for by inverting for the displacements on dislocation planes in an elastic half-space. A model incorporating two axial planes is preferred to a model with a single rupture plane in a normalized root mean square (NRMS) sense; however, the distribution of axial plane displacements does not correlate with the displacements on the main rupture plane in the way expected for a fault propagation fold. The results presented here indicate that the deformation associated with folding is too distributed to be resolved on a discrete plane, that the deformation occurs interseismically, or that one or both of the kink bands does not exist. A model of a single elevated plane, which is parallel to, but not coplanar with, the aftershocks, is better in a NRMS sense than the FPF model, indicating that anelastic deformation in the hanging wall may be distributed.
Introduction

The 1994 Northridge, California, earthquake occurred on a previously unrecognized and unmapped blind thrust fault in the Los Angeles basin. Geologists and geophysicists have come to agree that such faults are common in this basin. Earthquakes on blind thrust faults result in fault propagation folding in the fault's hanging wall. The recognition that such folds exist and are related to blind thrusts is not new: they have long been recognized in ancient exhumed fold and thrust belts and in reflection seismology profiles of tectonically active regions [e.g. Suppe and Medwedeff, 1990; Mitra, 1986]. However, the recognition that such folds grow seismically and that the occurrence of earthquakes on buried thrusts implies active folding, and vice versa, is a relatively recent idea [Stein and King, 1984].

Several observations indicate the possibility of folding during the Northridge earthquake. Ground breakage and surface folding in the region up dip of the fault plane have been attributed to coseismic folding [Jones et al., 1994; Yeats and Hufnle, 1995]. It has also been suggested that the swarm of aftershocks in the hanging wall of the Northridge fault plane were related to folding [Jones et al., 1994].

In this chapter a geologically based model is related to coseismic fold growth through surface deformation measured with Global Positioning System (GPS) (Figure 2.1). The model is based on a cross section through the hypocenter of the Northridge earthquake [Davis and Namson, 1994]. The fault propagation fold model is tested by inverting surface displacements for slip on the relevant deformation planes predicted by the fault propagation fold model and comparing the results with an inversion on just the main rupture plane.

There exists a discrepancy between the location of the best fitting rupture plane for the Northridge earthquake as determined from geodetic data and from seismic data. Although the aftershocks of the Northridge earthquake delineate a large, diffuse, and irregular plane, inversion of displacements on a main rupture plane that is constrained to lie in this plane produces a high postfit normalized root mean square (NRMS) [Hudnut et al., 1996; Shen et al., 1996]. The best fitting Northridge dislocation plane, as determined from nonlinear optimization of geodetic measurements, lies 2–3 kilometers above the aftershock plane [Hudnut et al., 1996].

This discrepancy has been observed in other earthquakes, such as the 1989 Loma Prieta, California [Eberhart-Phillips and Stuart, 1992; Arnadottir et al., 1992] and the 1989 Kalapana, Hawaii [Arnadottir et al., 1992] earthquakes. The magnitude of the discrepancies for the Loma Prieta and Kalapana earthquakes were reduced by using the full covariance matrix to invert the surface displacements [Arnadottir et al., 1992; Arnadottir and Segall, 1994], as well as by
Figure 2.1. Coseismic displacements from the Northridge earthquake measured using Global Positioning System (GPS). Epicenter of Northridge earthquake is shown by star.
incorporating lateral and vertical variations in the elastic properties of the crust into the deformation model [Eberhart-Phillips and Stuart, 1992; Du et al., 1994]. In order to minimize the bias in the data inversion, the full data covariance matrix is included in this inversion, as suggested by Arnadottir et al. [1992].

Here a homogeneous elastic crustal model is used to test specifically for evidence of coseismic fault propagation folding. Although the effect of horizontal variation in elastic properties has not yet been determined, previous work indicates that vertical layering does not affect the location of the best fitting dislocation plane: Shen et al. [1996] used nonlinear optimization of the fit to the GPS data with a vertically layered crustal rheology and found no significant effect on the vertical position of the best fitting Northridge dislocation plane.

The discrepancy may result from anelastic hanging wall deformation due to folding. Searching for a single dislocation plane when deformation has occurred on two planes will likely result in a best fitting plane that lies somewhere between the two. Using nonlinear optimization to find an additional dislocation plane, located in the hanging wall, while fixing location of the main rupture plane, greatly improves the fit [Shen et al., 1996]. However, allowing a secondary deformation plane to vary freely provides little insight as to why that plane is deforming.

**Model**

The model geometry (Figure 2.2) is taken from a cross section through the location of the Northridge earthquake, [Davis and Namson, 1994, Figure 2.2] which is based in part on thin-skinned models of fault propagation folding [Suppe, 1985]. The geology near the Northridge earthquake's location is complex; however, only structures related to slip on the Northridge fault plane are relevant to this model. Two folds are necessarily related to the Northridge fault plane: folds A and B (Figure 2.3). Fold A, with axial plane A (APA), results from the updip termination of the buried thrust fault. Fold B, with axial plane B (APB), results from the downdip junction of the Northridge fault plane with the Elysian Park Thrust. Although other folds and faults exist in the hanging wall, they are not required to deform due to displacement on the Northridge fault plane and are therefore not included in this model.

The main rupture plane and the axial planes of fault propagation folds are modeled as dislocations in an elastic half-space. As shown in detail below, although axial planes are not discrete planes of displacement, the deformation gradient tensor $F_0$ due to deformation associated with kink folding is the same as the deformation gradient tensor due to slip on a fault. The deformation gradient tensor is the basis of the derivation of surface displacement due to slip on a dislocation in an elastic half space [Okada, 1978, equation 1]. Like faults, kink folds can be
Figure 2.2. Schematic model of fault propagation folding on the blind thrust fault associated with the Northridge earthquake. The location of the Northridge earthquake is marked with a star. Boxes show locations of Figures 2.3a and 2.3b. EPT, Elysian Park Thrust; MRP, Main rupture plane. The fault geometry and the orientation of axial planes A and B (APA and APB) are taken from Davis and Namson [1994]. For simplicity, kink bands associated with other faults are not shown.
modeled as dislocations in an elastic half-space.

Assuming plane strain, a finite area surrounding an axial plane is analyzed to derive the approximate two dimensional deformation gradient tensor (Figure 2.3) and compare it with the deformation gradient tensor associated with faulting (Figure 2.4). The deformation gradient tensor \( F_{ij} \) is defined as

\[
F_{ij} = \frac{\delta u_i}{\delta x_j} = \begin{bmatrix}
\frac{u_i - u_n}{x_i - x_n} & \frac{u_j - u_n}{x_j - x_n} \\
\frac{u_i - u_n}{x_i - x_n} & \frac{u_j - u_n}{x_j - x_n}
\end{bmatrix}
\]

(2.1)

As the buried thrust fault propagates, deformation occurs along APA and APB. In APA, material moves from the footwall into the hanging wall, from right to left relative to the axial plane (Figure 2.3a). In APB, material moves from left to right up the fault ramp (Figure 2.3b). On one side of the axial plane, material undergoes a rigid body translation parallel to the mechanical layering. Across the axial plane, material undergoes the same rigid body translation plus a displacement \( \Delta u \) parallel to the axial plane. The deformation gradient tensor for an axial plane \( \Delta x F_{ij} \), where \( x_2 \) is parallel to the axial plane, is

\[
\Delta x F_{ij} = \begin{bmatrix}
0 & 0 \\
\frac{\Delta u_2}{\Delta x_1} & 0
\end{bmatrix}
\]

(2.2)

Similarly, for a fault (Figure 2.4), material on one side of the fault may undergo an arbitrary rigid body translation. Material on the other side of the fault undergoes the same rigid body translation plus a displacement \( \Delta u \) parallel to the fault. The displacement gradient tensor \( F_{ij} \), where \( x_1 \) is parallel to the fault, is

\[
F_{ij} = \begin{bmatrix}
0 & 0 \\
\frac{\Delta u_1}{\Delta x_1} & 0
\end{bmatrix}
\]

(2.3)

The deformation gradient tensors for these two scenarios are the same. The major difference between the models is that the deformation gradient tensor approaches a singularity at a fault plane, whereas during folding the formulation never approaches a singularity. The solution for \( \Delta x F_{ij} \) breaks down within \( \Delta u \) distance of the axial plane, where \( \Delta u \) is the amount of slip on the thrust fault. In addition, although the fold structure for the FPF model is an idealized kink fold, for which the fold hinge has zero radius of curvature, in nature even kink folds have a finite radius of curvature; within the fold hinge the equality between \( \Delta x F_{ij} \) and \( F_{ij} \) also breaks down (Figure
Figure 2.3. Detail of deformation in cross section across axial planes of active folds shown in Figure 2.2. The \( x_2 \) axes are parallel to the axial planes, shown as dashed lines. Heavy solid line indicates mechanical layering. The deformation of a solid rectangle passing through the axial plane is shown. Material points move parallel to mechanical layering plane during deformation. (a) Kink fold A. Material at points \( x_0 \) and \( x \) moves from right to left relative to APA. Upon crossing the axial plane, the direction of displacement changes from \( u_0 \) to \( u \). (b) Kink fold B. Material at points \( x_0 \) and \( x \) moves from left to right relative to APB. Upon crossing the axial plane, the direction of displacement changes from \( u_0 \) to \( v \). (c) Fold with a finite radius of curvature. Solid rectangle is continuously deformed as it passes through the fold hinge.
Figure 2.4. Detail of deformation in map view across a fault where the $x_2$ axis is parallel to the fault, shown as a solid line. Material at point $x_0$ moves with a rigid body translation $u_0$. Material at $x$ moves in the direction $u$, which is the sum of a rigid body translation $u_0$ plus a displacement parallel to the fault $\Delta u$. 
2.3c). Because the $A^p_{ij}F_{ij}$ is identical to $pF_{ij}$ everywhere except in the fold hinge, it is expected that at distances much greater than the radius of curvature of the fold hinge the surface deformation associated with the axial plane will be indistinguishable from that of a fault.

In three dimensions the possibility of out-of-plane displacements becomes relevant. It is expected that dip-slip or strike-slip displacement on a fault plane will result in similar displacements as material crosses associated axial planes. Provided that the strike-slip component is small compared to the dip-slip component, the displacement gradient tensor for both kink folds and faults should be similar.

**Method**

The fault propagation model was tested by using inverse theory to solve for slip on planes of dislocations embedded in a uniform elastic half-space. Three models are compared (Figure 2.5): a main rupture plane (MRP), a fault propagation fold (FPF), and a single elevated plane (SEP). The MRP model consists of one plane of 110 dislocations striking $122^\circ$, 20.0 km in along-strike length, and 21.8 km in downdip width. The main rupture plane was chosen such that it passes through both the main shock and the plane of aftershocks.

The FPF model includes the main rupture plane, as well as two planes of dislocations representing axial planes. The geometry of the axial planes was chosen by applying geological models [Suppe, 1985; Suppe and Medwedeff, 1990] to the Namson and Davis [1994] interpretation of the local geology. APA is required to adjoin the upper tip of the thrust fault and to bisect the interlimb angle of the fold observed at the surface. APB is required to bisect the angle between the Elysian Park Thrust and the main rupture plane. The main rupture plane (M) is constrained by the aftershock and main shock locations. The strike of all three planes is $122^\circ$. The main rupture plane and the two axial planes (APB and APA) have 110, 100, and 30 dislocations, respectively. They are all 20.0 km in along-strike length and 21.8, 20.3, and 6.38 km in downdip width, respectively.

Although the geometry of the FPF model implies a "thin-skinned" model, in which steeply dipping faults ramp down 10-15 km into a subhorizontal detachment fault, the model also tests a "thick-skinned" model, in which thrust faults continue at unconstrained angles into the lower crust. The resolved displacements on APB do not necessarily distinguish between the models: significant displacements on both axial planes would indicate thin-skinned deformation; however, insignificant displacements on either plane might indicate that deformation is accomplished interseismically, that the axial plane does not exist, or that deformation is not constrained to a narrow axial plane.
Figure 2.5. Geometry of models relative to aftershock (dots) and main shock (star) locations. M, main rupture plane; APA, axial plane A; APB, axial plane B; E, elevated plane. (a) Map view of dislocation models. (b) Cross-sectional view of models facing 302°.
Figure 2.5, continued.
The geological model of fault propagation folding predicts that slip on the axial plane, \( m \), is related to the main rupture plane displacement \( u \) by
\[
u = m \cos(\delta),
\] (2.4)
where \( \delta \) is half of the interlimb angle: 70° for APA and 80° for APB.

The SEP model, used for comparison, has a single plane with the same geometry as that of the best fitting dislocation plane *Hudnut et al.* [1996] found using nonlinear optimization methods. This model is used as a comparison because it represents the best fitting single dislocation plane: its geometry is determined solely by the fit to the surface displacements [*Hudnut et al.*, 1996]. The strike of this plane, 110°, differs from the planes used in the other two models. The plane has 130 dislocations, a length of 20.0 km, and a width of 26.0 km.

The planes are subdivided into dislocations that are approximately 2 km x 2 km in length and width. The dislocation dimensions are chosen such that further subdivisions do not significantly change the structure of the solution, while at the same time the inversion is easily computed. For each dislocation the Green's functions are calculated, relating a unit of each component of slip on a dislocation to the resulting surface displacements. An elastic Poisson half-space is assumed [*Okada*, 1985] with Lamé's parameters \( \mu = \lambda \). The Green's function matrix \( G_0 \) is formed to linearly relate unit dislocation displacements \( m \) to surface displacements \( d_0 \):
\[
G_0 m = d_0.
\] (2.5)
For example, in the FPF model, the Green's functions are calculated for two components of slip on 240 dislocations to get three components of displacement at 66 ground stations: a 198 x 480 matrix.

The surface displacements \( d_0 \) are obtained from GPS measurements made in the epicentral area between October 1992 and February 1994 and processed with GAMIT (Figure 2.1, this chapter and *Hudnut et al.* [1996]). A large number of institutions participated in these surveys (see acknowledgments).

In order to normalize the system of equations, the Cholesky decomposition of the inverse of the full data covariance matrix \( W_0 \) is used to weight the Green's function matrix and the data vector. The weighting matrix \( P \) comes from \( W_0 \), where \( P \) is defined such that
\[
W_0^{-1} = P^T P.
\] (2.6)
The weighted Green's function matrix \( G \) and the weighted surface displacements \( d \) are then
\[
G = PG_0,
\] (2.7)
\[
d = Pd_0.
\] (2.8)
The resulting system of equations,

\[ Gm = d, \]

has a unit variance. After this transformation, the residual vector, \( r = d - G\hat{m}_\text{est} \), is equal to the normalized summed squared residuals (NSSR) in the original system, so that the NSSR is

\[ \|r\|_2 = \|d - G\hat{m}_\text{est}\|_2 = \left\| \begin{bmatrix} d \end{bmatrix} - G \hat{m}_\text{est} \right\|_2 . \]

A penalty function with smoothness and positivity constraints is used to regularize this underdetermined problem. Applying a penalty function with positivity constraints is a robust method to stabilize and create a unique solution to underdetermined fault plane displacement inverse problems [Du et al., 1992]. \( H \), a finite difference approximation of the Laplacian, \( v^2 = \partial^2 / \partial x^2 + \partial^2 / \partial y^2 \), is applied to the slip estimate \( \hat{m}_\text{est} \) [Harris and Segall, 1987]. This measure of the smoothness is physically based, because it penalizes the fault traction variability [Matthews, 1991]. A positivity constraint is applied to the direction of slip vectors such that they are required to be within 45° of pure thrust motion: a reasonable constraint for a thrust event accompanied by folding. The focal mechanism of the main shock has a rake of 100° [Hauksson et al., 1995].

The set of equations to be solved is then

\[ \begin{bmatrix} G & \beta^4H \end{bmatrix} \hat{m} = \begin{bmatrix} d \end{bmatrix} , \]

where \( d_{\text{init}} = 0 \) is the assumed initial smoothness and \( \beta^2 \) is a penalty factor that weights the smoothness constraints.

The constraint that the direction of \( \hat{m}_\text{est} \) be within 45° of pure thrust motion is applied in the following way. The system of equations (2.11) is rotated 45° and is then solved with a positivity constraint applied to both components of slip such that the displacement is constrained to positive in both \( x \) and \( y \). The nonnegative least squares (NNLS) algorithm of Lawson and Hanson [1974] is used to solve the system of equations. The rotation of \( \hat{m}_\text{est} \) is then removed, such that the displacement is constrained to be within 45° of pure thrust.

The penalty factor \( \beta^2 \) was chosen from the trade-off between the roughness and the NSSR (Figure 2.6) for the MRP model. A value of \( \beta^2 = 0.005 \) was found to sufficiently minimize the NSSR without removing the detail in the slip distribution.

Calculation of the true NRMS is very difficult because of the nonlinearity due to imposing positivity constraints, yet the RMS alone is not sufficient to indicate whether one model is preferred to another model with fewer degrees of freedom. The RMS is normalized by the trace of the model resolution matrix, \( R \), which is an indication of the number of parameters resolved by the model. The calculation of the entire model resolution matrix is similarly difficult because of the
Figure 2.6. Trade-off curve between the roughness and the NSSR for the MRP model. The value of $\beta^2$ is plotted next to each point. $\beta^2=0.005$ is the penalty parameter used for the results presented.
previously mentioned nonlinearity; however, the trace of the model resolution can be calculated directly. A forward model is used to calculate the surface displacements due to each component of slip on each dislocation and invert these synthetic data for the predicted slip on that dislocation. The diagonal elements of $R$ are the amount of each predicted slip component for each dislocation.

**Results**

The results of the inversions for all three models are summarized in Table 2.1. The MRP model has a large NRMS of 5.77 and a moment of $1.56 \times 10^{19}$ N m, calculated assuming a shear modulus $\mu$ of $0.3 \times 10^{11}$ Pa. The moment is high compared to the moment of $1.12 \times 10^{19}$ N m determined from the moment magnitude of 6.7 [Hauksson et al., 1995]. The displacements on the fault plane are thrust, with a slight right-lateral component (Figure 2.7). The maximum displacement of over 5 m is above and to the northwest of the main shock location.

The FPF model has a NRMS of 3.87, lower than the NRMS of the MRP model. The total moment of the FPF model is high, $1.87 \times 10^{19}$ N m, but the moment on the main rupture plane only is $1.51 \times 10^{19}$ N m. The displacements on the main rupture plane are thrust, with a small right-lateral component (Figure 2.8a). The maximum displacement is over 4 m. The average displacement is 1.15 m. The displacements on APA differ from those on the main rupture plane (Figure 2.8b). The maximum displacement on APA is not located near the center of the dislocation plane, as it is on the main rupture plane; it is located in the northwestern corner of the axial plane. The strike-slip displacements on APA are somewhat randomly oriented: right-lateral in the northwest part of the dislocation plane, and left lateral in the southeast. In contrast, there are minimal displacements on APB (Figure 2.8c).

For comparison, the fault plane displacements are also inverted for using the SEP model. Although an inversion for this plane was presented by Hudnut et al. [1996], a different inversion method is used here, resulting in somewhat different results. The NRMS of 3.24 for this model is lower than that of the FPF model. The moment of $1.27 \times 10^{19}$ N m is also lower than for either of the other two models. The maximum displacement of over 3.5 m is located in the center of the dislocation plane. The displacements are closer to pure thrust than either of the other two models, with no systematic strike-slip component (Figure 2.9).

The residuals between the surface displacements observed with GPS and the surface displacements calculated for the MRP, FPF, and SEP models show the locations where the various models fail to fit the surface displacements (Figure 2.10). The MRP model fits the data poorly at stations near the fault, with the worst fit at stations in the region to the north and northwest of the dislocation plane. There is an improved fit to the stations near the dislocation planes in the FPF
model, but the fit to the stations to the northwest is not satisfactory. The overall fit to the data in the SEP model is comparable to that in the FPF model, except in the northwest, where there is a slight improvement.

### Table 2-1. Summary of Fit, Moment Estimates, and Displacement Estimates

<table>
<thead>
<tr>
<th>Model</th>
<th>$N SSR \times 10^3$</th>
<th>Trace of $R$</th>
<th>NRMS</th>
<th>Roughness</th>
<th>Moment $x 10^{19}$ N m</th>
<th>$u_x$ average, mm</th>
<th>$u_y$ average, mm</th>
</tr>
</thead>
<tbody>
<tr>
<td>MRP</td>
<td>6.19</td>
<td>11.9</td>
<td>5.77</td>
<td>0.0034</td>
<td>1.56</td>
<td>-344</td>
<td>1110</td>
</tr>
<tr>
<td>FPF</td>
<td>2.45</td>
<td>34.5</td>
<td>3.87</td>
<td>0.0036</td>
<td>1.87</td>
<td>(1.51)</td>
<td>(1070)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>-209</td>
<td>1070</td>
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<tr>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>(0.25)</td>
<td>526</td>
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<tr>
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<td></td>
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<tr>
<td>APB</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>-36</td>
<td>757</td>
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<tr>
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<td>1.87</td>
<td>19.9</td>
<td>3.24</td>
<td>0.0020</td>
<td>1.27</td>
<td>-36</td>
<td>757</td>
</tr>
</tbody>
</table>

Others have found it difficult to fit the surface displacements in the northwest. Shen et al. [1996] could not fit the GPS measurements to their single or two plane models, whereas Massonnet et al. [1996] found that a single plane model parallel to the SEP model could not explain the radar interferometry data. These two studies used an additional dislocation plane northwest of the main rupture plane to explain the displacements there. Massonnet et al. concluded that the additional dislocation plane was necessary to model the surface deformation due to a swarm of aftershocks under the Santa Susana Mountains.

Comparison of the calculated vertical displacements with leveling results is more useful than comparison of the vertical displacements measured with GPS. The MRP, FPF, and SEP model results are compared with six leveling lines in the Los Angeles basin (Figures 2.11 and 2.12). Although the error bars are not shown in these plots, a discussion of the leveling errors is given by Wald et al. [1996]. It is clear from Figures 2.11 and 2.12 that the SEP model fits the leveling data best and that the FPF model grossly overestimates the vertical displacement near the epicenter of the Northridge earthquake (e.g., Figure 2.12a).

### Discussion

Although the FPF model is better in a statistical sense than the MRP model, the displacements on the axial planes do not agree with the geological model's predictions. According to equation (2.4), a displacement on the main rupture plane of 1.15 m implies an average of 0.40 m of displacement on APA and 0.20 m of displacement on APB. The average displacements on APA are more than 1.5 times larger than those predicted from the amount of slip on the main rupture
Figure 2.7. Results from the inversion for displacements on the plane of the MRP model. Contours are at 0.5 m intervals. The perspective shown is looking horizontally from the hanging wall block. The displacements shown are hanging wall displacements. The location of the main shock is shown as a solid star.
Figure 2.8. Results from the inversion for displacements on the planes of the FPF model. Contours, perspective and displacements shown are as in Figure 2.7. (a) Displacements on the main rupture plane (M). (b) Displacements on APA. (c) Displacements on APB.
Figure 2.8, continued.
Figure 2.9. Results from the inversion for displacements on the planes of the SEP model. Contours, perspective, and displacements shown are as in Figure 2.7.
plane. The displacements on APB are negligible. Furthermore, it is expected that fault propagation folding accompanying slip on the main rupture plane will share many of the characteristics of the main rupture. The along-strike maximum displacement on the main rupture plane should correspond with the along-strike maximum displacement on the axial planes. Instead, the maximum displacement on APA is located in the northwest corner of the axial plane rather slightly southeast of the center, as on the main rupture plane. The orientations of slip on APA bear little similarity to those on the main rupture plane. The displacements vary across APA, from right-lateral in the northwest corner to left lateral in the southeast corner, whereas the displacements on the main rupture plane do not vary strongly along strike. These observations lead us to conclude that the displacements inferred on APA are not related to fault propagation folding and that there are no resolvable displacements on APB, fault related or otherwise.

Comparison of the FPF model results with leveling data leads to the conclusion that the model does not adequately explain the vertical surface displacements. An inversion including the leveling data was not done, as the evidence leads us to dismiss the FPF model.

There are a number of reasons why the FPF model might fail to adequately model the Northridge surface displacements. First, the FPF model's resolution may not be sufficient to distinguish slip on the main rupture plane from the axial planes. Second, the fault propagation folds may be too gentle to be modeled as kink folds. Third, deformation due to folding may be distributed throughout the hanging wall and may not be confined to the axial planes of the folds. Finally, the deformation on the axial planes may be accomplished interseismically, rather than coseismically. Each of these possibilities is discussed in turn.

The model resolution for displacements on distinct planes in the FPF model is adequate. This is determined by calculating the surface displacements due to a unit of thrust displacement on a single dislocation located near the center of each plane and inverting these surface displacements for the FPF model. Inverting for a unit of displacement on the main rupture plane resulted in a maximum slip on either axial planes of less than 0.05% of the unit slip, indicating that little slip due to faulting on the main rupture plane is projected onto the axial planes. Less than 0.12% of the displacement due to a unit of slip on either axial plane is resolved on to the main rupture plane. It should therefore be possible to distinguish between displacement on an axial plane and displacement on the main rupture plane.

As discussed in the Model section, the model depends critically on whether kink folds are involved in the deformation. The axial planes of folds with large radii of curvature will not behave like discrete planes of deformation. Although kink folds are used to model fault propagation folds, fold style depends on many factors, such as the mechanical properties of each layer [Suppe, 1985].
Figure 2.10. Residual coseismic surface displacements for MRP, FPF, and SEP models. Model results are subtracted from GPS observations.
Figure 2.11. Map of leveling stations and leveling lines shown in Figure 2.12. Epicenter of Northridge earthquake is shown as a star.
Figure 2.12. Comparison of vertical coseismic displacements measured with leveling (gray line with triangles) with calculated vertical displacements for MRP (solid line with circles), FPF (solid line with diamonds), and SEP (solid line with squares) models. (a) Projection of Interstate 5 leveling line data on a southwest oriented line. (b) Projection of State Route 101 leveling line data on a east oriented line. (c) Projection of State Route 118 leveling line data on a east oriented line. (d) Projection of State Route 126 leveling line data on a east oriented line. (e) Projection of State Route 27 leveling line data on a north oriented line. (f) Projection of Interstate 405 leveling line data on a north oriented line.
Figure 2.12, continued.
Figure 2.12, continued.
Within the hinge zone the deformation is distributed. Modeling a wide axial zone requires multiple planes of dislocations parallel to the mechanical layers which shear relative to each other, whereas a narrow axial zone requires only a single plane of dislocations parallel to the axial plane. It is possible that the FPF model failed to resolve deformation on the axial planes simply because deformation was not confined to a sufficiently narrow region. A model including layer-parallel dislocations could resolve this issue.

It is tempting to call upon distributed deformation throughout hanging wall to explain such phenomena as swarms of aftershocks in the hanging wall or the vertical discrepancy in the best fitting fault plane from geodetic versus seismic data. The geological model of fault propagation folding presented here requires that fold-related deformation occurs only in the hinge zones of folds. In this kinematic model, fold limbs do not deform during fault propagation folding: during deformation, mechanical layers move with rigid body translation, shearing only as they pass through the hinge zone and cross the axial plane. For this geological model, only the folds related to the buried thrust on which the Northridge earthquake occurred would deform in response to the earthquake. The discrepancy in the best fitting plane could not be related to layer parallel shear in the hanging wall.

Perhaps the most significant possibility for why the fault propagation fold model fails is that fault propagation folding occurs interseismically. The GPS network in the Los Angeles is being densified in order to locate potentially seismogenic buried faults. Interseismic surface deformation associated with fault propagation folding should resemble that of faults. Care must be taken in interpreting both seismic and interseismic surface displacements so that kink folds are not misinterpreted as faults, and vice versa.

One should use caution in finding a single best fitting dislocation plane, as the solution may represent the average distributed deformation due to both folding and faulting that will yield the incorrect orientation and location of faults. Instead, the dislocation geometry should be chosen based on aftershock locations and focal mechanisms and based on the geology.
References


Chapter 3

Three-dimensional block models of deformation of the Los Angeles region

Abstract

Recent earthquakes such as the 1994 Northridge, California, earthquake underscore the need to understand the kinematics and seismic potential on all faults in the Los Angeles basin, including buried and surface-breaking thrust faults, as well as strike-slip faults. Towards this end a kinematic model is used to relate observations of deformation on geological and geodetic time scales. In this model the crust is divided into a number of blocks, some infinite and some finite in extent, the boundaries of which are described completely as faults. This guarantees that the path integral constraint is satisfied. The blocks are rigid over geological time scales, with all deformation accommodated by slip on bounding faults. The interseismic surface velocities are determined by subtracting the coseismic displacements, divided by the earthquake recurrence time, from the geological rates. Two models are tested: a thick-skinned model, in which ramp faults root in the ductile lower crust, and a thin-skinned model, in which displacement on basal detachment faults is transferred to displacement on ramp faults. It is possible to use the existing geological fault slip data to create a kinematically consistent model. In inverting geological fault slip rates for block motions, the thin-skinned model is marginally preferred to the thick-skinned model in a statistical sense. However, comparison of predicted interseismic velocities to independent GPS observations favors the thick-skinned model.
Introduction

Beneath Los Angeles lies a complicated network of strike-slip and thrust faults, many of which pose a serious seismic threat. The recent Northridge earthquake and previous earthquakes have sent a warning to the scientific community about the potential for large earthquakes on blind thrust faults, as well as surface-breaking faults, beneath Los Angeles. Recent work indicates that there are faults with the potential for a moderately large earthquake, $M_w \sim 6.7$, or the size of the Northridge earthquake, at an average interval of 11 years, or a larger earthquake, $M_w \sim 7.4$, at an average interval of 140 years [Dolan et al., 1995]. These circumstances emphasize the need to understand the geological structures in order to plan for the hazards associated with them.

Geological estimates of location and slip rates exist for many faults in the Los Angeles region. However, as the Northridge Earthquake proved, not all potentially seismogenic faults have been recognized. The block model presented here is used to evaluate the consistency of geological data and to provide estimates for slip on fault segments for which there is no geological data.

In addition to geological evidence, over the past decade many scientists have worked together to build a Global Positioning System (GPS) network throughout southern California with which to measure surface displacements. The geodetic measurements are to be used to study interseismic displacements, as well as to study coseismic displacements as earthquakes occur. Toward this end, the model presented here relates the steady state geodetic observations to the predictions of this geologically based model.

The block model presented here is an extension of simpler fault models, in which interseismic velocities result from displacement below the locked section of a single fault plane or a set of faults [e.g. Savage et al., 1979; Feigl et al., 1993]. Fault models are used successfully to model interseismic deformation, particularly in regions where faults have simple geometries. However, because the path integral constraint is not satisfied, fault models are not in general kinematically consistent: they do not include all the deformation. The path integral constraint specifies that the relative velocity accumulated along a path between two points is independent of the path taken between those points. This problem is important in the application of this method to regions such as Los Angeles which have complicated fault geometries.

Previous purported block models differ from the model presented here in certain fundamental respects [e.g. Matsu-ura et al., 1986; Bennett, 1995; Feigl et al., 1993]. In some models blocks are not truly discrete: the blocks that border the region of interest are finite, as are the bounding faults. Most models either ignore the dip of the faults, specifying that all faults dip at 90°, or include the dip but do not require that faults link up to form discrete blocks. Gaps or
overlaps occur between dipping rectangular dislocations that are not coplanar. For these reasons these models are better for predicting interseismic velocities in the center of blocks and far from the edges of the model than at the model's edges or near fault junctions.

Some numerical models [e.g. Saucier and Humphreys, 1993; Peltzer and Saucier, 1996] are well suited to modeling the large scale tectonics of regions such as southern California, but do not account for subsurface fault geometry, and do not include buried faults, so that they lack resolution near faults. These models also allow permanent strain to accumulate in blocks. However, the permanent strain that accumulates in crustal blocks in these models is assumed to be elastic. This assumption may be valid for very short time, but should not be extrapolated from time scales of years to geological time scales.

Because the purpose of this study is to examine the implications of the geological data and of the assumptions of differing geological models, the model is required to include a path integral constraint and accurate fault geometry. In this block model of the Los Angeles basin the crust is divided into a number of blocks (Figure 3.1), some of which are infinite in extent. The choice of block geometries is discussed later. Block boundaries are described completely as faults. Over geological time scales, the blocks are assumed to be rigid, so that all deformation is accommodated by slip on faults bounding the blocks. This assumption guarantees that the path integral constraint is satisfied. Also, faults either end by truncating against other faults or by extending to infinity. The relative displacement on each fault is completely determined once the block motions are specified. In addition, the subsurface geometry of the fault boundaries is modeled as accurately as possible.

The deformation style in California lends itself well to analysis with block models, as deformation on geological time scales is largely confined to faults [King et al., 1994]. It is relatively straightforward to connect mapped faults to create the boundaries that delimit the blocks.

Interseismic velocities are assumed to be independent of the time since the previous earthquake. For more realistic viscoelastic models of deformation, crustal velocities depend on both geographic location and the stage in the earthquake cycle, with substantial variations for models with Maxwell times short compared to the earthquake repeat time and negligible variations for models with Maxwell times comparable to the time between earthquakes. There is little evidence for substantial variation in interseismic velocities, except in the immediate postseismic time. For example geodetic observations of velocity gradients near the San Andreas fault can be fit reasonably well by models that assume constant interseismic velocities [e.g., Feigl et al, 1993]. Although the assumption of constant interseismic velocity is ad hoc, a more realistic model would require greater knowledge of rheological parameters than currently exists.
Figure 3.1. Block model, superimposed on a fault map of southern California. P, Pacific block; CR, Coast Ranges block; SN, Sierra Nevada block; NA, North American block; M, Mojave block; SM Sierra Madre block; VB, Ventura basin block; SMM, Santa Monica Mountains block; W, Whittier block; Cl, Channel Islands block; PV, Palos Verdes block; NI, Newport-Inglewood block; EL, Elsinore block; SA, Salton block.
There is currently a controversy among geologists studying the Los Angeles basin as to whether the deformation is thick-skinned, in which case ramp faults root in the ductile lower crust, or thin-skinned, in which displacement on basal detachment faults is transferred to displacement on ramp faults (Figure 3.2). This controversy has important implications for seismic hazard. A given slip rate on a ramp fault implies a higher far-field convergence rate in the thin-skinned model than in the thick-skinned model. In addition, the thin-skinned model implies the existence of potentially seismogenic basal detachment faults beneath virtually all of the greater Los Angeles area. An attempt to shed light on this problem is made here by determining which model is more consistent with the geological and geodetic data.

Total deformation occurs by strain buildup during the interseismic period, followed by coseismic displacements. The average rate of interseismic strain accumulation is determined here by subtracting the coseismic displacements, divided by the earthquake recurrence time, from the geological rates. The coseismic displacements are calculated using the standard elastic dislocation model [Okada, 1985].

**Method**

Geological fault slip data are related to relative block motions through a simple transformation in this block model. The relationship between relative block motions and absolute block motions in a specified reference frame is linear. The fault slip data are inverted to determine absolute block motions. A set of block motion description vectors predicts relative block motions, or fault slip rates, between each pair of blocks at points on their shared boundary. In addition, the interseismic velocity is determined at positions within blocks by assuming that the interseismic velocity is the difference between the total, or geological, displacement and the coseismic displacement. The geological displacement is the block motion at those positions. The coseismic displacement is found by calculating the displacement at those positions due to the average yearly coseismic slip on all fault boundaries. The interseismic velocities are the geological rates minus the average yearly coseismic displacements [Hager, et al., *in prep.*].

**Inversion of Geological Data**

Geological fault slip rates \( d_0 \) are inverted to determine the rigid body motion \( \mathbf{m} \) of each block relative to the reference block using a weighted least squares method that minimizes the L2 norm of the difference between the observed fault slip rates and the calculated relative interblock velocities. Each finite block has three degrees of freedom and therefore three unique data are necessary to specify its motion. Only two unique data are required to solve for the motion of an
Figure 3.2. Schematic cross section through models. Each panel shows the horizontal ($v$ and $v \cos \delta$) component of velocity at surface, and the velocities along ramp fault ($v_r$) and detachment or relative block velocity ($v_f$). a) Thin-skinned model. b) Thick-skinned model.
infinite block because its rotation rate is zero. The block motion description vector \( \mathbf{m} \) is linearly related to the fault slip data \( \mathbf{d}_0 \) as explained below.

Geological fault slip rates are taken from data sets in review papers on the geology of southern California [Petersen and Wesnousky, 1994; Dolan et al., 1995; Hufnigle and Yeats, 1995]. Not all the geological slip rates included in these papers were used. A subset of the data was selected to avoid weighting the data by the number of studies done on any particular fault. The data subset was chosen to allow approximately equal weighting to all faults on which slip rates are known. Each slip rate datum is assigned to the midpoint of a representative fault segment on the block boundary.

The relative block velocities \( \mathbf{v} \) are equal to the geological fault slip data, \( \mathbf{d}_0 \) corrected for the dip, if necessary, and rotated from fault to global coordinates.

\[
\mathbf{v} = \mathbf{R} \mathbf{C} \mathbf{d}_0 \tag{3.1}
\]

where \( \mathbf{C} \) is the dip correction matrix and \( \mathbf{R} \) is the local to global rotation matrix.

A dip correction is applied in the thick-skinned model, but not in the thin-skinned model. In the thin-skinned model the fault-perpendicular component of the relative block velocity \( r \mathbf{v}_{\text{perp}} \) is equal to the segment’s fault slip rate \( r \mathbf{v}_{\text{perp}} \) in fault-based coordinates. The ramp velocity is equal to the detachment velocity, and the detachment velocity is equal to the relative block velocity (Figure 3.2a). In the thick-skinned model the fault-perpendicular component of the fault slip is corrected for the dip, because only the horizontal component of the slip rate is included in the relative block velocity (Figure 3.2b). The horizontal fault-perpendicular and fault-parallel components of the segment’s slip rate \( r \mathbf{v}_{\text{perp}} \) are related to the relative block velocity \( r \mathbf{v}_{\text{perp}} \) by

\[
r \mathbf{v}_{\text{perp}} = r \mathbf{v}_{\text{perp}} \cos(\delta). \tag{3.2}
\]

where \( \delta \) is the dip of the fault segment. The dip correction matrix is then used to rotate the fault slip rates \( r \mathbf{v} \) on each of \( M \) fault segments.

\[
\mathbf{C} = \begin{bmatrix}
1 & 0 & & \\
0 & \cos(\delta_1) & & \\
& \ddots & \ddots & \\
& & 1 & 0 \\
& & 0 & \cos(\delta_M)
\end{bmatrix} \tag{3.3}
\]

\( \mathbf{C} \) is diagonal and positive definite. In the solution for the thin-skinned model, \( \mathbf{C} \) is included in all equations, but is set equal to the identity matrix, \( \mathbf{I} \).
The fault-perpendicular and fault-parallel components of the relative block velocity \( r \nu \) for each segment are rotated from their local fault-based reference frame into the global reference frame. The velocity in the global coordinate system \( \nu \) is found from \( r \nu \) by removing the rotation of the strike, \( \alpha \), of the segment. The rotation matrix is applied to each of \( M \) segments.

\[
R = \begin{bmatrix}
\cos(-\alpha_1) & \sin(-\alpha_1) & 0 & 0 & \cdots & 0 & 0 \\
-\sin(-\alpha_1) & \cos(-\alpha_1) & 0 & 0 & \cdots & 0 & 0 \\
0 & 0 & \cos(-\alpha_2) & \sin(-\alpha_2) & \cdots & 0 & 0 \\
0 & 0 & -\sin(-\alpha_2) & \cos(-\alpha_2) & \cdots & 0 & 0 \\
\cdots & \cdots & \cdots & \cdots & \cdots & \cdots & \cdots \\
0 & 0 & 0 & 0 & \cdots & \cos(-\alpha_M) & \sin(-\alpha_M) \\
0 & 0 & 0 & 0 & \cdots & -\sin(-\alpha_M) & \cos(-\alpha_M)
\end{bmatrix}
\]  
(3.4)

The rotation matrix is tridiagonal and has the characteristic that

\[
R^T R = I.
\]  
(3.5)

The strike of the segment may not correspond to the exact strike of the fault at the location where the measurement was made. Fault slip data are not generally reported with fault strike. However, the strike of the segment is expected to be within 10° of the fault's strike.

The position matrix \( A \) relates the interblock velocities to the block motion description vector.

\[
A m = \nu
\]  
(3.6)

The \( A \) matrix is obtained from the equations that relate relative velocities between blocks to absolute velocities relative to some reference block. The relative velocity of block \( i \) relative to block \( j \) is equal to the velocity of block \( j \) relative to block 0, the reference block, subtracted from the velocity of block \( i \) relative to block 0,

\[
\nu_{ij} = \nu_{i0} - \nu_{j0}
\]  
(3.7)

or simply

\[
\nu_{ij} = \nu_i - \nu_j
\]  
(3.8)

The velocity \( \nu_j \) at a point \( x_m \) on a rigid block \( j \) can be expressed in a number of equivalent ways. The most obvious way is as a rotation at rate \( \omega_j \) about the Euler pole of the block \( j \),

\[
\nu_j = \omega_j \times (x_m - x_j^p)
\]  
(3.9)
where \( \mathbf{x}_j^p \) is the location of the Euler pole of block \( j \). This equation is nonlinear in that both \( \omega_j \) and \( \mathbf{x}_j^p \) are unknown. Alternatively, this same equation can be expressed as

\[
\mathbf{v}_j = \omega_j \times \mathbf{x}_m - \omega_j \times \mathbf{x}_j^p
\]

or

\[
\mathbf{v}_j = \omega_j \times \mathbf{x}_m + \mathbf{v}_0.
\]

The first term on the right side of equation (3.11) is the rotational velocity about a new axis located at the origin of the coordinate system, and the second term is a constant, the translational velocity at the origin. In other words, the velocity can be expressed as a rotation rate at \( \omega_j \) about the origin, plus a pure translation \( \mathbf{v}_0 \). In both equations (3.10) and (3.11) the rotation rate is the same. In many cases the origin is chosen either to correspond to the centroid of the block in question or to correspond to an Euler pole [e.g. Donnellan et al., 1993]. In this instance the origin of the global coordinate system is the rotation axis for all blocks.

The equation for each interblock velocity is then

\[
\mathbf{v}_j = \begin{bmatrix} 1 & 0 & -y_m \\ 0 & 1 & x_m \end{bmatrix} \begin{bmatrix} u_0 \\ v_0 \\ \omega \end{bmatrix} = \mathbf{A}' \mathbf{m}'
\]

where \( u_0 \) and \( v_0 \) are the components of the translational velocity of the block at the origin, \( \omega \) is the polar rotation rate about the origin, and \( \mathbf{m} \) is the block motion description vector for that block. Equation (3.12) is substituted into equation (3.8) and expanded along each boundary for which there is fault slip data. The \( \mathbf{A} \) matrix is formed by placing each submatrix \( \mathbf{A}' \) into \( \mathbf{A} \) in the three columns of \( \mathbf{A} \) that correspond to that block and the two rows that correspond to that equation for the relative block velocity.

There is some error in the formulation of \( \mathbf{A} \) in that it involves the position, \( x_m \) and \( y_m \), at which the fault slip data were recorded. The position is the location of the midpoint of the representative fault segment to which the fault slip data were assigned, which is accurate to within approximately 25 km. This error will affect the estimated rotation rate of the blocks because fault slip varies along the boundary of a rotating block.

The full block motion description vector \( \mathbf{m} \) has 3\( \times \)N elements, where N is the number of blocks. The \( \mathbf{m} \) vector is formed from the \( \mathbf{m}' \) matrices.
The full block motion vector is determined for finite blocks, including the translational velocity and the rotation rate. For infinite blocks only the translational velocity is determined. This avoids arbitrarily large fault-perpendicular displacements along infinite block boundaries. The rotation rates of infinite blocks are removed from the m vector, as are the corresponding columns of the A matrix.

Substituting equation (3.1) into equation (3.6) and rearranging yields

\[ C^{-1} R^T A m = d_0 \]  
\[ G_0 m = d_0 \]  

The diagonal weighting matrix is based on a diagonal covariance matrix of the fault slip data, reflecting the assumption that the data are independent. That is, each geological fault slip rate estimate is assumed not to be influenced by the rate estimates of surrounding faults. It is difficult to formulate an absolute covariance matrix given that the errors in each equation include uncertainties in strike of as much as 10°, possible uncertainties in dip of 5°, and uncertainty in position of up to 25 km. A relative covariance matrix is assigned to the fault slip rates by a simple process. An error of 1 mm/yr is assigned to each fault slip rate component that is available, and an error of 1000 mm/yr is assigned to each component that is not available. This is necessary because in places only one component of the fault slip data is known. The weighting matrix \( W \) is obtained from the covariance matrix \( V \) and is used to weight the system of equations.

\[ V^{-1} = W^T W \]  
\[ W G_0 m = W d_0 \]  

Although the covariance matrix represents relative rather than absolute uncertainties, units of mm/yr are used so that the covariance matrix can be rescaled to reflect absolute errors in the
estimate of the NRMS. The resulting system of equations is then solved using the generalized inverse [e.g., Menke, 1989].

\[ Gm = d \]  \hspace{1cm} (3.18)

**Coseismic Displacements**

The coseismic displacements at a given point are approximated using dislocations in an elastic half space to model the fault boundaries. The displacement on each dislocation is determined from the total yearly displacement on that boundary. Coseismic displacements are calculated based on the assumption that the upper 16 to 20 km deforms coseismically throughout the region of interest.

The dislocations form a three dimensional interlocking network of trapezoidal and triangular dislocations that closely approximate the fault geometry. The surface expressions of these dislocations form the block boundaries. Coseismic displacements on these dislocations are calculated using the approximation scheme described below.

There are two concerns in devising a method for approximating dislocations on rotating blocks with irregular boundaries. First, although the solution for the surface displacements due to slip on a rectangular dislocation is straightforward, the same is not true for other planar shapes, such as trapezoids or triangles. Second, the relative interblock velocity varies linearly along the length of any fault segment that comprises the boundary of a rotating block.

These two problems are addressed by separating each trapezoid into a rectangle and two triangles (Figure 3.3a and 3.3b). Triangular dislocations are approximated as rectangular subdislocations and the dislocation displacements are interpolated along the length of the triangle (Figure 3.3c). Each rectangular dislocation is subdivided into smaller subdislocations (Figure 3.3d). Dislocation displacements are linearly interpolated along the length of both triangles and rectangles.

The length of any subdislocation is set to 1.0 km or less, whatever size is necessary to fit an integer number of equal length subdislocations in each dislocation. The maximum error due to the approximation of triangles by rectangular dislocations is determined for a test case triangular segment dipping 45° in which the upper long edge intersects the surface. The maximum horizontal error is 8.3% of the horizontal slip rate, and falls off to less than 1% within one subdislocation length of the fault segment. The maximum error due to the along strike variation in displacement is calculated for a vertical surface-breaking dislocation. In this case, the maximum horizontal error is 9% of the difference between the displacement difference along the length of the fault. The error falls to less than 1% of the displacement difference within one subdislocation length from the fault.
Figure 3.3. Illustration of technique for approximation of trapezoidal dislocation. a) Fault segment. b) Fault segment trapezoid is separated into a rectangle and two triangles. c) A rectangle is subdivided into subdislocations to approximate the variation of the displacement along strike. d) Each triangle is subdivided into rectangular subdislocations to approximate the shape of a triangle, and to approximate the variation of displacement along strike.
Since velocities are not evaluated close to the faults, these approximations are adequate for the purposes of this study.

It is not possible to obtain a model that is consistent with the observed slip rates using a constraint that there are no fault-perpendicular displacements on vertical faults. In the calculation to determine the total displacements, fault-perpendicular displacements are allowed across vertical block boundaries. The inversion for block motions therefore allows fault-perpendicular displacements on vertical faults, displacements which are modeled with the fault-perpendicular component of the dislocation displacement \( u_3 \) (Figure 3.4).

Shortening across vertical faults may be accommodated by folding and minor thrust faulting parallel to and within a few kilometers of the vertical fault. The San Andreas fault, for example, is characterized by mountains along much of its length, probably due in part to such shortening.

The question of whether the fault-perpendicular component of the dislocation displacement on a vertical dislocation can adequately approximate shortening on minor parallel thrust faults was addressed by comparing two simple models of fault-perpendicular shortening. In the first model deformation is accommodated by the down-dip component of the displacement \( u_2 \) along two faults that meet at the surface and dip 45° in opposite directions. In the second model shortening is accommodated by the \( u_3 \) component of the displacement along a single vertical fault. In both models the faults have unit length and extend to unit depth, with a unit of horizontal fault-perpendicular displacement. Figure 3.5 shows the horizontal components of surface displacement for each model as well as the difference between the two. The maximum difference between each of these models is approximately 10%, and is greatest in the along-strike direction.

As the lengths of the dislocations increase, the differences between the horizontal components of the surface displacements decrease. For infinitely long faults the horizontal components are essentially identical for both models. Because the exact mechanism by which fault perpendicular shortening occurs on vertical faults is not known, and because the vertical faults in this model of the Los Angeles region are much longer than they are deep, the method used is found to be an adequate approximation of the \( u_3 \) component of the dislocation displacement, with two caveats. First, the error is largest where vertical faults bend or truncate against other faults. Second, although vertical displacements are not considered in this study, it should be noted that these two simple models have approximately equal and opposite vertical displacements.
Figure 3.4. The three components of displacement on each dislocation: strike-slip, $u_1$; dip-slip, $u_2$; and tensile, $u_3$ [from Okada, 1985]. Other symbols are fault plane width, $W$; length, $L$; depth, $d$; and dip, $\delta$. 
Figure 3.5. Comparison of horizontal displacements due to one unit of horizontal shortening on two symmetrically arranged thrust faults versus on one vertical fault. a) Displacements due to slip on two thrust faults which dip 45° away from each other but meet at the surface. Outline of dislocations is shown. b) Displacements due to shortening perpendicular to a single vertical fault, which is shown as a solid line. c) The difference between the displacements on a) and b).
Figure 3.5, continued.
Interseismic

The interseismic velocity at a position within a block is obtained by first calculating the total yearly displacement due to block motion, then subtracting the average coseismic surface displacements per year. In calculating the interseismic velocities for the thin-skinned model the fault-perpendicular component of the total displacement is adjusted for the following reason. The thin-skinned model of crustal deformation requires that the horizontal fault-perpendicular component of surface displacement decreases as crustal material moves through the fault ramp kink fold and up the fault ramp (Figure 3.2). Within the kink band the fault-perpendicular component of the total displacement is multiplied by the cosine of the dip of the fault.

\[ v_{pp} = v_{pp0} \cos(\delta) \]  

(3.19)

In a study of the coseismic deformation associated with the Northridge earthquake, no evidence was found for coseismic deformation associated with kink band deformation during that earthquake [Souter and Hager, 1997]. Although this is a single event, no other studies of this kind have been performed. It is therefore assumed that if deformation on kink bands occurs it occurs interseismically. In the absence of evidence for any variation in interseismic kink band displacement through the earthquake cycle, it is assumed to be constant in time. The location and width of kink bands used in the thin-skinned model are defined based on the geometry of fault ramps and flats (Figure 3.6).

Models

The scale of the problem is limited to the North American and Pacific plate boundary in the Los Angeles basin. This region is divided into fourteen blocks (Figure 3.1). The motions of the westernmost block, the Pacific block, and the easternmost block, the North American block are not inverted for. The North American block is used as the reference block: its motion is defined to be zero. The velocity of the Pacific Plate is assigned the NUVEL 1A plate velocity [DeMets et al., 1994].

The remaining blocks are defined based on the map traces of faults in the Los Angeles basin and southern California. The level of detail included in the model decreases away from the area of interest; although the block boundaries in the Los Angeles basin coincide largely with faults, block boundaries at distances of greater than 100 km from Los Angeles do not everywhere correspond to a mapped fault. For example, the North American block boundary south of the Garlock fault is represented as a single fault although a number of subparallel faults take up the deformation. Such simplifications are justified for the purpose of this model - to understand
Figure 3.6. a) Location of kink bands used in thin-skinned model. Kink bands are shaded gray. b) Cross sectional view illustrating how extent of kink bands are determined. For a ramp-flat system with a horizontal detachment, a ramp fault dipping at angle $\delta$ and having width $w$ and interlimb angle $2\alpha$, the width of the kink band is equal to the width of the fault plane.
Figure 3.6, continued.
Figure 3.7, continued.
deformation in the Los Angeles basin. The surface deformation in the Los Angeles basin depends little on whether deformation in the Mojave occurs on one fault or on several faults.

Block boundaries are chosen such that they connect as simply as possible wherever faults are not observed to intersect. All boundaries are required to either truncate against other block boundaries, or extend infinitely. The boundaries and faults are projected onto a two dimensional horizontal plane using a Lambert conformable conical projection. The projection is centered at longitude 241.50° and latitude 34.40°. The errors due to using a non-spherical coordinate system are small compared to the uncertainties in the fault slip rates and locations as well as the inherent errors in connecting boundaries.

As mentioned above, two scenarios for the deformation in the Los Angeles basin are addressed: a thin-skinned model and a thick-skinned model. In both models faults truncate against other faults, and do not cross. The geometry of the thin-skinned model (Figure 3.7a) is based on geological maps and cross sections [Dolan et al, 1995; Davis et al., 1989]. In this model most of the Los Angeles basin and regions to the south are underlain by subhorizontal detachment faults, and are modeled here as seismogenic. Detachment faults increase in depth from west to east, and from south to north, to a maximum depth of 20 km (Figure 3.8a). The San Andreas fault extends to a depth of 20 km everywhere except along the Brawley seismic zone, where it extends to 5 km. It is assumed that any fault below this depth is aseismic. For example, the basal detachment fault of the Sierra Madre fault is assumed to be aseismic, and therefore it is not included in the thin-skinned model. Fault slip rates (Table 3.1 and Figure 3.9a) are assigned to surface-breaking faults only; buried faults do not contribute to the inversion.

The faults in the thick-skinned model are nearly identical to the surface-breaking faults in the thin-skinned model (Figure 3.7b). There are no detachment faults in the thick-skinned model. The ramp faults extend to the ductile lower crust which is assumed to be 16 km deep in the south and west, and 20 km deep in the north and east (Figure 3.8b). The fault slip rates in the thick-skinned model are identical to those of the surface-breaking faults in the thin-skinned model (Figure 3.9a).

**Results**

The block motion description vectors are shown in Table 3.2 and Appendix I for both the thin-skinned and thick-skinned models. In these inversions, the NRMS of the thin-skinned model is 2.8, and of the thick-skinned model is 3.1. Because the errors are assigned in a relative sense, the NRMS of each model is scaleable by the true errors. If, for example, all fault slip rates have errors of 2 mm/yr, the NRMS for each model is divided by 2, yielding a NRMS of
Figure 3.8. Illustrative SW-NE cross section approximately along cross section A-A' of Figure 10c, a) thin-skinned b) thick-skinned model. Fault labels are as in figure 3.7.
### Table 1. Geologic slip rates and estimated slip rates for inversion for block motions.

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Figure 3.9. Map of observed and calculated slip rates. a) Observed geological slip rates used in inversion for block motions. Open squares indicate only fault-perpendicular slip rate is constrained. Open triangles indicate only fault-parallel slip rate is constrained. b) Calculated slip rates for thin-skinned model. Arrows show calculated slip rates at locations where input slip rates are used. Error ellipses are 95% confidence ellipses. c) Calculated slip rates for thick-skinned model. Black arrows and error ellipses are as in b).
Figure 3.9, continued.
Figure 3.9, continued.
1.4 and 1.6, respectively. The NRMS is a measure of the internal consistency of each model, relating how well the geological fault slip rates agree with the predictions of the block model, and how well the geological data can be integrated into a kinematically consistent model.

Only five blocks have large rotation rates in either the thin-skinned or thick-skinned model: the Mojave, Whittier, San Gabriel, Ventura basin, and Santa Monica Mountains blocks. Rotation rates are not large: the thin-skinned Santa Monica Mountains block has the largest rotation rate of 0.059 μrad/yr, or 3.4 °/Myr. In both models the Whittier block rotates counter clockwise, and the San Gabriel, Ventura basin, Santa Monica Mountains and Mojave blocks rotate clockwise relative to North America. The Whittier block is also the most poorly constrained. It has consistently high one-sigma errors in all three components of the block motion vector (see Table 3.2). The Mojave block is also poorly constrained.

Because the Coast Ranges block and the Channel Islands block do not rotate, the shortening across the Ventura basin is easily calculated by subtracting the motion of one block from the other. As expected, the thin-skinned model predicts higher convergence rates across the Ventura Basin than the thick-skinned model: 10.6 mm/yr versus 6.7 mm/yr, respectively. The azimuth of convergence is slightly more northward in the thin-skinned model, 335°, than in the thick-skinned model, 323°.

Displacement rates change sense along many faults but generally only where the magnitudes of displacement rate are less than approximately 1 mm/yr, such as along the San Gabriel fault. In general, the sense and magnitude of observed versus predicted geological slip rates agree within a few mm/yr (Figure 3.9 and Table 3.1). There are a few exceptions. Also of interest are the model predictions of strike-slip displacements on thrust faults, as well as shortening or extension on strike-slip faults.

Along the faults bounding the Los Angeles basin, model predictions of shortening are not consistently higher or lower than geological estimates. Along the Cucamonga fault both models predict generally higher amounts of shortening, 4.8 to 7.6 mm/yr, than the geological estimate of 4.5 to 5.5 mm/yr. However, on the Sierra Madre fault, which is separated from the Cucamonga fault by its intersection with the Santa Monica Mountains fault, the predictions of both models agree with geological estimates, at 3.1 mm/yr. The models also predict up to 2 mm/yr of left lateral strike-slip motion on the Cucamonga, whereas along the Sierra Madre fault the models predict up to 3.0 mm/yr of right lateral strike-slip motion. Along the Whittier fault the model predictions agree with the geological rates of 2.5 mm/yr of right lateral displacement. The models also predict 1.2 to 1.9 mm/yr of thrust motion on the Whittier fault.
Table 3.2. Block motion description vectors

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In the Ventura basin the model predictions of slip rates on the San Cayetano and Oakridge faults are slightly lower than geological observations. Small amounts of left lateral motion are also predicted on those thrust faults. The models also predict up to 1 mm/yr of right lateral motion on the Santa Susana fault.

On the western section of the Santa Monica Mountains fault, west of the intersection with the Palos Verdes fault, the geological estimates of slip are within 2 mm/yr of the amount of shortening predicted by the models. However, both models also predict significant right lateral slip, up to 4.9 mm/yr. On the section of the Santa Monica Mountains fault east of the Palos Verdes fault both models predict less than 1 mm/yr of left lateral motion, but agree with the geological estimates of dip-slip motion.

On the southern subparallel strike-slip faults the models generally predict lower strike-slip rates than the geological estimates. Along the San Jacinto fault the thick-skinned model predicts 6.2 to 8.6 mm/yr of strike-slip motion, and the thin-skinned model predicts 5.0 to 9.3 mm/yr, generally lower than the observation of 10 mm/yr. On the Elsinore fault geological slip rates of 5 mm/yr are again higher than the thick-skinned model predictions of 4.0 mm/yr of right lateral slip but agree with the thin-skinned model predictions. On the southern San Andreas fault the models both predict approximately 28 mm/yr, agreeing with geological estimates of 27 mm/yr.

On the Garlock fault the geological slip rate estimate is 8 mm/yr of left lateral motion, whereas the models predict over 9 mm/yr, as well as up to 5.3 mm/yr of thrust motion. On the eastern California shear zone the geological estimate of 10 mm/yr of right lateral motion is only slightly higher than both the thick-skinned prediction of 8.6 mm/yr and the thin-skinned model's prediction of 8.4 mm/yr.

The observed long term velocities relative to North America at GPS sites in and adjacent to the Los Angeles Basin (Figure 3.10a) are compared with calculated interseismic velocities for the thin-skinned and thick-skinned models to determine whether these models reasonably approximate the present day deformation as well as deformation observed over geological time. Residuals for the two models are calculated by subtracting the observed GPS velocities from calculated interseismic velocities (Figure 3.10b and 3.10c). The ambiguity in the model reference frame relative to the GPS observations is removed by subtracting the average residual of the Channel Islands block from the residuals. The normalized root mean square (NRMS) is calculated for each model relative to the GPS observations by

$$\| r \|_2 = \left\| W \left( v_{gps} - v_{est} \right) \right\|_2$$

(3.20)
Figure 3.10. GPS velocities. Error ellipses are 95% confidence ellipses. a) Observed interseismic velocities from GPS, relative to North America. b) Difference between GPS velocities and results of thin-skinned model with average Channel Islands block residual removed. c) Difference between GPS velocities and results of thick-skinned model with average Channel Islands block residual removed. Profile and stations in gray are shown in Figure 3.11a and 3.11b.
Figure 3.10, continued.
Figure 3.10, continued.
Where $W$ is a weighting matrix obtained from the covariance matrix $V$ of the GPS observations $v_{gps}$ by

$$V^{-1} = W^T W,$$  \hspace{1cm} (3.21)

and $v_{est}$ is the vector of interseismic velocities calculated at the GPS sites. In the reference frame with the average residuals of stations on the Channel Islands block removed, the NRMS of the thin-skinned model is 6.9, and the NRMS of the thick-skinned model is 4.9.

In order to obtain a more detailed picture of the two models, as well as how they compare to GPS observations, the interseismic velocities were calculated along a cross section that passes through the Los Angeles basin, crossing the San Andreas fault and the Mojave block (Figure 3.11a and b). This cross section crosses a number of faults, including the Palos Verdes, Newport-Ingleswood, Whittier, Cucamonga, and San Andreas faults. There are plans to install an number of continuous GPS stations in the SCIGN (Southern California Integrated GPS Network) along this profile.

**Discussion**

The results of the slip rate inversion for block motions favor the thin-skinned model. This is due in part to the way in which the errors are assigned to the fault slip data. In the inversion, the same errors are assigned to fault-perpendicular displacements which are smaller in the thick-skinned model than in the thin-skinned model. This results in larger relative errors in the thick-skinned model. In contrast, the comparison of calculated interseismic velocities with independent GPS data favors the thick-skinned model to the thin-skinned model. The comparison with GPS is independent of the models and their covariances, involving only the results of the inversions. In this comparison, the thick-skinned model significantly fits the data significantly better than the thin-skinned model.

An important aspect of this block model is that fault deformation acts as a system, in that changing the slip rate estimate on one fault affects estimates of slip rates on nearly every other surrounding fault, possibly in unforeseen ways. For example, requiring that fault-perpendicular displacement decreases from 8 mm/yr to 1 mm/yr along a boundary between two blocks requires large relative block rotations, which will in turn propagate to the other blocks, causing large variations in fault-perpendicular displacements on many faults.

The block model is also useful for understanding the relationship between the distribution of shear on subparallel strike-slip faults in the south and the distribution of shortening on thrust faults in the Ventura basin, Los Angeles basin, and Sierra Madre and Santa Monica Mountains (Los Angeles area for short). The strike-slip fault system can be seen in effect as feeding the thrust
Figure 3.11. Illustrative cross-sections through Los Angeles Basin comparing GPS velocities relative to North America with calculated velocities from the thin-skinned and thick-skinned models with the average Channel Islands block residual removed. The locations of stations used in this profile are shown in gray in Figure 3.10c. The solid line is the thin-skinned model. The dashed line is the thick-skinned model. a) Component of the interseismic velocity perpendicular to the Pacific-North American plate boundary. b) Component of the interseismic velocity parallel to the Pacific-North American plate boundary.
fault system. If, for example, all right lateral shear and no north-south shortening were taken up on the San Andreas fault, and if no right lateral shear were accommodated on the Palos Verdes, Newport-Inglewood, Elsinore or San Jacinto faults, then the total north-south shortening rate across the Los Angeles area would attain its highest possible value. If, on the other hand, right lateral shear were evenly distributed on subparallel fault strands, then the total north-south shortening would increase from east to west. The latter is closer to what is observed in our models. However, the complications of block rotations and shortening accommodated on the central section of the San Andreas fault reduce the effect.

Comparison of results of both models with GPS observations shows that both models yield surprisingly good predictions of the velocity field considering the simplicity of the models, and the possibility that the geological data reflect past deformation, not current slip rates. In the thin-skinned model there are systematic north and northeast directed errors in the Coast Ranges and Sierra Nevada blocks, respectively (Figure 3.10b). These probably result in part from higher far field displacements for a given amount of slip on a thrust fault in the thin-skinned model than in the thick-skinned model. The geological estimates of shortening on thrust faults in the Ventura and Los Angeles basins would therefore be too high in the thin-skinned model.

The residuals resulting from the thick-skinned model are generally within the 95% error ellipses (Figures 10c and 11). The systematic north and northeast residuals in the Coast Ranges and Sierra Nevada are smaller than in the thin-skinned model, but are still present. The misfit could be decreased in the Coast Ranges block by splitting it into two blocks with the addition of another fault parallel to the San Andreas fault. This would decrease the convergence across the Ventura Basin, causing it to agree with geological convergence rates [Donnellan et al., 1993].

In both models there are small systematic misfits near the Brawley seismic zone in the southeast. Not surprisingly, the residuals of both models are smallest within the Los Angeles and Ventura basins, where the block boundaries correspond most accurately to faults.

In the Mojave block, where the models lack detail, the fit is poor (Figure 3.10 and 3.11). As shown in Figure 3.11, within the Mojave block the component of the velocity parallel to the Pacific-North American plate boundary is resolved onto the San Andreas fault rather than distributed across the block. The models could be improved by subdividing the Mojave block, incorporating fault locations and slip data on faults north and east of the central section of the San Andreas fault.

Block rotation rates are generally consistent with previous paleomagnetic work. The Ventura Basin, the Santa Monica Mountains, and San Gabriel blocks rotate in a clockwise sense at 1-2°/Ma, although the rates of rotation are much lower than paleomagnetic estimates of 3.6-
7.2°/Ma [Hornafius, 1985; Homafius et al., 1986; Kamerling and Luyendyk, 1985]. The paleomagnetically determined rotation history of the Mojave block is complex, and involves both clockwise and counterclockwise rotations [Luyendyk et al., 1985; Ross et al., 1989]. In the thin-skinned model the rotation rate of the Mojave block is 3°/Ma in a clockwise sense. However, the errors on the Mojave block motion are high, suggesting again that it should be subdivided into smaller blocks to reflect its complex geology. Rotation rates for the other blocks are not easily related to the paleomagnetic results because the blocks in this model do not correspond to blocks in paleomagnetic studies.

Rotation rates may be small in these inversions due to the large block size. For example, it is necessary to assign fault slip data to more than one segment along the San Cayetano fault to avoid an unrealistically large rotation rate, which would result in normal motion in the west and thrust motion in the east. In addition, the Coast Ranges block, which here includes most of the western Transverse Ranges, could be subdivided by including the White Wolf fault. Smaller blocks that rotate at relatively higher rates than those predicted here do not cause large variations in fault-perpendicular displacement along their boundaries.
References


Chapter 4

Discussion

The inversion for the fault propagation folding model (FPF) presented in Chapter 2 is better in a statistical sense than the competing model of deformation confined to the main rupture plane (M), however the fault and axial plane deformation resolved by the FPF model does not agree with geological predictions for the distribution of shear on the axial planes. Three preferred possibilities are presented for the lack of evidence for the FPF model. First, the folds may be too gentle to be modeled as discrete planes. Second, the deformation on those planes may occur interseismically. Third, fault propagation folding may not occur. Of these three choices, the first two could be tested systematically.

Gentle folds, which have wide axial zones, are not well represented by the model developed in Chapter 2. In the future a model could be developed to test for deformation on finite width fault propagation fold axial zones. This could involve inverting for deformation on a model similar to that presented here, but on dislocations parallel to the fault, rather than to the axial planes. In gentle folds, fault-parallel planes are the planes across which the deformation occurs in the fold hinge. In contrast, on kink folds the deformation is confined to, and can be modeled as occurring on, the axial plane itself. The problem in implementing such a model is that it is underconstrained. Positivity constraints and smoothing alone would not necessarily suffice to regularize the problem, and more constraints would have to be found.

Other future work on this topic could focus on determining whether kink band deformation occurs interseismically. Dense geodetic coverage of a fault such as the Northridge earthquake fault over time would give us a more accurate picture of the entire seismic cycle, including the effects of fault related folds. As the number of GPS sites in southern California is only likely to increase in the years to come, the potential for in-depth study of earthquake-prone thrust faults is promising.

The results presented in Chapter 3 indicate that geological observations of fault slip data are not only consistent with a kinematic model, but that they are generally consistent with geodetic observations. That they are consistent with geodetic data is an indication that fault slip rates may not have changed significantly over time. Future refinements of the geologically based block model should focus on determining the time evolution of fault slip rates more precisely. Inversion of geodetic data for block motions is one of clearest means to this end.

It is tempting to look at the map of geodetic observations (Figure 3.10a) and infer that they can be inverted for block motions. However, preliminary inversions of geodetic data for block
motions indicate that there are fundamental problems in the inversion. GPS observations alone do not contain enough information to invert for current block motions. Coseismic displacements are poorly constrained by stations far from faults. Also, block rotations are currently poorly constrained because too few stations exist on each block.

It is recommended here that future work should include an inversion of geodetic data for block motions using geological constraints. A joint inversion using both data sets is also a possibility, however such an inversion would not give any information on the time evolution of deformation. Instead positivity constraints can be applied to well understood faults to ensure that fault displacements are constrained to have the correct sense of offset.

Additional geological information may be incorporated into the block model as it becomes available. Geological data from poorly represented regions such as the Mojave will have a positive impact on our understanding of the large scale geological framework. In addition, greater coverage of southern California with GPS stations will have a tremendous impact on the quality of future inversions. As our estimate of fault slip rates in southern California becomes more reliable, so will our ability to determine seismic hazard.

It is difficult to draw strong conclusions from the studies in Chapters 2 and 3 on whether deformation in the Los Angeles region is thick-skinned or thin-skinned. However, the weight of the evidence is tilted slightly in favor of the thick-skinned model. No evidence was found to support the occurrence of coseismic fault propagation folding in first study. In the second study, although inversion for block motions in the thin-skinned model was statistically better than in the thick-skinned model, the reason probably lies in the construction of the data covariance matrix. Comparisons of the results of the two models with independent geodetic observations is an indication of the how well each model predicts current deformation. The thick-skinned model is superior in this respect.
### Appendix I. Segment displacements for Thin-skinned and Thick-skinned models

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