Analysis of Postseismic Processes: Afterslip, Viscoelastic Relaxation and Aftershocks

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Thesis presented for the degree of
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January 2010
Declaration

I declare that this dissertation represents my own work unless referenced to the contrary in the text. All references and other sources used have been appropriately acknowledged in the work. No part of this Thesis has been submitted elsewhere for the purpose of academic examination, either in its original or similar form.
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Acknowledgements

My Ph.D study is financed by DAAD (Deutscher Akademischer Austausch Dienst). The facilities are provided by Deutsches GeoForchungsZentrum, Potsdam University and Ruhr University Bochum. I am very much indebted to these institutions and Prof. Dr. Jochen Zschau for granting me the opportunity to make my Ph.D.

I would thank my supervisor, Prof. Dr. Frank Roth, for inviting me to Germany and his important guidance in these years, and also to Dr. Rongjiang Wang for his technical supports and many guiding discussions.

I would sincerely thank Sebastian Hainzl and Bogdan Enescu, who have been of fundamental help for my Ph.D study. They encouraged me and helped me get through a difficult time in the beginning, and never end to help me solve the problems. With lots of patience, they always carefully read my first-version manuscript, and taught me the better way to do the scientific work. Without their help, this study would have hardly been completed. I am also very grateful to Yehuda Ben-Zion for many valuable discussions and advices in these years, and to Birger Lühr who patiently answered my various questions.

I would cordially express my thank to Junping Chen, Jürgen Klotz, Sinan Özeren, Zhengkang Shen and Shengli Ma for their kindly supports on the GPS data so that I could carry out the work on my Ph.D topic, to M. E. Pritchard and Francisco Lorenzo-Martín for their providing me the earthquake source models, to several reviewers who carefully read and helped to improve the papers I worked on. I am also very grateful to Heiko Woith, Thomas Walter, Claus Milkereit, Monika Sobiesiak, Onno Oncken for their constructive discussions, to many GFZ colleagues for their important supports in the three years.

I will never forget the help and supports of my friends, Silke Eggert, Domenico DiGiacomo, Matteo Picozzi, Gilbert Brietzke, Angelo Strollo, Samira Alipour, Manoochehr Shirzaei and many Chinese friends. Their friendship was so important for my easy and happy life in Germany.

I would like to thank my colleagues of CENC, who gave me lots of concern and took care of the routines in China.

Finally, I would express my heartiest thank to my parents and especially to my husband. They always stand by me and give me the power to move on. Very importantly, they took most of the responsibility in taking care of my daughter.

From the bottom of my heart, THANK YOU!
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Chapter 1  Introduction

The earthquake cycle is usually divided into three time intervals, interseismic, coseismic (the major earthquakes) and postseismic interval. In the interseismic interval, stress builds up around the locked fault zone and produces seismicity. The regional seismicity mostly accounts for only a small portion of the strain accumulated during the interseismic interval. The large portion is usually released by the mainshock itself and the postseismic processes. In the postseismic interval, besides abundance of aftershocks, aseismic deformation occurring in the lithosphere is also evident. It has been found that the seismic moment produced by postseismic deformation is sometimes comparable to or even higher than the main event itself [Pritchard and Simons, 2006, and references therein]. For example, it has been reported that the postseismic deformation in the first 50 days following the 1989 M7.4 Sanriku-oki earthquake in Japan released seismic moment close to the mainshock [Plank and Langmuir, 1998]; the GPS measured postseismic deformation released between 30% and 50% of coseismic moment release in the first 45 days and 5 months following the 2004 M9.0 Sumatra-Andaman earthquake [Hashimoto et al., 2006; Subarya et al., 2006]. Compared to the aftershocks, the seismic moment released by postseismic aseismic deformation is usually far higher. For example, the postseismic deformation in the first 80 days following the 1999 M7.4 İzmit earthquake released seismic moment 4 times higher than the aftershocks [Reilinger et al., 2000]; Chlieh et al. [2004] showed that about 60% (vs. 40%) of the postseismic moment release in 3.3 years following the 1995 M8.1 Antofagasta earthquake attributes to aseismic deformation (vs. aftershock seismicity). Therefore, postseismic relaxation has been recognized as a very important component during the seismic cycle.

The stress change produced by postseismic deformation might influence earthquake interactions and affect earthquake risk estimate in a wide spatial and temporal domain. For example, the postseismic deformation following the İzmit earthquake seems to have further loaded the fault segment of the Düzce earthquake that occurred after 87 days [Hearn et al., 2002]. The 1992 Landers earthquake has been thought to trigger the 1999 Hector Mine earthquake through postseismic stress transfer [Freed and Lin, 2001; Zeng, 2001; Pollitz and Sacks, 2002]. In the time scale of decades, it was indicated that postseismic relaxation builds up more stress along the North Anatolian fault (NAF) than tectonic loading, and might have increased hazard for a subsequent big event [Lorenzo-Martin et al., 2006a]. On large spatial scales, it was shown that two seismic gaps near southern Peru and north Chile are stressed by postseismic relaxation following the mega earthquakes in the past 60 years along the South American fore-arc [Casarotti and Piersanti, 2003]. Postseismic relaxation is also likely to affect the occurrence of the aftershocks. Postseismic creep downdip of the fault zone was thought to drive the aftershock occurrence [Perfettini and Avouac, 2004]; and viscous diffusion of stress in the upper mantle might lead to the observed diffusion of seismic activity with increasing time after the mainshock [Marsan and Bean, 2003].

Because of the aseismic character, postseismic relaxation can be observed only through
geodetic measurements. Postseismic deformation on the surface was firstly documented in 1968 [Smith and Wyss, 1968] following the 1966 Parkfield earthquake in California, and was measured by strainmeters and a small-scale geodetic network. In the recent 20 years, the highly developed space geodesy, particularly, Global Positioning System (GPS) and interferometric Synthetic Aperture Radar (InSAR), has captured valuable postseismic deformation on the surface. Detailed postseismic deformation has been reported after many continental earthquakes, such as, 1989 M7.1 Loma Prieta earthquake [Savage et al., 1994; Bürgmann et al., 1997; Segall et al., 2000]; 1992 M7.3 Landers earthquake [Shen et al., 1994; Savage and Svarc, 1997]; 1994 M6.7 Northridge earthquake [Donnellan and Lyzenga, 1998]; 1997 M7.6 Manyi earthquake [Ryder et al., 2007]; 1999 M7.1 Hector Mine earthquake [Pollitz, 2001]; 2001 M7.8 Kunlun earthquake [Zhang et al., 2007], and after earthquakes that occurred in the subduction zone, such as, 1960 M9.5 Valdivia earthquake [Khazaradze and Klotz, 2003; Lorenzo-Martín et al., 2006b]; 1994 M7.6 Sanriku-Haruka-Oki earthquake [Heki et al., 1997]; 1995 M8.1 Antofagasta earthquake [Chlieh et al., 2004; Pritchard and Simons, 2006]; 2003 M8.0 Tokachi-oki earthquake [Miyazaki et al., 2003], 2004 M9.2 Sumatra earthquake [Pollitz et al., 2008].

To date, the open question concerning the postseismic relaxation is which process is significantly involved in the observed deformation. Postseismic deformation has so far been proposed due to various mechanical origins, ranging from poroelastic rebound, afterslip along the strike or along the dip of the coseismic fault plane, to distributed or localized deformation in a ductile middle/lower crust or upper mantle (namely ductile creep or viscoelastic relaxation). The three mechanisms are likely predominant in different temporal and spatial domains [Segall, 2004]. Poroelastic rebound occurs mostly in a short time period of months after a large earthquake and/or in a small scale area, usually within 10-30 km from the fault [Peltzer et al., 1998; Jónsson et al., 2003; Freed et al., 2006]; whereas afterslip and especially viscoelastic relaxation dominate a much larger temporal and spatial space. For example, in the case of the 1992 M7.4 Landers earthquake, it has been reported that poroelastic rebound might contribute to the postseismic deformation [Peltzer et al., 1998; Fialko, 2004] in a region of 10 to 20 km from the fault. On the other hand, it was found that this large event was followed by afterslip in the first ~3 months influencing a larger area [Shen et al., 1994; Savage and Svarc, 1997]. In addition, Deng et al. [1998] and Pollitz et al. [2000] showed that distributed viscous flow in the lower crust and/or upper mantle can better explain postseismic deformation of the first three years following the Landers earthquake. As other examples, the postseismic deformation after 30 years following the M9.5, 1960 Valdivia earthquake in Chile was found to be well explained by viscoelastic relaxation [Lorenzo-Martín et al., 2006b]; Jónsson et al. [2003] presented that postseismic deformation of the first 1-2 months following the 17 and 21 June 2000 M6.5 earthquakes in southern Iceland were mainly related to poroelastic rebound; Ryder et al. [2007] showed that postseismic deformation for 4 years following the 1997 Manyi (Tibet) earthquake can be explained by afterslip or viscoelastic relaxation mechanism.

Most studies focused only on one of the three physical mechanisms. However, a comprehensive analysis of the relative contributions of each deformation source is necessary for understanding the whole postseismic process. This work aims to investigate the time evolution of the postseismic mechanisms and recognize the dominant postseismic process in different spatial and temporal domains based on the GPS data collected in several
years following the big events. In my work, I focus on afterslip on the extended coseismic fault plane and distributed viscoelastic relaxation processes that have significant and long-lasting contributions to the surface deformation.

Mainshock induced postseismic relaxation can be modeled deterministically as the response of a fault due to the coseismically produced stress. For example, Hearn et al. [2002; 2009] calculated coseismic shear stress change along the İzmit fault plane and adopted rate-strengthening friction law to model afterslip. They found an acceptable fit to the measured displacements following the İzmit earthquake. However, because of time-varying postseismic processes and also likely large uncertainties of the applied coseismic slip model, the forward modeling approach has difficulties to explain the detailed deformation near the rupture plane and the postseismic deformation in the long-term. On the other hand, inversions constrained by geodetic data can provide information where postseismic relaxation occurs. Repeated surveys and most importantly continuous GPS stations resolve also the time dependence of postseismic relaxation. In my work, I invert for afterslip on the extended fault plane in order to locate postseismic relaxation. Note that afterslip in this thesis, except at the places where friction-related slip is specified, refers to postseismic relaxation that occurs along the coseismic fault plane or in the narrow fault zone around the rupture plane.

As documented, viscoelastic relaxation that occurs in the ductile layer usually lasts for years or even decades. However, because of the deep location and temporal decay behavior of the relaxation source, the deformation observed on the surface is rather small, usually a couple of millimeters. The recent intensive GPS measurements in some seismically active regions provide now good opportunities to capture the weak signal transmitted from the lower crust and/or upper mantle and determine parameters of rheological models. The constructed rheological models provide useful information for characterization of seismic cycles and related hazard. So far, both linear [e.g. Pollitz, 1997; Wang et al., 2006] and non-linear [e.g. Freed and Bürgmann, 2004] rheological models have been proposed. The non-linear rheological modeling need additional geothermal information (e.g. geothermal gradient), which is often not available, as in the study areas analyzed in this work.

In this thesis, I take advantage of the PSGRN/PSCMP code [Wang et al., 2006] for forward modeling the postseismic viscoelastic relaxation. The best rheological model is selected from a large set of forward models. PSGRN/PSCMP code is a numerical tool based on elastic dislocation theory and implemented with different types of linear rheology for the calculation of elastic/inelastic response of a stratified medium to dislocations (earthquakes). In Chapter 2, I provide the mathematical descriptions of the inversion method and the PSGRN/PSCMP code.

In order to understand the postseismic behaviors in different tectonic settings, I investigated the 1999 M7.4 İzmit earthquake in Turkey (typical continental strike-slip event) and the 1995 M8.1 Antofagasta earthquake in Chile (thrust-fault event in the subduction zone) as study cases. The 1999 İzmit earthquake (40.76°N, 29.97°E, Delouis et al., 2002) has ruptured the west-end of the North Anatolian fault. Its occurrence has been proposed to largely increase the seismic risk in the Marmara Sea region, near the megacity Istanbul [Parsons et al., 2000]. For another case, the Antofagasta earthquake is a large event located immediately to the south of an important seismic gap of ~500 km, which did not rupture
since the occurrence of the South Peru (M9.1, 1868) and the Iquique (M9.0, 1877) megathrust events. Therefore, the detailed understanding of the postseismic deformation and the rheological model would provide valuable information for seismic hazard evaluation in the two regions. I present the results dealing with the postseismic modeling for the İzmit earthquake in Chapter 3 and for the Antofagasta earthquake in Chapter 4.

In addition to the observed postseismic aseismic deformations, aftershocks are significant marker of the relaxation processes following large earthquakes. Some similarities have been found between aftershocks and postseismic deformation. Firstly, both have generally consistent kinematic motion with the mainshocks [Bürgmann et al., 2002; Hsu et al., 2002; Bohnhoff et al., 2006]. Secondly, an Omori-type decay which is well-known for the aftershock sequences has also been noticed for the postseismic deformation recorded by continuous GPS measurements following e.g. the 1999 M7.6 Chi-Chi earthquake in Taiwan [Perfettini and Avouac, 2004; Savage et al., 2007] and the 2001 M8.4 Peru earthquake [Perfettini et al., 2005]. Some investigations also indicated spatial and temporal correlations between aftershocks and afterslip inverted from geodetic measurements. For example, Miyazaki et al. [2004] showed that the aftershocks of the 2003 Tokachi-oki earthquake overlap the afterslip area. In addition, it has been found that most of the postseismic deformation following the 2003 Altai earthquake can be explained by seismic moment release in aftershocks [Barbot et al., 2008]. These observations suggest that aftershock sequences and postseismic geodetic deformation are closely correlated and should be examined in more detail. In my Ph.D thesis, I build physical connections between aftershocks and aftershocks-induced component of geodetic deformation in the framework of damage rheology that was derived in laboratory rock experiments. This aftershocks-related deformation provides a candidate for interpreting the early postseismic relaxation detected (e.g., according to inversions constrained by the geodetic measurements) in the seismogenic zone, especially that overlapping the aftershock area, where rate-weakening friction is dominant in the framework of rate-and-state friction. Such early postseismic relaxation at the shallow depth is hard to be explained by either frictional-afterslip (rate-strengthening friction) or viscous flow (which is thermal-/depth-dependent). I present a new approach to separate aftershocks-related postseismic deformation from the other ductile contributions and take the İzmit earthquake as a study case. The results of this study are presented in Chapter 5.

Finally, I summarize the main results and draw the conclusions concerning my study on postseismic processes in Chapter 6.
Chapter 2  Methodology

The study in my thesis, aiming to investigate afterslip and viscoelastic relaxation mechanisms, uses a direct inversion method and forward modeling. This chapter deals with the mathematical description of the problems and methods I used. During the modeling, I use layered halfspace to take the local heterogeneity into account. The elastic parameters, which are adopted from the seismic reference models in the two study areas around the İzmit earthquake and the Antofagasta earthquake, will be given in Chapters 3 and 4, respectively.

2.1 Inversion for afterslip

2.1.1 Inversion method

Given the same fault geometric parameters as the coseismic slip model, the observed postseismic displacements are linearly related to afterslip by

\[ y = Gb, \]

where \( G \) is the Green’s function matrix, \( y \) and \( b \) represent the data and the discrete afterslip distribution, respectively. Equation system 2.1.1 describes a linear projection from the source space to the observation space. This linear problem can be solved by Least-Squares (LS) method using QR factorization or Singular Value Decomposition (SVD), if it is uniquely determined or over-determined. In many practical cases of slip inversion, however, the problem is underdetermined. In my study cases, GPS measurements at fewer than 100 sites are available, while the number of the unknowns (slip) exceeds 1000, when one tries to obtain a good resolution of the slip distribution and discretize the fault plane into patches of a few square kilometers. Therefore, \( a \ priori \) and/or artificial constraints are needed.

As in the traditional smoothing technique [e.g. Segall and Harris, 1987], the slip models are restricted to those with an appropriate roughness in the slip. Thus, the inversion becomes a minimization problem applied to an objective function (in the sense of mathematical optimization) consisting of both misfit and smoothing terms,

\[ F(b) = \|Gb - y\|^2 + \alpha^2 \|Hb\|^2, \]

where \( H \) is the finite difference approximation of the Laplacian operator, and \( \alpha^2 \) is the positive smoothing factor.

The linear optimization problem is solved by taking \( \frac{\partial F}{\partial b} = 0 \), i.e.,

\[ G^T(Gb - y) + \alpha^2H^Tb = 0, \]
The slip vector $\mathbf{b}$ can be obtained iteratively using steepest gradient method,

$$
\hat{\mathbf{b}}^{k+1} = \left[ I - \varepsilon(G^T \mathbf{G} + \alpha^2 \mathbf{H}^T \mathbf{H}) \right] \hat{\mathbf{b}}^k + \varepsilon \mathbf{G}^T \mathbf{y},
$$

where $\varepsilon$ is a parameter that dictates how far to move in the downhill gradient direction during the iterations. The method converges when a small enough $\varepsilon$, in particular satisfying [Press et al., 1986],

$$
0 \leq \varepsilon \leq \frac{2}{\max \text{eigenvalue}(G^T \mathbf{G} + \alpha^2 \mathbf{H}^T \mathbf{H})}.
$$

The smoothing factor $\alpha^2$ is selected using Cross Validation (CV) method [e.g. Árnadóttir and Segall, 1994]. This method determines the smoothing parameter based on the premise that a good model would successfully predict data that are not used in the inversion. Thus, for each given $\alpha^2$ value, we firstly get a group of slip models, each of which is obtained by omitting one datum,

$$
\mathbf{b}_i(\alpha^2) = \arg \min_{\mathbf{b}} \left( \| \mathbf{E}_i(G \mathbf{b} - \mathbf{y}) \|^2 + \alpha^2 \| \mathbf{H} \mathbf{b} \|^2 \right),
$$

where $\mathbf{E}_i = \mathbf{I} - \mathbf{e}_i \mathbf{e}_i^T$ is the matrix operator for removing one of the rows. Secondly, the cross-validation sum of squares (CVSS) are calculated by summing each squared residual between the omitted datum and that predicted by the model,

$$
CVSS(\alpha^2) = \frac{1}{m} \sum_{i=1}^{m} w_i \left( \mathbf{G}_i \mathbf{b}_i(\alpha^2) - \mathbf{y}_i \right)^2,
$$

where $\mathbf{G}_i = \mathbf{e}_i \mathbf{e}_i^T \mathbf{G}$, $w$ is the weighting factor given by the uncertainty of each measurement, and $m$ is the number of the observations. The smallest $CVSS(\alpha^2)$ provides the optimal $\alpha^2$. Thus, the CV method finally provides a model that does not rely too much on any individual observation.

During the inversion, we additionally apply two physical constraints. Firstly, because the postseismic deformation normally has similar kinematic motion as the main shock, we constrain the rake angle of afterslip in $[\psi - 20^\circ, \psi + 20^\circ]$, where $\psi$ is the rake angle determined for the coseismic slip. Secondly, it has been geologically proven that the slip on the fault plane tapers off gradually towards the fault tip [e.g. Scholz et al., 1993; Scholz and Gupta, 2000], so we constrain the slip as 0 at the two ends of the fault along strike and the slip increasing linearly inside the fault plane, if the slip is nonzero at the boundary during the iterations.

### 2.1.2 Reliability test of inversion method

Afterslip and viscoelastic relaxation mechanisms studied in this thesis have different characteristics. Afterslip is creep that occurs along the fault plane, most likely concentrates at the shallow depth (in the seismogenic zone), especially in the early days following the mainshock [e.g. Bürgmann et al., 2002; Langbein et al., 2006]. In comparison, viscoelastic relaxation is three-dimensionally distributed deformation in the deep inelastic layer, e.g. in the lower crust and upper mantle. In order to check if the inversion method can resolve the
shallow dislocation-like source from the three-dimensionally distributed deformation in the deep layer, we do a reliability test. To account for the situation of the small postseismic motion observed in a short time period or in the long-term, we use the synthetic data with low signal-to-noise ratio and take the İzmit earthquake as a test case.

We simulate two data sets, one related to afterslip on the fault plane and the other to viscoelastic relaxation below the upper crust. The surface displacements induced by afterslip are simulated by 1% of the 1999 İzmit coseismic slip (cf. Fig. 3-5) derived by Wright et al. [2001] in an elastic, layered model (cf. Fig. 3-4). Then, we simulate the displacement rate between 2003 and 2005 induced by viscoelastic relaxation in response to the İzmit earthquake [Wright et al., 2001] and the subsequent Düzce earthquake (M7.2, 87 days after the İzmit earthquake, Ayhan et al., 2001) using a rheological model. This model consists of an elastic upper crust and a Maxwell viscoelastic lower crust overlying a Maxwell mantle. The viscosities of the two inelastic layers are set to $2 \times 10^{19}$ and $7 \times 10^{19}$ Pa·s, according to the best estimates presented in Chapter 3. We superimpose a random component to the simulated displacement data to include the measurement uncertainty. The E-W and N-S random components are taken from two normal distributions, with standard deviations of 0.90 mm and 0.85 mm respectively, in accord with the measurement uncertainty of the displacement rate data collected between 2003 and 2005 [Ergintav et al., 2007].

For the two simulations based on afterslip and viscoelastic relaxation, the general displacement field, in the first case, is characterized by the slip on the fault plane in the elastic layer and has a shallow slip source; in the latter case, it is characterized by distributed deep viscous flow below the brittle upper crust. In both cases, we invert for the possible slip on the extended fault plane, and test if the inversion results reflect the deformation locations. The inversion is based on an elastic, layered medium. Note that, in the case of the non-stationary viscoelastic relaxation induced deformation, the inversion is based on a wrong model assumption, namely an elastic rheology. The inversion result might not be able to accurately reflect the effect of the laterally distributed viscous flow as given by viscoelastic lower crust/upper mantle. However, we will show below that the inversion can still basically resolve the depth of the deep deformation source.

The results are plotted in Fig. 2-1. Figure 2-1a and 2-1b show inversion results from the whole simulated data (for 78 GPS sites, cf. Fig. 3-1), based on the afterslip and viscoelastic relaxation model respectively. The RMS values for the two inversions are 0.24 mm and 0.25 mm. To investigate the influence of the number of data on the inversion, we use fewer data for a second set of models, namely those at the 35 ERG1 sites which recorded the displacements in the first 300 days (cf. Fig. 3-1) following the İzmit earthquake. The results are presented in Fig. 2-1c and 2-1d, respectively. The inversions have RMS values of 0.22 mm and 0.23 mm. The similarities of the inversion results based on different amount of data show that the inversion is stable.
Chapter 2. Methodology

Figure 2-1. Inversion results from synthetic data base on afterslip and viscoelastic relaxation. Afterslip is simulated by 1% of coseismic slip of Wright et al. [2001] (cf. Fig. 3-5a) and viscoelastic relaxation that occurred between 2003 and 2005 in response to the coseismic slip of the İzmit earthquake [Wright et al., 2001] and the Düzce earthquake [Ayhan et al., 2001] is simulated by the E-M-M model with viscosities for the lower crust/upper mantle of $2 \times 10^{19}/7 \times 10^{19}$ Pa·s. Inversion results from simulated data at 78 GPS sites for (a) the afterslip model and (b) the viscoelastic model. Inversion results using only part of the data (at 35 ERG1 sites) for (c) the afterslip model and (d) the viscoelastic model. The red stars mark the location of the İzmit earthquake.

The results also reflect the depth difference of the two modeled deformation sources, the slip in the elastic layer or the deep viscous flow. Because we used a discretized fault plane during the inversion, the obtained slip distribution for afterslip shows a smoother pattern than the input model. However, the large slip locus along strike and its depth location correspond well to the input model. The results for viscoelastic relaxation show that the resolved apparent slip concentrates below the depth of 20 km, which corresponds to the location of the assumed viscous flow source. The location of inverted large slip along strike is consistent with the location of the high coseismic slip, where a significant stress change was produced during the earthquake.

2.2 Static deformation in a multilayered half-space produced by internal sources

2.2.1 Description of the boundary-value problem

The partial differential equations governing the static deformation in an elastic medium are given by the equilibrium equation,

$$ \nabla \cdot \Gamma + f = 0, \quad (2.2.1) $$

where $\Gamma$ is the Lagrangian incremental stress tensor, $f$ is the force vector.

For an isotropic elastic medium, Hooke’s linear constitutive relation between stress and strain holds,
\[ \boldsymbol{\Gamma} = (\lambda \nabla \cdot \mathbf{u}) \mathbf{I} + \mu [\nabla \mathbf{u} + (\nabla \mathbf{u})^T], \]  

(2.2.2)

where \( \mathbf{u} \) is the displacement vector, \( \lambda \) and \( \mu \) are the Lamé constants, \( \mathbf{I} \) is the unit tensor, and \( (\nabla \mathbf{u})^T \) denotes the tensor transpose of \( \nabla \mathbf{u} \).

Displacement and stress satisfy the continuity conditions at an interior material interface,

\[ \mathbf{u} |^+ = 0, \]  

(2.2.3)

\[ \mathbf{e}_n \cdot \mathbf{\Gamma} |^+ = 0, \]  

(2.2.4)

where \( \mathbf{e}_n \) is the unit normal vector of the interface, and the symbol \( |^+ \) denotes the increment of the respective quantity from one side to the other side of the interface. The stress on the free surface and at an infinite distance is expressed as,

\[ \mathbf{e}_n \cdot \mathbf{\Gamma} = 0. \]  

(2.2.5)

The displacement at an infinite distance is expressed as,

\[ \mathbf{u} = 0. \]  

(2.2.6)

In the Green’s function approach, \( \mathbf{f} \) describes a point source, and can be replaced by equivalent body force in seismology.

The system of Eq. 2.2.1 with the boundary conditions (Eqs. 2.2.5-6) governs static stress/strain field in the multilayered half-space in response to internal sources, e.g. earthquakes. Inelastic response accounting for postseismic viscoelastic relaxation can be implemented by the complex shear modulus in frequency domain according to correspondence principal. It will be given in the next section. In my thesis, I use forward modeling to analyze the surface deformation caused by postseismic viscoelastic relaxation. The forward modeling is done using the software PGRN/PSCMP [Wang et al., 2006].

### 2.2.2 Implement of viscoelastic component

To account for inelastic response in the deep ductile layer, linear viscoelastic rheology is implemented by a complex shear modulus in the frequency domain according to the correspondence principle [Ben-Menahem and Singh, 1981]. Three kinds of rheological bodies are usually applied in previous studies, Maxwell rheology, Standard Linear Solid (SLS) rheology and Burgers’ rheology, among which the first two are special cases of the third rheology. The Burgers’ rheology is a Maxwell element in series with a Kelvin element, as shown in Fig. 2-2. The complex shear modulus for Burgers’ rheology is,

\[ \mu(i\omega) = \mu_1 + \frac{i\omega\eta_1[(1-\beta)\mu_1 + i\omega\beta\eta_1]}{(i\omega\eta_1 + \mu_2)(1-\beta)\mu_2 + i\omega\beta\eta_2} + i\omega\beta\mu_2\eta_2, \]  

(2.2.7)

where \( \omega \) is frequency, \( \mu_1 \) and \( \eta_1 \) are Kelvin shear modulus and Kelvin viscosity, \( \mu_2 \) and \( \eta_2 \) are Maxwell shear modulus and Maxwell viscosity, respectively. \( \beta(=\frac{\mu_2}{\mu_1 + \mu_2}) \) is the relaxation strength, a value between 0 and 1. Choosing different parameters in Eq. 2.2.7, we get the complex shear modulus for perfect elasticity, Maxwell rheology and SLS.
rheology, respectively.

For elasticity,

\[ \mu(i\omega) = \mu_2, \text{ when } \eta_1, \eta_2 \to \infty; \]  

(2.2.8a)

for Maxwell rheology,

\[ \mu(i\omega) = \mu_2 \frac{i\omega\eta_2}{i\omega\eta_2 + \mu_2}, \text{ when } \eta_1 \to \infty \text{ or } \beta = 0; \]  

(2.2.8b)

for SLS rheology,

\[ \mu(i\omega) = \mu_2 \frac{(1-\beta)\mu_1 + i\omega\beta\eta_1}{\mu_2 + i\omega\beta\eta_1}, \text{ when } \eta_2 \to \infty. \]  

(2.2.8c)

Replacing the complex shear modulus of Eq. 2.2.7 into Eq. 2.2.2, the inelastic response in frequency and wavenumber domain can be calculated. The final stress and displacement field is obtained by applying inverse transforms.

Figure 2-2. Schematic diagram of a Burgers’ element, a Maxwell element (marked with the dotted rectangle) in series with a Kelvin element (the dotted and dashed rectangle). The dashed rectangle marks a Standard Linear Solid element.
Chapter 3 Afterslip and viscoelastic relaxation following the 1999 M7.4 İzmit earthquake, from GPS measurements

Intensive GPS monitoring after the 1999 İzmit earthquake provides an opportunity to understand the postseismic behaviour of a strike-slip fault and the rheology below the brittle upper crust. Two data sets are available: displacements measured during the first 300 days after the İzmit earthquake and velocity measurements between 2003 and 2005. Using an inversion method and forward modeling, respectively, we investigate two mechanisms: (1) afterslip on and below the coseismic rupture plane; (2) viscoelastic stress relaxation in the lower crust and upper mantle described by a Maxwell or Standard Linear Solid (SLS) rheology. The inversion results show that the first several months following the İzmit earthquake were dominated by afterslip at depths shallower than 30 km and the slip amount decayed with time; after that, apparent afterslip has a very different spatial distribution and is located much deeper. For viscoelastic relaxation, a model with an elastic upper crust and a Maxwell viscoelastic lower crust overlying a Maxwell mantle (E-M-M) fits the data measured in the first 300 days better in the far-field than in the near-field. However, the observed far-field, 300-day displacement and the long-term (2003-2005) displacement, which might be dominated by viscoelastic relaxation, cannot be described by a Maxwell rheological model with constant viscosity: the effective viscosity increases over time. Therefore, we have built a refined rheological model: an elastic upper crust and a SLS lower crust overlying a Maxwell viscoelastic mantle (E-SLS-M). Our best solution yields a viscosity for the lower crust of $\sim 2 \times 10^{18}$ Pa-s, a relaxation strength of $2/3$, and a viscosity for the Maxwell mantle of $7 \times 10^{19}$ Pa-s. Finally, we explain the data using a composite model, consisting of the preferred E-SLS-M model and the afterslip model obtained from the residual displacement after correcting for viscoelastic relaxation. For the early time period, the residual displacements can be mainly explained by shallow afterslip whose magnitude decays with time, and whose spatial distribution is stable, while the residual displacements for the later time period require negligible afterslip. It indicates that the postseismic deformation in the later time period induced by a deep source can be almost entirely explained by the E-SLS-M model. The composite model can generally explain the data in the entire spatial and temporal space.

3.1 Introduction

Many large earthquakes were followed by postseismic deformation. Obvious changes of displacement field have been observed, e.g. after the M8.2, 1946 Nankaido earthquake,

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Chapter 3. Postseismic modeling for the 1999 M7.4 Izmit earthquake

Japan [Okada and Nagata, 1953] and after the M9.5, 1960 Valdivia earthquake, Chile [Lorenzo-Martín et al., 2006b]. Postseismic deformation, lasting days to years and covering areas with dimensions of several kilometers up to hundreds of kilometers, has been explained by afterslip, viscoelastic stress relaxation or poroelastic rebound. The three mechanisms might be predominant in different temporal and spatial domains [Segall, 2004].

Since 1939, a series of large earthquakes took place sequentially along the North Anatolian Fault (NAF). The most recent one was the devastating 17 August, 1999 İzmit earthquake (M7.4), which was followed 87 days later by the 12 November 1999, M7.2 Düzce earthquake. GPS monitoring was active before the 1999 İzmit earthquake, and was considerably intensified after the event. It provides a good opportunity to understand the postseismic behaviour of a strike-slip fault and the rheology below the brittle upper crust.

Based on the GPS measurements in this region, Hearn et al. [2002] studied the postseismic deformation of the first 80 days following the İzmit earthquake using the finite element method. They found that the surface deformation in this time period could be explained better by afterslip on the fault rather than viscous shear-zone creep or poroelastic rebound. Bürgmann et al. [2002] reported that the afterslip during the first 87 days occurred on and below the coseismic rupture. Ergintav et al. [Ergintav et al., 2009] used a logarithmic time function to model 7 years of GPS measurements after the İzmit earthquake, and derived afterslip in different time periods of the first 7 days, 6 months and 6 years after the mainshock, respectively. These studies focused mainly on the fault-zone mechanisms, and only a simple viscoelastic model based on Newtonian rheology was tested during the investigation for rheological properties below the brittle upper crust. In contrast to previous studies, we use GPS data of both short-term (300 days) and long-term (6 years) measured after the İzmit earthquake and focus on both viscoelastic relaxation and afterslip. We will show that the comparison between the two mechanisms provides new insight into the crustal rheology and postseismic behaviour of the strike-slip fault.

3.2 Data

3.2.1 GPS data measured after the İzmit earthquake

Two GPS data sets measured in different time periods are available for this research: displacements measured during the first 300 days (August 1999-June 2000) after the İzmit earthquake [Ergintav et al., 2002] and velocity measured between 2003 and 2005 [Ergintav et al., 2007]. Figure 3-1 shows the GPS stations that recorded the two data sets, which are referred as ERG1 and ERG2, respectively. In this study, we investigate only the horizontal displacements since the vertical displacements have large measurement uncertainties, which reach ~10 mm or even more. The GPS measurements in the NAF zone were presented in detail in many previous studies [e.g. Ergintav et al., 2002; Bürgmann et al., 2002]. Therefore, we only describe briefly the data used in this study.

There are 35 ERG1 stations, 10 continuous and 25 campaign sites, located in an area with a dimension of 100×200 km². Most of the stations are situated close to the fault (Fig. 3-1). Because the M7.2 Düzce earthquake (about 100 km away from the İzmit earthquake) occurred during our observation period, i.e., 87 days after the İzmit event, the coseismic
displacement of the Düzce earthquake is subtracted based on a coseismic slip model [Ergintav et al., 2002; Ayhan et al., 2001]. The 35 corrected displacement time-series in the first 300 days were modeled by Ergintav et al. [2002] using a superposition of a linear and an exponential component. The exponential function describes the postseismic displacement induced by the İzmit earthquake. They estimated that the postseismic displacement has an exponential relaxation time of 57 days. Based on their modeling, we use the displacements at 17 time points (days 9, 27, 45…) with an 18-day interval between 9 and 300 days after the İzmit earthquake. This data set is named as ERG1 in this text.

The 67 ERG2 GPS stations, which recorded the velocity data between 2003 and 2005, are shown by black filled and black open triangles in Fig. 3-1. The data were collected twice a year in June and October, with a total of five campaigns between June 2003 and June 2005. The velocity values plotted along a profile perpendicular to the İzmit fault trace are shown in Fig. 3-2. We will show below that, according to these velocity measurements, the postseismic deformation of the İzmit earthquake can still be detected 4-6 years after the event.

![Figure 3-1. Map of the İzmit region showing the distribution of GPS stations: red stars/squares mark continuous/campaign GPS sites on which data in the first 300 days after the İzmit earthquake are available (ERG1, Ergintav et al., 2002); black-solid triangles are GPS sites on which the velocity between 2003 and 2005 was recorded [ERG2, Ergintav et al., 2007]; blue-open triangles are GPS sites with measurements used for secular modeling (RLG, Reilinger et al., 2006). Black-open triangles mark GPS sites on which both ERG2 data and RLG data were recorded. Yellow stars are epicenters of the İzmit (N40.76°, E29.97°, Delouis et al., 2002) and Düzce (N40.82°, E31.20°, Milkereit et al., 2000) earthquakes; their rupture traces are shown by black bold lines [Wright et al., 2001] and a black dashed line [Ayhan et al., 2001]; thin dashed lines show the simplified NAF geometry [Lorenzo-Martín et al., 2006a]. The Anatolia block moves to the west by ~24 mm/y relative to the Eurasian plate. The two black circles mark the stations on which the displacements are shown as examples in Fig. 3-3.](image)

### 3.2.2 Modeling secular deformation

Postseismic modeling requires separation of the deformation induced by secular tectonic motion. In general, the secular tectonic deformation can be determined from the measurements before the large earthquake. However, most of the GPS stations used for this research were installed hours or days after the İzmit event. Therefore, the secular deformation on each GPS site is estimated through a secular model based on the GPS data

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13
of Reilinger et al. [2006], which were recorded between 1988 and 1999 before the İzmit earthquake in our study region (see RLG GPS stations in Fig. 3-1).

Under the collision of the Arabian and African plates against the Eurasian and Anatolian plates, the Anatolian block moves westwards with respect to the Eurasian plate, causing a nearly pure right-lateral, E-W striking fault, the North Anatolian Fault (NAF) zone. Therefore, the observed E-W secular velocity can be modeled by a fundamental function as shown in Eq. 3.1 [Savage and Burford, 1973]. It describes the fault-parallel interseismic surface displacement caused by an infinitely long dislocation slipping at rate \( s \) below a locking depth \( d \) in an elastic half-space,

\[
v(x) = A - \frac{s}{\pi} \times \arctg(x/d),
\]

where \( x \) is the distance perpendicular to the fault, measured on the surface; \( A \) refers to the velocity close to the rupture. In this study, the velocity/displacement is calculated relative to the Eurasia plate. We assume a fault plane along 40.72°N, according to an average NAF surface trace for the region between 26°E and 33°E. Leaving \( d \) free, we estimate from the data of Reilinger et al. [2006]: \( A = 10.5\pm0.2 \text{ mm/yr}, s = 26.0\pm1.3 \text{ mm/yr}, d = 19.8\pm2.7 \text{ km}, \) and a residual standard error of 1.5 mm/yr (black symbols, Fig. 3-2a). The estimated \( d \) value is close to the locking depth of ~18 km in the NAF zone [Hearn et al., 2002; Lorenzo-Martín et al., 2006a]. Based on the secular model, the Anatolian block in the İzmit region has an estimated westward movement relative to the Eurasian plate of ~26 mm/yr, which is close to the value of ~24 mm/yr derived in previous studies [McClusky et al., 2000; Flerit et al., 2003]. The N-S displacement rates are very small and have a weak linear trend, as shown in Fig. 3-2b. A linear model is considered to fit these data. Influenced by the complex geometry of NAF (Fig. 3-1) in the west and south areas, the N-S displacement measurements on the south side of the rupture show a stronger scatter than the E-W ones, and the linear fit model has a relatively large standard error of 2.3 mm/yr (black symbols, Fig. 3-2b).
3.2.3 Subtraction of secular movement from the displacement data

Because RLG and ERG1/ERG2 stations (see Fig. 3-1) have a similar coverage in the study area, the 2D secular model obtained based on RLG GPS data can approximate the interseismic deformation at each ERG1/ERG2 site. Therefore, the secular deformation can be subtracted from the GPS data measured after the İzmit earthquake. As examples, Fig. 3-3 shows the modeled secular displacements and the observed postseismic displacements after the correction for the secular component at two ERG1 sites in the first 300 days. These postseismic displacements have magnitudes up to ~7 cm.
Chapter 3. Postseismic modeling for the 1999 M7.4 İzmit earthquake

Figure 3-3. Calculated secular displacement (gray dotted) and secular-corrected postseismic displacement (black solid) observed at two GPS sites during the first 300 days following the İzmit earthquake. The two sites (circled in Fig. 3-1), which are located 1.64 km and 34.81 km normal to the fault trace, are marked by circles and stars, respectively.

Concerning the precision level of the ERG1 data, the continuous and campaign GPS measurements have uncertainties (1σ) of ~2 mm and ~4 mm [Ergintav et al., 2002]. Taking possible error due to the secular modeling into account, we assume a higher error level of the displacement data after the correction for secular component, using 3 mm and 5 mm for the continuous and campaign measurements, respectively.

As shown in Fig. 3-2a (blue symbols), the E-W velocity measurements between 2003 and 2005 (ERG2, Ergintav et al., 2007) are larger than the modeled interseismic velocity. The fit of Eq. 3.1 to the ERG2 data (blue curve in Fig. 3-2a) shows a maximum difference of ~4 mm/y from the secular model as displayed by the black curve in the figure. After the secular component is removed, the E-W velocity of 2003-2005 has distance-dependent values between 0.1 and 14.8 mm/y. Most of the displacement values exceed the measurement uncertainties of ~0.90 mm/y [Ergintav et al., 2007]. Therefore, the displacement measured in this time period is likely to be postseismic deformation of the İzmit earthquake.

3.3 Modeling results

The elastic parameters used in our models are adopted from the seismic reference model for the NAF region [Milkereit et al., 2000] that is shown in Fig. 3-4. The depth of the elastic upper crust is specified as 20 km, according to the depth distribution of the relocated aftershocks of the İzmit earthquake [Milkereit et al., 2000]. The depth of the lower crust is set as 35 km [Lorenzo-Martín et al., 2006a]. A coseismic slip model for the İzmit earthquake is given by Wright et al. [2001], derived from InSAR data and surface rupture. The parameters of the İzmit source model are listed in table 3-1. Because we need to consider viscoelastic relaxation caused by the Düzce earthquake, its source parameters,
from Ayhan et al. [2001], are also listed in table 3-1. Our study, aiming to investigate afterslip and viscoelastic relaxation mechanisms, uses a direct inversion method and forward modeling, as described in Chapter 2. We apply a grid search method to look for the optimal viscosities by minimizing the difference between models and observations, i.e. Root Mean Square error (RMS).

![Figure 3-4](image_url)

**Figure 3-4.** $V_p$, $V_s$ and $\rho$ profiles for the stratified medium used in this study [Milkereit et al., 2000]. The horizontal lines mark the three layers: the upper crust (0-20 km), lower crust (20-35 km) and mantle (>35 km).

<table>
<thead>
<tr>
<th>Table 3-1. Source parameters of the İzmit (İz) and Düzce (Dc) earthquakes [Wright et al., 2001; Ayhan et al., 2001].</th>
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<tr>
<td><strong>Fault Patch</strong></td>
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<tr>
<td>------------------</td>
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<tr>
<td>İz_1</td>
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<td>İz_2</td>
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<td>İz_3</td>
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<td>İz_4</td>
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<td>İz_5</td>
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<td>İz_6</td>
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<td>Dc</td>
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### 3.3.1 Spatial afterslip distribution

It has been reported that mainshock-induced afterslip might occur on the coseismic fault plane or on its extension to a deeper or shallower part of the Earth’s crust [Marone and Scholz, 1988; Perfettini and Avouac, 2004]. Therefore, we invert for afterslip on the extended fault plane down to a depth of 50 km. During the inversion, we use the fault geometry defined by Wright et al. [2001]. The fault segments and their derived coseismic
slip based on InSAR measurements are shown in the left panel of Fig. 3-5. We vary the rake angles around the ones determined for coseismic slip, and constrain the rake angle of afterslip in $[\psi - 20^\circ, \psi + 20^\circ]$, where $\psi$ is the rake angle determined for the coseismic slip. We discretize the fault plane into 2×2 km rectangular patches. We perform the inversions using the displacements in six time periods: the first 27 days, days 27-64, 64-101, 101-192, and 192-300 following the İzmit earthquake, and between 2003 and 2005. The E-W surface displacements of the six time periods have distance-dependent values between 0.2-38.5 mm, 0.7-23.7 mm, 0.2-14.0 mm, 0.6-26.7 mm, 0.7-25.0 mm, and 0.1-14.8 mm, respectively. Most of the displacements are much larger than the data uncertainty. The results are displayed in the left panels of Fig. 3-6. For clarity, the afterslip distributions are plotted on a vertical profile along the fault.

The inversion results show that: (1) afterslip in the first 27 days might have occurred with a maximum slip rate of ~3 m/y, and the most significant slip locates shallower than 30 km; (2) the slip distribution during the time periods of days 27-64 has a similar pattern as that inverted from the data in the first 27 days, and displays a decaying magnitude with time; (3) different from the first 100 days after the İzmit earthquake, the inverted slip during days 101-192 and days 192-300 is concentrated much deeper, mostly below 30 km. The slip concentration at the location between 100 and 130 km along strike might also be related to postseismic deformation of the Düzce earthquake; (4) The inverted slip displays a gradually deep-going trend and the time period of days 64-101 presents a transition from shallower slip to deeper apparent slip; (5) The inverted slip for the time period between 2003 and 2005 is much smaller than that in the time period immediately after the large earthquake, and is mainly deeper than 40 km.

The inversion results (table 3-2) indicate that the modeled displacements for the six time periods have similar fitting qualities to the data. As examples, we present in Fig. 3-7 the forward model for the time periods of days 27-64 and 101-192. The observed and predicted displacements are displayed by gray and black arrows, respectively. Because of the local heterogeneity, inadequacies in the model, etc., the data on some sites cannot be fit well by the inverted slip models. The displacements at several sites around the east end of the fault, which were poorly explained by the inversion result for days 101-192, might be

Figure 3-5. Two-dimensional plot showing coseismic slip of the İzmit earthquake derived by (a) Wright et al. [2001] and (b) our inverted coseismic slip using the GPS data of Reilinger et al. [2000]. The red stars mark the location of the İzmit hypocenter. I, II…VI label the six fault segments defined by Wright et al. [2001].

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related to the unmodeled postseismic deformation of the Düzce earthquake.

Figure 3-6. Two-dimensional plot showing inverted apparent slip rate along the fault plane using the postseismic GPS displacements (left panels) and the displacements after the correction for viscoelastic relaxation based on our preferred E-SLS-M model (right panels). The inversion results are for six time periods, respectively: the first 27 days; days 27-64; days 64-101; days 101-192; days 192-300 and between 2003 and 2005. The İzmit hypocenter is shown by the red stars. For the inversion, the fault plane was discretized to 2 by 2 km rectangular patches.

When comparing this result (in the left panels of Fig. 3-6) with the inversion results derived from the simulated data shown in Fig. 2-1, we find that the depth of the apparent slip for the time period of days 100-300 and 4-6 years after the İzmit earthquake basically agrees with the inverted deformation source for viscoelastic relaxation (cf. Fig. 2-1d). The apparent slip shown in Fig. 3-6 are similar with the apparent slip distributions east of the İzmit hypocenter shown in Fig. 2-1b and d; while the significant apparent slip concentration in the west derived in the synthetic test was not resolved from the real data. Notice that the
most significant shallow afterslip occurred west of the İzmit hypocenter in the first ~3 months. It suggests that the stress might be relaxed mostly by afterslip in the early time period, which leads to an absence of viscoelastic relaxation induced deformation in this area in the later time period. Therefore, deep viscoelastic relaxation was likely the dominant mechanism after ~100 days from the large event.

The inversion method resolves the presumed slip on the extended coseismic fault plane, which can best reproduce the observed surface displacement. According to the inversion results, the afterslip at shallower depth occurred in the first ~3 months following the İzmit earthquake and the slip magnitude decayed with time. Then, the later time periods (days 100-300 and between 2003 and 2005) were governed by a deeper deformation source. As we will show in section 3.4.1, according to the afterslip in the first 2 months, the decay of afterslip versus time is fast, and the function used to model the afterslip decay predicts an average slip rate between 2003 and 2005 of ~7 mm, much smaller than the dislocation inverted (~80 mm) from the surface displacements. Therefore, it is unlikely that the postseismic deformation between 2003 and 2005 was produced by the same afterslip mechanism as that in the first several months. In addition, Ergintav et al. [2007], using both GPS and gravity data (with the latter constraining well the vertical deformation), indicated that viscoelastic relaxation of the lower crust/upper mantle might occur between 2003 and 2005. Our inversion results also suggest that viscoelastic relaxation in the lower crust/upper mantle likely played an important role from ~100 days after the İzmit earthquake; while shallower afterslip dominated the first 2-3 months following the mainshock.

Table 3-2. RMS values (mm) of the best-fit models.

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<tbody>
<tr>
<td>ASP</td>
<td>6.45</td>
<td>3.84</td>
<td>2.22</td>
<td>5.54</td>
<td>8.48</td>
<td>4.18</td>
</tr>
<tr>
<td>EM&lt;sup&gt;W&lt;/sup&gt;</td>
<td>12.85</td>
<td>7.75</td>
<td>4.53</td>
<td>8.23</td>
<td>10.50</td>
<td>5.34</td>
</tr>
<tr>
<td>EM&lt;sup&gt;R&lt;/sup&gt;</td>
<td>12.90</td>
<td>7.83</td>
<td>4.60</td>
<td>8.75</td>
<td>11.78</td>
<td>5.40</td>
</tr>
<tr>
<td>EMM&lt;sup&gt;W&lt;/sup&gt;</td>
<td>12.50</td>
<td>7.28</td>
<td>4.37</td>
<td>8.14</td>
<td>10.36</td>
<td>4.63</td>
</tr>
<tr>
<td>EMM&lt;sup&gt;R&lt;/sup&gt;</td>
<td>13.53</td>
<td>7.49</td>
<td>4.41</td>
<td>8.62</td>
<td>10.68</td>
<td>4.68</td>
</tr>
<tr>
<td>CMP&lt;sup&gt;W&lt;/sup&gt;</td>
<td>6.50</td>
<td>4.00</td>
<td>2.62</td>
<td>6.89</td>
<td>9.86</td>
<td>4.54</td>
</tr>
</tbody>
</table>

Note:
<sup>W</sup> refers to the source model of Wright et al. (2001);
<sup>R</sup> refers to the source model of Reilinger et al. (2000);
ASP refers to the afterslip model;
EM refers to the E-M model;
EMM refers to the E-M-M model;
ESM refers to the E-SLS-M model;
CMP refers to the composite model.
Figure 3-7. The observed displacements (gray arrows) and modeled displacements (black arrows) for the time period of (a) days 27-64 and (b) days 101-192 based on the afterslip (ASP) inversion results. The gray dashed lines display the fault traces of the İzmit and Düzce earthquake and their epicenters are marked by yellow stars.

3.3.2 Forward modeling of the viscoelastic relaxation effect based on Maxwell rheology

Working from a simple model (with few parameters) to a more complex model, we started viscoelastic relaxation modeling using Maxwell rheology. We built a two-layer rheological model, made up of an elastic upper crust and a Maxwell half-space (E-M), and also a three-layer model consisting of an elastic upper crust and a Maxwell viscoelastic lower crust overlying a Maxwell viscoelastic mantle (E-M-M). The modeled elastic upper crust and lower crust are at the depths of 0-20 km and 20-35 km, respectively (cf. Fig. 3-4). The elastic parameters of each layer are obtained from the $V_p$, $V_s$ and density profiles shown in Fig. 3-4. We used the 6-segment coseismic slip model derived by Wright et al. [2001] as input and calculated the effective viscosities of the inelastic layers best fitting the observations for different time periods. Our results show that the best-fit viscosity of the E-M model has a similar increasing trend with time as that of the E-M-M model (see table 3-3). The best-fit viscosity of the former is in between the estimated viscosities for the lower crust and upper mantle of the latter. In comparison, the E-M-M model fits the data
better than the E-M model (see table 3-2) and is capable of describing more details of the medium structure. The E-M model is only a special case of the E-M-M model. Therefore, we show the detailed results based on the E-M-M model.

We estimate the viscosities using the grid search method and apply the method to the GPS measurements in the first 300 days after the İzmit earthquake. The stability of the estimates is evaluated by the best-fit results from 500 simulations, which are obtained by adding random noise to real data within the given data uncertainties. We calculate the best-fit viscosities using the displacements measured in several time periods, days 0-27, 27-64, 64-101 and 101-300, the same as those for afterslip inversion. The viscosity estimates are presented in table 3-3. The uncertainty values are specified by 25% and 75% quartiles of each set of 500 estimates. The misfits of the best-fit models are summarized in table 3-2.

The results show that the best-fit viscosities estimated from the displacements of the first month have rather low values of $\sim 5 \times 10^{17}$ Pa·s for the lower crust and upper mantle. The estimates, reflecting highly viscous flow below the brittle upper crust, are much smaller than the values obtained in previous studies [e.g. Motagh et al., 2007]. The best-fit viscosities of the lower crust and upper mantle increase with time, and the estimates of the upper mantle viscosity are comparatively unreliable with large uncertainties. Comparing with afterslip inversions, the rheological models with much lower degree of freedom generally have higher misfit values (cf. table 3-2).

Using the velocity data of 2003-2005, we estimate that the best-fit viscosities of the lower crust/upper mantle are $\sim 2 \times 10^{19}/7 \times 10^{19}$ Pa·s, respectively. A contour plot of the RMS values is shown in Fig. 3-8. It indicates that the viscosity estimate of the upper mantle is less constrained than that of the lower crust. However, the results suggest at least that the upper mantle has a higher viscosity than the lower crust, if the surface deformation of this time period was induced only by viscoelastic relaxation. We will show in section 3.4.2 that the surface deformation between 2003 and 2005 might be almost totally controlled by viscoelastic relaxation.

The 300-day displacements lead to the best-fit viscosities of lower crust/upper mantle far smaller than the estimates based on the velocity data of 2003-2005. The results reflect that a Maxwell rheology with constant viscosity is not sufficient to explain the GPS data over the entire time series. For the early months after the mainshock, the surface displacement might be influenced by significant afterslip at the shallower depth as shown in the previous section.

In addition to the applied source model of Wright et al. [2001], we also tested the source model derived by Reilinger et al. [2000], in order to investigate the influence of different source models on viscosity estimates. Instead of a constant slip on each fault segment, slip in Reilinger’s source model varies with depth. The results (table 3-2 and 3-3) show that the estimated viscosity for the lower crust based on the source model of Reilinger et al. [2000] increases with time, similar to that based on Wright’s source model. The estimates based on the former are overall smaller than those based on the latter. However, the difference between the two estimates is not significant. In this study, we stick to the simple coseismic slip model of Wright et al. [2001].
### Table 3-3. Best-fit viscosities (Pa·s) of the rheological models.

<table>
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<tbody>
<tr>
<td></td>
<td>$\eta_{lc} \times 10^{17}$</td>
<td>$\eta_{um} \times 10^{17}$</td>
<td>$\eta_{lc} \times 10^{18}$</td>
<td>$\eta_{um} \times 10^{18}$</td>
<td>$\eta_{lc} \times 10^{18}$</td>
<td>$\eta_{um} \times 10^{18}$</td>
</tr>
<tr>
<td>EM$^W$</td>
<td>5.0±0.2</td>
<td>1.7±0.3</td>
<td>2.7±0.4</td>
<td>4.2±0.5</td>
<td>5.0±0.8</td>
<td>3.0±1.2</td>
</tr>
<tr>
<td>EM$^R$</td>
<td>3.0±0.3</td>
<td>1.1±0.2</td>
<td>1.8±0.4</td>
<td>2.8±0.4</td>
<td>4.1±0.9</td>
<td>2.0±1.0</td>
</tr>
<tr>
<td>EMM$^W$</td>
<td>5±1</td>
<td>7±1</td>
<td>1±1</td>
<td>2±1</td>
<td>4±2</td>
<td>3±1</td>
</tr>
<tr>
<td>EMM$^R$</td>
<td>3±1</td>
<td>3±1</td>
<td>0.7±0.1</td>
<td>3±5</td>
<td>1±1</td>
<td>9±5</td>
</tr>
<tr>
<td>ESM$^W$</td>
<td>$\eta_{lc} = 2(±2) \times 10^{18}$; $\eta_{um} = 7\times 10^{19}$; $\beta = 2/3±1/7$</td>
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Note: $\eta_{lc}$ and $\eta_{um}$ refer to the viscosities of the lower crust and upper mantle, respectively; The other symbols have the same meaning as those for table 3-2.

**Figure 3-8.** Misfit (RMS) contour plot of the E-M-M model based on the velocity data measured between 2003 and 2005. The cross marks the location of the minimum. $\eta_{lc}$ and $\eta_{um}$ refer to the viscosities of the lower crust and upper mantle, respectively.

Compared with the afterslip mechanism, viscoelastic relaxation normally has more significant effect in the far-field or on long-term deformation [Pollitz et al., 2000]. Therefore, to further investigate if the E-M-M model is appropriate to explain our far-field or long-term data, we do an additional test and use two data sets: (1) the far-field (>30 km away from the fault), 300-day data to exclude significant contribution from afterslip in the near-field; (2) the data measured in the long-term (between 2003 and 2005). The far-field or long-term data normally have low displacement values. Therefore, we test the displacements of the first dataset between different times $t_1$ and 300 days, e.g. $D_{(9,300)}$, $D_{(27,300)}, D_{(46,300)}$..., so as to investigate possible variability of estimated parameters with time. During the calculation, we fix the viscosity of the upper mantle, which was less sensitive to the surface deformation of the study region, as 7×10$^{19}$ Pa·s, in agreement with our estimates of the E-M-M model based on the velocity data of 2003-2005. The results shown in Fig. 3-9 indicate that the viscosity estimates increase with time, which means that an E-M-M model with constant viscosities also cannot describe both the short-term, far-field deformation and the long-term deformation. Therefore, as we will test in section
3.3.4, another rheology like the Standard Linear Solid (SLS) is needed to describe the transient deformation.

![Figure 3-9](image)

**Figure 3-9.** Grid search results for the E-M-M model and the E-SLS-M model using the far-field (>30 km from the fault) displacements between \( t_1 \) and day 300 (\( D_{(t,300)} \), ERG1 data), together with the 2003-2005 velocity measurement. The estimates of lower crust viscosity (blue dashed line), based on the E-M-M model with mantle viscosity of \( 7 \times 10^{19} \) Pa·s, are plotted in (a). Red lines show estimated viscosity (a) and relaxation strength (b) of the E-SLS-M model versus \( t_1 \).

### 3.3.3 Misfit performance of the E-M-M model in space

Relative to an afterslip model, viscoelastic relaxation has a more significant effect in the far-field than in the near-field. To test if this can be seen in the data, we calculate the normalized misfit of the best-fit Maxwell rheological model on each ERG1 site,

\[
\tilde{\Delta}_i = \frac{\sqrt{(D_{i,x} - D'_{i,x})^2 + (D_{i,y} - D'_{i,y})^2}}{\sqrt{D_{i,x}^2 + D_{i,y}^2}}, \text{ for } i = 1, \ldots, 35 \tag{3.2}
\]

In Eq. 3.2, \( D \) is the observed displacement, and \( D' \) is the predicted displacement by the best-fit viscoelastic relaxation model. Small \( \tilde{\Delta} \) values mean that the misfit between model and observation is low, and the model can well explain the observation.

Because the rheological models for the different time periods have similar fitting quality to the data (cf. table 3-2), we only present in Fig. 3-10 the result for days 101-192 as an example. The displacements observed and predicted by the best-fit Maxwell rheological model are shown by gray and black arrows, respectively. The normalized misfit on each ERG1 site for this time period is displayed in Fig. 3-11. The normalized misfit \( \tilde{\Delta} \) values are plotted against the distance of each ERG1 site perpendicular to the simplified fault trace along 40.72°N. As the İzmit rupture is nearly a pure E-W strike-slip fault, we treat the \( \tilde{\Delta} \) values on the north and south sides of the fault equally, and plot absolute distance in Fig.
3-11. Some $\tilde{\Delta}$ values are larger than 1 (open triangles in the figure), implying that the movements do not agree with the assumed mechanism, and the data on these sites cannot be described by viscoelastic relaxation at all. Most of such large $\tilde{\Delta}$ values are on the sites near the fault. This inconsistency could relate to local heterogeneities that cannot be fully described by our coseismic slip model, or reflect a comparatively poor measurement on the individual sites. While neglecting the outliers, we notice that most of the data (except one) measured about 30 km and further away from the fault can be well explained by the rheological model with normalized misfits less than 0.6. The spatial scale of 30 km agrees with the theoretical results obtained by Hetland and Hager [2006]. They demonstrated that within several relaxation times, viscoelastic relaxation of the lower crust produces the most significant surface deformation at a distance about $\sim$2 times the elastic layer depth (20 km in our case) away from the fault.

![Figure 3-10](image1.png)

**Figure 3-10.** The observed (gray arrows) and modeled (black arrows) displacements for the time period of days 101-192 based on the best-fit E-M-M model. The gray dashed lines display the fault traces of the İzmit and Düzce earthquake and their epicenters are marked by yellow stars.

![Figure 3-11](image2.png)

**Figure 3-11.** Normalized misfit $\tilde{\Delta}$ of the best-fit E-M-M model for the time period of days 101-192 on each EGR1 GPS site versus the N-S distance of the site perpendicular to the fault trace along 40.72°N. Some outliers have $\tilde{\Delta}$ values larger than 1.0, which implies that the data on these sites cannot be described by the viscoelastic relaxation mechanism at all, are shown by open triangles.
3.3.4 Forward modeling of the viscoelastic relaxation effect based on Standard Linear Solid rheology

According to the results from section 3.3.2, a Maxwell rheological model with constant viscosity cannot describe the postseismic deformation which occurred in the far-field and long-term, where viscoelastic relaxation should be significant. Our result that the best-fit viscosity of Maxwell rheology increases with time, could be attributed to two possibilities: (1) even though the surface displacement of several months after the big event was dominated by deep viscoelastic relaxation, the shallow afterslip still contributed to the displacement field; (2) a Maxwell element is not appropriate in the description of the lower crust/mantle rheology, and a rheological model which can describe a transient process is necessary.

Therefore, we build a refined rheological model composed of an elastic upper crust and a SLS lower crust overlying a Maxwell viscoelastic mantle (termed as E-SLS-M). Unlike a Maxwell rheology, SLS has a long-term elastic strength and can model a transient process. To reduce the number of model parameters, we still apply Maxwell rheology to describe the mantle.

Using both the 300-day, far-field data (>30 km away from the fault) and the long-term data (between 2003 and 2005), we test the displacements between different time \( t \) and days \( 1 \) and days \( 300 \) (e.g. days 9-300, 27-300). We estimate two free parameters of the SLS element, \( \eta \) and \( \mu_1 (\eta_2 \rightarrow \infty) \) as displayed in Fig. 2-2, which govern the transient deforming process.

To obtain \( \mu_1 \), we estimate the relaxation strength \( \beta = \frac{\mu_2}{\mu_2 + \mu_1} \). The estimates (red lines, Fig. 3-9) show that after 100 days from the İzmit earthquake, the viscosity and the relaxation strength have stable values of \( 2 \times 10^{18} \) Pa·s and \( \beta = 2/3 \), i.e. \( \mu_1 = 0.5 \mu_2 \). The estimated SLS rheology has a relaxation time (\( \frac{\eta}{\mu_1 + \mu_2} \)) of 1 year. We calculate that the best-fit E-SLS-M model for the far-field data during 100-300 days and the long-term data (2003-2005) has RMS value of 4.89 mm.

If we base the E-SLS-M model only on the velocity field of 2003-2005, the best-fit model has a viscosity of \( 10^{19} \) Pa·s and \( \beta = 2/3 \), with RMS value of 4.61 mm. The misfit is only ~0.02 mm smaller than that of the E-M-M model with viscosities of \( 2 \times 10^{19}/7 \times 10^{19} \) Pa·s for the lower crust and upper mantle, or the E-SLS-M model with viscosities of \( 2 \times 10^{18}/7 \times 10^{19} \) Pa·s and \( \beta = 2/3 \). The small misfit difference reflects a similar long-term velocity field produced by the three models, and conversely, the best-fit E-SLS-M model with parameters of \( 2 \times 10^{18}/7 \times 10^{19} \) Pa·s and \( \beta = 2/3 \) can similarly describe the long-term postseismic deformation as the other two. However, it is superior to the other two models because it can describe both the long-term observations and the far-field, short-term observations. As an example, Figure 3-12a shows the displacements observed and predicted by the preferred E-SLS-M model for the time period of days 101-192. Theoretically, we expect the E-SLS-M model has a good fit in far-field and long-term as shown in Fig. 3-13a. The displacement time-series at two sites predicted by the preferred E-SLS-M model are
shown in Fig. 3-14. The clear difference between the postseismic displacements observed and predicted by the E-SLS-M model indicates that significant afterslip occurred in the early time period.

![Figure 3-12](image)

*Figure 3-12. The displacements observed (gray arrows) and predicted (black arrows) by (a) the estimated E-SLS-M (ESM) and (b) the composite (CMP) model for the time period of days 101-192. The gray dashed lines display the fault trace of the İzmit and Düzce earthquake and their epicenters are marked by yellow stars.*

### 3.3.5 Composite model

The investigations above indicate that the time periods before and after the first 2-3 months were dominated by two different mechanisms, shallower afterslip and deeper viscoelastic relaxation, respectively.

Now we analyze the inverted slip distribution after correcting for the postseismic surface deformation induced by viscoelastic relaxation based on the estimated E-SLS-M model. The results are shown in the right panels of Fig. 3-6. We find that after the correction for viscoelastic relaxation, the apparent afterslip focuses more clearly at the shallower depth and shows a consistent spatial pattern. The slip magnitude decays with time. After ~3 months from the mainshock, the slip is negligible. It reflects that the deep source induced surface deformation in the later time period was almost entirely removed based on the
preferred viscoelastic model. It further proves that the surface deformation of the later time period can be mostly modeled by the estimated E-SLS-M model.

Therefore, we build a composite model, consisting of our preferred E-SLS-M model and the afterslip model obtained from the residual displacement after correcting for viscoelastic relaxation. The misfits of the composite model are summarized in table 3-2. It shows that the composite model has RMS values far smaller than the best-fit E-M-M models, and is close to but not better than the pure afterslip inversions. This is because in the composite model, the far-field data are mostly explained by the viscoelastic relaxation model with only two free parameters; while for the pure afterslip inversion, because we permit slip to occur on an extended fault plane down to the depth of 50 km, this model has much higher degree of freedom to explain both near-field and far-field data. Consistently, we present the observed and predicted surface displacements for the time period of days 101-192 as an example in Fig. 3-12b. It shows that, compared to the preferred E-SLS-M model, which mostly explains the data in the far-field, the composite model obviously improves the fits in the near-field.

In addition, Figure 3-13 shows the spatial and temporal distribution of normalized misfit (cf. Eq. 3.2) between the predictions and the observations on each GPS site during the first 300 days. The two predictions shown are based on the estimated E-SLS-M model and the composite models, respectively. The misfit on each GPS site is calculated using the displacement that occurred over 72 days, with each value spaced by 36 days (i.e., a 36-day overlapped displacement used during the calculations). The results generally indicate that the viscoelastic relaxation mechanism can better explain the measurements in the far-field and long-term than the near-field and short-term. Some exceptions (e.g. the blue area in Fig. 3-13a, located at a distance of ~50 km from the fault and ~80 days from the İzmit earthquake) are likely to be related to additional postseismic deformation produced by the Düzce earthquake (e.g. possible afterslip on the rupture plane of this large event) that is not able to be explained by the present model. Fig. 3-13b shows that, except for some areas (e.g. at a distance of 60-70 km and days 80-120) where the data can be well explained by the E-SLS-M model and the composite model produces a poorer fit prediction than the single model, the composite model in general explains well the postseismic displacements in both near-field and far-field, short-term and long-term.
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Figure 3-13. Temporal and spatial distribution of the normalized misfit between the displacement observed in the first 300 days and predicted by (a) E-SLS-M model with parameters $2 \times 10^{18}/7 \times 10^{19}$ Pa·s and $\beta = 2/3$, and (b) the composite model. The misfit value on each ERG1 GPS site is calculated using the displacement that occurred over 72 days, with each value spaced by 36 days (i.e., a 36-day overlapping displacement was used during the calculations).

Figure 3-14. The observed postseismic displacements at two GPS sites (1.64 km and 34.81 km away from the fault) during the first 300 days following the İzmit earthquake are shown by two curves with black circles and black stars, respectively. The blue ones show the corresponding surface displacements at the two sites predicted by the E-SLS-M model with viscosities of $2 \times 10^{18}/7 \times 10^{19}$ Pa·s and $\beta = 2/3$, based on the stratified medium shown in Fig. 3-4 (solid blue lines) and a modified medium by 50% decrease of bulk modulus (dotted blue lines). The clear difference between the postseismic displacements observed and predicted by the E-SLS-M model indicates that significant afterslip took place in the early time period.
3.4 Discussion

3.4.1 Predominant postseismic mechanisms at different time scales

Our results indicate that the first 2 or 3 months were governed by afterslip; after that, viscoelastic relaxation played a major role. This is consistent with the results based on the FEM method [Hearn et al., 2009]. They indicate that the postseismic deformation after ~3 months from the İzmit earthquake needs viscoelastic relaxation to explain it; whereas in the first 3 months, it is in accord with frictional afterslip on and below the İzmit earthquake rupture.

Figure 3-14 shows that the contributions of viscoelastic relaxation during the first 1-2 months were very small relative to the overall surface displacement. Therefore, we neglect the effect of visco-elasticity in the first 2 months and estimate the temporal decay of afterslip. Assuming that afterslip on the fault plane follows an Omori’s law type decay, the cumulative afterslip has the form [Montési, 2004],

\[ S(t) = c_1 \ln(1 + \frac{t}{c_2}), \]

where \( c_1 \) and \( c_2 \) are constants. Normalizing the inverted afterslip on each patch in the first ~2 months to a value between 0 (when \( t = 0 \) day, the day of the mainshock) and 1 (when \( t = 64 \) days), Eq. 3.3 has the following parameter values: \( c_1 = 0.33 \pm 0.01 \) and \( c_2 = 3.47 \pm 0.06 \) day with a residual sum square of 6.10 (the red curve in Fig. 3-15a). The afterslip decay is similar to the aftershock decay. To demonstrate this, we analyze the aftershock catalogue of the Kandilli Observatory and Earthquake Research Institute (Istanbul, Turkey), with a completeness magnitude of ~2.5 [Daniel et al., 2006]. The result is shown by the black dotted curve in Fig. 3-15a. This consistency might reflect the decay behavior of the common stress field that governs the occurrences of both aftershocks and the afterslip.

Based on the afterslip decay function, the slip rate during the time period of days 100-300 is about 11.2% of that during the first 64 days following the İzmit earthquake. Meanwhile, we estimate the average slip rate between 2003 and 2005 to be ~7 mm, which is far smaller than the inverted apparent slip of ~80 mm presented in Fig. 3-6. This result confirms that the time period between 2003 and 2005 was controlled by a different postseismic mechanism, e.g. viscoelastic relaxation below the upper crust as we studied in this work. Alternatively, deep afterslip could have taken place due to a localized deformation induced by the coseismic stress change in the discrete shear zone [Kenner and Segall, 2003].

To investigate the latter case, we calculate the coseismic shear stress change down dip of the fault plane. Because the simple 6-segment source model of Wright et al. [2001] causes strong stress singularity at the edge of the rupture plane, we use the source models of Reilinger et al. [2000]. The results (Fig. 3-16) show that the locations of the high shear stress changes above the depth of 30 km are close to the location of significant afterslip along strike in the first ~3 months. Another slightly higher shear stress change is located at
the depth of ~36 km and ~50 km along strike. The location is basically consistent with the inverted apparent slip (Fig. 3-6, left panels) for the later time period (after the first 3 months) and the inverted results based on the synthetic data for viscoelastic relaxation (cf. Fig. 2-1b and d). However, the possible deep afterslip does not display, as expected, a monotonic decay with time, that is, the inversion results do not show a similar deep afterslip in the first 3 months. In contrast, after correction for viscoelastic relaxation, our inverted afterslip (Fig. 3-6, right panels) shows a generally decaying magnitude with time. Therefore, using our preferred E-SLS-M model with only two free parameters, we can explain both the far-field and long-term postseismic deformation, including that 4-6 years after the İzmit earthquake. Therefore, we conclude that viscoelastic relaxation is more likely the major mechanism for the postseismic deformation after the first 3 months following the İzmit earthquake.

Figure 3-15. Omori-law decays of afterslip and aftershocks following the İzmit earthquake. (a) The normalized afterslip on each subfault (black circles) at days 0, 9, 27, 46 and 64, the normalized cumulative number of aftershocks ($M \geq 2.5$, black curve), and the fitting curves to afterslip (red curve) and to number of aftershocks (black dotted curve) by the integrated form of Omori-law (Eq. 3.3). The normalized afterslip and aftershock number have values between 0 (at 0 day) and 1 (at 64 day). (b) M-t plot of the aftershock sequence ($M \geq 2.5$).
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3.4.2 Rheology of the lower crust and upper mantle in the İzmit region

In this study, we use an E-SLS-M model to approximate the rheology in the İzmit region. Maxwell rheology has commonly been used in previous investigations to describe the long-term rheological behavior of the lower crust and upper mantle [Rundle and Jackson, 1977; Savage and Prescott, 1978; Hetland and Hager, 2006]. However, it cannot account for a rapid transient deformation. A Burgers’ body, composed of a Maxwell body in series with a Kelvin body, was proposed in some studies [Hetland and Hager, 2006; Pollitz, 2003, 2005] to model the postseismic deformation. Being able to describe a transient deforming process after a stress disturbance and behaving viscously like a Maxwell body in the long-term, a Burgers’ rheology is a good alternative for description of the upper mantle. However, it involves additional free parameters, which cannot be properly constrained by the limited data. In addition to the linear rheology, Freed and Bürgmann [2004] reported that a stress- or time-dependent power-law visco-elasticity of the upper mantle could explain the postseismic deformation of the 1992 Landers earthquake and the 1999 Hector Mine earthquake. However, the very different spatial distributions of the apparent slip before and after ~100 days (the left panels of Fig. 3-6) suggest a variation of the dominant postseismic mechanism at this time. In addition, the non-linear rheological model with more free parameters can only be performed using numerical approaches like the finite element method. Therefore, we apply a linear rheological model, made up of an elastic upper crust, SLS lower crust and Maxwell mantle. This model, involving only 2 free parameters, can describe both the transient deformation and steady-state situation. Assuming a Maxwell material for the mantle and its viscosity of $7\times10^{19}$ Pa·s in agreement with the best-fit estimates of the E-M-M model based on the long-term observations, we find that a SLS body with a viscosity of $2\times10^{18}$ Pa·s and a relaxation strength of 2/3 can best fit the data.

Figure 3-16. Shear stress change downdip of the coseismic rupture plane produced by the source model of Reilinger et al. [2000].
Although some studies [e.g. Pollitz, 2005] reported a stronger lower crust and a weaker mantle, our results suggest that the lower crust has a lower viscosity than the upper mantle, in agreement with the results of some other investigations on postseismic deformation following large earthquakes (e.g. the 1999 Chi-Chi earthquake, Sheu and Shieh, 2004). This result (i.e., a weaker lower crust and a stronger upper mantle) is also consistent with the general view of the continental lithosphere, derived from the correlation between the cutoff depth of continental seismicity, the onset of thermally activated processes in crustal material [Sibson, 1982], and the seismic/geological evidence for a weak lower crust [Kay and Kay, 1981; Kusznir, 1991; Brocher et al., 1994].

Some uncertainties in our analysis could lead to an underestimation of the viscosity of the SLS lower crust. Although the far field (>30 km) deformation of the first 300 days was probably dominated by viscoelastic relaxation, it still involves a small unaccounted contribution of afterslip on the fault, which can cause an underestimation of the viscosity. On the other hand, our assumption of a Maxwell mantle rheology, which does not take possible transient deformation into account, could also lead to an underestimation of viscosity in the lower crust. However, the stability of our estimates over all observation periods indicates that neither bias is significant with respect to the data (the long-term deformation and the far-field, short-term deformation) used to estimate the E-SLS-M model parameters.

Some studies based on rock experiments reported that there could be a large difference between the static and dynamic bulk modulus due to systematic differences in the deformation process or due to cracks [e.g. Cheng and Johnston, 1981; Mockovciakova and Pandula, 2003]. Therefore, we test the viscoelastic relaxation using a different stratified medium modified by a 50% decrease of the bulk modulus. The results (Fig. 3-14) show that this correction produces only little change in the surface deformation. In summary, we find that our E-SLS-M model with viscosities of $2 \times 10^{18}/7 \times 10^{19}$ Pa·s and $\beta = 2/3$ can represent the rheology of the Izmit region at time scale of months to several years.

Based on a Maxwell half-space, Hearn et al. [2009] suggested the viscosity in the region has a value of $2-5 \times 10^{19}$ Pa·s, which is between the transient viscosity in the lower crust and the steady-state viscosity in the upper mantle of our E-SLS-M model; Motagh et al. [2007] used interseismic InSAR data to obtain a steady-state viscosity of $1.3 \times 10^{19}-3.6 \times 10^{20}$ Pa·s, which is comparable with our mantle viscosity of $7 \times 10^{19}$ Pa·s, when the SLS lower crust behaves elastically in the long-term. A similar result based on SLS rheology was obtained by Ryder et al. [2007]. Their model, consisting of an elastic layer and a SLS half-space with viscosity of $4 \times 10^{18}$ Pa·s and $\beta = 0.68$, best explains the 4-year postseismic displacements following the 1997 M7.6 Manyi (Tibet) earthquake.

### 3.4.3 Relations between coseismic slip, afterslip and aftershocks

Figure 3-5b shows our inverted coseismic slip of the İzmit earthquake using the GPS data of Reilinger et al. [2000]. As for the afterslip inversion, we use an extended fault plane [Wright et al., 2001] down to a depth of 50 km, and discretize the fault plane into 2x2 km rectangular patches. Consistent with the thickness of the elastic upper crust, the inverted
coseismic slip is confined well above this depth. Figure 3-5 shows that the slip amount and location of our inverted maximum coseismic slip are close to that of Wright et al. [2001]. Our slip distribution is also consistent with the coseismic slip derived by Reilinger et al. [2000].

In comparison, the inverted major afterslip (Fig. 3-6) on segments III (30 to 50 km along strike), IV (50 to 70 km along strike) and VI (~110 to 130 km along strike) took place in regions near or down-dip of large coseismic slip (Fig. 3-5b). Afterslip inversions of the 1999 Chi-Chi earthquake [Hsu et al., 2002] and the 2003 M8 Tokachi-oki earthquake [Miyazaki et al., 2004] demonstrate similarly that the afterslip tends to occur adjacent to areas with large coseismic slip. It suggests that stress changes caused by coseismic slip promote the afterslip. The result is consistent with the concept of velocity-strengthening frictional afterslip based on the rate-and-state friction law, which may occur on the earthquake rupture surface in areas of low coseismic slip, where the fault zone was loaded during the earthquake [Marone et al., 1991]. Our inverted afterslip concentrations between 30 and 70 km along strike are similar to those of Bürgmann et al. [2002] for the first 80 days, but the slip heterogeneity is highlighted in our inversion; the slip concentration between 100 and 140 km along strike in Bürgmann et al. [2002] was not detected in our inversion, possibly because of our poorer data coverage in this area.

Figure 3-17 shows the spatial distribution of the aftershocks (M≥2.5) in the first 2 months following the İzmit earthquake. Many aftershocks were concentrated from ~70 to 130 km along strike [Aktar et al., 2004]. Comparing Fig. 3-17 and Fig. 3-6, we cannot find clear spatial correlation between afterslip and aftershocks.

Figure 3-17. Two-dimensional plot along the fault plane showing the aftershocks (M≥2.5) in the first 64 days following the İzmit event. The hypocenter of İzmit earthquake is shown by a red star. The aftershock catalogue is managed by the Kandilli Observatory and Earthquake Research Institute (Istanbul, Turkey).

3.5 Conclusion

We used the GPS data recorded in the first 300 days following the İzmit earthquake and 4-6 years after the event to analyze the postseismic deformation induced by the strike-slip rupture. The comprehensive investigation of two postseismic mechanisms, afterslip and viscoelastic stress relaxation, provides an insight on the behavior of the fault zone and the visco-elasticity below the brittle upper crust after a large earthquake. The inversion results show that the postseismic deformation in the first 2 or 3 months was dominated by afterslip at the shallower depth. The magnitude of afterslip decayed with time. In the later time period, the inverted apparent slip locates deeper and has a very different slip pattern from
that in the first 2-3 months, which reflects the increasing effect due to viscoelastic relaxation in the lower crust/upper mantle.

Using both the long-term observations (between 2003 and 2005), and the far-field, short-term (the first 300 days) observations, we found that the viscosities of the best-fit Maxwell rheological model has an increasing trend with time. Therefore, we built a refined rheological model (E-SLS-M), made up of an elastic upper crust and a SLS lower crust overlying a Maxwell upper mantle. We estimated viscosities of $2 \times 10^{18}/7 \times 10^{19}$ Pa·s for the lower crust/upper mantle, and a relaxation strength $\beta = \frac{2}{3}$. The best-fit rheological model can better explain the surface displacement in the far-field and long-term than in the near-field and short-term. Finally, we explained all data using the estimated E-SLS-M model together with the afterslip model obtained from the residual displacement after correcting for viscoelastic relaxation. The composite model can generally explain the data in the entire temporal and spatial space.
Chapter 4 Modeling of postseismic and interseismic deformation following the 1995 M8.1 Antofagasta earthquake

Large earthquakes in the subduction zone are typically followed by significant aseismic deformation, lasting months to decades. The long-term postseismic deformation might be induced by afterslip and/or viscoelastic relaxation. To understand the postseismic relaxation processes occurring in a subduction zone, specifically the dominant mechanism in different time periods, we modeled the postseismic and interseismic deformation that occurred after the 1995, M8.1 Antofagasta earthquake in the Andean subduction zone based on displacement data recorded by campaign GPS measurements. We found a pronounced postseismic deformation, which can be explained in the first 1-2 years after the mainshock by (1) afterslip that occurred at the depth between 30 and 50 km, i.e. downdip of the main coseismic rupture, (2) deep slip at ~70 km depth, and (3) a slip concentration at ~22°S. In this time period, viscoelastic relaxation induced deformation is indiscernible. The spatial distribution of the inverted afterslip is mostly consistent with the positive Coulomb stress change produced by the Antofagasta earthquake. The inversion based on the later postseismic deformation that occurred between 1997 and 1999 basically shows a continuation of the slip obtained between 1996 and 1997, but with a smaller magnitude. However, comparing with the temporally fast-decaying afterslip, viscoelastic relaxation was detectable in this time period, and can explain the apparent slip at the depth of ~45 km. We utilize a 2-layer rheological model, made up of an elastic layer between 0 and 45 km depth and an inelastic halfspace, described by the Maxwell body. Our best-fit rheological model provides a viscosity estimate of ~3×10^{19} Pa·s. Based on the GPS measurements collected between 2002 and 2005, we obtain the best fit secular model in the fore-arc region as follows: a convergence rate of 68 mm/y between Nazca and South America plate, together with a locking depth of 45 km without transition zone or a locking depth of 40 km with a transition zone of 10 km. The GPS data between 1997 and 1999 indicate a crust-shortening rate of ~7 mm/y, and elastic strain accumulates by a rate of 8 mm/y in the back-arc region. In addition, our Coulomb stress modeling indicates that the Antofagasta event might slightly encourage the occurrence of the Tocopilla event.

4.1 Introduction

Convergence and coupling between the subducting Nazca plate and the overriding South America lead to pronounced interseismic accumulation of elastic strain and many large subduction earthquakes along the Andean subduction zone. The 1995 M8.1 Antofagasta earthquake (24.17°S, 70.74°W, Monfret et al., 1995) is one of the large events that ruptured

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the subduction interface over a length of 220 km. It locates immediately to the south of an important seismic gap of ~500 km length that had not ruptured since the occurrence of the South Peru (M9.1, 1868) and the Iquique (M9.0, 1877) megathrust earthquakes. About ten years later, the Antofagasta earthquake is followed by the 2007, M7.8 Tocopilla earthquake (22.33°S, 70.16°W, Delouis et al., 2009), which is ~150 km north to the Antofagasta earthquake. Therefore, this is one of the most interesting areas in the world for studying tectonic processes, such as seismic cycles, transient and permanent deformations.

Large plate boundary earthquakes are known to be followed by time-dependent postseismic deformation [e.g. Nur and Mavko, 1974; Miyazaki et al., 2004; Pollitz et al., 2008], which could be induced by afterslip and/or viscoelastic relaxation. Afterslip, a kind of stress relaxation process that occurred on the coseismic rupture plane or on its downdip extension [Marone and Scholz, 1988] was often utilized for postseismic modeling. It might be because afterslip model with a high degree of freedom is normally easier and better to explain the observations than a viscoelastic relaxation model. For the Antofagasta earthquake, its following postseismic deformation has been detected by the GPS and InSAR measurements [Klotz et al., 2001; Khazaradze and Klotz, 2003; Chlieh et al., 2004; Pritchard and Simons, 2006]. Most of the studies on the Antofagasta earthquake explained the postseismic deformations using the afterslip model. However, some studies on other earthquakes show that time-dependent viscoelastic relaxation might not be ignored. For example, clear postseismic deformation can still be detected after 30 years following the 1960 Valdivia M9.0 earthquake [Khazaradze and Klotz, 2003; Lorenzo-Martín et al., 2006b]. Post-glacial rebound studies [e.g. James et al., 2000] indicate that the mantle viscosity in a forearc/backarc environment can be of the order of 10^{19} Pa·s, where viscoelastic relaxation following a large subduction event might produce significant surface deformation.

In this study, we model the postseismic deformations from the point of view of both viscoelastic relaxation and afterslip based on GPS measurements collected in 10 years following the Antofagasta earthquake. This provides insight into postseismic mechanism following the subduction earthquake. Specifically, this study will find out which temporal and spatial domain was significantly influenced by afterslip/viscoelastic relaxation mechanism, and what the rheological parameter in the inelastic layer is. In addition to the investigation on postseismic mechanism, we build a new secular model for the Central Andes.

### 4.2 Data

The campaign GPS data are measured by the SAGA (South America Geodynamic Activities) network that is managed by GFZ German Research Centre for Geosciences. We utilize the measurements between 1996 and 2005. In this time period, the GPS data were collected in October of 1996, 1997, 1999, 2002, and 2005. While the displacement data based on two campaign measurements in 1996 and 1997 has been published [Khazaradze and Klotz, 2003], the remainder are preliminary data of the SAGA network, whose measurements and data processing are still on-going. These data provide us an opportunity to test our modeling approach, and investigate tectonic motion and the postseismic deformation following the
Antofagasta earthquake. We focus on the region between 21°S and 27°S (around the 1995 Antofagasta earthquake). The study region and the GPS stations are shown in Fig. 4-1. We define three areas, the central area (area B) of the Antofagasta earthquake, and the north/south areas (area A and C), which are marked by rectangles in the figure. The postseismic deformation of the Antofagasta earthquake in the central area (area B) is expected to be more significant than that in the exterior areas. Because of the large uncertainty of vertical measurements (normally higher than 1 cm), we focus on the horizontal measurements in this study. Figure 4-2 displays the horizontal velocities in the four time periods based on the five GPS campaign measurements. The velocities are projected to the direction of Nazca-South America convergence (N77°E). Notice that the measurements between 1997 and 1999 provide the longest E-W profile of ~800 km (between 71°W and 64°W).

![Figure 4-1. Map of the study region showing the GPS stations (triangles). The 1995 M8.1 Antofagasta earthquake [Monfret et al., 1995] and the 2007 M7.8 Tocopilla earthquake [Delouis et al., 2009] are shown by their focal mechanisms in the figure. The solid curve shows the location of the trench. The dotted curves display the Wadati-Benioff zone contours beneath the central Andes [Cahill and Isacks, 1992; Isacks, 1988; Tassara et al., 2006]. Three rectangles, labeled by ‘B’, ‘A’, and ‘C’, mark the central area and exterior areas of the Antofagasta earthquake.](image-url)
In the interseismic period, the velocity field is determined by the Nazca-South America convergence rate, and therefore, should be invariant and stable. The horizontal velocities along the direction of Nazca-South America convergence shown in Fig. 4-2 indicate a systematic variation of velocity versus time following the 1995 Antofagasta earthquake, especially in area B. It reflects that the velocity field was influenced by the transient postseismic deformation induced by the large event. In the time period between 1996 and 1997 (1-2 years after the mainshock), the postseismic deformation was significant, and produced a more pronounced westwards displacement field than in the later time periods. It leads to the lowest velocity measurements along the direction of Nazca-South America convergence between 1996 and 1997. A clear postseismic deformation can still be found in the GPS measurements collected between 1997 and 1999. The velocity field measured between 1999 and 2002 are close to that measured between 2002 and 2005. It indicates that stress relaxation was low after ~4 years from the Antofagasta earthquake and the difference between the possible postseismic displacements that occurred in the two time periods was indiscernible among the GPS measurements.

4.3 Preliminary secular model

The data presented in section 4.2 shows that the velocity fields in the time periods of 1999-2002 and 2002-2005 were close to each other. It reflects that the possible postseismic deformation of the Antofagasta earthquake was insignificant in the two time periods, especially between 2002 and 2005. Thus, based on the displacements measured between 2002 and 2005, we build a preliminary secular model in the study area. The dislocation
Chapter 4. Interseismic and postseismic modeling following the 1995 Antofagasta earthquake

model [Savage, 1983] is utilized for the fore-arc region, as other studies [e.g. Klotz et al., 2001; Pritchard et al., 2002]. A schematic map in Fig. 4-3 shows the fundamental tectonic movements in the study region.

![Schematic map of the Andes](image)

**Figure 4-3.** The schematic map showing the subduction zone in the central Andes with a locked zone and a transition zone. The eastwards movement is bordered to the stable Brazilian shield by a thrust fault.

The displacement field near the trench is determined mainly by 4 aspects, (1) the geometry of the subduction zone; (2) the convergence rate of Nazca-South America plate; (3) the locking depth; and (4) a possible transition zone. The geometry of the subduction zone in the study area can be constrained by the aftershock distribution of the Antofagasta earthquake [Husen et al., 1999; Sobiesiak, 2000; Sobiesiak et al., 2007] and the shape of the subducted Nazca plate [Isacks, 1988; Cahill and Isacks, 1992], as shown in Fig. 4-1. According to the aftershock locations, the subduction zone in this region has a dip angle of ~20° above 50 km depth level, and steepens at deeper depth. The dip angle at 200 km depth is about 33° according to the locations of Wadati-Benioff zone beneath the central Andes [Isacks, 1988; Cahill and Isacks, 1992; Creager et al., 1995; Gutscher, 2002; Yuan et al., 2000], which is recently supported by the gravity study [Tassara et al., 2006]. The 20° dip angle at shallow depth is close to the result based on the teleseismic data [Comte and Suarez, 1995].

Three convergence rates, 78 mm/y (NUVEL-1A, DeMets et al., 1990; DeMets et al., 1994), 68 mm/y [Norabuena et al., 1998] and 64 mm/y [Angermann et al., 1999] have been reported so far. For the locking depth, a depth range between 36 and 45 km has been given according to the distribution of earthquakes of M>6.0 [Tichelaar and Ruff, 1991]; a locking depth close to 50 km was determined based on the coseismic rupture plane of M>7.7 earthquakes and aftershock distributions [Suarez and Comte, 1993]; using the teleseismic data, Comte and Suarez [1995] documented a locking depth of 60±10 km; the results based on the geodetic data provided a locking depth of, e.g. 55 km [Chlieh et al., 2004] and 40 km [Gagnon et al., 2005]. In addition, different transition zones [e.g. Klotz et al., 2001; Chlieh et al., 2004] have also been documented. Summarizing most of the previous results, we test different models to find out the one fitting best the GPS measurements collected between 2002 and 2005.

During the calculations, we utilize PSGRN/PSCMP program [Wang et al., 2006]. The elastic parameters of a stratified medium are adopted from a seismic reference model (table 4-1) of the study area [Husen et al., 1999]. We search the locking depth between 20 and 70 km by a searching grid of 5 km, and test several values (0, 10, 20 and 30 km) for the width of transition zone. In the transition zone, the slip increases from 0 to the full slip, i.e. the convergence rate. The dislocation models are evaluated by RMS (Root-Mean-Square) errors.
Chapter 4. Interseismic and postseismic modeling following the 1995 Antofagasta earthquake

The results are shown in Fig. 4-4. The parameters for the best fit model are summarized in table 4-2. It provides a convergence rate of 64 mm/y and a locking depth of 30 km without transition zone. Figure 4-5 shows the velocity field between 2002 and 2005 observed (black) and predicted by the best-fit model (red). It indicates that, except several sites located at the north end of the study area, the obtained preliminary secular model can basically describe the data. The poor fitting at several stations is because of bending subduction zone along strike in the northern area, which is unaccounted in our model.

Table 4-1. P-wave and S-wave velocities of the applied stratified medium [Husen et al., 1999].

<table>
<thead>
<tr>
<th>Layer</th>
<th>Depth (km)</th>
<th>Vp (km/s)</th>
<th>Vs (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.0</td>
<td>5.21</td>
<td>2.99</td>
</tr>
<tr>
<td>2</td>
<td>2.5</td>
<td>5.37</td>
<td>3.09</td>
</tr>
<tr>
<td>3</td>
<td>4.5</td>
<td>5.55</td>
<td>3.19</td>
</tr>
<tr>
<td>4</td>
<td>6.5</td>
<td>5.72</td>
<td>3.29</td>
</tr>
<tr>
<td>5</td>
<td>8.5</td>
<td>5.89</td>
<td>3.39</td>
</tr>
<tr>
<td>6</td>
<td>10.5</td>
<td>5.98</td>
<td>3.44</td>
</tr>
<tr>
<td>7</td>
<td>15.0</td>
<td>6.80</td>
<td>3.75</td>
</tr>
<tr>
<td>8</td>
<td>20.0</td>
<td>6.81</td>
<td>3.88</td>
</tr>
<tr>
<td>9</td>
<td>25.0</td>
<td>6.95</td>
<td>3.94</td>
</tr>
<tr>
<td>10</td>
<td>30.0</td>
<td>6.98</td>
<td>4.05</td>
</tr>
<tr>
<td>11</td>
<td>35.0</td>
<td>7.11</td>
<td>4.11</td>
</tr>
<tr>
<td>12</td>
<td>40.0</td>
<td>7.41</td>
<td>4.18</td>
</tr>
<tr>
<td>13</td>
<td>45.0</td>
<td>7.69</td>
<td>4.30</td>
</tr>
<tr>
<td>14</td>
<td>50.0</td>
<td>8.05</td>
<td>4.39</td>
</tr>
<tr>
<td>15</td>
<td>60.0</td>
<td>8.48</td>
<td>4.73</td>
</tr>
<tr>
<td>16</td>
<td>70.0</td>
<td>8.48</td>
<td>4.78</td>
</tr>
</tbody>
</table>

Table 4-2. The parameters of the best-fit secular model.

<table>
<thead>
<tr>
<th></th>
<th>Locking depth (km)</th>
<th>Transition zone (width, km)</th>
<th>Convergence rate (mm/y)</th>
<th>RMS (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Preliminary</td>
<td>30</td>
<td>0</td>
<td>64</td>
<td>4.16</td>
</tr>
<tr>
<td>Refined</td>
<td>45</td>
<td>0</td>
<td>68</td>
<td>4.29</td>
</tr>
</tbody>
</table>
Figure 4-4. Root Mean Square (RMS) error of the preliminary secular models (for the fore-arc region) depending on the locking depth, width of the transition zone and the convergence rate between Nazca and South America plate. The filled triangle marks the best-fit result.

Figure 4-5. Velocity field between 2002 and 2005 observed (black arrows) and predicted (red arrows) by the best-fit preliminary model. The dip angle of the subduction zone between the trench (dashed curve) and 50 km depth are set as 20° according to the aftershock distribution of the Antofagasta earthquake [Husen et al., 1999; Sobiesiak, 2000; Sobiesiak et al., 2007], and the subduction zone steepens from 50 to 200 km. The stars indicate the epicenters of the Antofagasta and Tocopilla earthquakes.
4.4 Modeling of postseismic deformation

Based on the preliminary secular model presented in section 4.3, we subtract the interseismic deformation from the GPS measurements collected between 1996 and 1997, and between 1997 and 1999, respectively. The obtained postseismic displacement rates are shown in Fig. 4-8 by black arrows. Between 1997 and 1999, a large aftershock (M7.1, January 30, 1998) occurred near the mainshock. We subtract the coseismic displacements produced by this large aftershock using the source model derived by Pritchard and Simons [2006]. Using the corrected data, we investigate the postseismic deformations from the point of view of both afterslip and viscoelastic relaxation. During postseismic modeling, we utilize only the data measured west of ~67°W, basically consistent with the spatial coverage of the data (recorded between 2002 and 2005) used for the secular model for the fore-arc. The selected data with a spatial coverage of 400×600 km² around the Antofagasta epicenter should include most of the postseismic deformation following the large event.

4.4.1 Afterslip model

During afterslip modeling, the inversion method is utilized [Wang et al., 2009]. Firstly, we define the fault geometry for afterslip inversion as the fault plane with dip angle of 20° between the trench and 50 km depth as well as steepening at deeper depth. The defined fault plane is close to the coseismic rupture plane given by Pritchard and Simons [2006]. In their coseismic slip model, three fault segments along dip were used. Their coseismic slip distribution and the coseismic Coulomb stress change on the extended coseismic fault plane are shown in Fig. 4-6.

![Figure 4-6](image_url)  
**Figure 4-6.** Coseismic slip of the Antofagasta earthquake and coseismic Coulomb stress change (color coded) on an extended fault plane. The receiver fault has the strike of 5°, and a dip angle of 20°. The black contour lines with red numbers show the coseismic slip in meter (the slip of the hanging wall relative to the foot wall, with rake angle ~95°) of the Antofagasta earthquake [Pritchard and Simons,]
Chapter 4. Interseismic and postseismic modeling following the 1995 Antofagasta earthquake

2006]. The dashed curves mark the 1 and 2 meters contour lines of the coseismic slip of the 2007 Tocopilla earthquake [Delouis et al., 2009]. The stars mark the hypocenters of the Antofagasta and the Tocopilla earthquakes. The dashed straight black lines indicate the depth of the fault plane (30, 50, 70 and 90 km).

Some studies [e.g. Marone and Scholz, 1988; Perfettini and Avouac, 2004] have shown that the mainshock induced afterslip might occur on the coseismic fault plane or on its extension to the deeper part of the Earth crust. Therefore, we use an extended fault plane down to the depth of 80 km. The fault discretization is 10×10 km. The rake angles are confined in a range of ±20° around the rake angle determined for the coseismic slip [Pritchard and Simons, 2006]. The afterslip is obtained through a constrained least-square minimization [Wang et al., 2009]. The results are presented in Fig. 4-7 and Fig. 4-8. The derived two afterslip models have RMS errors of 4.3 mm/y and 8.1 mm/y, respectively.

**Figure 4-7.** The inverted afterslip rate using the postseismic velocities measured for the time periods between 1996 and 1997 (a), between 1997 and 1999 (b). The solid curve (2 meters contour line) and the dashed curves (1 and 2 meters contour lines) show the main ruptures of the Antofagasta and Tocopilla earthquakes, respectively. Their hypocenters are indicated by red and yellow stars, respectively. The black circles show the aftershocks in the first 40 days (12.8-28.9) after the Antofagasta earthquake [Husen et al., 1999; Sobiesiak, 2000]. The green circles show the larger aftershocks ($M_{w} \geq 5.8$, from Harvard Centroid moment tensor (CMT) catalogue). The white straight dashed lines mark the 30, 50 and 70 km depth levels.
Chapter 4. Interseismic and postseismic modeling following the 1995 Antofagasta earthquake

Figure 4-8. The postseismic velocities observed (black arrows) and predicted (red arrows) by the best-fit afterslip models for the time period between 1996 and 1997 (a) and for the time period between 1997 and 1999 (b). Blue rectangular shows the fault plane used for inversion. The dashed curve shows the location of the trench. The other symbols have the same meaning as in Fig. 4-5.

The inversion result based on the data measured between 1996 and 1997 shows several slip concentrations: (1) at the depth of ~30-50 km; (2) at the north end of the extended fault plane (~22°S); (3) at ~70 km depth. The slip distribution obtained using the postseismic data measured between 1997 and 1999 is similar to that obtained for the time period between 1996 and 1997, but has a smaller magnitude. The modeling results shown in Fig. 4-7 and 4-8 indicate that the afterslip models can well explain the postseismic displacements that occurred between 1996 and 1999.
Chapter 4. Interseismic and postseismic modeling following the 1995 Antofagasta earthquake

A comparison between the coseismic rupture, coseismic Coulomb stress change, aftershocks [Husen et al., 1999; Sobiesiak, 2000] and afterslip distribution indicates that: firstly, the significant afterslip occurred downdip of the main coseismic rupture; secondly, afterslip mostly occurred in the area where coseismic Coulomb stress change was increased; thirdly, the aftershocks are concentrated downdip of the main coseismic rupture, but in the area where the afterslip was relatively low (cf. Fig. 4-7).

4.4.2 Composite model including both viscoelastic relaxation and afterslip mechanisms

The occurrence of a large earthquake produces stress change in the volume around the rupture plane. In the framework of viscoelastic relaxation, the coseismic stress disturbance is relaxed after the event by viscous flow in the inelastic layer, e.g. in the lower crust or upper mantle. In this study, we utilize a simple 2-layer rheological model, made up of the elastic crust and an inelastic halfspace, described by a Maxwell body. As for secular modeling, we utilize the elastic parameters of a stratified medium obtained from seismic reference model of this study area [Husen et al., 1999].

During the analysis for viscoelastic relaxation, we mainly use the data recorded between 1997 and 1999. The GPS measurements collected between 1997 and 1999 provide a good spatial coverage and the longest E-W displacement profile. In this time period (e.g. 2-4 years after the mainshock), viscoelastic relaxation is still expected to produce significant deformation, which is not dominated by afterslip. It is different from the situation in the time period immediately after the mainshock, when the displacement produced by afterslip is so significant that the viscoelastic relaxation induced displacement is indiscernible [Wang et al., 2009]. We use the coseismic slip derived by Pritchard and Simons [2006] as source model, and test the rheological models with variant inelastic layer depths, 40 km, 45 km, 50 km, and 60 km, respectively. The velocity profiles (at 24°S) predicted by the models for the time period between 1997 and 1999 are shown in Fig. 4-9a.

We assume viscosity in the inelastic layer of $3 \times 10^{19}$ Pa·s. The results show that, for a fixed viscosity value in the inelastic layer, the rheological models with a deeper inelastic layer produce smaller surface displacements; and vice versa. The significant difference occurred at the distance of ~60 km from the trench. At the distance of ~80 km and further away from the trench, where the GPS measurements were collected, the difference between the models with variant inelastic layer depth is not distinguishable. Therefore, in this study, we focus on the rheological model with the inelastic layer depth of 45 km, as indicated by the aftershock distribution [Husen et al., 1999; Sobiesiak, 2000]. Figure. 4-9b shows the spatial distribution of the viscoelastic relaxation induced postseismic velocity that occurred between 1997 and 1999 based on the rheological model with the inelastic layer depth of 45 km and viscosity of $3 \times 10^{19}$ Pa·s.

According to Fig. 4-9, viscoelastic relaxation produces a surface displacement field positive to the West at a distance of ~110 km and further away from the trench (east of ~70.5°W), the same as the postseismic observations (cf. Fig. 4-8b). However, the viscous flow produces the displacements near the trench (less than 110 km distance or west of ~70.5°W) positive to the East, which is contrary to the observations. It indicates, at least, that viscoelastic relaxation mechanism alone is not able to explain the observed postseismic...
displacements. As modeled in the last section and also documented by some publications [Chlieh et al., 2004; Pritchard and Simons, 2006], the observed postseismic displacement at this stage was most likely produced by afterslip.

![Figure 4-9](image)

**Figure 4-9.** Postseismic velocities (mm/y) between 1997 and 1999 measured and predicted by the rheological models. (a) horizontal velocities measured (dots with error bars) in area B and predicted on the E-W profile at 24°S by the rheological models with viscosity of $3 \times 10^{19}$ Pa·s and the inelastic layer depths of 40, 45, 50 and 60 km, respectively. The black and gray error bars show the 1-sigma and 3-sigma measurement uncertainties, respectively. The thick line with arrows on the top indicates the source location relative to the trench. (b) the viscoelastic relaxation induced velocity (positive to N95°E) field based on the rheological model with the inelastic layer depth of 45 km and viscosity of $3 \times 10^{19}$ Pa·s. The star marks the Antofagasta earthquake, and the triangles show the GPS sites.

The two mechanisms, afterslip and viscoelastic relaxation, might dominate different spatial domains in the time period between 1997 and 1999. According to Fig. 4-9a, the viscous flow produced the most significant displacement at the distance of ~80-100 km from the trench (between 70.2°W and 70.6°W) during the time period between 1997 and 1999, and positive to the east. At the distance further away from the trench between 70.6°W and 68.0°W (see Fig. 4-9b), the displacement is nearly constant and positive to the west. In comparison, the GPS measurements collected in area B (the central area of the Antofagasta
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Earthquake) show the most significant postseismic displacement (positive to the west) at the
distance of ~240 km away from the trench; while the observed displacements close to the
trench are small. It indicates that the significant postseismic deformations far away from the
trench (at the distance of ~240 km) were likely produced by deep afterslip.

Therefore, to confine viscosity in the rheological model, we use the data measured
between 23°S and 25°S, where the displacement field was influenced by viscoelastic
relaxation (cf. Fig. 4-9b). We do not use all of the measured postseismic deformation data
here because too much additional signal, e.g. likely afterslip induced postseismic
deformation north of 23°S, may overwhelm the signal related to viscoelastic relaxation.
Using the selected data, we firstly derive an afterslip model, which has a RMS error of 9.8
mm/y. According to Fig. 4-9a, if viscoelastic relaxation occurred between 1997 and 1999,
the derived afterslip should include a certain amount of viscoelastic relaxation induced
deformation. Thus, we build a composite model, which is described as,

\[ y = \alpha \times y_{ASP}(u) + y_{VER}(\eta), \quad (4.1) \]

where \( y \) is the observed displacements, \( y_{ASP} \) is the displacements predicted by the
afterslip model with slip \( u \), and \( y_{VER} \) is that predicted by the viscoelastic relaxation
model with viscosity \( \eta \) of the inelastic layer. \( \alpha \) is the weight factor, should be a value
smaller than 1. Using the grid search method, we can obtain the optimum \( \alpha \) and \( \eta \) in the
composite model that best fits the data. The grid search results (Fig. 4-10a) show that the
model with an \( \alpha \) value of 0.85 and an \( \eta \) value of \( 3 \times 10^{19} \) Pa·s provides the lowest RMS
error of 9.1 mm/y, smaller than the RMS error of the afterslip model including viscoelastic
relaxation.

Now, we try another approach to investigate possible viscoelastic relaxation. Because
the displacements produced by viscoelastic relaxation are only determined by viscosity of
the inelastic layer in our rheological model, we can firstly remove the viscoelastic
relaxation induced displacements from the data, and then, invert for afterslip using the
residual displacements. This composite model is defined as,

\[ y = y_{VER}(\eta) + y_{ASP}(u | \eta), \quad (4.2) \]

where \( y_{ASP}(u | \eta) \) represents the surface displacement predicted by the afterslip \( u \)
derived from the data after correcting for viscoelastic relaxation based on the rheological
model with viscosity \( \eta \). The others have the same meaning as those in Eq. 4.1. The results
(Fig. 4-10b) provide that the composite model with viscosity of \( 3 \times 10^{19} \) Pa·s best fits the
data. In Fig. 4-10b, the gray curve shows the results when we apply this approach to the
data recorded between 1996 and 1997. The RMS errors do not show a localized minimum
viscosity, indicating that a composite model is not needed for this data set. It suggests that
the time period between 1996 and 1997 was dominated by afterslip mechanism, and the
viscoelastic relaxation induced deformation cannot be constrained due to the strong signal
produced by afterslip.

Based on the postseismic displacements measured between 1997 and 1999, both of the
investigating approaches obtain the viscosity in the inelastic layer in the Antofagasta region
of \( \approx 3 \times 10^{19} \) Pa·s. The afterslip model derived from the data after correcting for viscoelastic
relaxation based on the rheological model with viscosity of $3 \times 10^{19}$ Pa·s is shown in Fig. 4-11. Comparing with Fig. 4-7b, the apparent slip around the depth of ~45 km and at the latitude of ~24°S disappears after correcting for viscoelastic relaxation. It implies that this apparent slip at the depth of ~45 km might be induced by viscoelastic relaxation. The depth of ~45 km is consistent with the reported Moho depth [Yuan et al., 2000] and the depth of the inelastic layer applied in this study.

![Graph](image.png)

**Figure 4-10.** Plots for RMS errors (mm/y) of the composite models including both viscoelastic relaxation and afterslip mechanisms. (a) Contour plot for the RMS errors of the composite model with variant $\alpha$ and $\eta$ values (cf. Eq. 4.1). (b) RMS error versus viscosity plot, which is obtained from the second approach (cf. Eq. 4.2). In (b), the solid curve and dotted curve show the RMS errors of the composite models for the time period between 1997 and 1999, between 1996 and 1997, respectively. The triangles mark the optimal parameters.
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4.5 Refined secular model

In the refined secular model, we consider firstly the interseismic deformation in the fore-arc after correcting for viscoelastic relaxation that occurred between 2002 and 2005 based on our preferred viscoelastic relaxation model. We secondly model crustal shortening that occurred in the back-arc region. For this, two data sets are utilized: (1) the data recorded between 2002 and 2005 in the fore-arc region; (2) the measurements east of 67°W (in the back-arc region) collected between 1997 and 1999.

4.5.1 Refined secular model for the fore-arc region

Based on the rheological model derived in section 4.2, the viscoelastic relaxation induced displacement was still detectable between 2002 and 2005, especially near the trench, where a displacement rate of ~3 mm/y is predicted by the viscoelastic relaxation model. It can cause a bias in our preliminary secular model. Therefore, we correct the data recorded between 2002 and 2005 for viscoelastic relaxation, and then build a refined secular model for the fore-arc region. The same grid search method presented in section 4.3 is utilized. The results are shown in Fig. 4-12. It provides the best-fit model: the convergence rate of 68 mm/y, a locking depth of 45 km and no transition zone. Notice that several other models can explain the data similarly well as the best-fit model. These are the models with a locking depth of 40 km and transition zone of 10 km, the model with a locking depth of 35 km and transition zone of 20 km, and the model with a locking depth of 30 km and transition zone of 35 km. It indicates a tradeoff between the locking depth and transition zone when the secular model is only confined by geodetic data. Fig. 4-13 shows the result for the best-fit model (locking depth of 45 km and no transition zone).
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Figure 4-12. Root Mean Square (RMS) error of the refined secular models (for the fore-arc region) depending on the locking depth, width of the transition zone and the convergence rate between Nazca and South America plate. The filled triangle marks the best-fit result.

Figure 4-13. Map showing the velocity field between 2002 and 2005 observed (after correcting for viscoelastic relaxation produced by the Antofagasta earthquake; black arrows) and predicted by the refined secular model (red arrows). The other symbols have the same meanings as those in Fig. 4-5.
4.5.2 Modeling for crust-shortening in the back-arc region

The long E-W profile of the measurements collected between 1997 and 1999 provide us an opportunity to investigate the crust-shortening effect in the back-arc region. As demonstrated in Fig. 4-9a, a pronounced deviation from the secular model (relative to the stable South America) occurred at the distance of 500 km (east of ~67°W) and further away from the trench. That means, the eastward displacement rates observed in this region is smaller than that predicted by the secular model based on the measurements collected in the fore-arc region. After subtraction for the amount predicted by the preliminary/refined secular model, the residual velocity field is uniformly positive to the west, as shown by black arrows in Fig. 4-15b.

Geological evidence shows that crustal shortening in the past few million years has been concentrated in the Sub-Andean fold and thrust belt [Hindle et al., 2002; Oncken et al., 2006 and references therein]. Therefore, we model the crust-shortening motion utilizing the main Sub-Andean thrust fault that located in the Sub-Andean region [Oncken et al., 2006], which has strike of ~184°. We apply a typical continental stratified medium: the elastic upper crust (0-20 km), the inelastic lower crust (20-45 km) described by the Standard Linear Solid element and the viscous mantle (below 45 km) described by the Maxwell element at the steady state. We model the long-term and permanent crust-shortening deformation using a back-creep in the elastic upper crust and the inelastic lower crust. The back-creep is utilized to describe the backstop that the stable Brazilian shield imposed on shortening at the eastern border of the Andean chain. On the other hand, the seismicity data [Engdahl et al., 1998; Oncken et al., 2006] indicate that this area is characterized by relatively high seismic level. Thus, part of the tectonic motion is accumulated elastically in the interseismic time period. We model it by a back-slip in the elastic layer. The elastic strain accumulated by the rate of back-slip rate estimated from the data, will be released by the earthquakes. A schematic diagram for the model in the back-arc region is shown in Fig. 4-14.

![Figure 4-14](image.png)

**Figure 4-14.** A schematic diagram showing the model for the back-arc region.

During the modeling, the location of the fault is fixed according to the Sub-Andean fold and thrust belt (FTB). We adjust the dip angle of the fault and the slip amount on the fault to best fit the data. Using the data collected between 1997 and 1999 after subtracting the displacements predicted by the refined secular model, and the postseismic
displacements predicted by the afterslip model and the preferred rheological model, we obtain that the best-fit model for the back-arc region (Fig. 4-15): the fault with dip angle of 10°, the back-creep rate of 7 mm/y and the back-slip rate of 8 mm/y, with the RMS error of 7.2 mm/y. The model suggests a crust-shortening rate of ~7 mm/y in this area. The estimate of the shortening rate is basically consistent with 5-10 mm/y shortening rate obtained from geological observations spanning the last 25 Ma. The accumulation of elastic strain in the interseismic time period consists with the relatively high seismicity in this region [Engdahl et al., 1998; Oncken et al., 2006]. In addition, we also try to model the measurements in the back-arc region using the back-slip and back-creep separately. The modeling results are presented in the Supplement.

Figure 4-15. Modeling results for the back-arc region. In (a), the ellipses show the RMS error contour lines for the model with different dip angle of the fault. The symbols in the ellipses indicate the optimum
locations. The best-fit model provides that the dip angle of 10°, the back-slip rate of 8 mm/y and the back-creep rate of 7 mm/y. In (b), the measured velocity after subtracting the amount predicted by the refined secular model for the fore-arc region, the postseismic velocities predicted by the afterslip model and the preferred rheological model are shown by black arrows. The red arrows show the velocity field predicted by the best-fit model.

In addition to the model presented so far, we also test the other two models, the back-creep model and the back-slip model, respectively. The back-creep model is based on the assumption that all of the westward tectonic motion is converted to the permanent crust-shortening deformation. The back-slip model is based on the assumption that the westward tectonic motion from the convergence between the Nazca and South America plate are all stored elastically and will be released by an earthquake in the future. The elastic model has been commonly used in the previous studies [Norabuena et al., 1998].

Using the same data and the fault plane as presented in this section, we obtain that the best-fit back-creep model (in Fig. 4-16a), the dip angle of 20°W and slip amount of ~13 mm/y. The estimated slip rate is higher than the geologically estimated crust-shortening rate (5-10 mm/y). The results for the back-slip model (Fig. 4-16c) provide the dip angle of 10°W and the slip amount of ~14 mm/y. This model has a good fit to the data in the near field, but poorly explains the data in the far field (cf. Fig. 4-16d). Therefore, both of the models cannot well explain the measurements in the back-arc region.
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Figure 4-16. Modeling results for the back-arc region. Plots of the RMS errors for the back-creep model (a), and the back-slip model (c), where the maximum value for grid search is set as 20 mm/y, which is far higher than the geologically observed value of ~10 mm/y in this region. The circles in (a) and (c) mark the best-fit results; In (b) and (d), the measured displacement rates (black arrows) after subtracting for the those predicted by the refined secular model, the postseismic displacement rates predicted by the afterslip model and the preferred rheological model are shown by black arrows. The red arrows show the displacement rates predicted by the best-fit models based on the back-creep model (b), the back-slip model (d).

4.6 Discussion

4.6.1 Coseismic stress changes, afterslip and the 2007, M7.8 Tocopilla earthquake

To understand the mechanism of afterslip, we calculated the coseismic Coulomb stress change on the extended fault plane. During the calculation, we took the receiver fault with a strike of 5° and a dip angle of 20°, and set the effective friction coefficient to 0.4. The results (in Fig. 4-6) show that the significant slip concentration between 30 and 50 km depth obtained for the time period between 1996 and 1997 located downdip of the main coseismic rupture, where the Coulomb stress change was mostly positive.

The inversion results in this study also show significant deep afterslip at ~70 km depth. Even though it has been indicated that the fault geometry has influences on distribution of the inverted deep slip [Moreno et al., 2009], the geophysical data, e.g., Antofagasta aftershocks [Husen et al., 1999; Sobiesiak, 2000], tomography [Yuan et al., 2000] and gravity data [Tassara et al., 2006], show that our applied fault geometry is a reasonable approximation in the study region. Our results indicate that Coulomb stress change at ~70 km depth is comparatively lower than at 30-50 km depth. However, the Coulomb stress
change at the deep depth is far higher than the triggering threshold for aftershocks, which is normally believed to be 0.01 or 0.1 MPa [Reasenberg and Simpson, 1992; Stein et al., 1992; Harris et al., 1995]. On the other hand, the deep afterslip (see Fig. 4-7) could also be attributed to a stress relaxation that occurred in the serpentinized subduction interface, where the strong, dry rocks are chemically changed into weak, hydrous rocks composed of serpentine minerals, when water moves into cool mantle rocks. Serpentinized forearc mantle has been found in the Central Andes based on seismic studies [Giese et al., 1999]. The slippery serpentinite in the subduction zone suggests the high possibility for the occurrence of afterslip. Therefore, both pronounced Coulomb stress change and the possible low friction interface are likely responsible to the large deep afterslip.

In addition, the inversion results derived from the GPS measurements also provide apparent afterslip at the ~22°S, which is 100 km further north of the Antofagasta rupture. This afterslip might be overestimated when we notice that our secular model does not well describe the observed displacements in the area due to unaccounted along strike bending of the subduction zone. This causes a bias during correcting for the tectonic motion. On the other hand, the Coulomb stress change at ~22°S is low, but is still positive. Pritchard and Simons [2006] documented that along-strike variation of frictional property contributes to the spatial variation of afterslip. Therefore, a low friction interface in the northern area is possibly a reason for the afterslip in this area.

Comparing our inversion results with the afterslip derived by Pritchard and Simons [2006] based on the InSAR measurements, both of them have a similar spatial pattern. Their high slip magnitude north of the main coseismic rupture are likely caused by the large aftershocks that occurred in 1995 and 1996 (cf. green circles shown in Fig. 4-7); while the data we used were collected later, i.e. from October 1996, so our results do not resolve such a high slip in the north.

The M7.8 Tocopilla earthquake is another recent subduction event that occurred at the contact of the Nazca and South America plates. It ruptured the seismic gap over a distance of about 150 km and a width of less than 50 km, in the northern prolongation of the subduction zone ruptured during the Antofagasta earthquake. The Antofagasta earthquake produced positive Coulomb stress change in the rupture area of the Tocopilla earthquake. In addition, we also calculate the total Coulomb stress change produced by the mainshock together with the large aftershocks and the afterslip between 1996 and 1999 (without considering viscoelastic relaxation). Beside the strong M7.1 aftershock (January 30, 1998), we here include additionally seven larger aftershocks (M≥5.8, from Harvard Centroid moment tensor (CMT) catalogue), as shown in Fig. 4-17. It indicates that most of the coseismic rupture area of the Tocopilla earthquake, especially the hypocenter, is in the area where Coulomb stress change is positive. When viscoelastic relaxation process is additionally included, the positive Coulomb stress change near the Tocopilla hypocenter in 2007 is increased by more than 0.01 MPa. Therefore, the Antofagasta earthquake likely encouraged the occurrence of the Tocopilla earthquake.
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Figure 4-17. Coulomb stress change (color coded) produced by the Antofagasta mainshock, eight large aftershocks (Mw≥5.8, from Harvard Centroid moment tensor (CMT) catalogue) and afterslip between 1996 and 1999. The receiver fault is defined as strike of 5° and dip angle of 20°. The dashed curves indicate the main coseismic rupture of the Tocopilla earthquake and the dashed lines show the different depth levels. The red and yellow stars mark the hypocenters of the Antofagasta and the Tocopilla earthquake, respectively.

4.6.2 Postseismic mechanisms following the Antofagasta earthquake

Based on the preliminary GPS data, we have found that the postseismic deformation was significant in the first 4 years following the Antofagasta earthquake, and was still influential after 7-10 years following the big event. In this study, we investigated two postseismic mechanisms, afterslip and viscoelastic relaxation. After a thrust event in a subduction zone, e.g. the Antofagasta earthquake, afterslip produces a similar displacement field as the mainshock; while viscoelastic relaxation produces more significant deformation near the trench than that in the area further away, and the displacement near the trench is opposite to that produced by the mainshock/afterslip. These characteristics provide an opportunity to discern the two mechanisms in different temporal and spatial domains.

Our investigations based on the GPS measurements between 1996 and 2005 show that the postseismic deformation in the early time period of 1-2 years following the Antofagasta earthquake was dominated by afterslip. The viscoelastic relaxation induced postseismic deformation was indiscernible. The results for the time period between 1997 and 1999 basically display a continuation of afterslip that occurred between 1996 and 1997. However, comparing with the temporally fast-decaying afterslip [Wang et al., 2009], viscoelastic relaxation starts to be detectable. The apparent slip located at ~45 km depth that occurred between 1997 and 1999 was mainly explained by viscoelastic flow. Based on the postseismic measurements between 1997 and 1999, we estimated the viscosity of the inelastic layer in the Antofagasta region to be ~3×10^{19} Pa·s. This result is consistent with
the viscosity estimate \((4 \times 10^{19} \text{ Pa·s})\) based on the postseismic deformation of 30 years after the Valdivia earthquake \cite{Khazaradze2003}, but smaller than the estimate of \(10^{20} \text{ Pa·s}\) reported by Lorenzo-Martín et al. \cite{2006b} for the same event.

Based on the GPS data measured between 2002 and 2005 and the data after correcting for postseismic viscoelastic relaxation, we have obtained two different secular models for the fore-arc region, the preliminary model and the refined model. The preliminary model provides a convergence rate of 64 mm/y and a locking depth of 30 km; while the refined secular model provides a convergence rate of 68 mm/y and a deeper locking depth of 45 km. The difference indicates the pronounced contribution from postseismic viscoelastic relaxation between 2002 and 2005. A comparison between the preliminary and refined secular model is shown in Fig. 4-18. It indicates a difference in the area near the trench. The preliminary model with a shallow locking depth produces a sharp deformation profile and much higher deformation near the trench; while the refined secular model produces a flatter velocity profile and a smaller deformation. Comparatively, the refined model with locking depth of 45 km is closer to the previous reported interseismic modeling results. For example, Klotz et al. \cite{1999} documented a locking depth of 50 km; Chlieh et al. \cite{2004} reported an interseismic model fully locked down to a depth of 35 km with a transition zone between a depth of 35 and 55 km. In addition, our derived locking depth is consistent with the local seismicity distribution, which is above \(\sim 50 \text{ km}\) depth. The improvement of the secular model based on the data after correcting for viscoelastic relaxation, on the other hand, indicates that our viscosity estimate of \(\sim 3 \times 10^{19} \text{ Pa·s}\) is a reasonable approximation to the rheology in this area.

![Figure 4-18](image)

**Figure 4-18.** The velocity (in the South America fixed reference frame) profile (at 24°S) before (gray dots with error bars) and after (black dots with error bars) correcting for viscoelastic relaxation based on the rheological model with the locking depth of 45 km and viscosity of viscosity of \(3 \times 10^{19} \text{ Pa·s}\). The profiles predicted by the preliminary and refined secular model are shown by gray and black dotted curves, respectively.
4.6.3 Viscoelastic relaxation produced by the 2007 Tocopilla earthquake

The recent M7.8 Tocopilla earthquake occurred nearly immediately north of the subduction zone ruptured during the Antofagasta earthquake. Therefore, similar rheological properties are expected in the rupture area. Based on our preferred rheological model derived using the postseismic measurements following the Antofagasta earthquake, we can deduce the magnitude of the expected viscoelastic relaxation in response to the Tocopilla earthquake. While it is still an interesting topic to distinguish the possible postseismic mechanisms, such as afterslip and viscoelastic relaxation, this investigation is able to provide some hints concerning the optimal station distribution of the measurements in order to capture postseismic deformation, e.g. GPS measurements.

Figure 4-19 shows the displacement profile along 22°S, predicted by our preferred viscoelastic relaxation model taking the coseismic slip of the Tocopilla earthquake [Delouis et al., 2009] as input. The postseismic displacement that occurred in the first month, one year and two years following the event are shown by different curves, respectively. The results indicate that the maximal horizontal postseismic displacement in two years following the big event reaches ~8 mm, and is positive to the East. This significant postseismic deformation occurred at the distance of ~80 km from the trench.

Based on the InSAR measurements covering the coseismic period and 45 days postseismic period, Chlieh et al. [Chlieh et al., 2008] reported that the released postseismic moment could have been more than 30% of the coseismic moment release. It is much higher than that predicted by the rheological model. Therefore, the early postseismic deformation was nearly entirely produced by afterslip. According to the profiles, the GPS stations located in the area at the distance of ~80 km from the trench should be able to capture a clear postseismic signal produced by viscoelastic relaxation. Therefore, better data coverage in this area would be very helpful to distinguish the postseismic mechanisms and to build a refined rheological model in the region around the Tocopilla earthquake.
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4.7 Conclusion

For the Andean subduction zone between 21°S and 27°S, we investigated the interseismic deformation and the different postseismic deformation processes following the 1995, M8.1 Antofagasta earthquake. Pronounced postseismic deformation was observed in the first 4 years following the mainshock. During the investigation for postseismic deformation, we utilized an inversion method to obtain afterslip, and a linear rheological model, made up of an elastic layer overlying a Maxwell inelastic halfspace, to model viscoelastic relaxation induced by the event. Our investigations show that the two mechanisms dominated different temporal and spatial space. In the first 1-2 years, afterslip was dominant. We inverted several slip concentrations: downdip of the main coseismic rupture between 30 and 50 km depth; a deep slip at ~70 km depth and a slip near 22°S. Most of the afterslip locates in the area where coseismic Coulomb stress change was positive. In this time period, viscoelastic relaxation induced deformation is indiscernible. The inverted slip for the time period between 1997 and 1999 basically shows a continuation of the slip obtained between 1996 and 1997, but with a smaller magnitude. However, comparing with the temporally fast-decaying afterslip, viscoelastic relaxation was detectable in this time period and can explain the apparent slip at the depth of ~45 km. Our best-fit rheological model provides a viscosity estimate of ~3×10^{19} Pa·s. Coulomb stress modeling indicates that the co- and postseismic stresses induced by the Antofagasta earthquake encouraged the occurrence of the Tocopilla event.

Using the GPS measurements collected between 2002 and 2005, and correcting for viscoelastic relaxation based on our rheological model, we built a secular model for the fore-arc region with a Nazca-South America convergence rate of 68 mm/y and a locking depth of 45 km. This locking depth fits better to the seismicity depth distribution, and is
closer to the previous published secular models in this area, than the secular model based on the data without correcting for viscoelastic relaxation, which provides a locking depth of 30 km. The improvement of the model based on the data after correction reflects that our estimated viscosity of \( \sim 3 \times 10^{19} \) Pa·s is a reasonable approximation to the rheology in the Antofagasta area. Meanwhile, our secular model for the back-arc region provides a long-term crust-shortening rate of \( \sim 7 \) mm/y and elastic strain accumulates in the back-arc region by a rate of 8 mm/y.
Chapter 5  
Postseismic deformation induced by brittle rock damage of aftershocks

Large earthquakes are commonly followed by significant aseismic deformations, which were usually explained by viscous flow in the deep ductile zone and/or afterslip along the up-/down-dip extension of the coseismic fault plane due to rate-strengthening friction. Besides postseismic deformation, abundance of aftershocks occurs densely around the coseismic rupture zone. Laboratory experiments indicate that “microscopic” brittle rock failures (acoustic emission) are associated with a “macroscopic” damage-related inelastic relaxation. Utilizing basic relations between local brittle failures and gradual inelastic strain in the framework of damage rheology, we develop connections between aftershocks and the aftershocks-induced component of geodetic deformation. This provides a candidate for interpreting the early postseismic relaxation detected (e.g., according to inversions constrained by the geodetic measurements) in the seismogenic zone, specially that overlapping the aftershock area, which is hard to be explained by either rate-strengthening friction or viscous flow. Assuming the Omori-Utsu decay rate for aftershocks, we find that the temporal decay of the damage-related postseismic relaxation follows a generalized power law relation with the standard Omori-Utsu law as a limit case. Our results provide a way for estimating the separate contributions to observed postseismic deformation that stems from brittle failures in the seismogenic zone (aftershocks) and additional non-seismic (ductile) processes. Using the obtained theoretical expectations, we analyze postseismic displacements measured by GPS stations around the North Anatolian fault during the first \(~3\) months following the 1999 M7.4 İzmit earthquake. We find that the observed postseismic displacements decay slower than the aftershock seismicity. Detailed analyses indicate that the contributions to the observed geodetic deformation from elastic relaxation associated with the aftershocks and the ongoing tectonic motion are generally small. Based on our theoretical results, we conclude that up to 50\% of the measured surface displacements at near-fault sites can be attributed to aftershock-induced inelastic deformation in the seismogenic zone. The remainder postseismic deformation can generally be explained by the relaxation occurred at the deep layer based on the rheological model of Wang et al [2009].

5.1 Introduction

Postseismic deformation, following the occurrence of large earthquakes, has long been detected by means of geodetic measurements \cite[e.g. Shen et al., 1994; Savage and Svarc, 1997; Bürgmann et al., 1997; Ergintav et al., 2002; Ryder et al., 2007; Hearn et al., 2009]{Shen94}. Several physical models have been proposed to explain the postseismic relaxation processes. Montési \cite{Montesi04} suggested that the postseismic relaxation is, due to coseismically produced stress, nonlinear viscous flow in the deep ductile shear zone. He derived that the postseismic relaxation process follows power-law decay with time. The studies based on linear viscoelastic relaxation in the ductile zone found that the postseismic deformation rate has an exponential decay \cite{Shen94, Savage97}. Besides the mechanics origin at the deep ductile zone, rate-strengthening friction along up/down-dip extension of the coseismic fault plane (namely frictional-afterslip) was suggested to be responsible for the postseismic deformation \cite{Marone91}. According to rock experiments and theoretical investigations \cite{Blanpied91, Chester95}, rate-strengthening, which is thermally activated or related to the hardening mechanism (involving dilatancy) \cite{Scholz98}, may occur at shallow depth on fault cutting through loose sediments \cite{Marone88} or at the deep depth below the seismogenic zone. In this framework, the postseismic relaxation follows a time-dependent logarithmic function.

Besides these theoretical studies, inversions constrained by geodetic data provided information where postseismic relaxation occurs. To distinguish the frictional-afterslip, we name the slip from inversions as inverted-afterslip. It has been indicated that, following e.g., 1999 M7.1 Hector Mine \cite{Owen02}, and 2003 Chengkung M6.5 \cite{Hsu09} earthquakes, the inverted-afterslip is principally on the fault beneath the coseismic rupture, which is consistent with the viscous flow or rate-strengthening mechanism. On the other hand, many studies have shown the early inverted-afterslip locates at a shallower depth in the seismogenic zone following some big events, e.g. 1989 Loma Prieta earthquake \cite{Pollitz98}; 1999 M7.4 İzmit earthquake \cite{Buehrmann02, Wang09} and 2003 M6.9 Boumerdes earthquake (Algeria) \cite{Mahsas08}. The frequently observed early inverted-afterslip in the seismogenic zone is difficult to be strictly explained by either deep viscous flow or rate-strengthening friction. Particularly, many reported inverted-afterslip locates in areas of high seismic activity (see details below), where rate-weakening friction is expected. Therefore, the mechanism of the shallow postseismic relaxation should be carefully considered.

In addition to the observed aseismic deformations, aftershocks are significant marker of the relaxation processes following large earthquakes. Some similarities have been found
between aftershocks and aseismic deformation. Firstly, both have generally consistent kinematic motion with the mainshocks [Bürgmann et al., 2002; Hsu et al., 2002; Bohnhoff et al., 2006]. Secondly, an Omori-type decay which is well-known for the aftershock sequences has also been noticed for the postseismic displacements recorded by the continuous GPS measurements following e.g. the 1999 Chi-Chi earthquake [Perfettini and Avouac, 2004; Savage et al., 2007] and 2001 M8.4 Peru earthquake [Perfettini et al., 2005]. Some investigations also indicated spatial correlations between aftershocks and inverted-afterslip. For example, Miyazaki et al. [2004] showed that the aftershocks of the 2003 Tokachi-oki earthquake overlap the afterslip area constrained by the GPS measurements. Such spatial correlations have also been found following e.g., Chi-Chi earthquake [Yu et al., 2003] and 2005 M8.7 Nias-Simeulue Earthquake [Hsu et al., 2006]. In addition, it was reported that most of the postseismic deformation following the 2003 M7.3 Altai earthquake can be explained by seismic moment release in aftershocks [Barbot et al., 2008]. These observations suggest that aftershock sequences and postseismic geodetic deformation might be closely correlated.

Aftershocks are mainly located within the seismogenic volume affected by the mainshock-induced stresses. Geological and geophysical studies indicate that large faults are surrounded by tabular damage zones with reduced elastic moduli compared to the host rocks [e.g., Chester et al., 1993; Ben-Zion et al., 2003; Fialko et al., 2002]. In addition, recent seismological studies documented clear temporal changes in the elastic properties of fault damage zones following the occurrence of large earthquakes [e.g., Sawazaki et al., 2006; Peng and Ben-Zion, 2006; Rubinstein et al., 2007; Wu et al., 2009]. It is reasonable to assume that similar changes of elastic moduli accompany locally the occurrence of aftershocks.

Ben-Zion and Lyakhovsky [2006] analyzed aftershocks in a viscoelastic damage rheology model, based on nonlinear continuum mechanics and thermodynamics, for evolving elastic properties and related deformation patterns in rocks sustaining irreversible brittle deformation. The effective elastic moduli in the damage model are functions of an evolving non-dimensional state variable \(0 \leq \alpha \leq 1\) representing the local crack density [Lyakhovsky et al., 1997]. The evolution of the microcrack density during local brittle failures (e.g., aftershocks) changes the effective viscosity of the medium and leads to gradual inelastic deformation [Hamiel et al., 2004].

Based on stress-strain relations in the framework of damage rheology, we build in the present paper connections between aftershocks and damage-related postseismic relaxation, and derive the aftershock-induced components in the postseismic geodetic deformation. Assuming that the decay rate of aftershocks obeys the Omori-Utsu law, we show in section 5.2 that the time evolution of the damage-related postseismic deformation rate follows a
generalized Omori-type decay, with the standard Omori-Utsu law as a limit case. Based on
the theoretical expectation, the aftershocks-related postseismic deformation can be
separated from non-brittle ductile deformation (e.g., deep afterslip or viscoelastic
relaxation). We present in section 5.3 an application example, the 1999 M7.4 İzmit
earthquake around the North Anatolian fault in Turkey. The results are discussed in section
5.4 and summarized in section 5.5.

5.2 Time evolution of aftershocks-induced aseismic
deformation

5.2.1 Deformation in a viscoelastic damage model

Ben-Zion and Lyakhovsky [2006] have proposed a viscoelastic damage model in relation to
aftershocks [see also Ben-Zion, 2008]. We adopt in the present study the relevant
ingredients of the viscoelastic damage model. The total deformation field in the viscoelastic
damage model can be written, in analogy with Maxwell viscoelasticity, as the sum of three
strain components,

$$\varepsilon_{ij} = \varepsilon_{ij}^e + \varepsilon_{ij}^d + \varepsilon_{ij}^\alpha,$$

(5.1a)

where $\varepsilon_{ij}^e$ is elastic strain, $\varepsilon_{ij}^\alpha$ denotes a damage-related inelastic strain and $\varepsilon_{ij}^d$
represents a non-brittle ductile strain. In a simple model consisting of a brittle seismogenic
zone over a ductile substrate, the third term in Eq. 5.1a may be assumed to be zero in the
seismogenic zone, while the first 2 terms may be assumed to be zero in the ductile substrate.
For a simplified situation of uniform deformation in a single direction, Eq. 5.1a may be
replaced by the corresponding scalar version

$$\varepsilon' = \varepsilon^e + \varepsilon^d + \varepsilon^\alpha.$$

(5.1b)

For a 1D case of uniform deformation, the relation between the stress $\sigma$ and elastic
strain in the damage model is given [Ben-Zion and Lyakhovsky, 2006] by

$$\sigma = 2\mu_0(1-\alpha)\varepsilon^e,$$

(5.2a)

where $\mu_0$ is the initial rigidity of the undamaged solid, $0 \leq \alpha \leq 1$ is the damage state
variable and $\mu_0(1-\alpha)$ is the effective elastic modulus of the damaged solid.
Analysis of stress-strain and acoustic emission data of fracturing experiments [Hamiel et al., 2004] indicates that the inelastic strain rate in a gradual distributed damage process (e.g., aftershocks), not associated with large localized system-size instabilities (e.g., mainshocks), can be described by

\[ \dot{\varepsilon}^i = C_v \dot{\alpha} \sigma, \quad (5.2b) \]

where \( C_v \) is a material constant [Hamiel et al., 2004] and the product \( C_v \dot{\alpha} \) represents the inverse of an effective damage-related viscosity in the process of damage increase. The effective viscosity increases with decreasing damage rate (e.g., aftershock rate). When the damage rate declines to zero, the effective viscosity is infinite and the system transfers to a healing process. Therefore, Eq. 5.2b is a constitutive relation only for the ‘seismic’ period (degradation process), but not for ‘interseismic’ period, where healing process might play an important role.

Over relatively short time intervals after the occurrence of mainshocks, the total strain in a volume around the fault can be assumed to be approximately constant [Ben-Zion and Lyakhovsky, 2006] and uncoupled with the ductile layer. In this case, the rate of elastic strain relaxation in the seismogenic zone is equal to the rate of increasing inelastic strain in the volume containing the aftershocks. From Eq. 5.1b with zero ductile component in the seismogenic zone, this can be written as \( \dot{\varepsilon}^e = -\dot{\varepsilon}^i \), where the overdots denote time derivatives.

### 5.2.2 Omori-type decay for damage-related postseismic deformation rate

As described above, we apply a constant strain boundary condition, and obtain the inelastic strain rate,

\[ \dot{\varepsilon}^i = 2R \dot{\alpha} (1 - \alpha) \varepsilon^e, \quad (5.3) \]

with \( R = \mu_0 C_v [Ben-Zion and Lyakhovsky, 2006] \). The elastic strain rate is

\[ \dot{\varepsilon}^e = -2R \dot{\alpha} (1 - \alpha) \varepsilon^e. \quad (5.4) \]

The integration of Eq. 5.4 provides elastic strain

\[ \varepsilon^e = \varepsilon_0 A e^{R(1-\alpha)t}, \quad (5.5) \]
where \( A = e^{-R(1-\alpha_0)^2} \), with \( \alpha_0 \) and \( \varepsilon_0 \) of damage state and elastic strain at \( t = 0 \).

Replacing \( \varepsilon^i \) in Eq. 5.3 by Eq. 5.5, the inelastic strain rate is,

\[
\dot{\varepsilon}^i = 2R\varepsilon_0 A(1-\alpha)\dot{\varepsilon}e^{R(1-\alpha)^2}.
\]  

(5.6)

The damage state variable \( \alpha \), which represents the local microcrack density in rock experiment, should rely on the aftershock seismicity during the postseismic time interval. We assume a linear relationship between \( \alpha \) and seismic moment release of the aftershocks,

\[
\alpha = \alpha_0 + \varphi \sum_{i=1}^{N} M_i,
\]

(5.7)

where \( M \) is magnitude, \( N \) is the number of the aftershocks and \( \varphi \) is a constant. Assuming that the average seismic moment release is a constant regional value, we have \( \alpha = \alpha_0 + \varphi N \) with \( \varphi = \varphi'(M) \). This linear relationship is consistent with that of Ben-Zion and Lyakhovsky [2006]. Replacing \( \alpha \) in Eq. 5.6, we have the damage-related inelastic strain relying on aftershock seismicity,

\[
\dot{\varepsilon}^i = 2R\varepsilon_0 (1-\alpha_0 - \varphi N)\varphi\dot{N}e^{-2(1-\alpha_0)R\varphi N}(R\varphi N)^2.
\]

(5.8)

The time-dependency of inelastic strain in Eq. 5.8 can be obtained by applying the classical Omori-Utsu law of the aftershock number [Utsu et al., 1995],

\[
\dot{N}(t) = \frac{K}{(c+t)^p},
\]

(5.9)

where \( K, c \) and \( p \) are constants. The total number of aftershocks is

\[
N(t) = K \cdot g(t)
\]

with

\[
g(t) = \begin{cases} 
\ln(1+t/c), & p = 1 \\
(c + t)^{1-p} - c^{1-p}, & p \neq 1
\end{cases}
\]

(5.10)

When \( \varphi N << 1-\alpha_0 < 1 \), which is appropriate to the short time period of months following a large earthquake, we can neglect the quadratic term \( (\varphi N)^2 \) in the exponential function and only consider the linear dependence on \( \varphi N \),

\[
\dot{\varepsilon}^i \cong 2R\varepsilon_0 (1-\alpha_0 - \varphi N)\varphi\dot{N}e^{-2(1-\alpha_0)R\varphi N}.
\]

(5.11)

Defining
\[ m = 2R(1 - \alpha_0)\varphi K, \quad (5.12) \]

We have for \( p=1 \),

\[ \dot{\varepsilon}^i = m\varepsilon_0 e^m \left[ 1 - \beta \cdot g(t) \right] \frac{1}{(t + c)^{m+1}}, \quad (5.13) \]

\[ \varepsilon^i = \varepsilon_0 \left[ 1 - \frac{\beta}{m} \right] - \varepsilon_0 \left[ 1 - \frac{\beta}{m} - \beta \cdot g(t) \right] \left[ 1 + \frac{t}{c} \right]^{-m}. \quad (5.14) \]

with \( \beta = \frac{\varphi K}{1 - \alpha_0} \ll 1 \). Equation 5.14 provides that \( \dot{\varepsilon}^i = 0 \) for \( t=0 \). Notice that when \( \beta \) is a small value, the time evolution of inelastic strain rate in Eq. 5.13 is dominated by the last term, which provides a generalized Omori-type decay.

Figure 5-1 displays the normalized inelastic strain rate described by Eq. 5.8 (without approximation) for different \( m \) (or \( R \)) values, assuming constants \( \alpha_0=0.25, \varphi=10^{-4} \) [Ben-Zion and Lyakhovsky, 2006], \( K=133 \), and \( c=0.003 \) days [Enescu et al., 2007]. The three plots correspond to the cases of \( p=1, 0.5 \) and \( 1.5 \), respectively. As seen, the \( m \) (or \( R \)) parameter determines the decay behavior of the inelastic strain. Small \( m \) (or \( R \)) values produce a slow decay of inelastic strain, while large \( m \) (or \( R \)) values indicate a fast decay behavior. This is consistent with the aftershock simulations in Ben-Zion and Lyakhovsky [2006]. They presented that small (vs. large) \( R \) values, suggesting a more brittle (vs. viscous) material property, produce long (vs. short) aftershock sequences with slow (vs. fast) decay. For \( m \approx 0 \), the time evolution of inelastic strain is close to the Omori-Utsu aftershock relation. Figure 5-1 also indicates that the damage-related inelastic strain should decay faster than the seismicity. If we denote with \( p_d \) and \( p_s \) the exponents in the Omori-Utsu law for the damage-related inelastic strain and the aftershocks, respectively, the forgoing results imply that \( p_d > p_s \). This conclusion can also be drawn directly from Eq. 5.13 for the special case of \( p=1 \).

If the inelastic strain rate obeys Eq. 5.8, the damage-related postseismic deformation \( D(t) \) that is proportional to the inelastic strain (i.e., \( D(t) \sim \dot{\varepsilon}^i \)) [Montési, 2004], is determined by Eq. 5.8 as well. The scaling parameter between the damage-related postseismic deformation and the inelastic strain can be collectively included in the linear constant in Eq. 5.8 or in the Omori-Utsu law. In the following sections we investigate the decay behaviors of the aftershocks and postseismic deformation following the 1999 İzmit earthquake and analyze the contribution of aftershocks-related deformation in order to separate the brittle and non-brittle components among the entire postseismic GPS measurements.
5.3 Analysis of observed postseismic deformation following the İzmit earthquake

5.3.1 Previous studies on postseismic relaxation

The 1999 August 17, M7.4 İzmit earthquake (40.76°N, 29.97°E) was a devastating event that occurred on the North Anatolian Fault Zone (NAFZ) and caused many casualties. It was followed by the 12 November 1999, M7.2 Düzce earthquake after 87 days. The İzmit earthquake ruptured over 150 km long section of the NAFZ, from the Sea of Marmara in the west to Düzce in the east. Because the Marmara Sea region was identified as a seismic gap likely to generate large earthquakes [Toksöz et al., 1979], GPS monitoring was active before the 1999 event, and was intensified after the İzmit mainshock to track early postseismic deformation. This allowed capturing the time-varying postseismic deformation field by the GPS measurements (Fig. 5-2).

Some studies inverted for postseismic slip on the extended coseismic fault plane based on the GPS [Bürgmann et al., 2002; Çakir et al., 2003; Wang et al., 2009; Ergintav et al.,
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2009] and InSAR (interferometric synthetic aperture radar) [Çakir et al., 2003] measurements following the İzmit earthquake. Except for some details in the inverted slip patterns because of variant study time periods and likely different inversion methods, these studies generally show similar afterslip concentrations. The results commonly indicate that the postseismic relaxation focused at the shallow depth (in the seismogenic zone) in the first days or 1-3 months. Çakir et al. [2003] also showed that the afterslip in the first month is overlapping or near the high seismic area according to the relocated aftershock catalog.

The İzmit postseismic displacements have also been modeled by variant relaxation processes, frictional-afterslip [Hearn et al., 2002, 2009], non-linear viscous flow in the deep ductile creep zone [Montési, 2004] and 3-dimensionally distributed viscoelastic relaxation in the mantle [Hetland, 2006]. Very detailed studies are given by Hearn et al. [2002, 2009]. These authors modeled the postseismic deformation in the first several months based on rate-strengthening friction, in response to coseismic stress change, along down-dip extension of the coseismic fault plane above 2 km depth and below 10 km depth. Their obtained afterslip model provides more significant creep at the shallow depth than the deep one, indicating the strong early postseismic displacements near the rupture plane. Meanwhile, they reported that the fictional-afterslip model is not able to produce enough observed postseismic deformation in the vicinity of the İzmit rupture and can only explain 63% of the observed early postseismic displacements. Generally speaking, the physical models that focus on frictional-afterslip/thermally-activated deep postseismic relaxation can not fully explain the early afterslip detected in the seismogenic zone, especially overlapping the aftershock area. Thus, it is likely that a larger portion of the early İzmit postseismic displacements is related to the aftershocks in the fault zone and induced by damage-related inelastic deformation.

In addition, Ergintav et al. [2009] found that three logarithmic terms with characteristic decay time of 1, 150 and 3500 days are necessary to fit the displacement time series of 7 years following the İzmit event. The term with the decay time of 1 day is characterized by the strong early fault-parallel motion originated from the shallow depth; while the other two terms with longer decay constants describe either localized deep afterslip or viscoelastic relaxation, which produces more broadly distributed strain with a symmetric double-couple pattern. Their detected early relaxation phase with short decay time of 1 day is likely to be related to the aftershocks-related inelastic deformation.

Therefore, we apply our model to the postseismic displacements observed after the İzmit earthquake and investigate the relative contributions of the aftershocks-related component and the non-brittle ductile components of the observed geodetic field at different locations. We first model the time evolution of the observed aftershock rates and postseismic geodetic deformation, respectively, in order to compare their decay behaviors.
Using the theoretical results of section 5.2, we then try to separate the three deformation components described in Eq. 5.1, which contribute to the postseismic deformation observed on the surface.

**Figure 5-2.** Locations of the GPS stations (black triangles) and the spatial distribution of observed aftershocks (gray circles, $M \geq 2.5$) in the first 87 days following the İzmit earthquake. Aftershocks with magnitude larger than 4.0 are highlighted by black circles. White stars display the epicenters of the İzmit (40.76°N, 29.97°E) and Düzce (40.82°N, 31.20°E) earthquakes. Black solid lines show the rupture trace of the İzmit earthquake [Wright et al., 2001]. The gray dotted lines mark a simplified location of the North Anatolian fault [Lorenzo-Martín et al., 2006a]. The aftershocks catalogue is from the Kandilli Observatory and the Earthquake Research Institute (Istanbul, Turkey).

### 5.3.2 Omori-Utsu decay of the aftershocks

We mainly use the aftershock catalogue from the Kandilli Observatory and Earthquake Research Institute (Istanbul, Turkey). The magnitudes of the events were selected keeping, by order of availability, $M_w$, $M_L$, $M_S$, mb and $M_D$ [Daniel et al., 2006]. The catalogue probably misses numerous aftershocks in the first days after the mainshock, leading to inaccurate estimate of the onset time of the power law decay in the Omori-Utsu law [e.g., Lolli and Gasperini, 2006]. This catalogue also has a 1-day gap between 20 August (day 3 from the İzmit earthquake) and 21 August 1999. We fill the gap using the catalogue managed by the Earth Sciences Research Institute (ESRI) of the Marmara Research Center, TÜBİTAK. The ESRI deployed a very dense network of seismic stations for monitoring the aftershock activity following the İzmit earthquake [Aktar et al., 2004]. Based on the validity of the Gutenberg-Richter statistics for the observed moderate events (Fig. 5-3), we consider the catalogue complete for earthquakes with magnitude above $M_c = 2.5$. 

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Figure 5-3. Frequency-magnitude statistics of the employed aftershocks. The catalogue, from the Kandilli Observatory and Earthquake Research Institute, has a 1-day gap between August 20 and 21, 1999. The gap is filled up using the catalogue managed by the Earth Sciences Research Institute (ESRI) of the Marmara Research Center, TÜBİTAK.

The aftershocks of the İzmit earthquake extended over an area of 40 by 170 km (Fig. 5-2). Our investigations focus on the fault zone (approximately less than 10 km distance perpendicular to the fault according to the aftershock distribution), where the aftershocks densely occurred. We analyze the time period of the first 87 days after the İzmit mainshock, which is before the Düzce earthquake, using the modified Omori-Ustu law (Fig. 5-4). The obtained parameters of the Omori-Ustu relation for the aftershocks around the rupture zone are $K=86.95\pm0.39$, $c_s=0.74\pm0.07$ days and $p_s=0.87\pm0.05$. As noted in section 5.2, the subscript $s$ is used to distinguish the parameter estimates based on the aftershocks from those derived in the next section based on the postseismic displacements. Because of the incompleteness of the catalogue immediately after the large mainshock, the estimated $c_s$ value might be inaccurate and larger than the true value [e.g., Peng et al., 2006; Enescu et al., 2007].
5.3.3 Time-dependent postseismic deformation

We use continuous GPS measurements recorded at seven stations located mostly near the fault (black triangles in Fig. 5-2). Two GPS stations (TUBI, DUMT), which belong to MAGNET network, were operating before the İzmit earthquake. Four additional continuously recording GPS stations were installed in the 2 days following the mainshock. Stations UCGT and BEST are located less than 15 km north of the fault, and stations HAMT and MURT are approximately 5 km south of the İzmit rupture. In addition to the six stations near the fault, continuous GPS data at station KANT ~35 km from the fault is also included in the study.

The İzmit earthquake is a continental strike-slip event with dominant postseismic motion in horizontal directions. Instead of treating the E-W and N-S GPS measurements separately, we utilize the principal component of the motion defined as

\[ D = \sqrt{D_E^2 + D_N^2} \]

with an azimuth of \[ \beta = \arctan(D_E / D_N) \]. It has been shown that the azimuth of the principal displacement is relatively stable in the short time period of months following the Izmit earthquake, but can be different from site to site [Bürgmann et al., 2002]. The corresponding measurement uncertainty is calculated by \[ \sqrt{sd_E^2 + sd_N^2} \], where sd denotes the standard deviation of a given component. The postseismic deformations are given by \[ D_E = \tilde{D}_E + D_{E0} \] and \[ D_N = \tilde{D}_N + D_{N0} \], where \( \tilde{D}_E \) and \( \tilde{D}_N \) represent the recorded displacements measurements starting from some days after the mainshock, and \( D_{E0} \) and \( D_{N0} \) represent the postseismic displacements that occurred immediately after the mainshock and have not been captured by the GPS measurements (e.g., HAMT station started to run...
from about 1 day after the İzmit mainshock). With these definitions the principal
displacement is given by
\[
D = \sqrt{\left(\tilde{D}_E + D_{E0}\right)^2 + \left(\tilde{D}_N + D_{N0}\right)^2}
\]
where \( \tilde{D}_E \) and \( \tilde{D}_N \) are known.
The unknowns \( D_{E0} \) and \( D_{N0} \) values are determined by assuming that \( D_E \) and \( D_N \) are 0
at \( t = 0 \) (the time of the mainshock), and that the postseismic displacement rate decays
following the Omori-Utsu law. Figure 5-5 presents two examples showing the observed
displacements (\( \tilde{D}_E \) and \( \tilde{D}_N \)) and corrected values (\( D_E \) and \( D_N \)). We investigate the time
evolutions of the postseismic displacements using the corrected data.

As mentioned, we consider for the analysis only the measurements in the time period of
~3 months after the İzmit mainshock (before the Düzce earthquake). At this short time scale,
the tectonic loading effect is negligible compared with the significant postseismic relaxation
[Wang et al., 2009]. For a simple comparison of the decay behaviors of aftershocks and
aseismic postseismic deformation, we may assume that the later has the same \( c \) value as the
former (i.e. \( c_d = c_s \)), and use the exponent \( p_d \) in the Omori-Utsu law as a free parameter for
estimation. In this case, the two postseismic activities can be evaluated only by their decay
exponent values. Since \( c_s \) is probably overestimated due to the incompleteness of the
aftershock catalogue immediately after the mainshock, we confine in the analysis \( c_d \) to be in
the interval \((0, c_s]\).

The inversion results are presented in Table 5-1 and Fig. 5-6. To evaluate the
confidence interval of the estimates, accounting for both the measurement uncertainty and
inversion method, we additionally estimate the parameters using 100 simulated data sets
with means of the real displacement values and standard deviations of 1-standard deviations
of the GPS measurements. The confidence intervals of the estimates are obtained through
25% and 75% quartiles of the 100 estimates. The results in Table 5-1 (see also Fig. 5-6)
show that with confined \( c_d \) in \((0, c_s]\), the estimates of \( p_d \) and \( K_d \) base on the real
displacements are in the range of the estimates based on the synthetic data with the
consideration of measurement uncertainty. It indicates that the estimations are stable and
reliable. The estimated \( p_d \) values with \( p_d < p_s \) for all of the seven GPS stations indicate that
the postseismic displacements decayed slower than the aftershock seismicity after the İzmit
earthquake. For convenience, we refer to the derived results (Fig. 5-6 and Table 5-1) of the
Omori-type decay of the postseismic displacements as \( D_{net}(t) \). In the following sections,
we investigate the contributions to the observed postseismic displacements from elastic
strain produced by aftershocks, aftershocks-related inelastic relaxation and ductile
relaxation. Additionally, we consider a simplified interseismic model for the study region to
show that the tectonic motion has small effect over the studied time domain.
Figure 5-5. Two examples showing the observed ($\bar{D}_E$ and $\bar{D}_N$, dots with error bars) and corrected ($D_E$ and $D_N$, circles) postseismic displacements in the first 87 days following the İzmit earthquake. The corrected displacements are defined as $D_E = \bar{D}_E + D_{E0}$ and $D_N = \bar{D}_N + D_{N0}$, where $D_{E0}$ and $D_{N0}$ represent the postseismic displacements that occurred immediately after the mainshock and not captured by the GPS measurements. The solid and dashed curves show, respectively, the fitting results for the observed and corrected displacements based on the Omori-Utsu relation for postseismic deformation.
Figure 5-6. Principle geodetic displacements (dots with error bars) observed at seven GPS sites, and the modeling results based on the Omori-Utsu relation for postseismic deformation (white curves). The modeling assumes that the $c_d$ parameter is confined in the interval $(0, c_s]$ associated with the aftershock seismicity. See text for more explanations.

Table 5-1. Estimated parameter values ($K_d$, $c_d$, $p_d$) for the Omori-Utsu type inelastic deformation with $c_d$ in $(0, c_s]$.

<table>
<thead>
<tr>
<th>Station</th>
<th>$K_d$</th>
<th>$c_d$(day)</th>
<th>$p_d$</th>
<th>RMS(mm)</th>
<th>Dist (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>HAMT</td>
<td>3.05</td>
<td>0.81</td>
<td>0.58</td>
<td>2.68</td>
<td>5.5</td>
</tr>
<tr>
<td></td>
<td>[2.88, 3.62]</td>
<td>[0.81, 0.81]</td>
<td>[0.55, 0.63]</td>
<td></td>
<td></td>
</tr>
<tr>
<td>MURT</td>
<td>4.69</td>
<td>0.81</td>
<td>0.60</td>
<td>3.93</td>
<td>5.6</td>
</tr>
<tr>
<td></td>
<td>[4.29, 4.92]</td>
<td>[0.81, 0.81]</td>
<td>[0.56, 0.62]</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TUBI</td>
<td>1.89</td>
<td>0.01</td>
<td>0.70</td>
<td>3.26</td>
<td>7.4</td>
</tr>
<tr>
<td></td>
<td>[1.67, 2.81]</td>
<td>[0.01, 0.06]</td>
<td>[0.65, 0.80]</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
5.3.4 Different contributions among the whole postseismic displacement

5.3.4.1 Contribution from tectonic motion

Here we investigate the contribution from the tectonic motion. Consistent with the employed GPS data, we firstly obtain the secular velocities at the seven GPS sites (Fig. 5-7) relative to the fixed Eurasian plate according to Wang et al. [2009]. Secondly, we subtract the tectonic displacements from the observed postseismic measurements, and fit the corrected data (denoted $D_t(t)$) using the Omori-Utsu law. Thirdly, the contribution of the tectonic motion is determined by $(1 - [D_t(t)/D_\infty(t)]) \times 100\%$. The results (dashed-dotted curves in Fig. 5-8) indicate that the tectonic motion has negligible effect on the total postseismic displacements in the first 87 days following the İzmit earthquake.

![Figure 5-7](image)

**Figure 5-7.** E-W component (a) and N-S component (b) of the tectonic motion at the seven GPS sites based on Wang et al. [2009]. Because the N-S displacements at the seven GPS sites are small and close to each other, only two examples are shown in (b).
5.3.4.2 Contribution from elastic deformation induced by aftershocks

To investigate the elastic relaxation process produced by the aftershocks, we consider the large events with $M \geq 4.0$ indicated by the black circles in Fig. 5-2. We calculate the surface displacements produced by these aftershocks and examine their contributions to the measured postseismic displacements. According to the Centroid Moment Tensor (CMT) solutions of the large aftershocks, most of the aftershocks between $29.3^\circ$E and $30.5^\circ$E have similar focal mechanism as the İzmit mainshock. We thus treat the $M \geq 4.0$ aftershocks as strike-slip events, similar to the mainshock, and calculate the surface displacements at the seven GPS sites produced by these aftershocks. The calculations are done with the Okada’s code, using slip values based on the empirical relation with the magnitudes [Wells and Coppersmith, 1994].
The calculated displacements at the seven GPS sites and used aftershocks are shown in Fig. 5-9. The results indicate that the aftershock seismicity produced influential surface displacements at some GPS sites, e.g., HAMT, where the E-W displacement reaches ~7 mm in the first 87 days. After correcting for the elastic effect of the aftershocks, the postseismic displacement at HAMT site is decreased by ~20%. The contributions of the aftershocks at the other stations, \(1 - \left[ \frac{D_a(t)}{D_{noa}(t)} \right] \times 100\%\) (shown by dashed curves in Fig. 5-8), are generally small (less than 5%), as has been documented [Reilinger et al., 2000]. Here \(D_a(t)\) denotes the fit of the Omori-Utsu law to the data which are corrected from the elastic effect of the aftershocks.

![Figure 5-9: M-t plot (M≥4.0) and the calculated E-W (a) and N-S (b) displacements associated with elastic relaxation produced by the aftershocks at the seven GPS sites.](image)

### 5.3.4.3 Contributions from damage-related inelastic relaxation

According to the theoretical results of section 5.2 and Fig. 5-1, the damage-related aseismic deformation decays faster than the aftershock seismicity, and the decay rate for aftershock seismicity provides an upper limit for the associated inelastic relaxation rate. Therefore, the maximum damage-related aseismic deformation can be determined by setting \(p_d = p_s\) and \(c_d = c_s\), leading to

\[
D_{dam}(t) = \frac{K_d}{1 - p_s^d} \left[ (c_s + t)^{-p_s} - c_s^{-p_s} \right].
\]  

We adopt the values of the \(K_d\) parameter from the numbers summarized in Table 5-1, because \(K_d\) quantifies the magnitude of the measured postseismic displacements at different sites and is insensitive to \(c\) and \(p\) parameters in the Omori-Utsu law. The percentage of the damage-related deformation among the total postseismic deformation, calculated by
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$D_{dam}(t)/D_{tot}(t) \times 100\%$, is shown by the solid curves of Fig. 5-8. The results indicate that the damage-related inelastic relaxation relative to the total postseismic deformation is different from site to site. However, the aftershock related inelastic relaxation contributes generally up to 50% of the total geodetic deformation in the first 87 days.

5.3.4.4 Contribution from ductile relaxation

The stress perturbations generated by large earthquakes are also likely to produce some non-brittle stable inelastic deformation. Such deformation, represented here collectively by the ductile strain term of Eqs. 5.1a and 5.1b, may have contributions from distributed deformation below the seismogenic zone [e.g., Ben-Zion et al., 1993; Deng et al., 1998; Pollitz et al., 2000; Wang et al., 2009], localized deformation below the brittle section of the fault [e.g., Montési, 2004; Perfettini and Avouac, 2004] and poroelastic relaxation [e.g., Peltzer et al., 1998; Jónsson et al., 2003; Fialko, 2004]. The results of the previous sections can now be used to estimate the amplitude of the combined ductile component of the observed postseismic deformation (i.e., deformation not associated with the damage-aftershocks process). Figure 5-10 shows the components of the postseismic measurements that may be attributed to non-brittle ductile relaxation at the different sites. The results indicate that ductile relaxation was responsible for less than 50% of the postseismic deformation in the first 87 days.
Figure 5-10. The percentage of the residual displacement (excluding aftershocks-related elastic/inelastic deformation and tectonic motion) among the observed postseismic displacements (solid curves). The dashed curves show the percentage of the estimated distributed ductile relaxation in the lower crust and upper mantle based on the rheological model of Wang et al. [2009].

5.4. Discussion

5.4.1 Damage-related postseismic relaxation and frictional afterslip

Both aseismic deformation and aftershocks are significant markers following large earthquakes. Their relationship can give us a hint about the role of the two different mechanisms, frictional-afterslip and damage-related postseismic relaxation.

Frictional afterslip within the framework of rate-and-state dependent friction models (RS) is an important candidate to explain the postseismic deformation [Marone et al., 1991]. It models the stable sliding, due to coseismically produced stress, on an existing surface.
According to the RS law, an earthquake nucleates in the unstable rate-weakening field, while the stable sliding occurs in the rate-strengthening field. The transition between rate-weakening and rate-strengthening is largely temperature/depth dependent. For the continental crust, the transition occurs at 15-20 km depth, corresponding to the temperature of ~300°C. That means the rate-strengthening occurs mainly below the seismogenic zone. The zone near the surface (less than 3-4 km depth) is also likely governed by rate-strengthening because of poorly consolidated material [Marone et al., 1991]. Thus, the spatial distributions of aftershock seismicity and frictional-afterslip should be negatively correlated in the RS framework.

Damage rheology developed from rock experiment in laboratory [Lyakhovsky et al., 1997] has been applied to studies on fault-zone properties [e.g., Finzi et al., 2009], spatio-temporal patterns of earthquakes [e.g., Ben-Zion and Lyakhovsky, 2002] and the seismic cycle [e.g., Ben-Zion and Lyakhovsky, 2006; Lyakhovsky et al., 2005]. A theoretical investigation [Lyakhovsky et al., 2005] has shown that the RS friction and damage rheology can be unified to model the healing period (i.e. decreasing damage) and weakening period (i.e. increasing damage). However, different from RS friction, damage-related deformation occurs in a volume around the main fault plane rather than only on the plane. The elastic properties in the volume are spatially heterogeneous and related to the time-varying damage state that is characterized by the local density of microcracks. Therefore, the damage-related postseismic relaxation occurs in the volume around the fault, where aftershocks densely occur.

It is difficult to distinguish the damage-related relaxation and frictional-afterslip based only on inversions of geodetic measurements, as been recognized for the hard distinction of viscoelastic relaxation in the ductile layer from deep afterslip [Thatcher, 1983; Savage, 1990]. The spatial distribution of the relocated aftershocks might provide useful information to distinguish the two postseismic mechanisms. As mentioned above, the high seismic area corresponds to rate-weakening regime. Therefore, the inverted-afterslip overlapping the aftershock area suggests the damage-related relaxation, instead of frictional-afterslip (due to rate-strengthening). For the İzmit earthquake, thousands of aftershocks occurred densely around the fault plane after the İzmit earthquake. The inverted-afterslip according to InSAR measurements in the first month after the big event locates in the seismogenic zone, especially overlapping or near the area with high seismicity [Çakir et al., 2003]. Therefore, it is reasonable to believe that a large portion of the measured postseismic displacements is due to aftershocks-related relaxation.
5.4.2 Postseismic relaxation in the deep ductile zone following the İZmit earthquake

Using basic stress-strain relation in the framework of damage rheology [Ben-Zion and Lyakhovsky, 2006] and assuming a linear connection between seismic moment release and damage variable during uniform deformation over relatively short time intervals, we obtain a relation between the decay rates of aftershocks and the related postseismic deformation. This enables us to attribute the remainder (non aftershocks-induced) portion of the observed postseismic deformation to non-brittle processes represented collectively by the ductile strain of Eq. 5.1.

Using both short-term (in one year after the mainshock) postseismic displacements in the far-field (>30 km from the fault) following the İZmit event and long-term (3-5 years after the mainshock) displacements, which might be dominantly produced by the deep ductile relaxation, Wang et al. [2009] built a rheological model (E-SLS-M) utilizing Standard Linear Solid element and Maxwell elements for the inelastic lower crust and upper mantle, respectively. They found that over this time scale the lower crust has viscosity of $2 \times 10^{18}$ Pa s and relaxation strength of $2/3$, and the upper mantle has viscosity of $7 \times 10^{19}$ Pa s. It has been shown that afterslip inverted from the postseismic data after correcting for the relaxation component predicted from the E-SLS-M rheological model concentrates mainly in the elastic layer. Thus, the E-SLS-M model can describe the relaxation in the deep ductile zone that occurred during the postseismic interval.

Therefore, we adopt here the E-SLS-M model [Wang et al., 2009] to quantify the remainder non-brittle relaxation component. We calculate the contribution of the postseismic relaxation in the deep ductile zone to the observed postseismic deformation at the seven GPS sites. The results shown by the dashed curves in Fig. 5-10 indicate that the residual postseismic displacements, left after subtracting the contributions from the aftershock-induced elastic/inelastic deformation and the tectonic motion, can be attributed at most GPS sites to the relaxation described by the E-SLS-M model. Only at sites MURT, BEST and KANT, the observed deformation cannot be completely explained by a combination of aftershock-induced and ductile effects, which might indicate some local rheological heterogeneity.

In the eastern section of sites MURT and BEST, tomography studies show comparatively higher $P$ wave velocities and higher $Q$ values suggesting colder and more brittle material [Nakamura et al., 2002; Koulakov et al., 2009]. Based on the results of Ben-Zion and Lyakhovsky [2006], cold brittle material is expected to produce long Omori-type aftershock sequence with relatively low $p$ value and high event productivity.
Therefore, when we applied the constraint $p_e = p_s$ (in section 5.3.4.3 a homogenous $p_s$ value has been determined for the entire study region) for the geodetic deformation, we likely underestimated the aftershock related aseismic deformation in this area. On the other hand, the deformation summation due to aftershock related inelastic relaxation and distributed ductile relaxation overestimates the postseismic displacement at site KANT. We note that KANT is far from the fault damage zone and the surrounding aftershock seismicity is rather low. Thus, our simple procedure overestimated the damage-related deformation at this site (cf. Fig. 5-9) and produced modeled postseismic displacement that is higher than the observed displacement (cf. Fig. 5-10).

5.5. Conclusion

We analyzed aftershocks-related relaxation in the framework of damage rheology, which can explain the inverted afterslip in the seismogenic zone, particularly in the zone overlapping the aftershock seismicity. Assuming a linear connection between seismic moment release of aftershocks and damage variable during uniform deformation over relatively short time intervals, we derived that the damage-related inelastic deformation rate follows a generalized Omori-Utsu type decay, with the standard Omori-Utsu law as a limit case (section 5.2). The obtained theoretical results can be used to separate between brittle and ductile contributions to postseismic deformation, which are dominated in the shallower seismogenic zone and the deeper ductile region, respectively.

We applied the obtained theoretical relations to the observed GPS displacements at the sites around the rupture zone of the 1999 İzmit earthquake. The results can be summarized as follows (section 5.3). (1) The entire postseismic displacement measured on the surface decayed slower than the aftershock seismicity in the first 87 days after the İzmit earthquake. (2) The aftershocks related aseismic relaxation can maximally account for up to 50% of the total deformation. (3) The remainder postseismic displacement can be mainly explained by the relaxation at the deep depth based on the rheological model and parameters derived by Wang et al. [2009]. (4) The contribution from aftershocks-induced elastic relaxation is generally less than 10% of the observed postseismic deformation, but it can be influential at some individual sites. (5) The tectonic motion is negligible in the examined time interval. It is important to examine in future works whether the theoretical results and partitioning of observed surface geodetic deformation obtained in this work apply in a similar way to other cases of postseismic deformation.
Chapter 6  Summary and conclusions

Aseismic deformation is frequently observed following large earthquakes, which might be related to processes of afterslip, viscoelastic relaxation and/or poroelastic rebound. However, the relative importance of the different processes as well as their relation to aftershocks is still an open scientific issue. Because of its localized nature, postseismic deformation has great potential to shed light into the processes underlying earthquake interactions and to deduce local rheological parameters of inelastic layers. For this goal, I analyzed the postseismic deformation following two large earthquakes, the 1999 M7.4 İzmit earthquake in Turkey and the 1995 M8.1 Antofagasta earthquake in Chile in this Ph. D thesis.

Making use of the GPS data measured in several years following the İzmit and Antofagasta earthquake, I used an inversion method and forward modeling to investigate two mechanisms: (1) afterslip on and below the coseismic rupture plane; (2) viscoelastic stress relaxation in the ductile zone. The results show that, in the case of both big events, the early postseismic deformation was dominated by afterslip along the coseismic fault plane. The inverted slip pattern shows a monotonous distribution and the slip amount decays with time. Because of the faster decay behavior of afterslip, viscoelastic relaxation process increasingly plays an important role in the later time period (see Chapter 3). For the İzmit earthquake, viscoelastic relaxation became important after ~100 days following the big event, while for the Antofagasta earthquake, viscoelastic relaxation became detectable about 2 years after the mainshock and contributed to a small portion (15%) of the total measured postseismic deformation. This different scale of afterslip-dominance might be related to the different sizes of the mainshocks (M7.4 for the İzmit earthquake and M8.1 for the Antofagasta earthquake), or different tectonic settings.

Both case studies provided that the Maxwell viscosity in the upper mantle is on a scale of $10^{19}$ Pa·s. The estimates are basically consistent with the results in previous studies. However, different postseismic behaviors have also been found for the two earthquakes, a typical continental strike-slip event and a thrust event in subduction zone, respectively. Viscoelastic relaxation following the İzmit earthquake produces similar kinematic motions as the mainshock (cf. Fig. 3-7), and explains the data measured shortly after the big event better in the far field than in the near field. It suggests that the deep deformation can be attributed to viscoelastic relaxation. In comparison, viscous flow following the Antofagasta earthquake produces more significant motion in the near field, which, different from that in the far field, is opposite to the coseismic displacement. Viscoelastic relaxation can mainly
explain the deformation near the Moho depth (~45 km). The inversion results derived from the postseismic displacement following the Antofagasta earthquake show apparent deep afterslip (at ~70 km depth) below the Moho depth level. It might be related to chemical reaction (e.g. serpentinization) in the subduction zone that leads to low friction on the interface.

In recent years, the Finite Element Method (FEM) has been more and more used for interseismic, coseismic and postseismic modeling. Because this method can take lateral heterogeneities into account, FEM models are expected to be closer to reality than the layered halfspace models. However, a study in Himalaya region [Vergne et al., 2001] has shown that dislocations in an elastic half-space model is able to well describe surface displacement and stress field, close to the results based on the FEM model, although the latter took other additional geological information, such as local thermal structure and rock material in the lithosphere, into account. Another study on postseismic deformation following the 1946 Nankaido earthquake in Japan [Yoshioka and Suzuki, 1999] indicated that when a reasonable depth of continental elastic layer in the subduction zone and viscosity in the inelastic layer are applied, the layered half-space model predicts a similar displacement field as the 3D FEM model, which considers additionally the oceanic subduction slab. For the 1960 M9.5 Valdivia earthquake, Lorenzo-Martín et al [2006] presented that the layered 2D model can explain the GPS data as well at the 3D FEM model [Khazaradze and Klotz, 2003]. Therefore, the analysis based on layered halfspace model in my thesis should provide useful physical information in the two study regions.

One-dimensional physical models describing postseismic creep/slip that occurs on or downdip of the coseismic fault plane have been formulated based on friction law or creep rheology [Marone et al., 1991; Montési, 2004; Perfettini and Avouac, 2004]. These models are helpful for our understanding the postseismic relaxation process, especially deep afterslip. However, the relationship between postseismic aseismic deformation and aftershocks, which occur simultaneously and follow a similar Omori-type decay with time, has less been addressed. In this thesis, I built physical connections between the two postseismic activities in the framework of damage rheology. This provides a candidate for interpreting the early postseismic relaxation detected (e.g., according to inversions constrained by the geodetic measurements) in the seismogenic zone, specially that overlapping the aftershock area, which is hard to be explained by the previous mechanisms, e.g., rate-and-state friction or viscous flow (see details in 5.4.1). Utilizing basic relations between local brittle failures and gradual inelastic strain in damage rheology and assuming the Omori-Utsu decay rate for aftershocks, I find that the temporal decay of the damage-related postseismic relaxation follows a generalized power law relation with the standard Omori-Utsu law as a limit case. The theoretical results provide a new approach to separate damage related deformation from the other ductile deformation. The study on the
İzmit earthquake indicates that up to 50% of the surface displacements at near-fault sites measured in the first 3 months following the big event can be attributed to aftershock-induced inelastic deformation in the seismogenic zone. The remainder postseismic deformation can generally be explained by the postseismic relaxation in the deep visous layer.

After finishing this study, some general conclusions concerning postseismic modeling can be drawn. The postseismic mechanisms (e.g. afterslip and viscoelastic relaxation) dominate different time and spatial spaces. Afterslip seems to be the most important relaxation process that releases a large portion of the total postseismic moment. To understand the entire postseismic process, however, a comprehensive investigation including different postseismic mechanisms is necessary. It has become clear that only one set of displacement measurements for a certain time is not enough to resolve afterslip and viscoelastic relaxation. More datasets collected in different time periods, especially displacement time-series recorded by continuous GPS measurement are important for understanding the time-varying postseismic mechanism and for investigating the relationship between aftershocks and postseismic aseismic deformation. In particular, model simulations can help to set-up optimally distributed GPS-networks for recognizing the postseismic mechanisms. Specifically, for thrust events in the subduction zone, a dense distribution of GPS sites in the near field, e.g. less than 250km from the source, is needed to capture deformation produced by viscoelastic relaxation process. The measurements with good spatial and temporal coverage in such areas are important to resolve afterslip and viscoelastic relaxation mechanism following a thrust event, and to build a good regional rheological model. While for strike-slip events, a good coverage in the middle/far field is important to resolve which postseismic relaxation is dominant. A better characterization of postseismic processes with its significant seismic moment release is crucial for understanding the whole seismic cycle and improving seismic hazard analysis in seismically active regions.
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