Chapter 4

A joint analysis of GPS motions and InSAR to infer the coseismic surface deformation of the Izmit, Turkey Earthquake.

4.1 Introduction

Over the past decade, space geodetic observations of surface motions have been widely used to model the (a)sismic surface deformation of the Earth (e.g. Bennett, 2003; Bos et al., 2003a; Shen et al., 1996; Shen-Tu et al., 1998, 1999; Snay et al., 1996; Ward, 1998). These models have provided an increased understanding of the complex spatial distribution of crustal deformation which is particularly useful for understanding the dynamics causing deformation and for seismic hazard analysis. For the latter a detailed analysis of the coseismic surface deformation field can provide important information on the relaxation of the long-term deformation field during an earthquake, as well as on the complexity of the earthquake source and crustal structure. Although space geodetic data have been used in studies of several major earthquakes (e.g. Landers, Hector Mine; Fialko et al., 2001; Hudnut et al., 1994; Massonnet et al., 1993), kinematic modeling of the coseismic surface deformation has been prohibited due to the absence of fault slip as a parameter in the various analysis methods available. Coseismic GPS and InSAR data have been utilized to model the slip distribution of earthquakes assuming a linear elastic behavior of the deformation field (Hudnut et al., 1994; Massonnet et al., 1993). For Fialko et al. (2001) it proved possible to estimate the surface displacement field

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from both an ascending and descending InSAR interferogram. Due to decorrelation of the InSAR images near the fault trace, the displacement field was interpreted directly without estimations of fault slip or the displacement gradient field. The recently developed kinematic inversion method of Spakman and Nyst (2002) inverts geodetic observations for the displacement gradient field and fault motion simultaneously which allows us to distinguish between coseismic slip on the fault and strain release in the blocks bounding the fault. I utilize this method to determine the kinematic coseismic surface deformation field of the 1999 Izmit, Turkey earthquake.

Figure 4.1: Tectonic setting of the Izmit earthquake plotted against topography. Black line indicates the fault trace utilized in this study. Black names between vertical lines state fault segment names, while the grey square outlines the area contained in the InSAR interferogram. The stars show the epicenter locations of the Izmit and Düzce earthquakes and the corresponding focal mechanisms are provided. For a color version of this figure see figure C.11 of appendix C.3.

On August 17, 1999 a $M_w = 7.5$ earthquake struck the Marmara region of northwestern Anatolia (Turkey). The earthquake was the most recent in a westward migrating sequence of earthquakes along the North Anatolian fault which commenced in 1939 and subsequently ruptured a 1200 km section of the fault (Ambraseys, 1970; Barka, 1992). The August 17 event occurred near the town of Izmit and caused surface rupture with dextral displacements of over 5 m (Barka et al., 2002) along a 110 km segment of the North Anatolian fault. Rupture was induced on at least 5 major fault segments (from west to east: Yalova, Gölcük, Sapanca, Sakarya and Karadere; figure
4.2 Data

4.2.1 GPS data

The GPS evidence for the coseismic displacements of the Izmit, Turkey earthquake is provided by Reilinger et al. (2000). For modeling the spatial distribution of the coseismic surface deformation field I utilize a subset of 41 stations immediately surrounding the fault trace of this event, as well as the November 12 Düzce earthquake (figure 4.2). The dataset comprises 3 continuous stations that were operating at the time of the earthquake (named sites in figure 4.2) and 38 survey-mode sites for which coseismic displacements could be determined. To obtain the coseismic displacements, the campaign data have been corrected by Reilinger et al. (2000) for elastic strain accumulation prior to the event and post-seismic afterslip up to the time of remeasurement.

4.2.2 InSAR interferograms

The InSAR analysis is performed on both the ERS 35-days coseismic interferograms available, namely an ERS1 interferogram spanning the period 12 August/16 September 1999 (orbit numbers 42229 and 42730, frame 811) and the corresponding ERS2 pair 13 August/17 September 1999 (orbit nrs. 22556 and 23057, frame 811). The in-
terferograms are shown in figure 4.3, where they have been masked for extremely low coherence values (< 0.1). For the chosen frame they have both a perpendicular baseline component of about 70 meters. With these baselines, the interferograms are in general not particularly sensitive to topography. However for such a mountainous area the topographic component of the interferometric signal is still significant. The topography was therefore estimated and extracted by using the tandem pair 12/13 August.

Although their perpendicular baseline component is approximately the same, the two interferograms show considerable differences in the fringe patterns, as already mentioned in Reilinger et al. (2000) and Feigl et al. (2002). The difference in the fringe pattern between the two differential interferograms is shown in figure 4.4. In particular, in the lower part of the ERS1 interferogram the fringes still seem to follow the topography, probably due to atmospheric effects wrapped around the relief. As for the upper part, the presence of a somewhat regular pattern of fringes might indicate an orbital error in one of the interferograms. However, I use very precise orbits (Scharroo and Visser, 1998), which in my experience never gave such a huge trend. The hypothesis that this could be a orbit error seems therefore very unlikely to me. At least part of the effect could in my opinion still be due to atmospheric effects.
Figure 4.3: InSAR interferograms: a) ERS1 interferogram spanning the period 12 August/16 September 1999 (orbit numbers 42229 and 42730, frame 811), b) ERS2 pair 13 August/17 September 1999 (orbit numbers 22556 and 23057, frame 811). Both interferograms have been masked for coherence < 0.1. The topographic component of the interferometric signal was estimated and extracted by using the tandem pair 12/13 August. For a color version of this figure see figure C.12 of appendix C.3.
4.3 GPS data inversion

In order to model the surface deformation field of the Izmit earthquake from the GPS displacements, I adopt the inversion method of Spakman and Nyst (2002). Although formulated for the analysis of relative motion data, it can as well be applied to relative displacement data. For the latter application, the method relates the relative displacements ($\Delta u_{ij}$) between stations $i$ and $j$ to the displacement gradient field ($\nabla u$; strain and rotation) and fault slip ($s$) by:

$$\Delta u_{ij} = \sum_{l=1}^{K+1} \int_{l}^{l_{ij}} \nabla u(r) \cdot dr + \sum_{k=1}^{K} \alpha_k s_k (r_{ij}^{h})$$

(4.1)

where $L_{ij}^{l}$ is an integration path between the stations $i$ and $j$, $K$ denotes the number of fault segments crossed, $r_{ij}^{h}$ is the location where the integration path $L$ crosses a fault, and $\alpha_k = \pm 1$ depends on the orientation of the fault with respect to the integration path $L_{ij}^{l}$. Equation 4.1 is exact in practice and does not involve any knowledge of crustal/fault rheology.

I parameterize fault slip by assuming constant slip on fault (sub)segments. I implement a single fault trace that approximates the mapped coseismic surface rupture. To avoid unrealistic detail the fault trace has been simplified and step-over features smaller than 2 km length have been removed (figure 4.5). The Yalova and Gölcük step-overs
remain an integral feature of the model fault trace. I have extended the fault towards the east to include the Düzcė earthquake rupture area and towards the west in the Marmara Sea. Thus the fault divides my model area in a northern and a southern domain. In total the fault consists of 11 fault segments for which I resolve 54 fault slip vectors on fault subsegments. The displacement gradient field is parameterized by subdivision of the study area into triangular domains spanned by model nodes. I assume a linear variation of the displacement gradient components within the triangular domains. The triangles are not allowed to cross the fault. I utilize a densification of the triangulation towards the fault trace in order to accommodate the large differences in displacement (and strain release) over relatively short distances (from mm at Istanbul to 2m at Gölcük near the epicenter; figure 4.5). However, a too dense triangulation leads, given the limited dataset, to heavily overparameterized models. I adopt a parameterization of 678 triangles spanned by 382 model nodes, leading to a total of $4 \times 382 + 2 \times 54 = 1636$ model parameters for the inversion of $\Delta \mathbf{u}$.

Substitution of the model parameterization in equation 4.1 leads to a linear system of equations. In order to assure internal consistency between $\nabla \mathbf{u}$ and fault slip in constituting the total deformation field, I extend the set of equations by defining 2 additional integration paths between all station pairs. A further set of equations derives from the fact that $\nabla \times \nabla \mathbf{u} = \mathbf{0}$ within regions bounded by faults. This constraint is defined for each triangle and a weight ($\alpha_r$) is assigned to tune the relative importance

Figure 4.5: Final parameterization of the İzmit models. Thick lines indicate fault segments, black dots are the triangle nodes and grey dots are the site positions. Note that in my choice for the triangle nodes I am not restricted to the locations of the observation sites. Triangles do not intersect faults. Nodes at the fault are doubled to allow the velocity gradient field to be discontinuous across the fault.
Table 4.1: Aspects of the inversion parameterization and average results for inversions I and II. Key: $i$, solution; $\sigma_r$, standard deviation of the $\nabla \times \nabla \mathbf{v} = 0$ equations; $\alpha_0$, $\alpha_i$, $\alpha_i^d$, and $\alpha_d$, the regularization parameters; $\tilde{r}_m = \frac{1}{M} \sum_{i=1}^{M} R_{ii}$, the average resolution, with $R_{ii}$ the diagonal elements of the resolution matrix and $M$ the number of model parameters; $\tilde{\sigma}_m^\prime = \frac{1}{M} \sum_{i=1}^{M} \sqrt{C_{ii}}$, the average standard deviation for the components of $\nabla \mathbf{v}$, with $M = 4T_n$ the number of components of $\nabla \mathbf{v}$; $\tilde{\sigma}_m^I = \frac{1}{M_I} \sum_{i=1}^{M_I} \sqrt{C_{ii}}$, the average standard deviation for the components of $f_k$, with $M_I = 2K$ the number of slip components.

<table>
<thead>
<tr>
<th>$i$</th>
<th>$T_n$</th>
<th>$K$</th>
<th>$\sigma_r$</th>
<th>$\alpha_0$</th>
<th>$\alpha_i$</th>
<th>$\alpha_i^d$</th>
<th>$\alpha_d$</th>
<th>$\chi^2$</th>
<th>$\tilde{r}_m$</th>
<th>$\tilde{\sigma}_m^\prime$</th>
<th>$\tilde{\sigma}_m^I$</th>
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<tbody>
<tr>
<td>I</td>
<td>382</td>
<td>54</td>
<td>2.0</td>
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<td>5.0</td>
<td>2.5</td>
<td>6.0</td>
<td>1.02</td>
<td>0.84</td>
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<tr>
<td>II</td>
<td>382</td>
<td>54</td>
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<td>3.56</td>
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For this application it proved unnecessary to regularize the fault slip parameters. The formal least squares solution to the regularized system is given by (see Spakman and Nyst (2002) for details):

$$
\mathbf{m} = (\mathbf{A}^T C_d^{-1} \mathbf{A} + \alpha_0^2 \mathbf{I}_0 + \alpha_i^2 \mathbf{I}_1 + \alpha_i^d \mathbf{D}^2 \mathbf{D})^{-1} \mathbf{A}^T C_d^{-1} \mathbf{d}
$$

(4.2)

where $C_d^{-1}$ is the data covariance matrix. The a posteriori model covariance is given by: $\mathbf{C} = (\mathbf{A}^T C_d^{-1} \mathbf{A} + \alpha_0^2 \mathbf{I} + \alpha_i^2 \mathbf{I} + \alpha_i^d \mathbf{D}^2 \mathbf{D})^{-1}$ and the model resolution kernel is $\mathbf{R} = \mathbf{C} \mathbf{A}^T C_d^{-1} \mathbf{A}$. The model depends on the tuning of four parameters: $\alpha_r$, $\alpha_0$, $\alpha_i$, and $\alpha_i^d$. For the tuning of these parameters, I primarily focus on obtaining a solution with acceptable data fit (by inspection of the normalized $\chi^2$ data misfit values), model covariance and resolution. The respective weights of the $\nabla \times \nabla \mathbf{u} = \mathbf{0}$ and the regularization equations, as well as an overview of some aspects of the solutions are provided in table 4.3.
4.4 Joint analysis

Presently, the computer implementation of the general method of Spakman and Nyst (2002) is restricted to the analysis of horizontal velocity/displacement vector fields. Although this implementation can be extended to include InSAR interferograms, the significant influence of atmospheric effects on the Izmit interferograms (see section 4.2.2), as well as the restriction to horizontal deformation made me adopt a different strategy for the joint analysis of GPS and InSAR data. Once the displacement gradient and fault slip has been inferred from the GPS data, equation 4.1 can be used to predict the horizontal displacement field at arbitrary points. The change in range \( \Delta u \) of the distance measured along the line-of-sight between the satellite and the ground point due to a relative displacement is \( \Delta u = -\mathbf{u} \cdot \mathbf{s} \), where \( \mathbf{u} \) is the displacement vector and \( \mathbf{s} \) is the unit vector pointing to the radar satellite. Assuming the absence of vertical coseismic displacements the predicted field \( \mathbf{u} \) is used to construct a synthetic interferogram.

I have implemented the forward approach for a 1x1 km regular grid spanning the modeling area contained within my InSAR image. I determine a predicted displacement field on the grid with respect to each GPS station. Subsequently, these 41 fields are made independent of the reference station. At each grid point the 41 resulting displacements are checked for outliers. Generally, the displacements are found to be consistent on a 95 % confidence level. From the 41 displacement fields a single, consistent displacement field is determined. Projection of this field on the line-of-sight direction provides me with a 1x1 km grid of range changes which is wrapped into a predicted interferogram.

The next step is a comparison of observed and predicted interferograms. The comparison is done mainly on fringe pattern and total amount of deformation, since error sources such as orbital errors, tropospheric delays, post-seismic deformation, as well as the absence of the vertical displacements will already introduce a misfit between the two interferograms. A quantitative analysis of the two dimensional and three dimensional range changes at the GPS stations showed that the restriction to a horizontal displacement field will introduce a misfit of up to 150 mm which corresponds to over 5 fringes in the wrapped interferograms. The largest vertical displacements are observed in the vicinity of the fault trace. Away from the fault the vertical displacements become small and the misfit rapidly reduces to less than 1 fringe. The influence of the tropospheric delays ranges from two to four fringes (Reilinger et al., 2000). Post-seismic displacements are very small compared to the coseismic and might introduce a misfit of less than 1 fringe in the model (Reilinger et al., 2000).

The first order differences between the observed and predicted interferograms are attributed to differences in fault slip between the inverted (from GPS solely) and actual slip at depth. These differences may exist as a result of lack of coseismic GPS data to fully constrain both the \( \nabla u \) field and fault slip (Spakman and Nyst, 2002). By trial and
error, the fault slip distribution is changed (and imposed a priori on the GPS inversion) to optimize the fit between the predicted and observed interferograms.

4.5 Coseismic surface deformation

4.5.1 Solution I: Model based on GPS data only

I commence by inverting the GPS data only leading to solution I. The model (figure 4.6) can fit the data very well on a 95% confidence level ($\chi^2 = 1.0$) with acceptable model covariance and resolution. The maximum data misfit does not exceed 0.013 m.

Figure 4.7 shows the $3\sigma$ model standard deviations ($\hat{\sigma}_i = \sqrt{C_{ii}}$) of each gradient component plotted as a percentage of the displacement gradient estimate of the component in each model node, as well as the diagonal elements of the model resolution kernel in both the longitudinal ($\phi$) and latitudinal ($\theta$) directions. I find that the interior of the model is well resolved with relatively small errors compared to the model amplitude estimate. Since the amplitude estimates of $(\nabla u_{\phi\phi})$ and $(\nabla u_{\phi\theta})$ are about a factor two larger than the amplitude estimates of $(\nabla u_{\theta\phi})$ and $(\nabla u_{\theta\theta})$ a significantly larger part of the model estimates exceeds the model errors (figure 4.7). The resolution of the model is good, though toward the boundary of the model the resolution deteriorates significantly. This is generally due to the lack of data and the increased regularization imposed towards the boundary. However, in the Marmara Sea and along the Göløyaka segment the reduced resolution may also be due to a trade-off between the displacement gradient field and fault slip, which is caused by a lack of data in the proximity of the fault (Nyst, 2001; Spakman and Nyst, 2002). The amplitude estimates of the displacement gradient field toward the model boundary decrease as expected for a coseismic strain field. However, the amplitudes are also smaller than the formal standard deviations and care should be taken in the interpretation of this part of the model.

The model standard deviations of the fault slip estimates are relatively small (averaging $\sim$14 mm). The largest standard deviations ($\sim$25 mm) are observed on the Göløyaka segment, whereas the smallest standard deviations ($\sim$7 mm) are found on the Marmara Sea segment.

The strain contribution of solution II (figure 4.6a) shows four distinct quadrants of extensional and contractional strains reflecting the source mechanism of the earthquake (figure 4.1). However, the contractional quadrant south of the fault is significantly smaller than the extensional quadrant. The transition between the quadrants north and south of the fault is characterized by left-lateral shear. The strain field along the Sakarya fault shows minor left-lateral shear on both sides of the fault. Significant contractional strain is observed north of the Göløyaka segment, with reduced left-lateral shear strains south of the fault segment. The strain contribution (figure 4.6a) also shows a band of right-lateral shear strain along the Bay of Izmit, extending onland eastward to Sapanca Lake. The band is accompanied by significant clockwise rotations (figure 4.6b). I note...
4.5 Coseismic surface deformation

Figure 4.6: The displacement gradient contribution of the model based on GPS data only (Solution I): a) Strain field. The contouring denotes the effective strain \((1/2e_{ij}e_{ij})^{1/2}\). The arrows denote the principal strains: contraction (black) and extension (white). b) Rotation field in degrees. For a color version of this figure see figure C.14 of appendix C.3.
that the band is located slightly to the north of the modeled fault trace in the Bay of Izmit. The rotation contribution further shows large anticlockwise rotations on both sides of the fault. This large scale anticlockwise rotation results from the unloading of seismic strain. However, the anticlockwise rotations are not uniform (figure 4.10b). The maximum anticlockwise rotations are found south of the Gölcük segment and north of the Sakarya and Karadere segments.

The fault motion contribution of solution I (figure 4.8) shows increasing right-lateral slip from the Marmara Sea eastward. Maximum slip (2.9 m) is reached on
the Sapanca segment. The fault motion subsequently decreases eastward to less than 1 m on the Gölyaka segment. Significant fault slip (~0.7 m) is also observed on the Marmara Sea segment consistent with source rupture analysis estimates (Güllen et al., 2002). The modeled fault slip of 1-1.5 m on the Yalova segment is consistent with estimates of rupture source analysis (Güllen et al., 2002) and with estimates by slip distribution modeling (Reilinger et al., 2000). Both the Hersek and Gölcük step-overs show significant normal motion. The fault motion is reasonably consistent with the observed surface ruptures (figure 4.8). However, despite a dense parameterization of the fault trace, the fault motion can not fit all the details of the surface rupture data (e.g. the large ruptures observed at Gölcük and on the Sakarya segments or the absence of rupture in the Akyazi Gap). I note that the fault slip modeled probably represents an average over the top few km of the crust and therefore does not necessarily coincide with the surface rupture data. The normal motion on the Gölcük step-over is less than observed, while the right-lateral motion is significantly higher. This difference may be induced by the smoothing of the fault trace, changing for instance the strike of the Gölcük step-over from N135°E to N107°E.

The synthetic interferogram determined from solution I (figure 4.9a) shows a south-
Figure 4.9: Synthetic interferogram of a) the model based on GPS data only (Solution I), and b) the model based on the joint analysis of GPS and InSAR (solution II). For a color version of this figure see figure C.15 of appendix C.3.
eastward turning of the fringes north of the fault, whereas the InSAR interferogram shows more E-W orientated fringes in this area. I attribute this difference to an under-estimation of fault slip on the segments along the Bay of Izmit. This is caused by a trade-off between fault slip and the displacement gradient field as a result of absence of GPS stations close to the fault (Nyst, 2001; Spakman and Nyst, 2002). The GPS stations are located along the shores of the Marmara Sea and Bay of Izmit whereas the fault trace is located along the southern shore. South of the fault the GPS stations are closer to the fault and the data better constrain the displacement gradient field. This trade-off may cause the band of right-lateral strain and clockwise rotations observed in my surface deformation model. I further note that the data density at Gölcük and along the western Sapanca segment is poor on the north side of the fault. To improve the fit between predicted and observed interferograms I initially a priori increase (by trial and error) the right-lateral fault slip on the Gölcük and Sapanca segments.

4.5.2 Solution II: Model based on a joint analysis of GPS and InSAR

Figure 4.8 shows the final fault motion obtained which provides an acceptable fit to both the GPS and InSAR data. The slip estimation shows a second peak of almost 4m on the Gölcük segment. The fault motion on the Yalova step-over has been increased due to the constraints imposed on the Gölcük segment. During the trial and error procedure the fault slip on the Yalova and Marmara Sea segments has been constrained to the previous unconstrained slip estimates of solution I to prohibit the occurrence of significant left-lateral shears, induced by the trade-off with an overestimate of right-lateral fault motion, within the Marmara Sea. The fault slip contribution complies reasonably well with the observed surface rupture data (figure 4.8). However, slip on the Sakarya segment seems to be underestimated, though there is a large scatter in the data, while slip on the Karadere segment is overestimated. This may be attributed to the presence of a thick sedimentary layer at the surface along these fault segments. Rupture observed within these sediments need not be representative of the rupture in the first few km of the Earth’s crust. The InSAR interferogram along these faults is very decorrelated and therefore can not further constrain my slip estimates.

The fit to the GPS data is slightly deteriorated ($\chi^2 = 3.56$) compared to solution I. However, the maximum misfit does not exceed 0.018 m. I attribute this deterioration to either the coarseness of my parameterization which does not allow enough detail to perfectly accommodate the trade-off in my displacement gradient field or strong changes in the displacement gradient field near the fault. The synthetic interferogram of solution II (figure 4.9b) compares to first order reasonably well with the InSAR image. North of the Bay of Izmit the predicted interferogram contains about 7 fringes fewer than the InSAR interferogram. This implies a difference of 196 mm in deformation in the range change direction. Especially in the near field, the density of the fringes is less than observed in the InSAR interferogram. This should be mainly attributed
Figure 4.10: The displacement gradient contribution of the model based on the joint analysis of GPS and InSAR (Solution II): a) Strain field. The contouring denotes the effective strain \( \left( \frac{1}{2} e_{ij} e_{ij} \right)^{1/2} \). The arrows denote the principal strains: contraction (black) and extension (white). b) Rotation field in degrees. For a color version of this figure see figure C.16 of appendix C.3.
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to the absence of vertical displacements in the vicinity of the fault trace (see section 4.4). The slight difference in tilt between the two interferograms north of the Bay of Izmit between Gölcük and Izmit can be attributed to tropospheric delay errors, although the smoothing of the Gölcük step-over may also have its influence. The horizontal fringes north of the fault along the Bay of Izmit are well reproduced. Further west, the synthetic fringes start to bend southward slightly east of where the InSAR fringes bend. Directly south of the fault the two interferograms comply rather well. However, going southward atmospheric effects obscure the deformation signal contained in the InSAR image and correlation between the two interferograms is lost (see section 4.2.2). West of the Bay of Izmit decorrelation in the InSAR interferogram, makes interpretation of the fringe pattern impossible. The model covariance and resolution have hardly been affected by the implementation of the a priori constraints. The average standard deviations on the fault slip estimates remain 0.014 m

The strain contribution of solution II (figure 4.10a) no longer shows the band of right-lateral shear strains observed in solution I. North of the Yalova step-over WNW- ESE extension is observed, where north of the Gölcük segment NNE-SSW contraction dominates. South of these segments I obtain WNW-ESE contraction and NNW-SSE extension, respectively. The strains are not just antisymmetric over the fault, but also very asymmetric. Around Gölcük the extensional strains south of the fault are significantly larger than the contractional strains north of the fault ($\sim 5.46 \cdot 10^{-5}$ vs. $\sim 2.43 \cdot 10^{-5}$, respectively). The same holds for the Gölyaka segment, where I find $\sim 4.07 \cdot 10^{-5}$ of contractional strain north of the fault and $\sim 1.92 \cdot 10^{-5}$ of extensional strain to the south. In the rotation field (figure 4.10b) the band of clockwise rotations in the Bay of Izmit has been replaced by anticlockwise rotation. Some minor clockwise rotation is still observed along the Sapanca segment. In all other aspects the strain and rotation fields of solution I and II are comparable.

4.6 Discussion

4.6.1 Influence of fault geometry

Though my deformation model reflects the four quadrants of the earthquake source, the sizes of the quadrants are not consistent and are not centered around the epicenter of the earthquake (figure 4.11). The transition between the contractional and extensional quadrants south of the fault is shifted slightly westward to the longitude of the Yalova step-over, whereas north of the fault the transition is shifted eastward and is located at the longitude of Sapanca Lake (figure 4.10 and 4.11). The amplitudes of the strains and rotations of my surface deformation model are distinctly asymmetric across the Gölcük, Gölyaka, Sakarya, and Karadere segments (section 4.5.2; figure 4.10). The dense surface displacement field predicted by my model shows even more profound asymmetry in both the east and the north component (figure 4.12). At Gölcük the east and north
displacements south of the fault exceed those north of the fault by a factor 1.5 and 3.5, respectively. Along the Sapanca segment the absolute east displacements north of the fault are significantly larger than south of the fault (1.7 m vs. 1.3 m, respectively). Similarly, north of the Karadere segment I obtain an east displacement of 1.3 m and a north displacement of 0.7 m, whereas south of the fault the absolute displacements are 0.4 m and 0.3 m in the east and north, respectively. Wang et al. (2003) used a Green’s function approach to compute the displacement components of the Izmit earthquake for both a homogeneous and a layered earth model. In both cases they obtain asymmetry
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in the north and east components of the displacement field similar in pattern to my displacement field, though different in amplitude. However, the displacement field due to a strike-slip earthquake on a vertical fault plane should be purely antisymmetric across the fault. This implies that a deviation from this specific situation causes asymmetry to occur. I stipulate that asymmetry can be induced by 1) non-linear elastic deformation of the upper crustal rocks, 2) deviations of the earthquake rupture plane from the vertical, 3) lateral variations in rigidity across the fault, and 4) a complex fault rupture due to fault geometry.

For the 1997 $M_{w} = 7.6$ Manyi (Tibet) earthquake Peltzer et al. (1998) obtained significant asymmetry of the surface displacements which was attributed to non-linear elastic deformation of the upper crustal rocks. Compared to linear elastic deformation, non-linear elastic deformation will predict relatively larger displacements in nominally extensional quadrants. Except for the Gölcük segment, the relatively larger displace-

Figure 4.12: A dense surface displacement field determined from solution II on a 1x1 km regular grid for my model region. Top figure: The east component of the derived displacements. Bottom figure: The north component of the derived displacements.
ments observed for the 1999 Izmit earthquake occur in the contractional quadrants of the event which is opposite to that predicted by the non-linear elastic deformation. Neither can non-linear elastic deformation explain the significant shift of the transition between the quadrants observed in the strain field (figure 4.10 and 4.11).

Deviations of the earthquake rupture plain from the vertical were suggested by Fialko et al. (2001) to explain the asymmetry observed in the displacement field of the 1999 $M_w = 7.1$ Hector Mine earthquake. For a non-vertical strike-slip dislocation in an elastic half-space the surface displacements on the foot wall side of the fault decay faster with distance from the fault than displacements on the hanging wall side. For the Izmit earthquake this implies a south-dipping Yalova step-over, Gölcük segment and Gölcük step-over and a north-dipping Sapanca, Sakarya, Karadere and Gölyaka segment. The earthquake focal mechanism of the Izmit earthquake indicates a north-dipping fault with a dip of $87^\circ$. DeLouis et al. (2002) estimated a northward dip of $85^\circ$ for the ruptured segments based on broadband teleseismic modeling, whereas Reilinger et al. (2000) inferred an optimal fault geometry from GPS data and obtained a vertical dip for the Yalova, Gölcük, Sapanca and Sakarya segments and a $63^\circ$ northward dip for the Karadere segment. On November 12 1999, the $M_w = 7.1$, Düzce earthquake ruptured the Gölyaka segment of my fault trace. Based on the GPS displacements of this event Bürgmann et al. (2002) inferred that the Gölyaka segment is a $54^\circ$ north dipping oblique normal, right-lateral fault. Therefore, the deviating dips of the Gölyaka and Karadere segments may explain the asymmetry observed in the surface displacements across these segments. The deviations from vertical for the other segments are insignificant and do not explain the observed asymmetry.

Theoretically, Rybicki (1978) and Mahrer and Nur (1979) showed that an asymmetry in the displacement field across a fault may be induced by a lateral variation in rigidity. Their results portray that given a rigidity contrast across the fault, strain localizes on the low-rigidity side of the fault. In my model strain localizes south of the Gölcük and north of the Gölyaka segments. Assuming that the stress level on both sides of the fault is comparable, I can deduce a rigidity contrast of 2.25 and 0.47, respectively, between the region north of the fault and the region south of the fault along these segments. However, no significant asymmetry in amplitude of the strain field is observed across the Sapanca or Karadere fault segments, while a significant asymmetry in the displacement field is observed. Therefore, a rigidity contrast across the fault may not be the proper explanation for the asymmetry observed.

The surface trace of the North Anatolian fault in the Marmara Sea region is characterized by several step-over features and gaps between the various fault segments. Based on an inversion of teleseismic and near-field data, Yagi and Kikuchi (2000) deduced that the rupture at the epicenter started just west of the Gölcük step-over and propagated westward along the Gölcük segment. During the first 5 seconds of the event, rupture to the east was obstructed by the extensional step-over. Subsequently, eastward rupture was triggered on the Sapanca segment. An asymmetric rupture of the
4.6 Discussion

Figure 4.13: Schematic representation of the two-stage main rupture: failure is initiated near the Gölcük step-over and propagates westward, while rupture eastward is obstructed by the step-over feature. This induces an increased extensional quadrant south of the fault and contractional strain north of the fault (top left figure). After approximately 5 sec. eastward rupture on the Sapanca segment is triggered and propagates at super-shear velocity releasing the accumulated strain (top right figure). This increases the extensional quadrant north of the fault, locating the transition between the extensional and contractional quadrant near Sapanca Lake (see figure 4.11). Bottom two figures give a schematic representation of the strain against the distance relative to the fault. Solid line indicates the strain accumulated prior to the 1999 event, dashed line indicates strain accumulated at the moment eastward rupture commenced. As the earthquake strikes strain west of the epicenter is released immediately, while during the first 5 sec. significant additional strain accumulates east of the epicenter.
main source is also visible in the time-evolution of the slip distribution model of De-
Louis et al. (2002) and in the source time function of the earthquake (Li et al., 2002,
figure 4.11). Due to the two stage main rupture strain west of the epicenter is released
immediately, while strain release east of the epicenter is obstructed by the absence of
eastward rupture (temporarily prevented by the step-over feature). During this short
period (5-7.5 sec.) contractual strain due to the earthquake accumulates north of the
step-over as extension is induced south of the step-over (consistent with the earthquake
focal mechanism; figure 4.11 and 4.13a). The subsequent eastward rupture releases
the strain and causes an initially very rapid eastward movement of the northern block
(figure 4.13b). This is consistent with the super-shear rupture velocity (4.8-4.9 km/s)
oberved along the Sapanca segment, whereas rupture along the other segments prop-
agated with a velocity of ~3.5 km/s (Bouchon et al., 2002; DeLouis et al., 2002; Yagi
and Kikuchi, 2000). This process explains the presence of the maximum east displace-
ments north of the fault at Sapanca Lake, as well as the shift observed in the transition
between extensional and contractional quadrants. The rapid eastward motion of the
northern block at the fault induces extension within the western half of the block, while
causing contraction in the eastern half of the block (figure 4.11). South of the fault, the
initial absence of eastward rupture and the subsequent sudden release of the central and
eastern part of the block induces the smaller westward shift of the transition between
the quadrants (figure 4.11).

I conclude that the dip of the Gölçük and Karadere segments may explain the
asymmetry observed in the surface displacement field across these fault segments.
However, a relative rigidity contrast across the Gölçük segment of 0.47, where the
region north of the fault is relatively weaker than the south, can also explain the local-
ization of deformation observed across the fault. The rigidity contrast may be attribut-
to the presence of volcanic rocks south of the fault. A combined effect is most prob-
able. The significant asymmetry in the surface displacements across the Sapanca and
Gölçük fault segments is a direct result of a two-stage main rupture process with initial
failure westward along the Gölçük segment, while eastward rupture was obstructed by
the Gölçük step-over followed approximately 5-7.5 sec after the onset of the event by
the triggering of failure along the Sapanca segment eastward.

4.6.2 Double source rupture process

In both solutions the largest strains are observed in the northwestern, southwestern,
northeastern and southeastern parts of my model (figures 4.6 and 4.10). In solution II
(figure 4.10) I observe a patch of NNE-SSW contraction north of the Gölçük segment
with twice as large NNW-SSE extension south of the fault segment. The extension
is associated with significant anticlockwise rotations. Along the Sakarya segment the
strain field portrays left-lateral shear. North of this segment and the Karadere segment
significant anticlockwise rotations are observed. This observation may be the result
of the trade-off between fault slip and the displacement gradient field caused by an overestimation of the fault slip. However, significant surface breaks are observed along this segment which actually exceed the model estimates.

Studies of the source process of the Izmit earthquake (Güllen et al., 2002; Li et al., 2002) reveal a complex source process. General consistency exists on the occurrence of at least three consecutive source ruptures (figure 4.11). The main event was located south of Izmit and induced a bilateral, asymmetrically propagating rupture (the two stage main rupture). After 20-25 sec, a second source initiated near Hersek. This induced a relatively slow rupture of 1-1.5 m and had a rather low moment release. About 30 sec after the initial rupture along the Gölcük segment, rupture on the easternmost segments was triggered at the location of the Akyazi Gap. The moment of this third rupture was a little less than half of the moment of the main source.

The source located at the Akyazi Gap provides a reasonable explanation for the observed shear around the Sakarya segment. The contractional strain induced north of the segment by the main shock is located in the extensional domain of the focal mechanism of this consecutive event (figure 4.11). Similarly, the extensional strain south of the segment due to the main rupture is located in the contractional domain of the second source. The superposition of the contributions of both events result in the shear strains observed in my model. The contractions and extensions due to the main source around the Karadere and Gölyaka segments are enhanced by the third source. This explains the relatively large strains observed in the eastern region of my model. Along similar lines, the second source at Hersek could be responsible for the strain field observed around the Gölcük segment. However, there would be a strong interaction with the complex main source. Therefore, the exact extent of the influence of the second source on the deformation field is much more difficult to distinguish. The significant strains in the western part of my model are a direct result of the main and second source.

4.6.3 Implications on slip distribution modeling

The complex source rupture and the influence of the fault geometry on the rupture initiation and interaction have important implications on traditional slip distribution modeling based on geodetic data. The elastic dislocation theory (Okada, 1985) utilized in these models assumes a homogeneous elastic half-space to relate surface displacement observations to slip on the fault. This assumption only allows lateral asymmetry of the displacement field to occur due to fault geometry and fault dip. However, the strike of the Izmit rupture is rather consistent for all but the Karadere segment, while the dip is near-vertical for all but the Karadere and Gölyaka segments. The complicated and very asymmetric displacement field deduced for the Izmit earthquake west of the Karadere segment would therefore lead to significant misfits to the data on both sides of the fault. The relatively large displacements on one side of the fault will be
underestimated, while the relatively smaller displacements on the other side of the fault will be overestimated, a pattern indeed observed in the slip distribution models of the Izmit earthquake of Reilinger et al. (2000) and Feigl et al. (2002). Therefore, I conclude that even though slip distribution models may provide important information on the mechanical behavior of the upper crust and about the mechanics of the earthquake rupture, they are based on simplifying assumptions which may not account for all the asymmetry observed across the fault. So, a careful assessment of the surface deformation of an earthquake is necessary to identify the influence of complex fault geometry and crustal structure on the displacement field.

4.6.4 Relaxation of the long-term strain field

Since the Marmara Sea region was identified as a seismic gap (Nalbant et al., 1998), a network of GPS observation sites has been established in the area (McClusky et al., 2000) and the interseismic surface deformation field was studied (Ayhan et al., 2002; Kahle et al., 2000). The surface deformation models are dominated by right-lateral shear strain rates all along the fault trace. Based on these shear strain rates an interseismic slip rate of 11-25 mm/yr was deduced for the North Anatolian fault (Ayhan et al., 2002). Except for the peninsula north of the Bay of Izmit the principal extensional strain rate axis dominates the shear strain rates.

The 1999 Izmit earthquake has released strain build-up over many years. Taking the interseismic slip rate of 11-25 mm/yr and the maximum fault slip obtained in my model (≈4 m), strain may have been accumulating over the past 160-360 years. The last main earthquake on this section of the North Anatolian fault occurred in 1719 (Barka et al., 2002), 280 years ago. The interseismic surface deformation models obtain an average dilatational strain rate of ≈ 1 \times 10^{-7} \text{ strain/yr} (Ayhan et al., 2002; Kahle et al., 2000) which over 280 years leads to a total amount of ≈ 2.8 \times 10^{-5} \text{ strain} that has accumulated. This is in good agreement with the average strain release I deduce in my model (≈ 2.3(±1.0) \times 10^{-5}). Therefore, I conclude that the Izmit earthquake has released almost all strain which has been accumulated since the last large earthquake on this stretch of the North Anatolian fault in 1719.

4.7 Conclusions

In a joint analysis of GPS and InSAR data the surface deformation field of the August 17 1999 Izmit (Turkey) earthquake has been determined in terms of strain, rotation and fault motion. The incorporation of fault motion is a unique feature of the implemented method (Spakman and Nyst, 2002) and is of prime importance for the purely kinematic estimation of a coseismic surface deformation field. The InSAR information was specifically useful to minimize the trade-off between fault slip and the displacement gradient in the inversion. This effect is most profound north of the Bay of Izmit where
4.7 Conclusions

GPS stations are relatively far from the fault. On the Gölcük segment of the fault trace slip up to ~4 m is necessary to properly fit the InSAR interferogram, whereas the GPS inversion led to only ~2 m of slip.

The strain contribution of my model shows four distinct quadrants of extensional and contractional strains reflecting the earthquake focal mechanism. The transition between the quadrants is not centered on the epicenter, but shifted eastward north of the fault, to the longitude of Sapanca Lake, and westward south of the fault, to the longitude of the Yalova step-over. Along the Sakarya segment I obtain left-lateral shear strains, whereas my fault slip estimates underestimate the surface ruptures. The strain field is distinctly asymmetric across the Gölcük and Gölyaka segments. The rotation field shows dominant anticlockwise rotations resulting from the unloading of seismic strain. Significantly larger anticlockwise rotations are observed south of the Gölcük and north of the Sakarya and Karadere segments.

My surface deformation model can be directly related to the very complex source rupture of the earthquake. I identify a two-stage main rupture process near İzmit and two consecutive source ruptures located at Hersek and the Akyazi Gap, respectively. The two-stage main rupture initiated near the Gölcük step-over and propagated westward along the Gölcük segment, while rupture eastward was obstructed by the step-over feature. This initial stage induces an increased extensional quadrant south of the fault and contractional strain north of the fault. Approximately 5-7.5 seconds into the main event eastward rupture on the Sapanca segment was triggered and propagated at super-shear velocity along the segment, releasing the accumulated strain. This process increased the extension in the quadrant north of the fault, locating the transition between the extensional and contractional quadrant near Sapanca Lake. Thus, throughout this main rupture the step-over features in the fault geometry had a very significant influence on the rupture initiation and propagation.

The two consecutive sources strongly interact with the deformation field of the main source and are responsible for the left-lateral shear strains observed along the Sakarya segment and the complicated pattern of deformation observed along the Gölcük segment. On the easternmost segments (Karadere and Gölyaka) the dip of the fault plane starts introducing asymmetry across the fault. However, the presence of strong volcanic basalts south of the Gölyaka segment may also introduce part of the localization of deformation observed north of this segment.

Finally, I deduce that the İzmit earthquake has been responsible for releasing almost all the strain which has been accumulating since the last main event on this section of the North Anatolian fault in 1719.