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**ANALYSIS OF THE CRUSTAL DEFORMATION CAUSED
BY THE 1999 İZMİT AND DÜZCE EARTHQUAKES USING
SYNTHETIC APERTURE RADAR INTERFEROMETRY**

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**1999 İZMİT VE DÜZCE DEPREMLERİNİN NEDEN OLDUĞU
KABUK DEFORMASYONUNUN YAPAY AÇIKLIK RADAR
İNERFEROMETRİSİ İLE İNCELENMESİ**

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**This thesis is dedicated to the memory of my supervisor Aykut Barka, who passed
away during this study**

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ABBREVIATIONS

CMT	: Centroid Moment Tensor
CNES	: Centre National d'Etudes Spatiales, France
CNRS	: Centre National de la Recherche Scientifique, France
DEM	: Digital Elevation Model
DMSP	: Defense Meteorological Satellite Program, USA
E	: East
ERD	: Earthquake Research Department, Afet İşleri
ERS	: Earth Resource Satellite
ESA	: European Space Agency
GFZ	: GeoForschungsZentrum Potsdam
GPS	: Global Positioning System
GTOPO30	: Global Topographic data 30 second posting, USGS
HVD	: Harvard University
InSAR	: Synthetic Aperture Radar Interferometry or Interferometric Synthetic Aperture Radar
JPL	: Jet Propulsion Laboratory
LOS	: Line Of Sight
M	: Magnitude
Ms	: Magnitude (from surface waves)
Mw	: Moment magnitude
Mo	: Seismic moment
N	: North
NAF	: North Anatolian Fault
NASA	: National Aeronautic Space Administration, USA
Nm	: Newton meter
NOAA	: National Oceanographic and Atmospheric Administration, USA
RMS	: Root Mean Square
S	: South
SAR	: Synthetic Aperture Radar
SLC	: Single Look Complex
SLR	: Side Looking Radar
SNR	: Signal to Noise Ratio
SPOT	: Satellite Pour l'Observation de la Terre, France
SVD	: Singular Value Decomposition
TUBITAK	: Türkiye Bilim ve Teknik Araştırma Kurumu
UGGS	: United States Geological Survey, USA
UTM	: Universal Transverse Mercator
USA	: United States of America
W	: West
3D	: Three Dimensional

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LIST OF SYMBOLS

a, b	: Variables of reverse exponential function
A	: Amplitude, fault or rupture area
α	: Baseline orientation angle, fault strike
B	: Baseline
B_c	: Critical baseline
B_h	: Horizontal component of baseline
B_r	: Frequency bandwidth
B_v	: Vertical component of baseline
B	: Parallel component of baseline
B_⊥	: Perpendicular component of baseline
β	: Angle between equator and nadir track
c	: Speed of light
D	: Antenna width
δd	: Elevation change
δR	: Range difference
δR_a	: Ground azimuth resolution
$\Delta\sigma_f$	: Coulomb stress change
δR_g	: Ground range resolution
$\Delta\sigma_n$	: Normal stress change
ΔR	: Scalar range change
$\Delta\tau$	: Shear stress change
ϕ	: Phase difference
H	: Height of radar instrument
h	: Height of target
h_a	: Altitude of ambiguity
φ	: Look angle (near range)
ψ	: Look angle (far range)
η	: Incidence angle
λ	: Wavelength
I	: In-Phase
L	: Antenna length
l	: Fault length
μ'	: Shear modulus, effective coefficient of friction
r	: Rake
R	: Range
R_n	: Near range
R_f	: Far range
R_m	: Mid range
s	: Slip
$\hat{s}$	: Unit vector pointing to satellite
S_w	: Swath width
Q	: Quadrature

θ : Look angle (mid range), fault dip
 τ_p : Pulse duration
 $\vec{u}$: Displacement vector
 $\mathbf{V}_s$: Velocity of radar instrument (Satellite or aircraft)
 $\mathbf{V}_x$: Velocity of target
 z : Fault depth
 w : Wave's frequency, fault width



ANALYSIS OF THE CRUSTAL DEFORMATION CAUSED BY THE 1999 MARMARA EARTHQUAKE SEQUENCE USING SYNTHETIC APERTURE RADAR INTERFEROMETRY

Summary

In this study, interferometric synthetic aperture radar (InSAR) technique is used to deduce the source parameters of the August, 17th 1999 Izmit and November, 12th 1999 Düzce earthquakes. InSAR is a method in which radar images collected by imaging radar systems on board, airplane or satellite platforms are combined in order to map the elevations, movements, and changes of the Earth's surface. A radar image contains both the amplitude and the phase of the electromagnetic signals reflected from targets within the imaging area. InSAR technique uses the phase information in two SAR images by calculating the phase difference between each pair of corresponding image points after precisely aligned to a fraction of a pixel width. The resulting new image is called interferogram. The interferogram is an interference pattern of fringes due to relative phase difference. Relative phase difference occurs as a result of slightly different viewing angles, changes on the Earth's surface, and tropospheric delays in the radar signal. Therefore, when the phase difference due to the different viewing geometry is removed from the interferogram, the remaining phase difference will practically show surface change only, assuming that no atmospheric artefacts exists. To measure the movements of the Earth's surface by InSAR, therefore, two images of an area taken at different times are required. When the necessary conditions, such as good orbital separation and same target reflectivity, are met, the surface movements that occurred in the time interval between the two data acquisitions will be captured by InSAR technique very accurately (sub-centimeter) with a fine resolution and high spatial distribution (100x100 km for ERS satellites). The movements on the Earth's surface can be due to earthquakes, plate movements, volcanoes, glaciers, landslide, salt diapirism, groundwater and petroleum extraction and land watering.

Here in this study, InSAR technique is used to map the surface deformation caused by 1999 Marmara earthquakes. Coseismic interferograms are calculated using radar images of the Earth Resource Satellites (ERS) operated by the European Space Agency. Two packages of SAR processing software are used; DIAPASON developed by CNES (Centre National d'Etudes Spatiales) of France and ROI_PAC developed jointly by California Institute of Technology and Jet Propulsion Laboratory (USA). First, the source parameters of the earthquakes are deduced by modelling the coseismic interferograms, together with the measurements of Global Positioning System (GPS), using elastic dislocation on rectangular planes embedded in homogenous and isotropic half space. Modelling is performed using both a forward approach and a simple inversion procedure with iterative linear least squares. Then, using the source parameters deduced from InSAR and GPS modelling, analyses of Coulomb failure stress are made to investigate stress interactions of the 1999 earthquake sequence together with some previous large earthquakes that occurred

in the region, and to study the seismic hazard in the Marmara region.

To determine an improved model of the slip associated with the 1999 Izmit earthquake, which ruptured the North Anatolian fault at the eastern end of the Sea of Marmara, SAR data is used in combination with tectonic field observations. The leading goal is to understand the main features of the coseismic and postseismic deformation, which are captured together in the SAR data. To achieve this, the ERS1-2 SAR data are analyzed carefully, which allows atmospheric artefacts to be identified and removed. Detailed field mapping and measurements of the earthquake surface rupture are also used. Dislocations in elastic half-space and a forward modelling strategy allow a slip model to be obtained by steps. A trial-and-error approach is combined with conventional inversion techniques to determine the slip in the different regions of the fault. The SAR data are well explained with three main zones of high slip along the fault, releasing a total moment of 2.3×10^{20} Nm (M_w 7.6) that is higher than the seismological estimates (1.7 - 2.0×10^{20} Nm). The inhomogeneous slip distribution correlates with fault segments identified at the surface. The Izmit rupture appears to have extended 30 km west of the Hersek peninsula into the Sea of Marmara with slip tapering from 2 m to zero. The western end of the rupture is located 40 km SSE from Istanbul. Using a careful approach and the available GPS data, a slip model that represents the coseismic slip alone is obtained, which suggests that the moment release during the main shock was 1.9×10^{20} Nm (Newton meter) (equivalent of moment magnitude $M_w=7.5$), coincident with the seismological estimates. It is concluded that the SAR data include the effects of 2 m of fast after-slip during the month following the main shock, within a zone of the fault located 12-24 km below the epicentral region. Near the hypocenter at a depth of 18-km, the fault appears to have experienced dynamic slip of 1 m associated with the main shock, followed by 2 m of rapidly decelerating postseismic shear during the following month. This study suggests that the distribution of heterogeneous slip and loading along the different fault segments may be important factors controlling the propagation of large earthquake ruptures along the North Anatolian fault.

Analysis of the coseismic interferograms of the Izmit earthquake reveals that the interferograms include also atmospheric artefacts correlated with topography. The phase-elevation ratio decreases with increasing elevation, reaching up to 6 cm of relative phase delay. Although the phase-elevation ratio also varies laterally, a simple, horizontally uniform model of atmospheric artefacts is calculated using a digital elevation model. Correction of the observed interferograms using this model reveals that some of the anomalies in the fringe pattern, which were previously interpreted as triggered slip, are also associated with atmospheric artefacts. The model not only explains the wide spread noise, deflection and bending in fringes, but also reveals some large scale artefacts previously undetected.

About three months after the Izmit earthquake, the Düzce earthquake occurred, extending the Izmit rupture about 50 km further east. The two adjoining earthquakes produced a surface rupture over 200 km of long along the North Anatolian fault. One interferogram is calculated to reveal the coseismic surface deformation caused by the earthquake. To deduce the source parameters of the Düzce earthquake the interferogram is modelled separately and jointly with the coseismic GPS measurements after removing atmospheric artefacts correlated with topographic elevation from the interferogram. Taking into account the slip distribution and the location and the length of the surface rupture observed in the field, an iterative linear inversion technique is used in the modelling. Focal mechanisms of the Düzce earthquake show that the fault dips strongly (59° - 64°) to the north with a dominant strike slip and a minor normal component. Field observations, however, show that the displacement is almost pure strike slip along most of the fault rupture and the

oblique slip is restricted only along the westernmost portion of the rupture. Preliminary modelling also supports the seismological observations in that both the SAR data and the GPS data, too, require a north-dipping fault. First, an optimal fault dip is therefore found by joint and separate inversions. Then using a fault with an optimum plunge, found to be 62° to the north, in the first stage of modelling, a simple rupture with single-fault geometry is used to deduce subsurface slip distribution. The InSAR and GPS derived models found through these inversions are very similar to each other showing a simple slip distribution in which the coseismic slip is maximum in the center and tapers off towards the both ends of the fault, in consistent with the surface slip distribution observed in the field. Including the joint inversion model, all the models explain both the InSAR data and the GPS data within the resolution of the geodetic data set (with RMS [root mean square] being well below 2 cm). Seismic moment deduced from the joint inversion is 6.1×10^{19} Nm ($M_w=7.2$), consistent with seismological estimates. In the second stage of modelling, a more complicated rupture geometry is used in the modelling because accommodating a significant horizontal motion in the long term via a fault that has such a significant fault dip is mechanically difficult to explain. Thus, a fault with multiple-fault-rupture geometry consistent with the spatial distribution of the aftershocks and tectonic observations is constructed and used in the modelling of the geodetic data set. Two intersecting faults are assumed to have been ruptured, one being the vertical Düzce fault and the other being a north dipping preexistent structure at depth. In this case, the Düzce earthquake is assumed to have been nucleated on the preexistent fault with north dip and propagated sideward and upward, triggering the vertical Düzce fault. Models obtained from the inversions of the geodetic data set using such a rupture geometry also explain both the InSAR data and the GPS data very well. Thus, spatial distribution of the geodetic data set allows both types of models to be reasonable as both of them explain the observation nearly equally well. Additional data, particularly in the near field, are required to better constrain the rupture geometry. In the absence of such a new data set, considering the long term activity of the North Anatolian fault (NAF) present in the region and the regional tectonic settings, a multiple-fault-rupture model is preferred here. Accordingly, the Düzce earthquake is thought to have been associated with multiple fault breaks, as commonly observed elsewhere in the world. Although there are some differences, the coseismic slip distribution on this complex fault geometry is similar to that found on the fault with single-fault-rupture geometry. Slip models with multiple-fault-rupture predict higher slip, particularly at depth and hence give slightly higher seismic moment ($6.3\text{--}6.5 \times 10^{19}$ Nm, $M_w=7.2$, which is still comparable with the seismological estimates). Both the GPS and InSAR data suggest a longer (~ 15 km) fault rupture to the east than the one mapped in the field, which explains why the magnitude of the Düzce event is surprisingly higher when considering its short rupture length observed in the field.

After determining the source parameters of the 1999 Izmit and Düzce earthquakes, the Coulomb stress changes caused by these earthquakes are calculated and interpreted in terms of static stress interactions. Because the Coulomb calculations are based on the elastic dislocation theory, source parameters of an earthquake play a key role in the accuracy of the Coulomb stress changes caused by that earthquake. Thus, unlike the previous studies, in which generally a rough slip distribution and a simple fault geometry were used, here in this study the source parameters determined from InSAR and GPS modelling are used in the calculations of the Coulomb stress changes. The Coulomb stress changes resolved on the Izmit rupture and the stress distribution in the earthquake area prior to the earthquake are calculated using four large earthquakes ($M > 7$) that occurred previously along the North Anatolian fault. Calculation of the Coulomb stress changes resolved on the Izmit rupture due to the previous earthquake shows that the Izmit earthquake

occurred in an area of stress increase. The effect of Coulomb stress changes on the Düzce earthquake due to the Izmit event is studied by calculating the resolved static stress changes on the Düzce rupture. Maps of the shear, normal, and the Coulomb stresses resolved on the Düzce rupture with a multiple-fault-geometry show that the Izmit earthquake, on the contrary to the previous events, promoted the Düzce earthquake by raising the static stress by 3–6 bars in the hypocentral area of the Düzce earthquake. Coulomb modelling shows the Sea of Marmara region is currently located in an area of enhanced stress increase due to the large earthquakes ($M > 7$) since 1912 Ganos event. The 1999 and 1912 events, in particular, increased the static stress over 5 bars on the submarine fault system in the east and west, respectively. The faults in this region therefore pose a serious seismic hazard particularly for Istanbul. Although it is under debate, detail studies based on the high resolution bathymetry data and deep seismic profiles suggest that the NAF in the Sea of Marmara is fragmented into three segments. If this is assumed to be the case and one or two segments simultaneously may break, then the question is whether the future earthquake will occur in the eastern or west Marmara. Considering the westward migration of earthquakes since the 1939 Erzincan event, one can suggest that the earthquake will likely occur in the eastern Marmara region. However, location of the historical large earthquakes that occurred in the Sea of Marmara before 1912 must be known better to answer this question with confidence.



1999 MARMARA DEPREMLERİNİN NEDEN OLDUĞU DEFORMASYONUN SENTETİK AÇIKLIK RADAR İNTERFEROMETRİSİ YÖNTEMİ İLE ARAŞTIRILMASI

Özet

Bu çalışmada 17 Ağustos 1999 İzmit ve 12 Kasım 1999 Düzce depremlerinin kaynak parametreleri sentetik açıklık radar interferometrisi (InSAR) yöntemi ile ortaya çıkartılmaktadır. InSAR topografik yükseklik, yeryüzü değişimi ve hareketlerinin alansal olarak ortaya çıkartılabilmesi için, uydu, uçak ve benzeri araçlara monte edilmiş radar sistemleri tarafından elde edilen radar görüntülerinin işlendiği bir metottur. Bir radar görüntüsü, görüntülenen bir alandan geri yansıyan elektromanyetik dalgaların hem genliğinden hem de faz değerlerinden meydana gelir. Ancak InSAR yönteminde sadece faz değerlerinden faydalanılır. Bu metotta iki radar görüntüsü hassas bir şekilde üst üste çakıştırıldıktan sonra bir görüntünün her bir pikseline ait faz değeri diğer görüntüde karşılığı olan pikseldeki faz değerinden çıkartılır. Bu şekilde ortaya yeni görüntü çıkar ve buna interferogram adı verilir. Bir interferogram iki görüntüye ait faz farklarının neden olduğu ve frinç olarak isimlendirilen renk bantlarından oluşan bir girişim desenidir. Faz farkı genel olarak, iki görüntünün farklı açılardan görüntülenmesi, yeryüzünde meydana gelen değişiklikler ve troposferdeki radar sinyali gecikmeleri sonucunda meydana gelir. Troposfer kaynaklı gecikmelerin olmadığı var sayıldığında, farklı bakış açılarından kaynaklanan faz farkı interferogramdan çıkartıldığı zaman geriye kalan sinyaller pratikte böylece sadece yeryüzünde meydana gelen değişimleri gösterecektir. Dolayısıyla, InSAR yöntemi ile yeryüzünde meydana gelen hareketleri ortaya çıkarmak için bir bölgeye ait farklı zamanlarda çekilmiş iki görüntüye ihtiyaç vardır. İki uydu arasındaki mesafenin uygun olması ve hedef noktalarının yansıma özelliklerinde fazla değişiklik olmaması gibi koşulların mevcut olması durumunda, iki görüntü çekimi arasında kalan zaman içerisinde meydana gelen yer hareketleri bu yöntemle çok geniş bir alanda (ERS uyduları için 100x100 km) hassas bir şekilde (cm altında bir hata payı ile) ortaya çıkartılabilir. Bu yöntemle ortaya çıkartılabilen başlıca yeryüzü hareketlerine örnek olarak depremler, levha hareketleri, volkanlar, buzullar, heyelanlar, tuz diyapirleri, yeraltı suyu ve petrol boşaltımı, ve tarla sulamaları verilebilir.

Bu çalışmada, InSAR yöntemi 1999 Marmara depremlerinin sebep olduğu yüzey deformasyonunu haritalamak için kullanıldı. Bunun için, Avrupa Uzay Kurumu ESA'ya ait ERS (Earth Resource Satellite) görüntüleri kullanılarak eşsismik interferogramlar hesap edildi. Interferogramların işlenmesinde Fransız Uzay Araştırmaları Merkezi CNES tarafından geliştirilen DIAPASON ve California Institute of Technology ve Jet Propulsion Laboratory (ABD) tarafından ortaklaşa geliştirilen ROI_PAC isimli yazılımlar kullanıldı. İlk olarak eşsismik interferogramlar GPS (Global Positioning System) verileri ile birlikte modellendi. Interferogramlarda gözlenen deprem deformasyonunun modellenmesinde homojen bir yarı uzayda olduğu varsayılan dikdörtgen düzlemlerde meydana gelen elastik yer değiştirme teorisi

kullanıldı. Modellemede hem düz çözüm, hem de küçük kareler hesaplamasına dayalı basit bir lineer ters çözüm yöntemi kullanıldı. Daha sonra, modellemeler sonucu elde edilen deprem kaynakları kullanılarak, Coulomb yenilme gerilmesi çalışmalarıyla 1999 Marmara depremlerine bölgede daha önce meydana gelen depremlerin etkisi, İzmit depreminin Düzce depremine olan etkisi ve İstanbul'da deprem tehlikesi irdelendi.

Kuzey Anadolu fayı üzerinde doğu Marmara bölgesinde oluşan 1999 İzmit depremi kınığına ait iyi bir kayma dağılımı modeli elde edebilmek için SAR verileri arazi gözlemleri ile birlikte değerlendirildi. Burada temel hedef, SAR verilerinin içerdiği eşsismik ve deprem sonrası deformasyonların önemli özelliklerini anlamaktır. Bunu başarmak için ERS1 ve ERS2 SAR verileri detay bir şekilde analiz edildi ve böylece bu verilerde bulunan ve atmosferdeki gecikmelerden kaynaklanan bazı heterojen hatalar (atmosferik hatalar) ortaya çıkartıldı. Ayrıca, yüzey kırığı üzerinde yapılan detay arazi ölçümlerinden ve haritalardan da yararlanıldı. Elastik yarı uzayda yer değiştirme ile düz çözüm stratejisi bir kayma modelinin adım adım elde edilmesine izin verdi. Fayın farklı bölgelerinde meydana gelen kaymaları ortaya çıkarmak için deneme yanılma yöntemi geleneksel ters çözüm yöntemleri ile birlikte kullanıldı. SAR verileri, fay boyunca üç ana bölgede konsantre olan yüksek kaymalarla çok iyi bir şekilde açıklanabilmektedir. Böylesi bir kayma modeli 2.3×10^{20} Nm (Newton metre) sismik moment vermektedir ve bu miktar sismik yöntemlerle elde edilen sismik moment değerlerinden ($1.7-2.0 \times 10^{20}$ Nm) yüksektir. Homojen olmayan bu kayma dağılımı yüzeyde gözlenen fay segmentleri ile oldukça iyi bir korelasyon göstermektedir. Modellemeler Hersek civarında yaklaşık 2 metre olan kaymanın batıya doğru azalarak yaklaşık 30 km boyunca devam ettiğini göstermektedir. Kırığın batı ucu İstanbul'un 40 km güney-güneybatısında yer almaktadır. Dikkatli bir yaklaşım ve mevcut olan eşsismik GPS verileri kullanılarak, salt eşsismik kaymayı gösteren bir model elde edildi. Bu modele göre, ana şok süresince açığa çıkan sismik moment, sismik verilerle elde edilenlere uyumlu olarak, 1.9×10^{20} Nm'dir ($\approx$ Moment Magnitude, $M_w = 7.5$). Buradan, SAR verilerinin, deprem merkezi altındaki bölgelerde ve fayın 12-24 km derinliklerinde deprem sonrası bir ay boyunca meydana gelen ve 2 metreye varan çok hızlı deprem-sonu-kayma (after-slip) hareketlerine ait yüzey deformasyonlarını içerdiği sonucu çıkartılmıştır. 18 km derinliklerde, deprem odak merkezi (hypocenter) civarında, depremin hemen ardından fayın 1 metrelik bir dinamik deprem-sonu-kayma gösterdiği ve depremi takip eden bir ay boyunca hızla azalan bir ivme ile 2 metreye ulaştığı düşünülmektedir. Bu çalışma, fay boyunca meydana gelen heterojen kayma dağılımının ve fay segmentlerinin heterojen bir şekilde yüklenmesinin Kuzey Anadolu fayı boyunca oluşan büyük depremlerin ilerlemesinde önemli rolleri olabileceğini işaret etmektedir.

İzmit eşsismik interferogramlarının detaylı bir şekilde incelenmesi göstermektedir ki, bu interferogramlar topografik yükseklikle korele olan atmosferik hatalar da içermektedir. Faz-yükseklik oranı artan yükseklikle azalmakta ve sinyal gecikmesi bazı bölgelerde 6 santimetreye ulaşmaktadır. Faz-yükseklik oranının yanal olarak da değişmesine rağmen, basit ve yatay yönde değişmeyen bir atmosferik hata modeli bölgenin bir sayısal arazi modeli kullanılarak hesaplandı. Interferogramların bu model kullanılarak düzeltilmesinden, interferogramlardaki frinçlerde bulunan ve daha önceden tektonik kökenli tetiklenmiş kaymalardan kaynaklandığı düşünülen bazı anomalilerin atmosferik hatalardan kaynaklandığı anlaşılmaktadır. Oluşturulan bu model sadece interferogramlardaki frinçlerde gözlenen büküm, kıvrım ve gürültüyü açıklamakla kalmayıp, daha önceden gözden kaçmış bazı büyük ölçekteki hataları da ortaya çıkarmaktadır.

İzmit depreminden yaklaşık üç ay sonra Düzce depremi meydana geldi ve bu İzmit depreminin ucunu yaklaşık 50 km daha doğuya uzattı. Bu depremler iki ucu bitişik

ve uzunluđu 200 kilometrenin üzerinde bir yer yüzü kırığı meydana getirdi. Düzce depreminin yer yüzünde meydana getirdiđi deformasyonu haritalamak için bir adet interferogram hesap edildi. Depremi kaynak parametrelerini ortaya çıkarmak için öncelikle bu interferogramda topografyadan kaynaklanan atmosferik hatalar modellenip giderildi ve sonra bu interferogram tek başına ve eşsismik GPS verileri ile birlikte ters çözüm yöntemi ile modellendi. Fay kırığının konumu ve uzunluđu ve üzerinde ölçülen kayma miktar ve dağılımları gibi arazi gözlemleri de dikkate alınarak, modellemelerde iterasyonlu lineer bir ters çözüm yöntemi kullanıldı. Deprem odak mekanizması çalışmalarından fayın kuzeye doğru oldukça eğimli olduđu (59° - 64°) ve kuvvetli sağ yanal atımın yanı sıra normal atıma da sahip olduđu ortaya çıkmaktadır. Ancak arazi gözlemleri, fayın büyük bir kısmı boyunca hareketin hemen hemen salt doğrudan atımlı olduđunu ve verrev kaymanın kırığın sadece en batı ucunda kısa bir mesafe içinde meydana geldiđini göstermektedir. Ön modellemelerden elde edilen sonuçlara göre, hem SAR hem de GPS verileri, sismik çalışmaları destekler bir şekilde, kuzeye doğru eğimli bir fay göstermektedir. Bu nedenle, ilk önce jeodezik verilere en uygun (optimum) bir fay eğimi iki ayrı veri setinin tek tek ve beraber ters çözümü yapılarak ortaya çıkartılmıştır. Daha sonra, 62° olarak bulunan optimum eğimli bir fay düzlemi kullanarak, modellemenin ilk aşamasında basit, tek faylı bir kırık modeline dayanan bir kayma modeli ters çözüm yöntemiyle ortaya çıkartıldı. Bu ters çözümlerle elde edilen InSAR ve GPS modelleri bir birine çok yakın basit kayma dağılımları vermektedir. Arazi gözlemlerine paralel bir şekilde en fazla kayma fayın ortasında olup fayın her iki ucuna doğru azalmaktadır. Ortak (joint) ters çözüm modeli dahil, bütün modeller hem InSAR verilerini hem de GPS verilerini çok iyi bir şekilde, bu verilerin hata payı mertebesinde açıklayabilmektedir (ortalama kare kök [RMS] hatası 2 santimetrenin altında). Ortak ters çözüm modeli 6.1×10^{19} Nm'lik bir sismik moment vermektedir, ki bu da sismolojiden elde edilen sismik momentlerle uyumlu bir değerdir. Modellemenin ikinci aşamasına daha karmaşık bir fay kırığı modeli kullanıldı, çünkü bu plaka sınırında uzun vadede yatay hareketin böylesine eğimli bir düzlem ile taşınmasını mekanik olarak açıklamak çok zordur. Bu nedenle tektonik ve sismik gözlemlere uygun ve birden fazla fayın aynı anda kırıldığı kırık modelleri oluşturuldu ve jeodezik veriler bu tip fay kırıkları kullanılarak modellendi. Kırık modeli bir birini kesen iki fay düzleminde oluşturuldu. Bunlardan birisinin düşey eğimli Düzce fayı, diğeri ise kuzeye doğru eğimli ve daha önceden var olan derinlerdeki yapısal bir süreksizlik düzlemi olduđu var sayıldı. Bu durumda Düzce depreminin eğimli düzlem üzerinde başladığı ve buradan yanlara ve yukarıya doğru yayılarak Düzce fayını tetiklediđi farz edilmektedir. Böylesi çoklu fay kırıkları kullanılarak elde edilen kayma dağılım modelleri de hem InSAR hem de GPS verilerini çok iyi bir şekilde açıklayabilmektedir. Bir diğeri deyişle, her iki fay modeli de jeodezik verileri bir birine yakın hata payları ile açıklayabilmektedir. Bu durum, jeodezik verilerin alansal dağılımının yetersizliđinden kaynaklanmaktadır. Bu nedenle, kırık geometrisinin daha iyi bir şekilde belirlenebilmesi için, özellikle fay kırığına yakın noktalarda, ilave veriye ihtiyaç vardır. Böylesi bir veri setinin yokluğunda, bu çalışmada, Kuzey Anadolu fayının uzun dönemdir bölgede mevcut olan aktivitesi göz önünde tutularak, çoklu fay kırığı modelleri yeğlenmektedir. Dolayısıyla, dünyanın bir çok yerinde daha önceden gözlemlendiđi gibi, Düzce depreminde de birden fazla fayın kırıldığı düşünölmektedir. Bazı farklara rağmen, böylesi karmaşık kırık modelleri kullanılarak elde edilen kayma dağılımları basit kırık modelleri kullanılarak elde edilen kayma dağılımlarına benzerlik göstermektedir. Çoklu-fay-kırığına dayanan kayma dağılım modelleri genellikle derinlerde daha yüksek kaymalar göstermekte ve dolayısıyla sismik moment daha yüksek çıkmaktadır (6.3 - 6.5×10^{19} Nm, $M_w=7.2$, ki bu değer halen sismik gözlemlerle uyumludur). Bütün modeller göstermektedir ki, hem InSAR hem de GPS verileri arazi gözlemlerinden yaklaşık 15 km daha uzun bir fay kırığını gerektirmektedir. Bu da, kısa yüzey kırığına rağmen neden Düzce depreminin bu

kadar büyük bir magnitüde sahip olduğunu açıklamaktadır.

1999 İzmit ve Düzce depremlerinin kaynak parametreleri elde edildikten sonra, bu depremlerden kaynaklanan Coulomb gerilim değişiklikleri hesaplandı ve bunlar statik gerilim etkileşimleri açısından yorumlandı. Coulomb gerilim hesaplamaları elastik yer değiştirme teorisine dayandığı için, bir depremin kaynak parametreleri o depremin meydana getirdiği Coulomb gerilim değişimlerinin doğruluğu konusunda en önemli rolü oynamaktadır. Dolayısıyla, içinde genellikle kaba kayma değerlerinin ve fay geometrisinin kullanıldığı daha önceki çalışmalara karşın, burada Coulomb gerilim hesaplamalarında InSAR ve GPS verilerinin modellenmesi sonucu elde edilen detay ve güvenilir deprem kaynak parametreleri kullanıldı. Marmara bölgesi civarında Kuzey Anadolu fayı üzerinde daha önceden meydana gelmiş dört büyük deprem ($M > 7$) kullanılarak, İzmit depremi öncesinde bölgedeki statik gerilim dağılımı ve bunun InSAR modellenmesinden elde edilen İzmit kırık düzlemi üzerine düşen (resolved) miktarı hesaplandı. Hesaplamalar, İzmit depreminin gerilmenin arttığı bir bölgede meydana geldiğini göstermektedir. İzmit depreminin Düzce depremine olan etkisini araştırmak için bu depremin çok faylı Düzce kırığı üzerine neden olduğu gerilim değişimi hesaplandı. Hesaplamalardan, İzmit depreminin Düzce depremi odağı civarında gerilimi 3-6 bar artırarak bu depreminin meydana gelmesine yardımcı olduğu sonucu çıkartılmaktadır. Şu anda, 1912 den buyana oluşan büyük depremler nedeniyle Marmara bölgesi gerilimin en fazla arttığı bir alanda bulunmaktadır. Özellikle 1912 ve 1999 depremleri Marmara denizi içerisinde bulunan Kuzey Anadolu fay sisteminin her iki ucunda gerilimi 5 barın üzerinde arttırmıştır. Bu bölgedeki faylar, dolayısıyla, özellikle İstanbul için ciddi bir deprem tehlikesi oluşturmaktadır. Tartışmalı olmasına rağmen, yüksek çözünürlüklü deniz tabanı verileri ve derin sismik profillere dayanan çalışmalar göstermektedir ki, Kuzey Anadolu fayı bu bölgede üç parçadan meydana gelmektedir. Bunun doğru olduğu varsayıldığında ve gelecekteki depremin bir veya en fazla iki parçalı kırktan meydana geleceği farz edildiğinde, şu anda cevap bekleyen soru depremin batıda mı yoksa doğuda mı meydana geleceğidir. 1939 Erzincan depreminden buyana Kuzey Anadolu fayı üzerindeki büyük depremlerin batıya doğru olan istikrarlı göçü düşünüldüğünde, bir sonraki depremin Batı Marmara'da meydana geleceği düşünülebilir. Ancak, bu soruya daha emin bir cevap verebilmek için Marmara içindeki faylar üzerinde 1912'den önce meydana gelen büyük tarihi depremlerin yerlerinin daha iyi bilinmesi gerekmektedir.

1. INTRODUCTION

In order to reduce the loss of human life and property in earthquakes, we need to learn as much as possible from every earthquake that occurs. In this context, this thesis focuses on studying the crustal deformation induced by the 1999 Marmara earthquake sequence using Synthetic Aperture Radar interferometry (InSAR), a relatively new and powerful remote sensing technique. SAR interferometry measures the phase difference between two images taken from different positions (by one antenna on different satellite passes or by two antennas mounted on an aircraft), corresponding to the change in the round-trip length of radar waves to the same ground point.

The displacement field induced by an earthquake is a direct manifestation of what happens at depth. Thus, from surface observations, one can derive the source parameters of an earthquake, such as the slip distribution along the fault, the geometry and location of the rupture. This information enlightens our understating of earthquake processes and can be used in seismic hazard assessments. Accordingly, after a brief summary and history of InSAR, and how it works, first, the coseismic surface deformation of the 17 August 1999 Izmit and 12 November 1999 Düzce earthquakes are mapped by InSAR. Then, in order to deduce the source parameters of the two earthquakes, the InSAR observations are modelled together with tectonic field observations and GPS measurements using elastic dislocations on rectangular planes embedded in homogenous and isotropic half space. And in the end, these source parameters are interpreted and used in Coulomb failure analysis to study stress interactions and earthquake hazard in the Sea of Marmara region.

1.1 Side Looking Radar (SLR) Imaging

1.1.1 Real aperture radar

Unlike optical and infrared imaging sensors, which relay on reflected or radiated energy (i.e. the Sun), radar is an active sensor, which provides its own illumination in the form of microwaves. Microwaves are electromagnetic waves in the 1-1000 GHz region of the electromagnetic spectrum (Fig. 1.1).

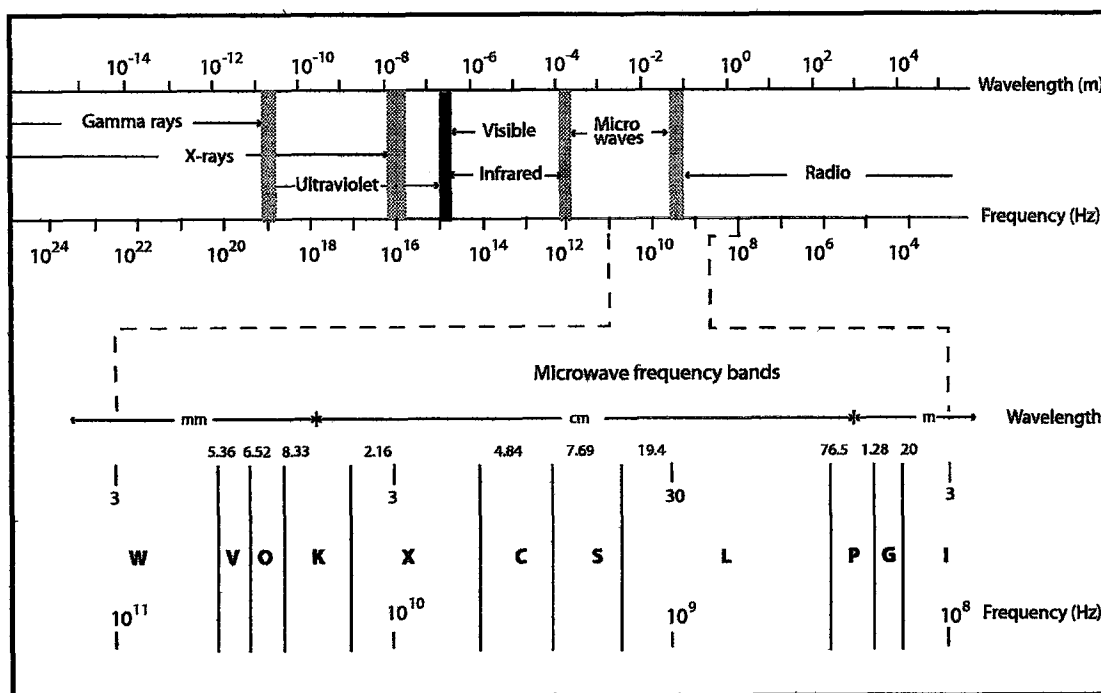


Figure 1.1 The electromagnetic spectrum and microwave frequency bands.

In radar imaging systems, a long antenna is mounted on a platform (aircraft [airborne system] or satellite [spaceborne system]) with longitudinal axis parallel to the flight direction (azimuth direction) (Fig. 1.2). Looking perpendicular to the flight direction (i.e. range direction), the antenna emits pulses of electromagnetic waves towards the Earth's surface and the echoes scattered from the targets are collected (Figure 1.2). The backscattered signal is stored in complex format. The term "complex" refers to complex numbers with their "real" and "imaginary" components. For example a wave might be described in complex format by $A \cdot [\cos(wt) + i \sin(wt)]$, where "w" represents the wave's frequency and "A" its amplitude (Fig. 1.3). The cosine value would describe the wave's real component, sine the imaginary component, and the two would combine as vectors to provide the wave's overall phase and amplitude. The real component and the imaginary component are sometimes called as the "I" (In-Phase) and "Q" (Quadrature, the 90 degrees shifted) data streams, respectively. The signal's phase is calculated as $\arctan(Q/I)$ whereas, the signal's amplitude is given by $\sqrt{I^2 + Q^2}$ (Fig. 1.3).

Both the cosine and sine components of backscattered SAR signals are measured and digitized on-board. The two resulting data streams are then transmitted to a ground station for further processing. Amplitude or phase images of the scene can then be constructed from the complex data.

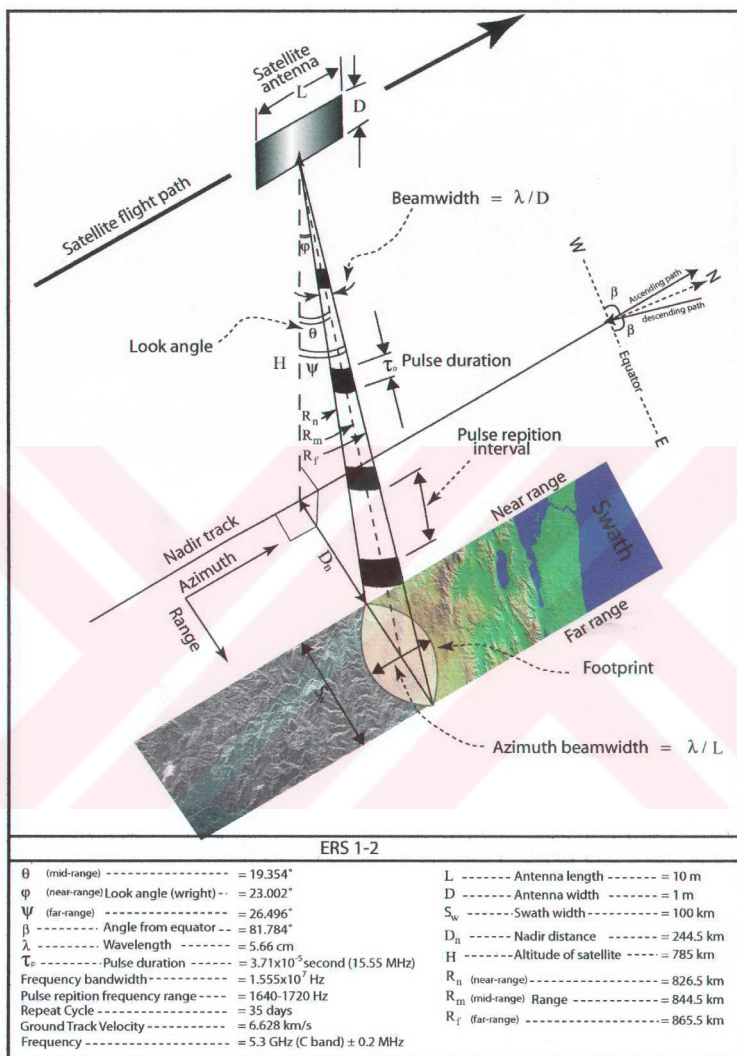


Figure 1.2 Geometry of a spaceborne side looking radar (parameters defined are for ERS satellites). A side-looking spaceborne radar system can map a continuous swath tens of kilometer (100 km kilometers, in the ERS1-2 case) in width as the satellite progresses along its orbit track.

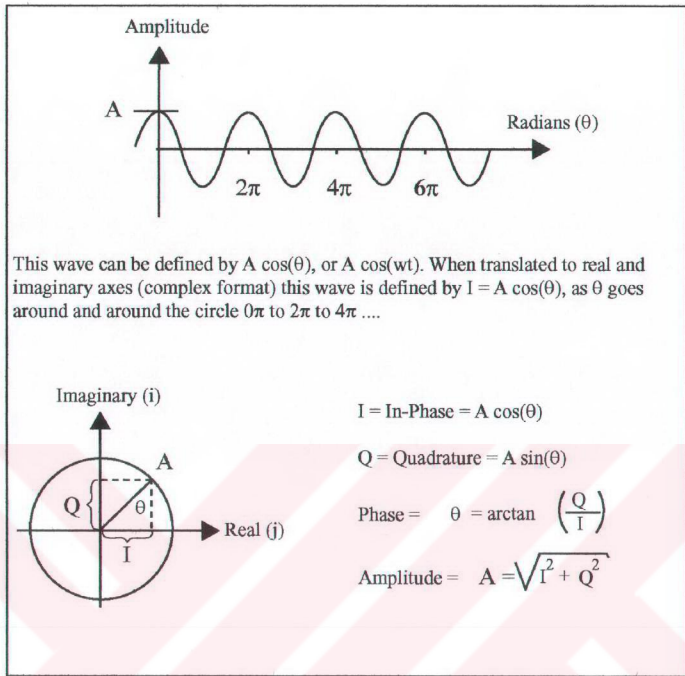


Figure 1.3 Representing waves in complex format.

Resolution of such radar images, defined as the minimum distance on the ground at which two objects can be imaged separately, depends on the antenna properties (length, width, and incidence angle), altitude of satellite, and radar pulse duration (Fig. 1.4). Resolution in the range and azimuth directions is not the same. The range resolution is limited by the pulse duration (Fig. 1.4a). The shorter the pulse duration the higher the resolution can be, but a shorter pulse duration would not produce a sufficient echo signal to noise ratio (SNR). Ground range resolution δR_g is

$$\delta R_g = \frac{c \tau_p}{2 \sin \eta} \quad (1.1)$$

where "c" is the speed of light (2.9979×10^5 km/s), " τ_p " is the pulse duration, and " η " is the incidence angle.

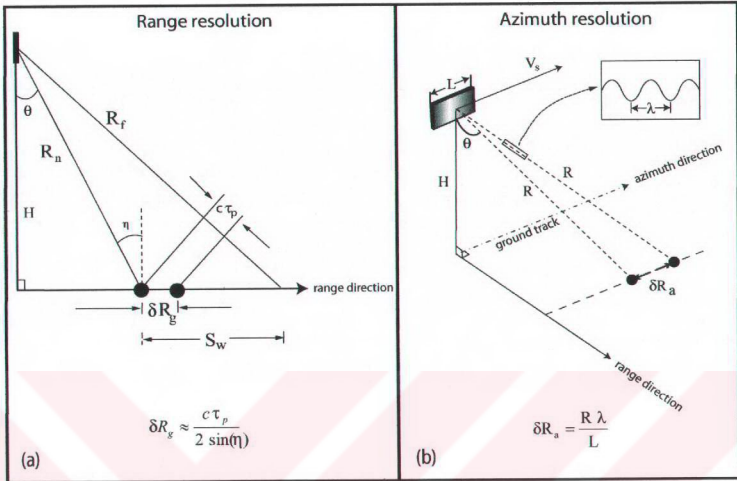


Figure 1.4 Resolution of a side looking radar in the range (a) and azimuth (b) direction. "H" is the satellite height, " R_n " is near range, " R_f " is far range, " η " is the incidence angle " c " is the speed of light, " τ_p " is the pulse duration, " δR_g " is the ground range resolution, " S_w " is the swath width, " V_s " the velocity of satellite, " θ " is the look angle, " λ " is the phase length, " R " is range, and " δR_a " is the azimuth resolution (adapted from Curlander and McDonough, 1991).

On the other hand, the azimuth resolution δR_a is

$$\delta R_a = \frac{R \lambda}{L} \quad (1.2)$$

where " R " is range, " λ " is wavelength of the radar signal, and " L " is the antenna length (Fig. 1.4b). Thus, the azimuth resolution is directly proportional to the length of the antenna; the longer the radar antenna the better the resolution.

Using a pulse duration of 37.1 microsecond, look angle of 23° , wavelength of 5.66 cm and mid range of 844 km (Figure 1.2), from the equations 1.1 and 1.2 the range resolution and the azimuth resolution for ERS radars on the ground are found to be 14.2 km and 4.8 km, respectively. In other words, the minimum pixel size of a real aperture radar image to resolve an object from ERS data will be 14.2×4.8 km, which is too big.

To improve both the range resolution and SNR to a level enough to be useful, the pulse compression technique is used together with a type of processing of the returned signal known as matched filtering (Curlander and McDonough, 1991).

The ground range resolution δR_g with this technique is

$$\delta R_g = \frac{c}{2 B_r \sin \eta} \quad (1.3)$$

where “ B_r ” is the frequency bandwidth of the transmitted radar pulse. Given a frequency bandwidth of 1.555×10^7 Hz and an incidence angle of 23° for ERS data, the equation 1.3 gives a ground range resolution of 24.6 m, 577 times more than the initial resolution (14.2 km) without signal processing.

1.1.2 Constructing a synthetic aperture radar

As shown above (equation 1.2), the length of the radar antenna determines the resolution in the azimuth direction of the image (along the flight direction): the longer the antenna, the finer the resolution in this dimension. For example, to obtain an azimuth resolution of 4 m with a radar range of 844 km and a wavelength of 5.66 cm, a ~12-km-long radar antenna is required (equation 1.2). Constructing such a long antenna is, of course, impractical. Here where the advanced signal processing techniques come in to the play and a long antenna is synthesized by combining signals (echoes) received by the radar as it moves along its flight track. Willy (1954) was the first to realize that the Doppler spread of the echo signal could be used to synthesize a much longer aperture to greatly improve the resolution of a side-looking radar. Aperture means the opening used to collect the reflected energy that is used to form an image. In the case of a camera, this would be the shutter opening; for radar it is the antenna. A synthetic aperture is constructed by moving a real aperture or antenna through a series of positions along the flight track (Fig. 1.5).

Taking into account the Doppler effect, the azimuth resolution δR_a will be

$$\delta R_a = \frac{L}{2} \quad (1.4)$$

This implies that the better resolution is in fact obtained with a smaller antenna, not what one would expect as this is the opposite to the rule for real aperture radar. Thus, the ground azimuth resolution is equal to half the antenna size and hence is 5 m for ERS SAR images. The expression (1.4) is not strictly true since a constant

Doppler frequency shift is assumed. After advance signal processing (range compression, range migration and azimuth compression), range and azimuth resolutions of ERS images are reduced to 20 m and 4 m on the ground, respectively.

Further technical and mathematical details of SAR and SAR processing can be found in Curlander and McDonough (1991), and in Rodriguez and Martin (1992).

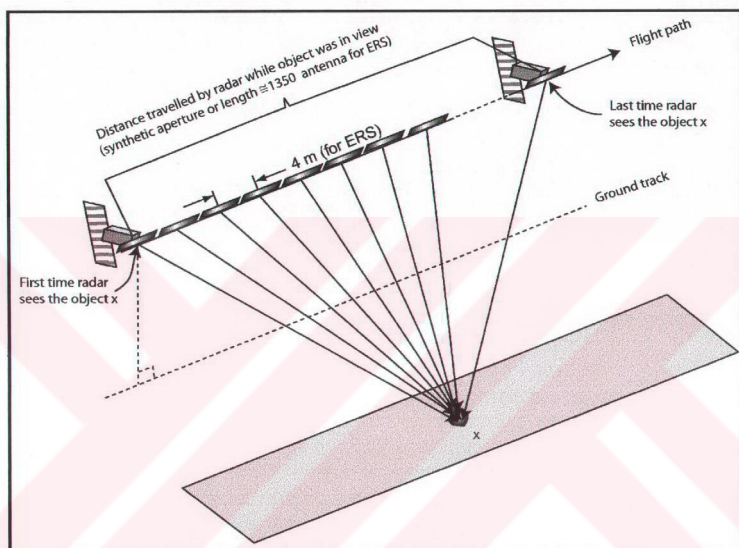


Figure 1.5 Schematic illustration of forming a synthetic aperture. As the spacecraft passes over a target, a ground target appears to be first in front of, then next to, then behind the spacecraft. In the meantime, the SAR has sent out many (for ERS, about 1350) pulses and therefore recorded a specific target's radar backscatter response about 1350 times. The response to each one of the 1350 pulses will be somewhat different, depending on changes in target-sensor geometry and Doppler effects. Just as the Doppler effect changes the sound one hears as a car passes, the motion of the spacecraft relative to the target increases or decreases the signal's frequency. By analyzing the return signals from those 1350 pulses, the target's Doppler history can be determined. This information allows a target's backscatter to be analyzed as if it had been seen by 1350 different antennas, or correspondingly of a synthesized antenna with length equal to the distance the spacecraft passed through while it was able to get backscatters from that target. This large synthesized antenna length significantly improves the resolution.

1.2 Synthetic Aperture Radar Interferometry

1.2.1 Introduction

The term interferometry is derived from the word interference. Interference is a phenomenon that occurs when one has waves of any kind such as, sound, light, ocean, electromagnetic, seismic waves etc. Interference occurs whenever two waves come together. InSAR is a method in which the phenomenon of interference is combined with synthetic aperture radar. A radar interferometer can be formed by relating the signals from two spatially separated radar antennas. The two antennas may be mounted on a single platform (for aircrafts usually) (single pass interferometry), or a synthetic interferometer may be realized by utilizing a single antenna on a satellite in a nearly exact repeating orbit, in which case the baseline is formed by relating radar signals on repeat pass over the same scene (repeat pass interferometry).

Conventional SAR systems provide a two-dimensional map of the radar reflectivity of the illuminated scene. While complex data are collected and processed to produce the SAR image, one of the final steps is to reduce a complex image (containing both magnitude and phase information) to a purely magnitude image, with the phase information being discarded. InSAR technique, on the other hand, uses the phase information in two SAR images by calculating the phase difference between each pair of corresponding image points after precisely aligned to a fraction of a pixel width. The resulting new image is called an interferogram. The interferogram is an interference pattern of fringes due to relative phase difference. It is effectively a contour map of the change in the distance from the radar to the ground surface. Each cycle of phase, or fringe, in the resulting interferogram corresponds to a change in range distance from the satellite or aircraft to the ground surface equal to one-half of the radar wavelength ($5.66/2=2.83$ cm for ERS). Since the launch ERS1 by European Space Agency ESA in 1991, the topic of Interferometric processing became a popular topic issue because SAR images collected by imaging radar systems on board, airplane or satellite platforms can be used to map the elevations, movements, and changes of the Earth's surface very accurately.

1.2.2 Historical background

Although precise measurements using the phenomenon of interference were started in the late 19th century and widely used in physics and astronomy as from the beginning of the last century, it was not until 1960s that this phenomenon was first

applied to radar (acronym for RAdio Detecting And Ranging) images, two decades after its discovery by Robert Alexander Watson-Watt. In 1881, Albert A. Michelson showed through mathematics and experiments that the interference of two beams of light from a distant source provides detail far greater than simply focusing the light with a single lens or mirror (Fig. 1.6). Michelson was an American physicist who received the Nobel Prize in 1907 for his measurements of the speed of light using his interferometer. In 1920, he and his colleagues used interferometry to measure the diameter of the first star outside our solar system. In 1947, radio telescopes first used interferometry to reveal details of astronomical objects. This was possible with radio wavelengths long before visible wavelengths because the required instrument-precision is related to the size of the wavelength. Radio wavelengths are

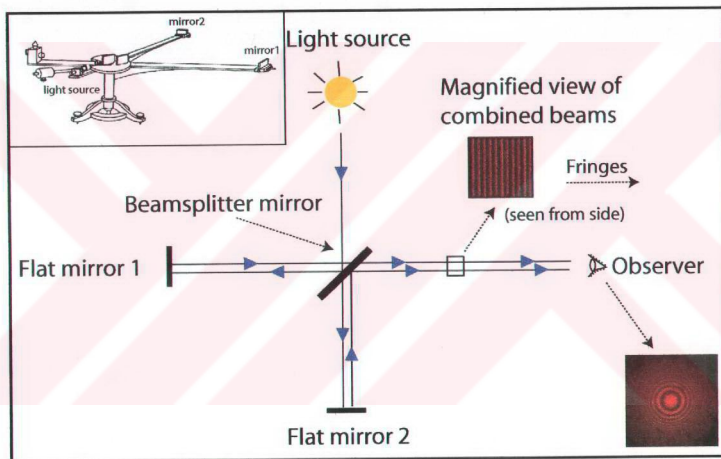


Figure 1.6 Schematic illustration of the first interferometer designed by Albert A. Michelson, a German American physicist (1852-1931). Inset is a drawing of the original interferometer. The light travels to a beam-splitter mirror where it is split into two beams, generating a path difference between them. One beam is transmitted through the beam-splitter to mirror-1 and the other beam is reflected at 90 degrees to mirror-2. Both mirrors should then reflect their respective beams back to the beam-splitter and strike the beam-splitter at the original incident beam's position. Part of mirror-1's reflected beam will then be reflected by the beam-splitter to the observer and part of mirror-2's reflected beam will be transmitted by the beam-splitter to the observer. Because of the introduced path difference, both beams have different phase when they interfere and thus fringes will form all along the optical axis of the combined beams, oriented perpendicular to this axis and will appear to stand still, even though the beams are traveling at the speed of light. To the eye, the fringes appear as alternating small rings of light and dark surrounding the central images of the light source.

approximately 200,000 times larger than optical wavelengths. Therefore, necessary accuracies of a tenth of an inch in radio astronomy convert to accuracies of a millionth of an inch at optical or visual wavelengths. Optical interferometry had to wait for the development of fast computers and lasers to measure and correct for the millionth-of-an-inch changes in distance.

First applications of SAR interferometry were used to study the surface of Venus and the Moon (Rogers and Ingalls 1969, Campbell et al., 1970; Shapiro et al., 1972; Zisk 1972a,b). Graham was the first to introduce the technique for topographic mapping in 1974 by using an airborne SAR system configured as a cross-track or vertical interferometer. He used two vertically separated antennas to receive simultaneously backscattered signals from the terrain. Vectorial addition of these signals produced a pattern of nulls corresponding to predetermined depression angles, which, when used in conjunction with range information, yielded elevation information. He recorded data optically from two channels: one was the normal SAR data; the other, the interferometer output containing the null patterns. He showed that since the multiple nulls were ambiguous, the elevation of at least one point within the scene must be determined by an alternate means to resolve the absolute elevation.

A decade after Graham's stimulating work, first detail and accurate topographic mapping was realized by Zekber and Goldstein (1986). A year after this work, extending the technique described by Raney (1971), the same scientists (Goldstein and Zekber, 1987) were also able to map the outgoing tide flowing out of San Francisco Bay to a velocity resolution of $\sim 4 \text{ cm s}^{-1}$ using two along-track parallel antennas mounted on an aircraft.

In 1988, A revolutionary idea came from Goldstein et al. who, for the first time, showed that repeat pass interferometry can be used to measure topography very accurately. Using two SEASAT images three days apart, Goldstein et al. (1988) mapped the topography of the Cottonball Base in Death Valley, California. The resulting topographic map was in good agreement with the published maps.

Soon after this work, Gabriel et al. (1989) introduced the differential synthetic aperture radar technique for mapping subtle surface changes. Using 3 Seasat images, Gabriel et al. (1989) were able to detect land swelling due to watering in Imperial Valley, California with sub centimeter accuracy and high spatial resolution over a wide area. Later on after the launch of ERS1 satellite in 1991 by European Space Agency (ESA), this pioneering research has inspired many scientists to use InSAR for mapping the movements of the Earth's surface such as earthquakes,

volcanoes, glaciers, landslides, salt diapirism, groundwater and petroleum extraction, and underground explosions (owing to ESA's policy of supporting scientific research). Although the satellite was designed for sea surface observations only, numerous papers on InSAR applications with the ERS1 data were published. After the launch of ERS2 satellite in 1995 (a replica of ERS1 satellite), the opportunities of spaceborne SAR interferometry was further broadened by the use of ERS1 and ERS2 in the tandem mission. This provided interferometric data acquired only one day apart until the ERS1 satellite died in March 2000 operating 6 years longer than its designed life time of 3 years.

In 1993, Massonnet et al. (1993) pioneered the use of InSAR to measure co-seismic deformation caused by the 1992 Landers, California earthquake. Since then, many earthquakes have been studied using SAR interferometry (Table 1.1).

1.2.3 Principles of InSAR: how it works

Here, some basic principles of InSAR are provided without getting into comprehensive mathematical and technical details for those who are unfamiliar with the method. Further details can be found in a number of review papers on the theory and applications of InSAR (Gens and Vangenderen, 1996; Bamler and Hartl 1998, Massonnet and Feigl, 1998, Bürgmann et al., 2000; Rocca et al., 2000; Bamler, 2000).

Interferometric data can be acquired in two ways; (1) by arranging two antennas on an aircraft or a space shuttle, or (2) by utilizing a single antenna on a satellite in a nearly exact repeating orbit (Fig. 1.7). The former is called single-pass interferometry and the latter repeat-pass interferometry. In single-pass interferometry, antennas can be separated parallel to the flight direction to map water currents and moving objects such as glaciers and ice-sheets (along-track interferometry) (Goldstein and Zekber, 1987; Goldstein et al., 1989; Laurence et al., 1994; Bao et al., 1997, 1999), or alternatively the antennas can be mounted perpendicular to the flight direction to measure local or global topography (across-track interferometry) (Zekber and Goldstein, 1986, Shuttle Radar Topography Mission in 2000).

Providing that the platform carrying the radar system flies along nearly the same flight path with a slightly different viewing geometry and the flight path is known very precisely, only one antenna can be sufficient to form a radar interferometer as well

Table 1.1 Earthquakes studies using SAR interferometry

1992	Landers, California, Ms=7.3	Massonnet et al., 1994, 1996a; Zebker et al., 1994; Hernandez et al., 1997; Hernandez et al., 1999
1992	Fawnskin, California, Ms=5.1	Feigl et al., 1995
1992	Little Skull Mountain, Nevada, Ms=5.3	Lohman et al., 2002
1993	Eureka Valley, California Mw=6.1	Massonnet and Feigl, 1995a; Peltzer and Rosen, 1995
1994	Northridge, California Mw= 6.7	Massonnet et al., 1996b; Murakami et al., 1996
1995	Grevena, Greece, Ms=6.6	Meyer et al., 1996
1995	Corinth, Greece, Ms=6.2	Bernard et al., 1997
1995	Hyogo-ken Nanbu, Japan Mw= 7.2	Ozawa et al., 1997
1995	Dinar, Turkey, Ms=6.1	Wright et al., 1998
1995	Neftegorsk, Russia, Ms=7	Tobita et al., 1998
1995	Nuweiba Aquaba, Israel, Ms=7.1	Baer et al., 1999; Klinger et al., 2000
1996	Pyrenean, France, Ml=5.2	Rigo and Massonnet, 1999
1997	Manyi, Tibet, Mw =7.6	Peltzer et al., 1999
1997	Kagoshima, Japan, Mw=6.1	Fujiwara et al. 1998
1997	Umbria-Marche, Italy, M=5.7 & 6.0	Salvi et al., 1999; Stramondo et al., 1999
1998	Fandoqa SE Iran, Ms=6.6	Berberian et al., 2002
1998	Iwate, Japan, Ms=6.1	Fujiwara et al., 2000; Nishimura et al., 2001
1999	Izmit, Turkey, Mw=7.4	Armijo et al., 2000, 2000; Cakir et al., 2001; Wright et al., 2001a; Delouis et al., 2002; Feigl et al., 2002
1999	Athens, Greece, Mw=5.9	Kontoes et al., 2000
1999	Düzce, Turkey, Mw=7.2	Bürgmann et al., 2002
1999	Hector Mines, California, Mw=7.1	Sandwell et al., 2000; Fialko et al., 2001; Peltzer et al., 2002
2000	Iceland, Ms=6.6	Pedersen et al., 2001

(i.e. repeat-pass interferometry). As the satellites are more stable and their position can be determined very accurately compared to aircrafts, the repeat-pass method is thus more suitable for satellites. (Fig. 1.8). The distance between two flight paths (i.e. orbits for satellites) is called base line (Fig. 1.8). For every SAR imaging system, there is a critical baseline length, above which interferometry becomes impossible (Zekber an Villasenor, 1992). The critical base line B_c is

$$B_c = \frac{\lambda R}{2 \delta R_g \cos \theta} \quad (1.5)$$

Taking 5.66 cm of wavelength, 20 m of ground range resolution, 23° of look angle, and 844 km of range, for ERS radar data the equation (1.4) gives a critical baseline of ~ 1300 m.

The geometry of repeat-pass interferometry is illustrated in Figure 1.8. Because the satellite does not fly along exactly the same orbital path it followed in the previous pass, a shift in the imaging swath occurs, and thus a target on the ground surface is viewed from two different points of view, providing a stereo couple of radar images. This, in turn, results in phase difference between two rays paths from satellite to the same target on the surface. The phase difference ϕ is given by

$$\phi = \frac{4 \pi}{\lambda} \delta R \quad (1.6)$$

where " δR " is the range difference. In other words, the phase difference is equal to 2π times the round-trip distance difference in wavelengths. Using law of cosines, path difference δR can be express in terms of the imaging geometry as follows

$$(R + \delta R)^2 = R^2 + B^2 - 2R B \sin(\theta - \alpha) \quad (1.7)$$

where " B " is the baseline length, and " α " is the base line orientation angle (Fig. 1.8). If we neglect the term of order $(\delta R)^2$ the expression (1.7) will be

$$\delta r \approx B \sin(\theta - \alpha) + \frac{B^2}{2R} \quad (1.8)$$

In the case of spaceborne geometries, the second term on the right is not significant and thus can be ignored as well. Then the expression (1.7) will be

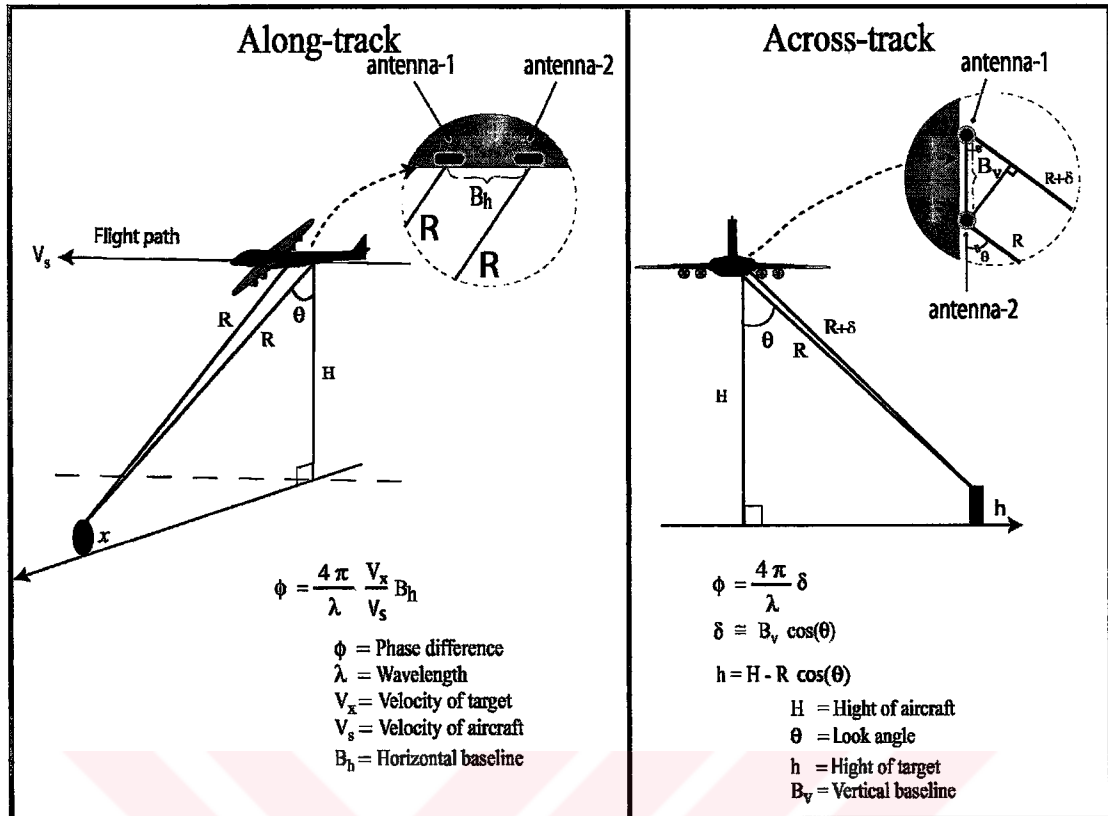


Figure 1.7 Geometry of single-pass interferometry. (a) Along-track geometry (perspective view) used for mapping velocity field of ocean and sea waves and mapping moving objects such as ice-sheets and glaciers. (b) Across-track geometry (in a plane perpendicular to the flight direction) used to construct digital elevation models. While the phase difference occurs due to movement of the object during the measurement in the along-track interferometry, the phase difference in the across-track interferometry is due to the elevation differences in the imaged area.

$$\delta R \approx B \sin(\theta - \alpha) \quad (1.9)$$

As shown in Figure 1.8, the component of baseline parallel to the look direction $B_{||} = B \sin(\theta - \alpha)$. Thus the path difference is

$$\delta R \approx B_{||} \quad (1.10)$$

Equations (1.6) and (1.10) show that the phase difference is the parallel baseline component in wavelength multiplied by two for round-trip travel. From the geometry shown in Figure 1.8, elevation of a target "h" can be given by

$$h = H - R \cos \theta \quad (1.11)$$

where "H" is the altitude of the satellite. ERS scenes are about 100 km of length along azimuth and acquired by the satellite in about only 15 seconds. Thus,

because the altitude of the satellite should remain almost the same in such a short time and the look angle does not change, it is clear that the instrument is sensitive to the topography. Topographic sensitivity of the instrument, or so called “the altitude of ambiguity” h_a is defined as the round-trip range variation equal to the radar wavelength (Massonnet and Rabaute, 1993) and given by

$$h_a = \frac{R \lambda \tan(\theta)}{2 B_h} \quad (1.12)$$

where “ B_h ” is the horizontal component of the baseline (Fig. 1.8). Thus, the topographic sensitivity depends on the horizontal distance between the two orbits. Altitude of ambiguity is simply the elevation difference that leads to a full wavelength (2π) change in phase, i.e, the elevation difference between adjacent fringes in an interferogram. For example, if the altitude of ambiguity is 100 m, there will be 10 fringes in an interferogram of a region with elevation up to 1000 m. Therefore, the smaller the altitude of ambiguity the more number of topographic fringes that an interferogram contains.

Equation (1.6) assumes that the imaged area and the atmospheric conditions, which effect the signal propagation, remain exactly the same between the two data acquisitions. If the same atmospheric conditions prevail, but an elevation change δd occurs due to phenomena such as earthquakes and subsidence in the interim of two acquisitions, the phase difference will be

$$\phi = \frac{4 \pi}{\lambda} \delta R + \frac{2 \delta d}{\lambda} \quad (1.13)$$

Because this study deals with surface change detection, hereafter the topographic phase component to the phase difference (i.e. the first term on the right in the equation 1.13) will be considered as an artefact that should be removed from interferograms. There are two ways of removing the topographic phase (Massonnet and Feigl, 1998). The first method uses a single interferogram, and thus it is called two-pass interferometry. In this method, the topographic phase is simulated using a

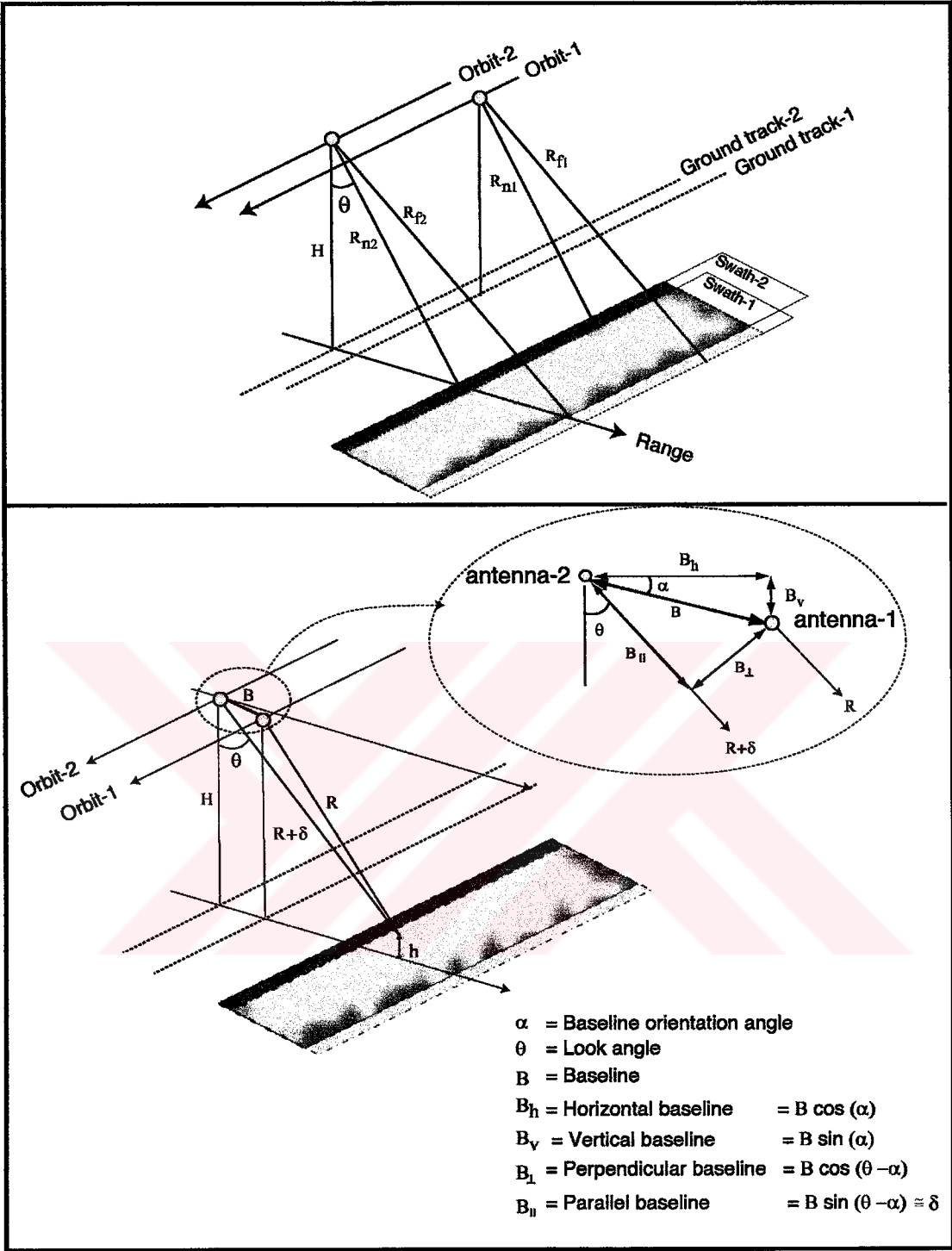


Figure 1.8 Geometry of spaceborne repeat-pass interferometry. Upper panel illustrates imaging geometry during two satellite passes. Note the exaggerated shift in the swath due to the different viewing geometry resulting from orbital deviations. Lower panel illustrates how the same target on the surface is imaged during two different passes. The blown up inset figure shows the baseline separation (in a plane perpendicular to the orbit of satellite) between two orbits and its various components commonly used elsewhere and in the text.

digital elevation model (DEM) and subtracted from the interferogram (Massonnet et al., 1993). The second method utilizes a second interferogram, and thus it is called three-pass interferometry (Zekber et al., 1994). In the three-pass interferometry, a second interferogram which contains no significant surface change is subtracted from the interferogram after scaling the data by the ratio of the parallel components of the baseline. As discussed in detail by Massonnet and Feigl (1998), each method has some advantages and disadvantages over each other.

1.2.4 InSAR processing: producing interferograms

In this study, the two-pass interferometry method is used, and thus only the InSAR processing with this technique will be addressed here. A flow chart in Figure 1.9 shows the steps in constructing interferograms. Two complex SAR images are required to calculate an interferogram. The two images can be ordered as raw data or single-look complex (SLC) images. For proper interferometric combination, the two images must be focused identically (Massonnet and Feigl, 1998). For example, the Doppler centroid used in focusing the both images must be the same. Therefore, it is safer to construct complex SAR images from raw data, instead of using preprocessed SLC images. To increase the signal-to-noise ratio, the pixel size of the images are often increased by averaging the neighboring pixels, a process called complex multi-looking. By doing so, a square shape of pixel can be obtained. For example processing 2 looks along range and 10 looks along azimuth will lead to a pixel size of 40x40 m for an ERS SAR image that has a pixel size of 4 m along azimuth and 20 m across (as performed in this study). The shifts and distortions between the two complex radar images, which arise mainly from different viewing angle, starting time and inherent system limitations, must be evaluated (Gabriel and Goldstein, 1988; Massonnet and Feigl, 1998). Each pair of corresponding image points must be precisely aligned to a small fraction of a pixel width (< 1 m), both in the range and in the azimuth direction. To achieve this, several correlation techniques are performed using usually the conventional amplitude images. This way, one of the images (slave image) is resampled and registered on the geometry of the other unchanged image (master). The master image can also be geocoded by correlating it with an amplitude image simulated from the digital elevation model, which allows to some data filtering (Massonnet and Feigl, 1998). Once the two images are co-registered very accurately, the next step is to form the interferogram by differentiating the phase component of each complex pixel of the master image from the corresponding pixel in the slave image. The phase in the interferogram records the topography and the deformation accumulated between two observation

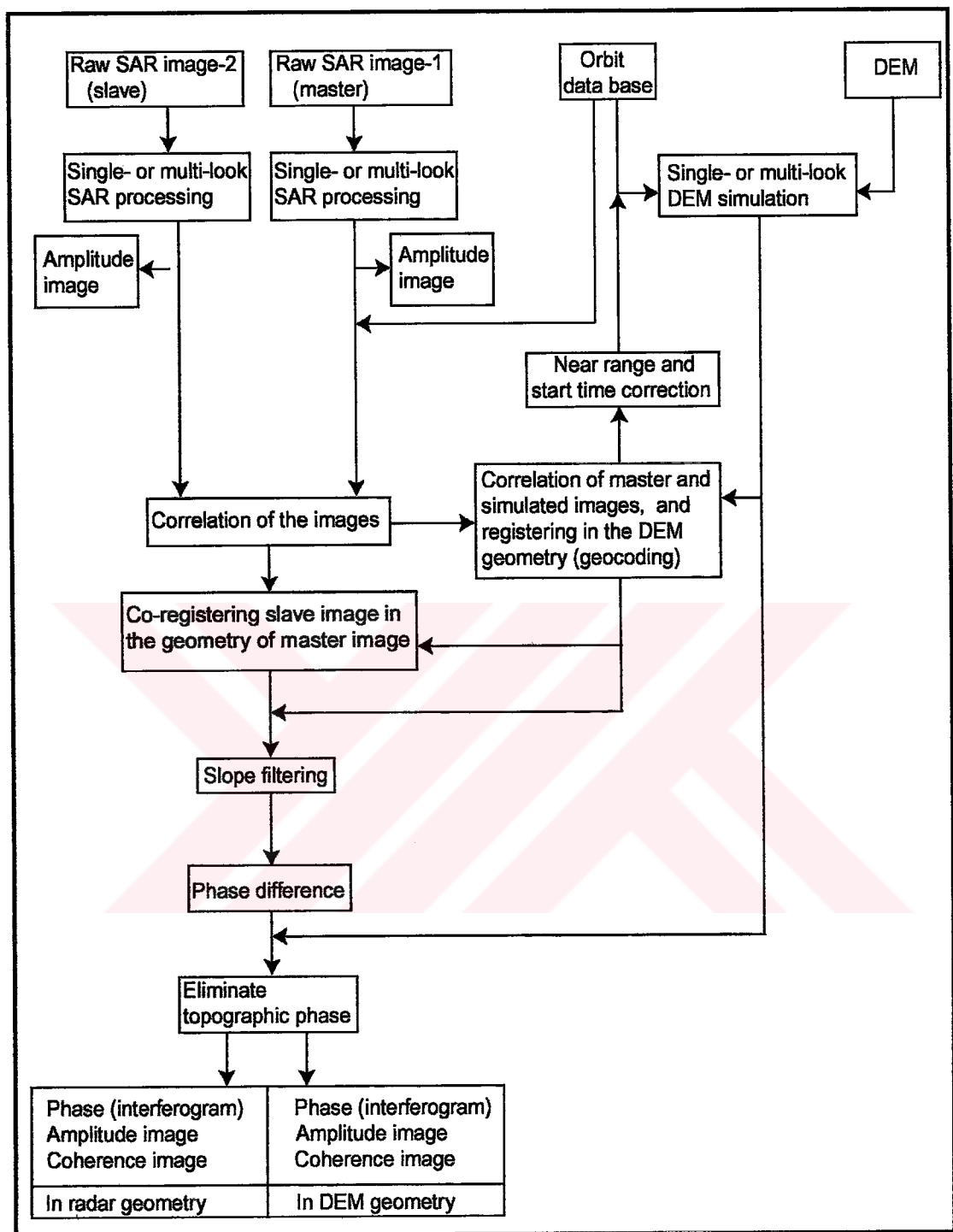


Figure 1.9 Schematic steps in two-pass SAR interferometric processing to map surface changes. This is chain of InSAR processing performed by the CNES's DIAPASON software package. In other InSAR processing packages, the order of steps such as geocoding may differ and some steps like slope filtering may not be performed.

times. Thus, the topographic contribution must be removed from the interferogram to reveal the deformation. To do this, the topographic phase is simulated using a DEM and orbit data, and then it is subtracted from the interferogram. The remaining signal in the interferogram is practically a contour map of surface deformation, expressed with fringes (a complete cycle of color, i.e., from blue to blue), each showing a full cycle of phase change (i.e. one wavelength, 5.66 cm for ERS). Because the round-trip range is measured, each fringe represents half a wavelength surface change (2.83 cm for ERS) along the radar line of sight.

What is measured by radar is not the absolute range change, but only the relative phase change between 0 to 2π . Thus, two targets will appear at the same phase if their ranges differ by an integral number of wavelengths. It is therefore necessary to find the multiples of 2π that disappeared or "wrapped". For example, if there are 10 concentric fringes due to a land subsidence in an interferogram, the maximum LOS subsidence in the center will be 28.3 cm (10×2.83) for ERS (Fig. 1.10). This assumes that the outermost fringe is actually the first fringe, beyond which there is no clear deformation. In other words, to ensure that we multiply the fringes with their correct integer number and thus to measure the absolute deformation, there must be an area of nominal deformation in the interferogram. When unwrapped an interferogram shows an obsolete and continues measure of LOS changes across the imaged area, providing again that the first fringe seen in the interferogram is the actual first fringe. If the scale of deformation to be measured is much higher than the scale of the interferogram, the first fringe may not be determined certainly, in which case the interferogram can still be unwrapped, but the measurements will not be absolute and thus additional observations (e.g. GPS) are needed to adjust the data. Unwrapping can be performed manually by digitizing the fringes and multiplying by 2π (contour line unwrapping) (Wright et al., 1998), or automatically by using several unwrapping algorithms (regional 2D unwrapping) (as discussed in detail by Gens and Vangenderen, 1996).

After unwrapping, fringes are sampled for modelling. As mentioned earlier, InSAR measures only the scalar change between the radar and the ground surface, which is equal to the component of the displacement in the LOS direction. Thus, the scalar change ΔR is

$$\Delta R = \bar{u} \cdot \hat{s} \quad (1.14)$$

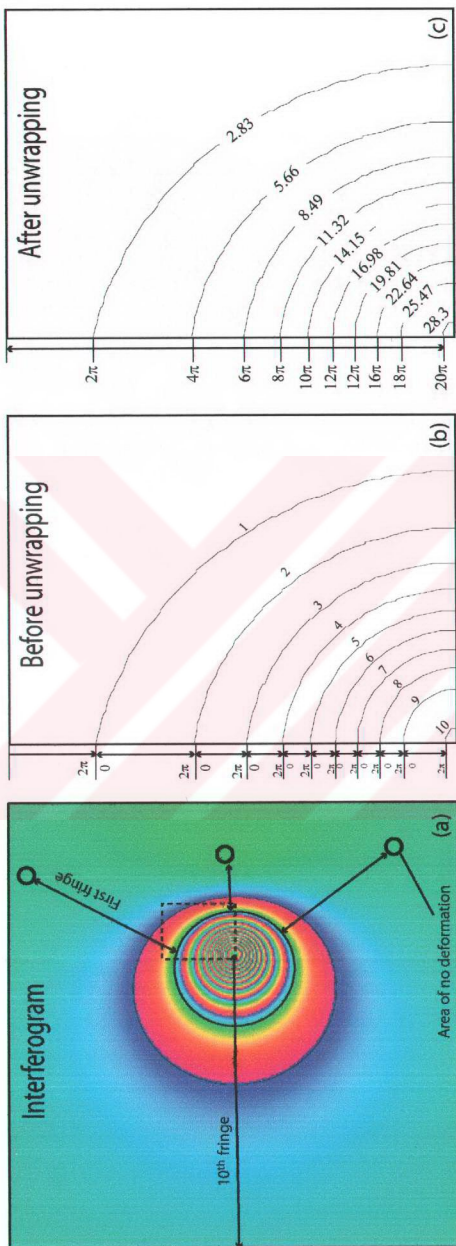


Figure 1.10 Unwrapping interferograms. An interferogram assumed to be obtained after InSAR processing is shown in (a). As seen in the interferogram, the area of no or nominal deformation is seen in the far field and is in green, which means that the fringes can be counted from green to green. Thus, the first fringe starts in the far field and ends in the next green towards the center of fringes. This way, 10 fringes can be counted in the interferogram (b). The interferogram is wrapped, meaning that, as illustrated in (b), each fringe shows a full cycle of phase (0 to 2π , i.e. one wavelength) that corresponds to half a wavelength (2.83 cm for ERS) relative surface change along the radar line of sight (LOS). In other words, two places of the same deformation will appear at the same phase if their ranges differ by an integral number of wavelengths. It is therefore necessary to find the same deformation will disappear or "wrapped". When unwrapped, an interferogram shows an obsolete and continues measure of LOS changes across the imaged area (c), providing that the first fringe seen in the interferogram is the actual first fringe. If the first fringe cannot be determined certainly (which might happen when the interferogram covers only a small area in the deformed region), the interferogram can still be unwrapped, but the measurements will still be relative.

where " $\vec{u}$ " is the displacement vector, and " $\hat{s}$ " is the unit vector (east, north, up) pointing from ground to satellite. The point vector depends on the viewing geometry, which can be calculated from orbit data.

For example, the unit vector is (-0.445 0.092 0.891) for ascending orbits around the Izmit area. As seen, the radar is almost insensitive to the displacement in the North-South direction, which is because InSAR cannot see any motion parallel to the flight direction that is nearly N-S. It is dominantly sensitive to vertical motion (~ 90%). Assume that the GPS measurements of a site near Izmit show that the land has moved 2 cm in the east-west direction, 2 cm north-south direction, and 4 cm in the vertical direction (2 2 4). Projecting this vector into radar look direction will give 2.81 cm displacement $(-0.445 \cdot 2 + 0.092 \cdot 2 + 0.891 \cdot 4)$. As the round-trip range measured by the radar, the phase shift will be two times of this change, i.e., 5.62 cm, producing about one fringe in an ERS interferogram.

1.2.5 Factors influencing InSAR measurements and data selection

Several factors control the accuracy of the measurements and the quality of the interferometric data. Some of the important factors are atmosphere, orbits, temporal decorrelation, baseline decorrelation, and topography. As will be discussed in detail in Chapter 3, the most limiting factor to InSAR measurements is the phenomenon of atmospheric artefacts. While avoiding and/or removing some sources of error is possible in some cases, some errors can not even be detect or solved in other cases.

1.2.5.1 Orbits

As described above, a radar interferometer is realized by two antennas that illuminate the same target on the ground surface from two different points. The spatial separation between the antennas (i.e, baseline) leads to a shift in the phases coming from the same targets. The phase difference can be calculated if the parallel component of baseline is known (Equation [1.6]). Therefore, precise knowledge of orbits is necessary. Although the position and the velocity of the satellite at a particular time are included in the headers files of SAR data, more precise orbit information based on additional observations are usually needed. The precise orbits are available at ESA and the Delft Institute for Earth-Oriented Space Research. If the orbits are not determined precisely and the image shift due to baseline separation is more than half a wavelength (i.e. one fringe) per pixel, hundreds of orbital fringes will occur and thus no interferometric effect will be seen. If the orbital

error is less than 1 m, then fringes running roughly parallel to the flight direction will appear in the interferogram. These orbital fringes have a constant gradient, and thus they can be removed using a plane of fringes with an opposite sign.

1.2.5.2 Atmosphere

Relative delay in the radar signal will occur if atmospheric conditions such as pressure, temperature and, principally, water vapor, are not the same at time of the acquisition of the two radar images. The delay, in turn, leads to phase shifts, and thus atmospheric fringes, or so called atmospheric artefacts, contaminate the interferogram. Atmospheric artefacts due to heterogeneous atmosphere appear locally and are difficult to detect as they interfere with the deformation signal. If the changes are horizontally homogenous, as will be discussed in Chapter 3, the artefacts can be revealed and removed. It is advisable to check meteorological data before ordering SAR data as the atmospheric effects can severely corrupt the interferograms.

1.2.5.3 Temporal decorrelation

Temporal changes in the physical properties of the targets in the scene lead to changes in the reflectivity characteristics of the targets. This, in turn, gives rise to decorrelation between the phases of the two images. Decorrelation or so called “incoherence” demolishes the quality of the interferograms, leading to random phase changes in each pixel in a given area where fringes are destroyed. Some of the main sources of decorrelation are the random displacements of the targets due to erosion, vegetation grow, cultivation, and in particular changes in the water content of the targets. Decorrelation can occur in the time period ranging from days to years depending on the surface characteristics. Correlation can be preserved for a long time (6-7 years) in the regions with bare rock, sparse vegetation, (such as volcanoes and deserts), arid climate and urban settlements. Complete decorrelation however may occur even within one day in some regions such as forests, marsh, cultivated lands and glaciers. Therefore, choosing a good temporal baseline (i.e. the time difference between the two SAR images) depends on the surface characteristics of the area interested and varies for different applications. While for digital elevation modelling a short temporal base line (usually ERS tandem pairs, i.e. one day apart) is required to avoid or minimize the decorrelation, long temporal baselines are necessary to reveal the accumulation of deformation in regions with tectonic or volcanic activity. Interseismic studies along the NAF, which are not included in this thesis, show that coherence is generally lost after two years in east-northeast of

Turkey. In regions with considerable seasonal changes like Turkey, it is usually better if both images were acquired in the same season, and preferably in summer to minimize the temporal decorrelation.

1.2.5.4 Baseline decorrelation

As mention above, there is a critical baseline length, above which interferometry becomes impossible (equation [1.5]). Decorrelation increases with increasing baseline. Because the baseline determines the topographic sensitivity and hence the altitude of ambiguity of an interferogram (equation [1.12]), choosing good baselines depends on the applications. In digital elevation modelling, image couples with a small altitude of ambiguity (50-100 m) are necessary to reveal the details of the topographic variations. In surface change studies with two-pass interferometry, on the other hand, couples with a high altitude of ambiguity are preferred so that residuals of the topographic fringes due to the errors in the DEM used are not significant. For example, a 100 m of error in the DEM used to remove the topographic contribution will cause to one full fringe as a topographic artefact if the altitude of ambiguity is 100 m. Small altitude of ambiguity of the image pair in two-pass interferometry will lead to reduction and sometime lost of coherence if the DEM is very rough.

1.2.5.5 Topography

In deformation studies with the two-pass method, as mentioned before, one of the radar images is geocoded by using various the correlation techniques and a DEM is simulated to remove the topographic phase from the interferogram. In the process of co-registration of the radar and simulated images, the area imaged must have some clear topographic features in order to obtain a good correlation and thus more accurate results. Misregistration of these images may happen in flat areas, which leads to decorrelation. Interferometry in areas with strong topography is very good for precise image registration, but steep slopes also lead to decorrelation. Therefore, if possible, regions with smooth topography should be chosen, which, as will be discussed in Chapter 3, also minimizes the possibility of the presence of atmospheric artefacts correlated with topography.

2. KINEMATICS OF THE AUGUST 17, 1999 IZMIT EARTHQUAKE DEDUCED FROM SAR INTERFEROMETRY

2.1 Introduction

The disastrous August 17, 1999 Izmit earthquake (Mw 7.4 from long-period waves) ruptured a portion of the plate boundary between Anatolia and Eurasia along the North Anatolian fault (NAF) (Fig. 2.1). The event was preceded by a sequence of six large earthquakes that ruptured progressively the NAF in the 20th century (Barka, 1996; Barka and Kadinsky-Cade, 1988; Stein et al., 1997; Nalbant et al., 1998). The Izmit event was also followed, three months later on November 12, 1999, by another destructive earthquake (Mw 7.2) that ruptured the neighboring Düzce fault, east of the Izmit fault (Akyüz et al., 2002). Within the next few decades, similar large earthquakes are expected to rupture the submarine fault system that extends west of the Izmit fault under the Sea of Marmara, adjacent to the city of Istanbul (Barka, 1999; Hubert-Ferrari et al., 2000; Parsons et al., 2000; Atakan et al., 2002). Determining accurately the coseismic deformation associated with the Izmit event and its rupture parameters appear fundamental to model the stress evolution and to better understand the progression of the ongoing earthquake sequence in this densely populated region.

Rapidly after the event, the Izmit fault rupture was mapped in the field by an international team (Barka et al., 2002). This allowed the surface fault geometry to be determined and the variation of slip along strike to be measured with accuracy, due to the presence of numerous good markers of human origin -still unrepaired- which were offset across the fault (roads, rail-ways, canals, walls, fences). However, the exact length of the rupture remained undetermined, especially because some tens of kilometers of its western extension under the eastern Sea of Marmara could not be observed directly.

Several studies using various data sets (near-field strong motion records, far-field body waves, GPS measurements and SAR interferometry) have attempted to characterize the coseismic slip distribution, leading however to significantly differing

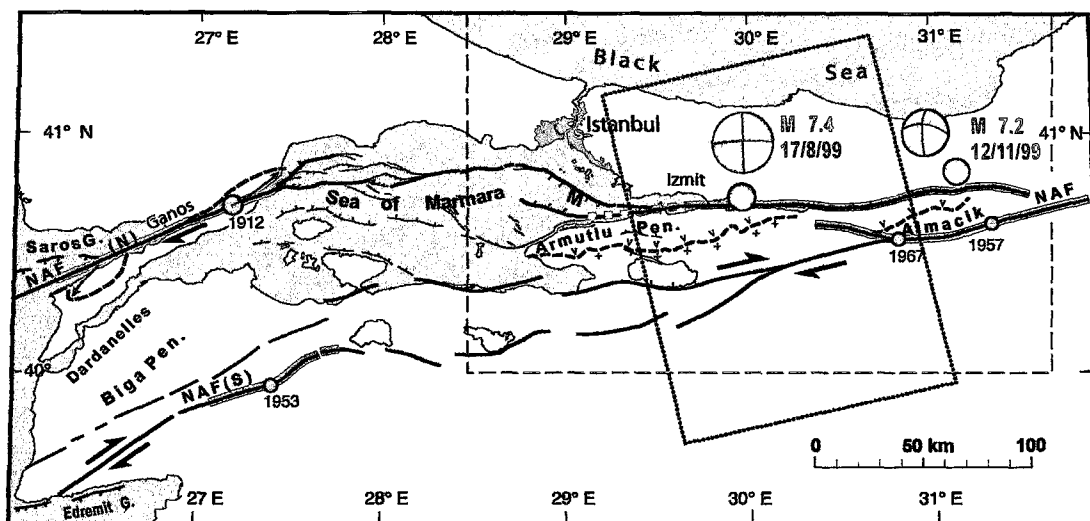


Figure 2.1 Active faulting in the Marmara pull-apart region (from Armijo et al. 1999, 2002). The North Anatolian fault (NAF) splays westwards into a northern (N) branch and a southern (S) branch 100 km apart. Faults associated with recent earthquake breaks are outlined with colors. The 1999 events occurred along a prominent fault splay east of Marmara. Fault-plane solutions are from the USGS catalogue. The dashed and dotted rectangles outline respectively the area enlarged in Figure 2.2 and the location of interferograms in Figure 2.3.

results (Reilinger et al., 2000; Bouchon et al., 2000; Yagi and Kikuchi, 2000; Wright et al., 2001a; Feigl et al., 2002; Delouis et al., 2002; Bürgmann et al., 2002a). The SAR interferograms obtained with ERS data of the Izmit earthquake contain artefacts due to a heterogeneous troposphere. These artefacts were encountered in previous studies and so the SAR data were considered less reliable than other independent data sets (e.g., Reilinger et al., 2000; Delouis et al., 2002). Using the available two tandem ERS1-ERS2 pairs, the topography and the meteorological data some of the atmospheric artefacts can be identified with confidence and removed.

Together the corrected SAR data and the tectonic observations provide the most accurate and complete description of the surface deformation associated with the Izmit earthquake. Combining the two data sets allows one to determine the coseismic slip distribution with depth across the different segments that ruptured with reasonable confidence. A trial-and-error approach is used by steps to explore solutions consistent with the tectonic information, then an inversion technique to improve the fit to the SAR data. The most significant features of the slip distribution and the fault segmentation appear well resolved, including the ruptures western end, close to Istanbul. This approach tends to reduce or to explain the discrepancies with models deduced from other data sets.

2.2 Tectonic Background, Field Observations of the Surface Break and Distribution of Aftershocks

Unlike the previous earthquakes of the twentieth century sequence, which broke 700 km along the linear eastern and central parts of the NAF, the 1999 Izmit and Düzce events ruptured a fault splay at the entrance of the more complex Sea of Marmara pull-apart region (Fig. 2.1). In this region, the NAF divides into a number of fault branches involving significant subsidence and crustal extension (Barka and Kadinsky-Cade, 1988; Parke et al., 1999; Armijo et al., 1999; 2002). The 1999 earthquakes occurred where two previous events had already ruptured in 1957 and 1967 contiguous fault segments south of the Almacık block (Fig. 2.1). Together the Izmit and Düzce earthquakes ruptured almost completely the more sinuous fault branch north of the Almacık block, so this block is now surrounded by the recent breaks. A prominent fault bend characterises the surface rupture near the city of Akyazı (Fig. 2.2). It may be explained by the long-term counter-clockwise rotation of the Almacık block with respect to Eurasia (Armijo et al., 1999; 2000). The long-term evidence also indicates that the earthquakes ruptured the largest branch of the NAF entering the Sea of Marmara at the Gulf of Izmit. This branch becomes gradually more extensional westward, as larger and larger fault step-overs and deeper pull-apart basins filled with sediment occur along it (e.g., Barka and Kadinsky-Cade, 1988; Armijo et al., 2002; see Fig. 2.2).

Detailed observations and maps of the Izmit earthquake surface rupture are reported by Barka et al. (2002). Here the main results relevant to this study are summarized. The rupture was readily observed on land over a total length of 110 kilometers. It is composed of a series of segments with overall E-W strike and mainly right-lateral slip. Seen in more detail, the strike of the rupture changes gradually to N80°E as it enters the Sea of Marmara and bends to a N70°E strike as it reaches the Almacık block.

Four main strike-slip segments are distinguished along the Izmit rupture, from West to East (Fig. 2.2): the Gölcük, the Izmit-Sapanca, the Sapanca-Akyazı and the Karadere segments. Two clear extensional step-overs separate the first three segments, at the Izmit Bay immediately east of Gölcük and at the Sapanca Lake. The Karadere segment forming the eastern end of the rupture has ENE strike and

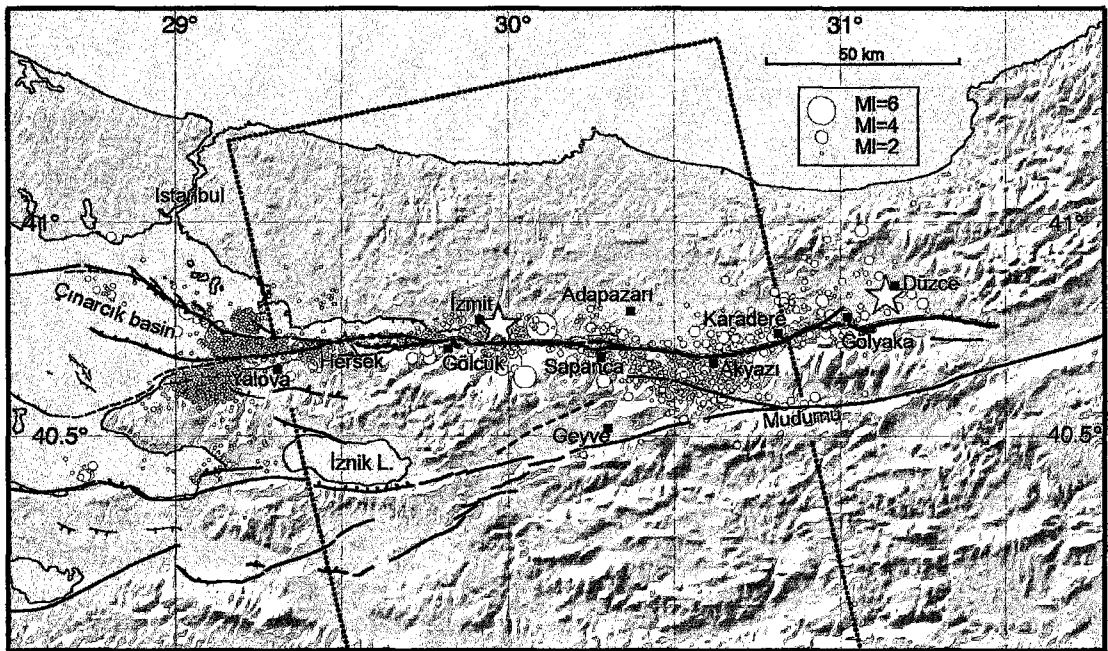


Figure 2.2 Fault segments and the 1999 earthquakes' breaks. Breaks of the Izmit (17/8/99) and Düzce (12/11/99) events are highlighted in red and purple, respectively. Stars denote epicenters of main shocks. Yellow circles are ML > 2 aftershocks recorded between 20/8/99 and 20/10/99 by the Tubitak permanent network (Özalaybey et al., 2002) and by a temporary array (Karabulut et al., 2002). The background DEM image is from GTOPO 30. The Izmit break has 110 km length on land but secondary features and aftershock distribution suggest that it extends 50 km west of Gölük, beyond the Hersek peninsula and offshore Yalova (dashed red lines).

its connection with the main rupture at the Akyazı fault bend is unclear. Another step-over is at the very eastern end of the rupture near Gölük and the Eften Lake. This area experienced up to 20 cm of right-lateral slip during the Izmit (August) event and much more lateral and normal slip during the subsequent Düzce (November) event (Akyüz et al., 2002; Hartleb et al., 2002; Ergintav et al., 2002).

Coseismic slip and its variation along strike could be measured with high precision along the fault trace. Significant slip variability was thus observed at the scale of the superficial complexities along the break (multiple mole-track branches and small stepovers), within a generally narrow fault zone (less than 10-50 m). Consequently, measurements of small markers sample fractions of the total deformation and usually underestimate the actual slip across the fault zone (Rockwell et al., 2002). Slip appears much less variable whenever large man-made markers crossing the fault (such as roads, railways and canals for irrigation) could be precisely surveyed. Such surveys integrate the deformation across the fault zone much better and are thus more reliable than the local measurements of smaller markers, to which they

clearly provide upper bounds. Figure 2.6 incorporates the best estimates of coseismic slip obtained from these surveys.

Maximum right-lateral slip exceeding 5.5 m was measured in two areas, east of the Sapanca Lake and in the city of Gölcük. Vertical slip was generally minor, but it reached locally 2.3 m over the oblique NW-striking normal faults that bound the Izmit Bay extensional step-over east of Gölcük. The hypocenter where the earthquake appears to have nucleated is to the East of the Izmit Bay and 5-10 km east of the region of maximum slip in Gölcük (Figures 2.2 and 2.5).

West of Gölcük (Fig. 2.2), the rupture continued with unknown extent under water, possibly along the edges of -and/or across- the elongated pull-apart feature seen in the bathymetry between Gölcük and the Hersek peninsula (Kuşçu et al., 2002). The Gölcük strike-slip segment must be short (less than about 5 km), because it is immediately flanked both to the East and to the West by significant step-overs with normal and oblique slip. Many large slumps that occurred all along the coastal area between Gölcük and Hersek were interpreted as lateral spreading effects of the submarine part of the earthquake rupture. However, no evident surface break was observed across the Hersek peninsula. Only some minor cracks were noticed in the ground near the tip of the peninsula, where the long-term morphology indicates the passage of a large strike-slip fault. This has led to the inference that the surface break of the Izmit earthquake ended somewhere east of the Hersek peninsula, with a total rupture length limited to 130 km.

However, the distribution of well-located aftershocks suggests a longer rupture (possibly including the Yalova-Hersek segment, west of Hersek (Özalaybey et al., 2002; Karabulut et al., 2002; Ito et al., 2002) (Fig. 2.2). Apart from aftershocks outlining the overall surface rupture, three regions hosted significant swarms; south of Akyazı, around the epicenter at Izmit, and north and west of Yalova. In fact, three swarms of aftershocks are located definitely west of the Hersek peninsula (Karabulut et al., 2002).

High-resolution bathymetric data acquired recently indicate that west of the Hersek peninsula and north of Yalova the submarine fault system splays apart into two main branches that veer towards a NW strike, as the depth to the sea-bottom increases dramatically (Fig. 2.2) (e.g., Armijo et al., 2002). These two fault branches have significant long-term antithetic normal component of slip. They run at the base of the two large escarpments that bound the 1150-m-deep Çınarcık Basin, which appears to be one of the largest pull-apart basins in the Sea of Marmara (Barka and

Kadinsky-Cade, 1988; Armijo et al., 2002). In a later section the bathymetry, the aftershock and the SAR information are used to determine the probable extent, the geometry and the slip distribution of the Izmit rupture in this submarine region.

2.3 InSAR Data

2.3.1 Data processing

Several interferograms that span the 1999 Izmit earthquake both in the descending and ascending modes of the European Space Agency's ERS1 and ERS2 satellites were calculated using DIAPASON software developed by French Aero Space Agency, CNES (Centre National d'Etudes Spatiales). Of these, only two ascending interferometric pairs have high coherence and give a good image of coseismic surface deformation. The two coseismic interferograms are formed by combining two pairs of tandem images of ERS1 and ERS2 acquired several days before the event in August (orbits ERS1-42229, ERS2-22556, 12-13 August) and about a month after in September (orbits ERS1-42730, ERS2-23057, 16-17 September) (Table 2.1, Fig. 2.3).

Here, the 2-pass method (Massonnet et al., 1993) is used, that is, the topographic contribution to the interferogram is removed using a digital elevation model (DEM). The ERS2 interferogram is constructed using precise orbits calculated by the University of Delft and thus it is assumed not to contain orbital residuals significant enough to be removed (more than 1 fringe across the image). However, the precise orbits were not available for the ERS1 interferogram and therefore it may contain orbital fringes. This problem can be solved by removing the difference between the two interferograms attributable to orbital precision. In this case a plane that contains two and a half fringes running roughly parallel to the satellite direction of flight is used to correct the orbital fringes.

The interferograms are filtered using a weighted power spectrum algorithm (Goldstein and Werner, 1998) and then coherent fringes (about 87%) are automatically unwrapped with Residu-Cut-Unwrap method (Goldstein et al., 1988) and sampled to be used for modelling. To have a clear view of the surface deformation (shape, gradient and the number of fringes), the interferograms are presented in Figure 2.3 re-wrapped with fringes, each fringe representing a range change of 5.66 cm (one wave length) along the radar line of sight. Because the interferograms span a time interval ending about one month after the event, they may contain some centimetres of range change due to post-seismic deformation as

deduced from the GPS measurements (Reilinger et al., 2000; Ergintav et al., 2002; Feigl et al., 2002).

Table 2.1 ERS data and interferograms used in this study (track 157 frame 815). H_a is the altitude of ambiguity (the magnitude of unmodelled topography required to create one fringe).

Orbit1- Track-Frame	Date	Orbit-2 Track-Frame	Date	H_a (m)	Interval (day)	Interferogram
42229-157-815	12-Aug-1999	42730-157-815	16-Sept-1999	3147	35	ERS1
22556-157-815	13-Aug-1999	23057-157-815	17-Sept-1999	329	35	ERS2
42229-157-815	12-Aug-1999	23057-157-815	17-Sept-1999	42	36	ERS1-ERS2
22556-157-815	13-Aug-1999	42730-157-815	16-Sept-1999	38	34	ERS2-ERS1
42229-157-815	12-Aug-1999	22556-157-815	13-Aug-1999	41	1	August tandem
42730-157-815	16-Sept-1999	23057-157-815	17-Sept-1999	40	1	September tandem

2.3.2 Analysis and interpretation of the data

Despite the rough topography and dense vegetation cover, the coherence is fairly good in large parts of the interferograms. It is lost in areas close to the fault (blank areas within the image frame). This may be partly due to the steep slip gradient in these areas. However, clear fringes can be observed within a few km from the surface rupture along the southern side of the Izmit-Sapanca segment. Decorrelation also occurs in the flat areas in the central and western parts of the interferograms, which is probably due to changes in the water content in the soils.

Because the coseismic displacement associated with the right-lateral strike slip occurs mostly in the E-W direction, that is nearly parallel to the radar line of sight, the fringes are mostly symmetric about the fault trace. The symmetry of fringes running parallel to the fault also suggests that the fault is very steep. However, in the central part of the interferograms the fringe gradient appears steeper on the northern side of the fault than on its southern side. This feature may indicate that the fault dips steeply to the North, in agreement with the focal mechanism of the main shock (Harvard CMT). The area of Izmit and Gölcük appears surrounded by elliptical-shaped fringes with high gradient, consistent with the large amounts of slip observed there. Fringes with high gradient are seen east of Gölcük towards the Hersek peninsula, and more spaced fringes continue tens of kilometres eastwards beyond Hersek. Several fringes appear deflected in two particular places: along the

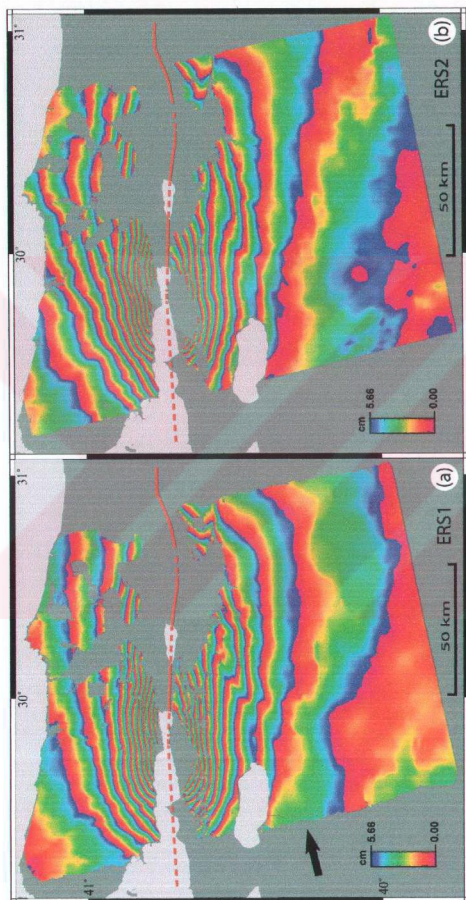


Figure 2.3 Interferograms of the Izmit earthquake. Data is from ESA satellites ERS-1 and ERS-2 acquired during ascending orbits. Surface rupture of Izmit earthquake is outlined in red. Each fringe (one full color cycle) represents 5.6 cm of range change along the radar line of sight (see text) whose horizontal projection is indicated by a black arrow. Positive changes indicate that distance to satellite has increased. (a) ERS-1 interferogram (12 August-16 September 1999). (b) ERS-2 interferogram (13 August-17 September 1999).

Mudurnu valley southeast of Akyazı and along the north-western edge of the Geyve Basin, south of Sapanca. These features appear to be along known faults but also appear to correlate with sharp topographic features. The possibility that these features result from motion on secondary faults dynamically triggered by motion on the main fault has been explored (Armijo et al., 2000; Wright et al., 2001a; Feigl et al., 2002). Detail studies reported in Chapter 2 indicate that these anomalies in the fringe pattern are most likely to be the product of atmospheric artefacts correlated with the topography.

The main difference between the two interferograms is on the northern side of the fault. There the fringes in the ERS2 interferogram trend more NE on the western side and more NW on the eastern side, making a broad concave-southwards cusp. Subtracting one interferogram from the other shows that the ERS2 data contain at least 2 more fringes –or a maximum of 14 cm range change- in this cusp region (Fig. 2.4a). This difference is significant and requires explanation before starting the modelling of the earthquake faulting process. There appears to be no correlation between the fringes in Fig. 2.4a and topography, in contrast with examples of similar features studied elsewhere (Delacourt et al., 1998; Beauducel et al., 2000). Therefore, these fringes are very likely a consequence of a heterogeneous troposphere (Feigl et al., 2002). It is not possible, however, to use a perfect "pair-wise logic" (Massonnet and Feigl, 1998) to determine if one of the radar images contains most of the atmospheric artefact. This is because the orbital separation combining ERS1-ERS2 pairs are not suitable for obtaining coherent interferograms (see Table 2.1). Nevertheless, the two one-day ERS1-ERS2 tandem pairs can be processed. With such short time period, both tandem interferograms should be very coherent and contain practically no surface deformation. However, coherence is almost completely lost in the southern and northwestern regions in both interferograms. While the signal is negligible in the August tandem interferogram (that is not shown here), the September tandem interferogram contains in its coherent part organized signal of up to three fringes (Fig. 2.4b). These fringes have an elliptical shape over a wide region that is the same as that of the cusp seen in the difference between the two coseismic interferograms (Fig. 2.4a). The features seen in the September tandem pair correspond thus very likely to local atmospheric artefacts of meteorological origin, at the scale of the interferogram, much like similar features described elsewhere (cf. Massonnet and Feigl, 1998).

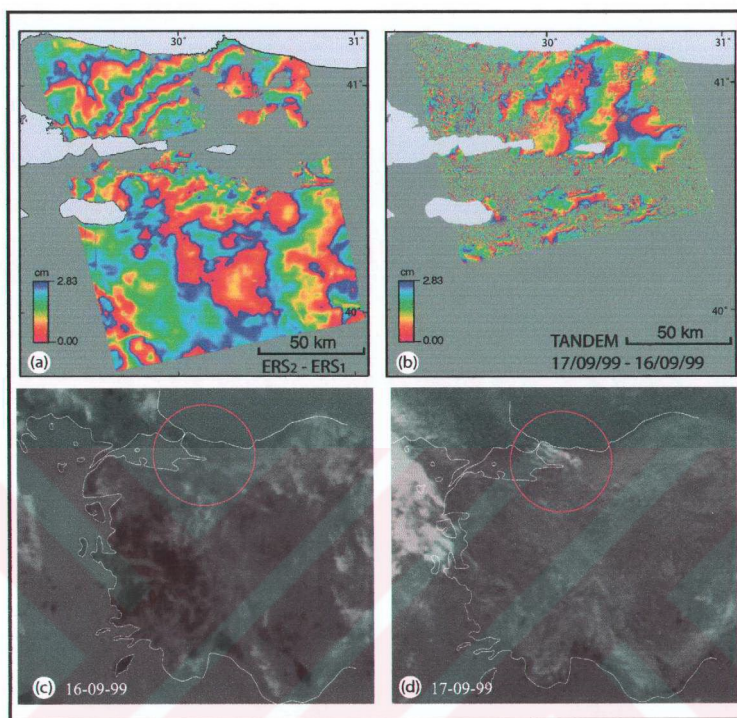


Figure 2.4 Identification of artefacts. Each fringe represents 2.8 cm of ground shift away from satellite along the radar line of sight. (a) Phase difference between the ERS1 (Fig. 2.3a) and ERS2 (Fig. 2.3b) interferograms. The five fringes seen in the northern part of the interferogram are not correlated with topography and a likely consequence of heterogeneous troposphere. (b) Tandem interferogram calculated from ERS2 and ERS1 images acquired on 17 and 16 September 1999, respectively. The fringes seen in the northern part of the interferogram confirm the occurrence of tropospheric artefacts. (c, d) Meteorological data; NOAA DMSP images acquired on 16 (c) and 17 (d) September 1999. The circle locates the study area. The tropospheric artefacts encountered in the ERS2 interferogram are related to the presence of clouds just before the acquisition of the 17 September ERS2 data.

The available meteorological data is checked to see if atmospheric changes have occurred during the time interval when the SAR tandem data were acquired, in September 1999. Two of the NOAA satellite images acquired in the days of ERS data acquisitions are shown in Figures 2.4c and 2.4d. The sky is clear in the Sea of Marmara region about 13 hours before the ERS1 data acquisition (16 September) (Fig. 2.4c). However, clouds cover the area to the North and to the Northeast of the Gulf of Izmit approximately 7 hours before the acquisition of the ERS2 data (17

September) (Fig. 4d). This sudden change suggests that the atmospheric and weather conditions were rapidly degrading from the 16th to the 17th September. Accordingly, the atmospheric artefacts seen in the September tandem interferogram are most probably included in the ERS2 image, which explains why the ERS2 coseismic interferogram has more fringes than ERS1 one. In addition, the good atmospheric conditions prevailing on the 16th September are similar to those seen in the NOAA data covering the 12th and 13th August 1999 (that is not shown here), when the first tandem pair was acquired.

Therefore, the ERS2 interferogram is not used to deduce the source parameters of the Izmit event, in contrast with earlier published work (Delouis et al., 2002; Wright et al., 2001a).

2.4 Modelling the Slip Distribution Using Elastic Dislocations

The purpose of modelling procedure used here is to determine a set of source parameters explaining both the tectonic observations and the InSAR data. As in other examples elsewhere in the world, the InSAR data set appears to be the best to deduce an overall image of the static rupture at seismogenic depth. Although with similar accuracy (within error of less than 1 cm), the GPS measurements sample discrete observation points with generally no comparable spatial coverage. For the Izmit event the GPS data have been used to model slip (Reilinger et al., 2000; Bürgmann et al., 2002a). The main implications of this data set in view of the results found in this study are discussed in section 2.7.

The quality of the SAR data is generally poor close to the rupture trace, due to lack of coherence. Pixel offsets across the fault trace in the SAR amplitude images can be used to determine the surface slip (Michel et al., 1999; Peltzer et al., 1999). However for the Izmit earthquake the results obtained using this technique are too scattered and thus of little use. A similar technique using SPOT satellite images provided good results only along one of the segments of the Izmit rupture (the Izmit-Sapanca segment; Michel and Avouac, 2002). The particularly precise measurements of offset markers, gathered in the field after the event, provide an overall coverage of the surface rupture (Fig. 2.6) and the slip observed is consistent with the SPOT data where the latter are available.

Surface deformation due to shear or tensile opening on rectangular planes embedded in an elastic half-space is calculated using the formulas given in closed analytic form by Okada (1985). A set of 7 fault parameters is used for modelling. The parameters are 1) fault length, 2) coordinates of the fault, 3) upper and lower

depth, 4) strike, 5) dip, 6) slip and 7) rake (Fig. 2.5). One such set defines a single dislocation to describe a single fault patch. Three components of displacement (East, North and Vertical) at a point or a regular grid on the surface, due to dislocations on multiple fault patches can be calculated simply by summing the displacements resulting from each patch. Using the position of orbits surface displacement can then be projected in the direction of the radar line of sight in order to compare it with the SAR data (See Chapter 1.2.4).

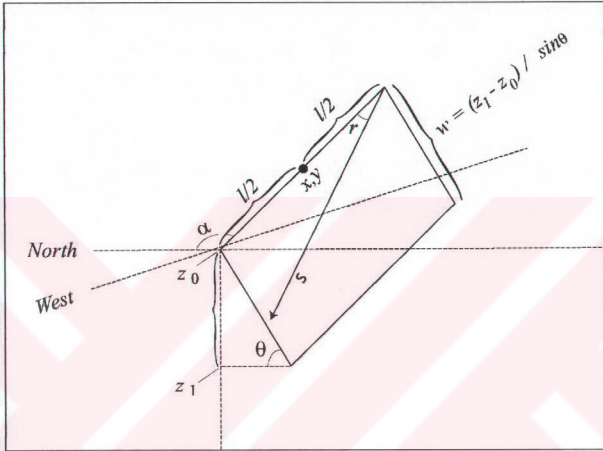


Figure 2.5 Fault parameters that define a dislocation on a single fault in modelling. "l" is fault length, "w" is fault width, "x" and "y" are coordinates of the fault center, "z₀" is upper depth of the fault, "z₁" is the lower depth, "α" is fault strike, "θ" is the fault dip, "s" is slip, and "r" is rake.

The seismic moment M_0 can be calculated by multiplying the area of fault patch ($A=l \cdot w$) by the amount of slip (s) and shear modulus (μ), that is, " $M_0=\mu \cdot s \cdot A$ ". Shear modulus " μ ", which varies typically between 30-36 GPa is taken 33 GPa in this study. For multiple fault patches, the total seismic moment will be the summation of moment calculated for each patch. Moment magnitude, M_w , can be calculated using the empirical formula " $M_w = 2/3 \log M_0 - 6$ " (Kanamori, 1977).

To model ERS1 interferogram, forward modelling strategy is used to obtain a first-order model that is then refined by steps combining a trial-and-error approach with a conventional inversion technique. The procedure seeks to fit the SAR data which are sampled uniformly where the interferogram is coherent, in this case 14000 samples of range change measured in the ERS1 interferogram.

2.4.1 Forward modelling

The first-order model (model I) is obtained with a simplified fault divided into vertical patches 5-km-long along strike, consistent with the geometry of the observed surface rupture on land and with the features seen in the bathymetry of the Sea of Marmara. Slip is purely right-lateral strike-slip, consistent with the measurements gathered at the surface for each patch and extrapolated uniformly down to 18 km depth (Fig. 2.6, I). This is the overall depth for which forward modelling gives the best fit to the SAR data (RMS of 2.4 cm in range). The resulting geodetic moment is 2.5×10^{20} Nm, which is equivalent to moment magnitude Mw 7.6. The implied fault rupture is about 160 km long from Gölyaka to approximately 30 km west of the Hersek peninsula. How the western end of the rupture is constrained by the SAR data is discussed more precisely in Chapter 2.5. The good fit to the SAR data (root mean square [rms]=2.4) obtained with this very simple approach suggests a possible correspondence between significant features of the slip distribution observed at the surface and the slip distribution on the fault at depth. Regions of relatively higher slip surrounded by regions of lower slip would coincide with well-identified strike-slip segments like the Karadere, the Sapanca-Akyazı and the Gölçük segments.

The distribution of residuals between the model and the SAR data is represented in Figure 2.7a. Residuals in the far field are flat, of small amplitude and generally negative (except in the NW corner of the scene), indicating that the dislocation model slightly overestimates the overall far-field effect of the earthquake deformation. The broad positive residuals in the NW corner of the scene could correspond to a minor atmospheric artefact that could not be removed. Closer to the fault trace the residuals are somewhat larger (up to 6 cm in range) and more conspicuous, both positive and negative. This may imply either local complexities of the actual deformation or local atmospheric artefacts, which require more detailed analysis.

2.4.2 Inverse modelling

A second modelling stage (models IIA and IIB) explores more refined slip distributions over the fault at depth using an inversion procedure with the first-order model as a starting solution. The previous fault is now divided into patches with

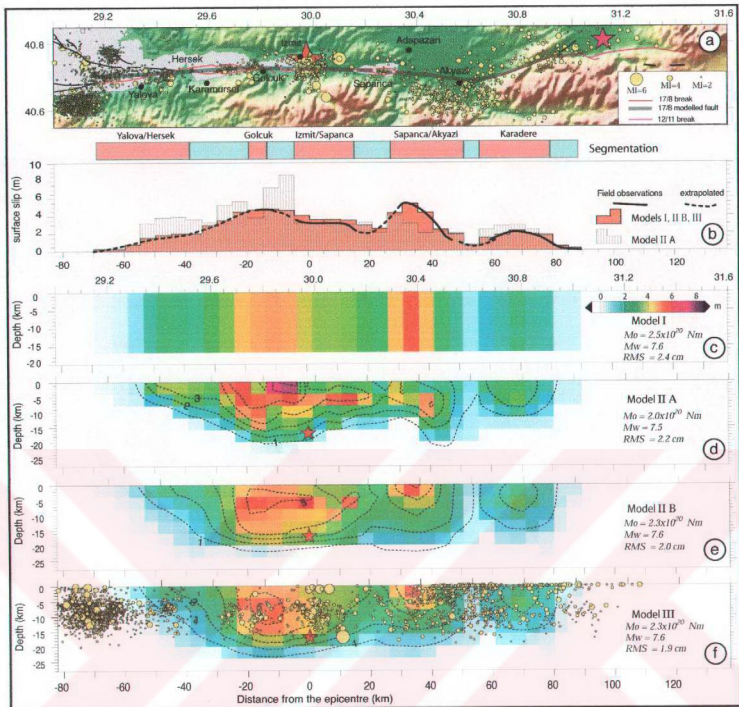


Figure 2.6 The Izmit earthquake fault trace, the coseismic surface slip and the modelled slip distribution at depth. (a) Shaded topographic map with fault segments and simplified fault trace. Surface breaks (thick line) of the Izmit (red) and Düzce (purple) events are indicated. Locations of mainshocks (stars) and of aftershocks (yellow circles) as in Fig. 2.2. (b) Surface right-lateral slip projected along the fault trace. The sinuous curve in bold integrates the most robust field measurements (dashed where extrapolated). The toothed graph in orange represents the discrete values used for modelling slip at depth in 32 fault patches with 5 km length along the fault strike (model I). The hatched graph is the surface slip obtained by inversion in IIA. Segments and inter-segment regions are indicated on top, by pink and light blue stripes, respectively. (c) First order forward modelling of slip distribution. Slip is purely right-lateral (in m). Slip variation along strike is consistent with the tectonic observations and is extrapolated in vertical fault patches to variable depth to fit the SAR data. Best fit is for uniform fault depth of 18 km. (d) Model obtained by inversion of slip in a fault with 224 vertical patches 5 km x 4 km. Slip is inverted using model I as initial slip distribution. Lines of equal slip are in metres. Resulting slip in the patches near the surface (0-4 km depth) is represented in b (hatched graph). (e) Inversion model with slip fixed for the patches near the surface (0-4 km depth). (f) Model obtained by inversion of slip in a fault slightly dipping north, using initial slip close to model IIB. Izmit-Sapanca and Sapanca-Akyazi segments have uniform 85° dip to the North and 176° rake, other segments are vertical. Mainshock (red star) and aftershocks (yellow circles) projected along fault strike. Scalar moment (Mo) Magnitude (Mw) and RMS are indicated for each model.

vertical width of 4 km down to 28 km depth (224 patches 5 km x 4 km). A simple iterative linear least squares inversion procedure (e.g., Ward and Barrientos, 1986) is used. Slip is the only free parameter; all other parameters are fixed. For each independent dislocation the problem is linear, but the solution is non-unique and unstable because the solution on the patches are not independent and the distribution of the data is heterogeneous (Du et al., 1992).

Figure 2.6 (IIA and IIB) illustrates two different alternatives and Figure 2.7 (IIA and IIB) the corresponding residuals. If slip is left free everywhere, the solution is very unstable (Fig. 2.6, IIA). RMS is reduced to 2.2 cm. Slip tends to be more homogeneously distributed and the regions of higher slip spread towards the sides. Slip in the patches near the surface (0-4 km depth) is inconsistent with the observed surface slip (see Fig. 2.6b). The most striking inconsistencies are along the Sapanca-Akyazı segment, where modelled slip is much less than observed, and near İzmit, where a patch modelled with very high slip (more than 8 m) seems artificial.

If slip is fixed at the observed values in the patches near the surface (which are otherwise poorly resolved by the SAR data) and left free elsewhere (Fig. 2.6, IIB) the solution is more stable. The fit to the data is better (RMS of 2.0 cm) and the regions of higher and lower slip remain well identified. Residuals are generally smaller close to the fault trace (compare Figures 2.7, IIA and IIB) but relatively large positive residuals remain on the northern side of the fault near İzmit. It is probably not coincidence that these residuals are located where high near-surface slip is indicated by the previous inversion (Fig. 2.6, IIA and Fig. 2.6b). Also the position of these residuals is close to the location of the largest aftershock (M_s 5.8; M_0 4×10^{17} Nm) that occurred on September 13, during the time covered by the SAR images. However, the source of this event is deep (16-18 km; Örgülü and Aktar, 2001) and it is unlikely that it may have modified significantly the fringe pattern.

In a third modelling stage (model III), small changes in the fault dip and rake is introduced to reduce residuals close to the fault trace. After trying different models a model with a minor down-to-the-north normal movement is retained. This model is consistent with the apparent asymmetry of the fringe gradients in the central part of both the ERS1 and the ERS2 interferograms (as discussed in Section 2.3), with the fault plane solution of the main shock (Fig. 2.1), and with the leveling measurements that show significant down-to-the-north normal movement between İzmit and Sapanca lake (Ergintav, 2002 pers. com). Then, slip is inverted using an initial solution similar to that obtained in the previous modelling stage (IIB) and keeping

the near-surface slip fixed. In model III (Figure 2.6), all the fault patches of the two central segments (Izmit-Sapanca and Sapanca-Akyazi) have dip of 85° to the North and a small component of normal faulting (176° rake). The resolution of slip in patches of this model is very similar to that in the vertical model. RMS is now 1.9 cm and the residuals close to Izmit are nearly erased (Fig. 2.7, III).

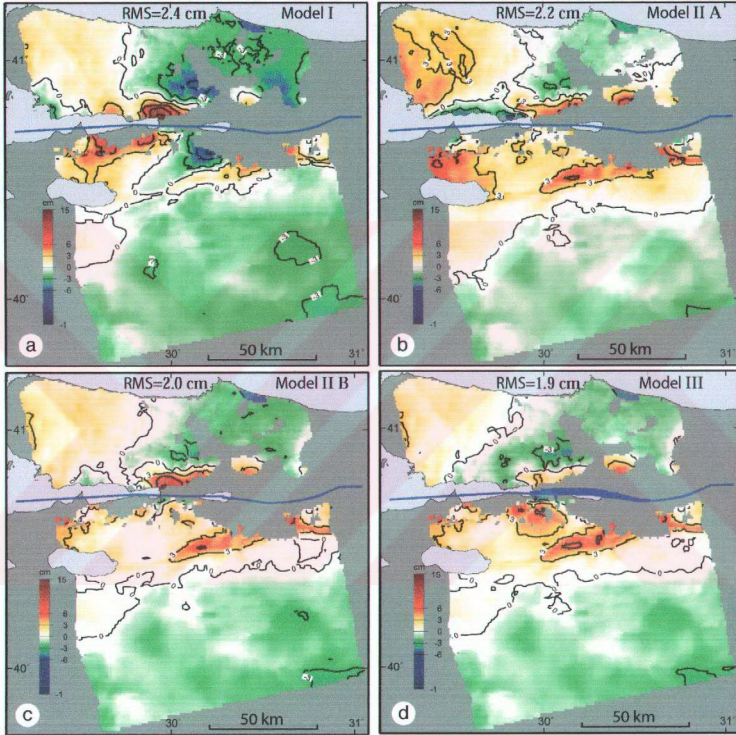


Figure 2.7 Residual (observed minus modelled) interferograms. The order of the interferograms follows the modelling stages discussed in the text. Residuals are expressed as positive (yellow to red scale) and negative (green to blue scale) values indicating that models respectively over- or under- estimate the surface deformation (the coseismic change in range between the two ERS1 scenes). (a) Residuals are for (I) the first-order forward model (Fig. 2.6, I); (b) (IIA) the inversion of slip in all the patches (Fig. 2.6, IIA); (c) (IIB) the inversion model with slip fixed in the patches near the surface (Fig. 2.6, IIB); (d) (III) the inversion of slip in a fault allowing for a minor down-to-the-north normal component of slip between Izmit and Akyazi (Fig. 2.6, III).

The regions with high slip in model III are very similar to those in the first-order model (I) but the progressive fit to the SAR data has caused a cut-off of these regions at different depth. Slip centered in the Sapanca-Akyazı and the Karadere segments appears mostly concentrated in the first 8 km near the surface. By contrast, the region of very high slip centred in Gölcük , immediately west of the hypocenter, seems more deeply rooted (down to 20 km; Fig. 2.6, III). Total seismic moment and moment magnitude (2.3×10^{20} Nm; $M_w=7.6$) are slightly lower than those in model I, but the fraction of moment released by the Gölcük and nearby segments appears very significant (about 2/3 of the total moment).

Figure 2.8 represents the synthetic fringes corresponding to model III, the resulting residual fringes and some selected profiles across the ERS1 data and the model. Overall the synthetic fringes reproduce very accurately the observations (compare Fig. 2.8a with Fig. 2.3a). The excellent fit to the data is also seen in the profiles (Fig. 2.8c). The maximum amplitude of range change across the fault is about 180 cm (profile 3) corresponding to 4.7 m of horizontal displacement parallel to the fault, exactly as measured in the field at Gölcük. The obtained RMS of 1.9 cm corresponds to an error of about 1%. However, it is clear that the very good fit to the data describing the deformation of the earth's surface corresponds to a much larger uncertainty in the estimates of slip across the fault at depth. Using the slip data collected in the field at the surface improves the stability of the solutions and reduces the slip uncertainty. In Figure 2.9, the projection of the Model III along the radar line of sight is also illustrated in a 3D perspective view so that the reader can easily visualize how the ERS radar sees the surface deformation.

2.4.3 Resolution analysis

To evaluate the resolution of the slip on the different regions of the fault, a truncated singular value decomposition approach (e.g., Du et al., 1992) and smoothed solutions are used. Introducing artificially small slip perturbations it is estimated that smoothed models can resolve slip of less than 0.5 m in the regions of the fault near the surface (12 km depth) and less than 1 m in the regions between 12 km and 24 km depth. The models cannot resolve slip of less than 1 m in regions of the fault at depth greater than 24 km. The resolution of the slip in some regions of the fault at shallow depth is bad however, because of the poor coherence of the interferogram in areas close to the fault trace (Fig. 2.10).

2.5 The Western End of the Izmit Rupture

The SAR data can be used to resolve the western end of the Izmit earthquake rupture. Figure 2.11 is an enlargement of the data together with two alternative rupture models for this region. Clearly models with rupture extending significantly

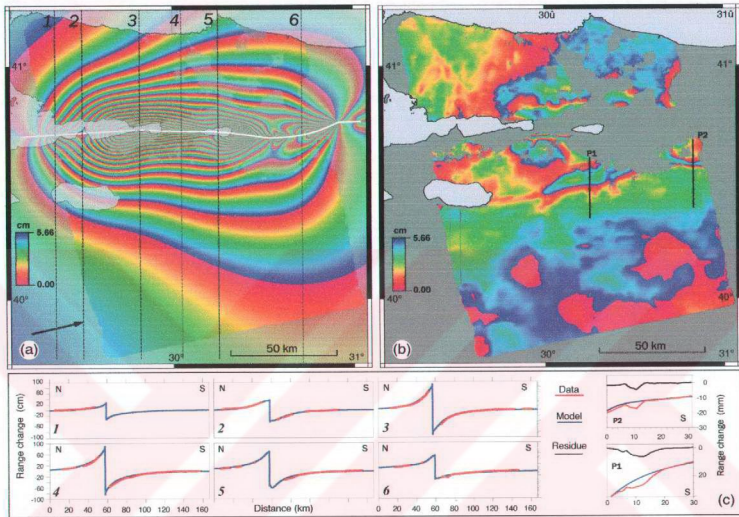


Figure 2.8 Detailed analysis of model III. (a) Synthetic interferogram. Fringes are emphasized with brighter colors in the coherent parts of the ERS1 interferogram to facilitate comparison with the data shown in Figure 3a. The numbered lines indicate the position of the six N-S sections displayed in c. (b) The residual interferogram is the same as in Fig. 2.7 (III) but expressed in color cycles (same color scale as in a). The residuals covering the southern region of the interferogram appear closely correlated with the topography. (c) Observed (red) and modelled (blue) profiles of range change across the fault. Significant small-scale misfits are seen across profiles 5 and 6 in the Geyve and Mudurnu regions (boxes labeled P1 and P2, location given in b). These are the two regions where deflected fringes are seen both in the ERS1 and in the ERS2 interferograms (see Fig. 2.3). The right panel in c gives enlarged profiles of P1, of P2 and of the corresponding residuals (data minus model; black lines). Detail analysis of these anomalies is discussed in Chapter 3.

westward beyond the Hersek peninsula (Fig. 2.11a) fit much better the data than models with rupture ending at the Hersek peninsula (Fig. 2.11b). In model III (Fig. 2.8a) an idealized Yalova-Hersek fault segment geometry is adopted taking into account the aftershock distribution and the position of prominent fault traces in the high-resolution bathymetry (Armijo et al., 2002).

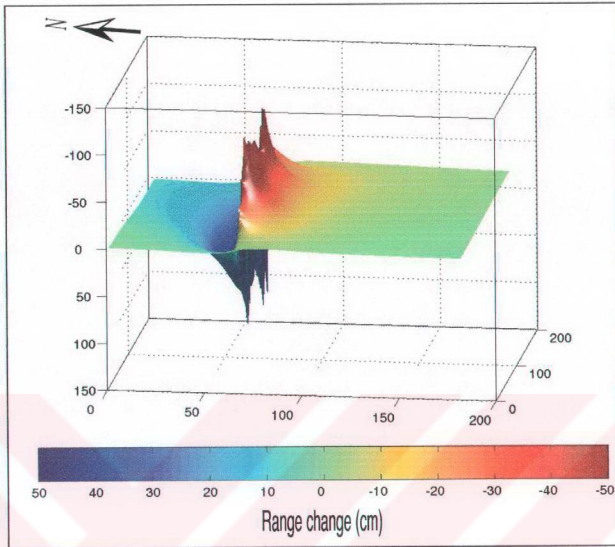


Figure 2.9 A 3D view of the Model III (Fig. 2.8) projected along the radar line of sight. As the displacement is mainly right lateral, looking from west to east the radar detects the surface deformation as range increase (green-to-blue) in northern side of the fault, and as range decrease (green-to-red) in the other side.

The modelled fault coincides with an aftershock cluster with almost planar and vertical distribution of hypocentres (Karabulut et al., 2002) that possibly defines the average position of the strike-slip fault segment connecting the İzmit fault with the more extensional faults bounding the Çınarcık basin. The interferogram does not contain information close to the fault to better constrain complexities of its geometry and kinematics. For instance, there is possibly some normal fault component of slip as the fault enters more and more into the Sea of Marmara. However, the overall symmetry of the observed fringes indicates that no significant normal faulting has occurred and therefore a vertical fault with pure right-slip is kept. Slip across the modelled rupture decreases over 30 km, from 4.5 m in Gölcük to 2 m in Hersek. Then it tapers over the next 30 km (Yalova-Hersek segment), from 2 m to zero (Fig. 2.6b). Thus, a significant average slip of 1-2 m is required down to a depth of 10-15 km across the first 15 km of the latter segment, immediately west of the Hersek peninsula. However, no clear surface break was observed after the İzmit earthquake across the Hersek peninsula (Barka et al., 2002). In addition, no fresh

surface break has yet been detected on the sea bottom during the recent surveys devoted to map the submarine part of the fault, west of Hersek (Le Pichon et al., 2001; Polonia et al., 2002; Armijo et al., 2002). Thus, the inferred rupture of the Yalova-Hersek segment may have not reached the earth's surface, although the moment released would have been 1.5×10^{19} Nm, equivalent to an event Mw 6.8.

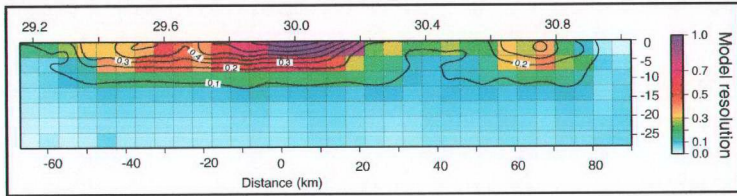


Figure 2.10 Model resolution from singular value decomposition (SVD). Resolution decreases with increasing depth. Areas of relatively low resolution are also seen at some shallow depths where the SAR data is absent in the near field.

2.6 Differences with Previous Models

The best model (model III, Fig. 2.6, 2.8) has differences from previous models (Figure 2.12) (Reilinger et al., 2000; Yagi and Kikuchi, 2000; Wright et al., 2001a; Feigl et al., 2002; Bouchon et al., 2002; Delouis et al., 2002). The atmospheric artefacts in the ERS data, which were elucidated and removed, explain some of the discrepancies with other models using the geodetic data (SAR and GPS). However, the most significant improvement comes from the use of a precise fault map and of the slip data collected in the field, which reduce the range of possible solutions. However, the model III is consistent with the model proposed by Yagi and Kikuchi (2000), which is derived solely from seismic data (near-field strong motion and teleseismic body wave data) (Figure 2.12).

Another important difference with other approaches concerns the moment release. The moment release in the best model (2.3×10^{20} Nm) is somewhat higher than that deduced from the seismic records ($1.7\text{--}2.0 \times 10^{20}$ Nm; Toksöz et al., 1999; Tajima et al., 1999; Yagi and Kikuchi, 2000). The difference may be due to the longer period of time (35 days) that is sampled by the SAR interferograms. As stated earlier, the SAR interferograms -and thus the models- may contain significant post-seismic deformation. This hypothesis is explored using the published GPS data.

The GPS data include four permanently operated stations and observations collected in several stations around the fault during many epochs before and soon after the earthquake (Reilinger et al., 2000). Reilinger et al. (2000) have used this

data set to retrieve the horizontal coseismic displacements reproduced in Figure 2.13, obtained by removing at each non permanent station the part of the motion attributed to interseismic and to post-seismic deformation. This set of GPS vectors can be used to calibrate a coseismic model derived from the "longer-period" model III. The same approach (fixing as for model III the same characteristics of the fault

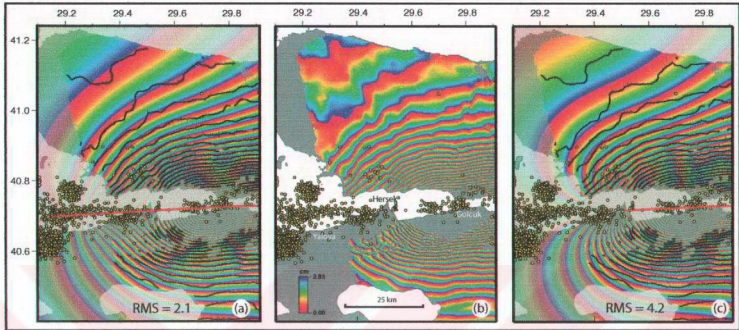


Figure 2.11 The western end of the Izmit rupture. Yellow circles are aftershocks recorded between 20/8/99 and 20/10/99 as in Fig. 2.2. Each fringe represents 2.8 cm of range change, as in Fig. 2.8. Observed interferogram in the middle panel (b) for comparison with the two alternatives. RMS calculated for this part of the interferogram are given in cm. (a) Synthetic fringes for a rupture extending 30 km west of the Hersek peninsula with the slip distribution of model III (Fig. 2.6). The simplified Hersek-Yalova fault segment roughly coincides with a cluster of aftershock with almost planar, vertical distribution (Karabulut et al, 2002). (c) Synthetic fringes for a rupture ending at the Hersek peninsula. Modelled fault trace in red. Black contour lines overprinted in (a) and (c) were obtained by automatic unwrapping of the observed interferogram. The difference between (a) and (c) is equivalent to $M_0 1.5 \times 10^{19}$ Nm, or an event Mw 6.8 which would have ruptured the Yalova-Hersek segment.

and slip near the surface) is used to fit the coseismic horizontal displacement at the GPS stations and to obtain the corresponding slip distribution on the fault at depth. The resulting "purely" coseismic model can be compared with model III (Figure 2.13). Overall the observed horizontal vectors are correctly reproduced by the "purely" coseismic model (rms=4 cm), with the exception of a few stations close to the fault, which may be affected by spurious surface effects. Both the predicted coseismic displacement and the GPS vectors in the far field (specifically to the North and South of Izmit, at 10-80 km distance from the fault trace) appear systematically smaller (3-6 cm) than the corresponding horizontal displacement vectors predicted by the "longer-period" model III (Fig. 2.13a). Similarly, the modelled coseismic slip on the fault at depth is smaller than the slip in model III (Fig. 2.13 b, c) and the

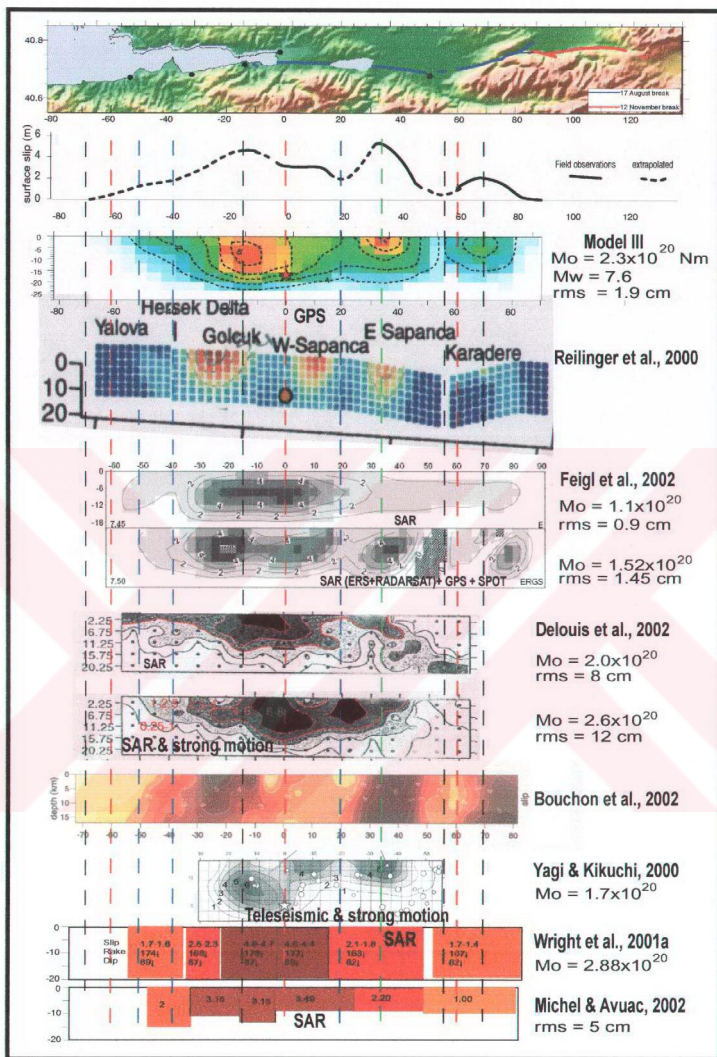


Figure 2.12 Previous models of slip distribution inferred for the August 17, 1999 Izmit earthquake using various data sets.

coseismic moment of 1.9×10^{20} Nm is close to the seismological estimates. The difference in slip between the two models (Fig. 2.13d) represents the after-slip that may have occurred in the month following the earthquake. There is some "noise" possibly due to some GPS stations close to the fault and to second-order defects of the models. However, zones with positive slip (0.8 m) emerge above the noise. Some of these zones are located at 4-16 km depth under regions of low coseismic slip and they are outlined by aftershock activity. Examples are below the bend area of Akyazı in the central part of the rupture, and below the Karadere segment at the eastern end of the rupture. Altogether, these shallow after-slip regions represent a small moment release (1×10^{19} Nm). They may be interpreted to occur in velocity strengthening regions of the fault (Tse and Rice, 1986). It is clear however, that the most significant and well-resolved after-slip is found in the more deeply seated region of the fault below İzmit and Gölcük. In the interferogram, this zone of large after-slip corresponds to a set of paired lobes of fringes enclosing slopes with opposite sign, which are symmetrically arranged on both sides of the fault trace (Fig. 2.14). It is very improbable that such a complicated feature could have resulted from an atmospheric artefact and thus a tectonic origin is favored. Thus, the excess of slip in model III strongly suggests that after-slip reaching 2 m has occurred during the month following the main shock, within a zone of the fault located at 12-24 km depth below the epicentral region. The corresponding moment release (0.3×10^{20} Nm) is equivalent to an event Mw 7.0 and represents about 14% of the total moment in model III. Therefore the difference in moment release between the "longer-period" model III and the seismological estimates appears explained by the occurrence of aseismic after-slip, deeply seated across the fault zone below the epicentral region.

2.7 Slip Heterogeneity, Fault Segmentation and Significance of the Rapid After-slip

Although broadly corroborating the inferences made earlier by Reilinger et al. (2000), the coseismic model calibrated with the same GPS measurements in this study, is simpler and appears more robust. It is consistent with the well-resolved features of the SAR interferometry and the tectonic observations. The modelling approach also allows one to discuss the geometrical relation between the fault segments, the location of the hypocentre and the slip distribution, either coseismic or the after-slip, and to draw simple mechanical inferences which differ from earlier inferences in some important aspects (Reilinger et al., 2000; Ergintav et al., 2002; Bürgmann et al., 2002a; Hearn et al., 2002).

Both the “longer-period” model III and the “purely” coseismic model indicate heterogeneous slip with three main zones of higher slip. Two of these zones correspond unequivocally to individual fault segments that are well identified in the surface morphology, namely the Sapanca-Akyazı and the Karadere segments. However, the third and largest zone of high slip, centred in the Gölcük segment but comprising also the Yalova-Hersek and the İzmit-Sapanca segments, is different: It stretches out through significant fault stepovers. This larger zone has released 2/3 of the total moment in the “longer-period” model III and it is interesting to note that the hypocentre of the main shock is located at its edge. In terms of slip distribution, the İzmit rupture has smoothed out the fault complexities (velocity strengthening regions) within the zone with maximum moment release around Gölcük. This feature is consistent with the idea that under Gölcük a large slip deficit and possibly a large stress concentration existed prior to the earthquake. It is also consistent with the occurrence of small events in this region in the years before the main shock (Barış et al., 2002). Nucleation of large events at the edges of zones with high stress release have been described elsewhere (Archuleta, 1984).

For the three main zones of higher slip the lower cut-off in the coseismic slip occurs at about 15-km depth. However, the after-slip zone appears to extend well into the lower crust, down to 20-25 km depth, directly under the zone of highest moment release and the main shock hypocentre. Thus, the most important after-slip does not appear to be concentrated under segments with relatively little coseismic slip, as suggested by Reilinger et al. (2000). Although the resolution in the depth estimate for the after-slip is poor (that of the SAR data modelling, discussed earlier), the inference that significant after-slip has occurred down to at least this depth range suggests static stress changes triggered by the earthquake over the same depth. This depth range seems also in keeping with the lateral extent (along strike) of the Gölcük high-slip region. Thus, the rapid localised after-slip under Gölcük requires an elastic response of the mid-lower crust and accelerated aseismic shear across the fault zone. The Gölcük after-slip zone encompasses a region of the fault having substantially slipped coseismically, including the hypocentre. This alternating behaviour suggests that the after-slip zone is located at the transition between an upper region of the fault dominated by stick slip (seismogenic) and a lower region dominated by plastic shear (aseismic). The results suggest that the rapid after-slip has well penetrated into the latter.

The SAR data must include all the post-seismic deformation during the first 29 days following the earthquake and the results found here can be checked for consistency

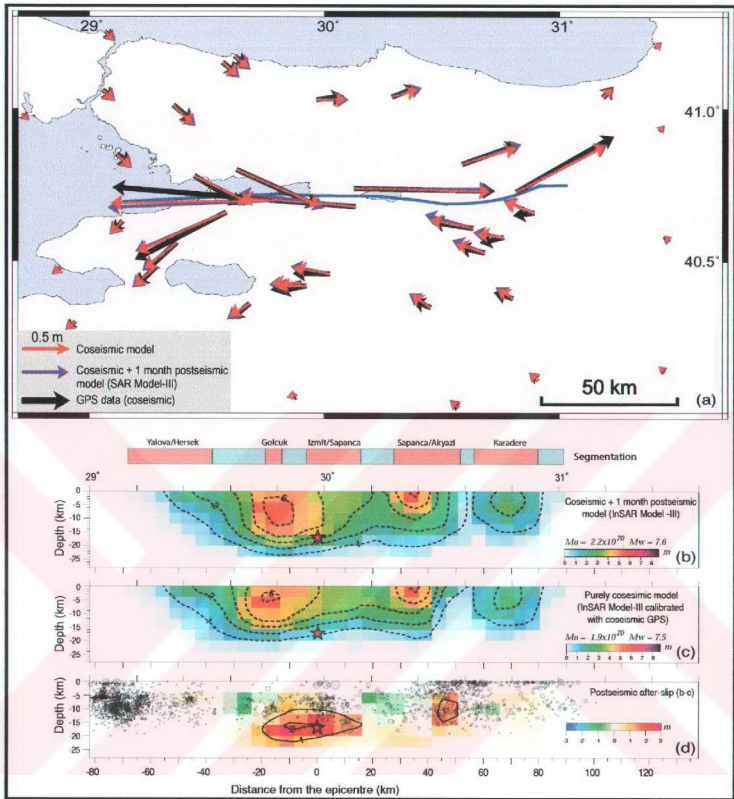


Figure 2.13 Separating the early after-slip from the "purely" coseismic slip. (a) Horizontal displacement is represented at the GPS stations. The coseismic GPS observations (black arrows) are from Reilinger et al. (2000). The vectors in violet are predicted from the "longer-period" model III represented in (b), which includes the 29 days of postseismic deformation captured by the ERS1 data. The vectors in red correspond to the coseismic model represented in (c), which is derived from the same geometry and kinematics as model III, but calibrated to fit the coseismic GPS data. The blue line is the simplified fault trace. The difference between (b-c) corresponds to the after-slip shown in (d). The red stars represent the main shock hypocenter. Aftershocks are in gray.

with models derived from the GPS data, which have less complete coverage in space and time. The zone of fast after-slip under Gölcük and İzmit deduced from the interferograms has been roughly depicted by Reilinger et al. (2000) using the post-seismic GPS data (Figure 2.12). Yet, more recent analyses of the GPS data suggest that the highest amount of after-slip has occurred below the Karadere fault segment, at the eastern end of the rupture (Bürgmann et al., 2002a; Hearn et al., 2002). The most important discrepancy with inferences found in this study is that none of the models derived from the GPS data predicts more than 0.4 m of after-slip during the 75-80 days following the main shock. Another difference concerns the depth to which the after-slip has penetrated. Bürgmann et al. (2002a) and Hearn et al. (2002) suggested after-slip of 10-40 cm down to depths of 40 km. Here it is found that a region with slip of 1-3 m at 18-km depth is well-resolved in the SAR models, while neither the SAR nor the GPS data can resolve slip of a few tens of centimetres in regions of the fault at more than 24-km depth. The deformation field associated with the region of high after-slip below Gölcük during the first 29 days after the main shock can be directly compared with the corresponding deformation deduced from the published GPS records (Bürgmann et al., 2002a). From the 13 permanent GPS stations available for the concerned region only three were in operation prior to the main shock and have thus captured without interruption all the post-seismic deformation (KANT, TUBI, DUMT; Fig. 2.14). The total horizontal displacements observed for these stations during this critical period of time (17 August to 16 September) are consistent with the vectors predicted by the large after-slip below Gölcük (Fig. 2.13). The remaining 10 GPS stations started to be installed in the near field of the fault in the days following the earthquake. The earliest reliable daily solutions are available only two days after the main shock. For these stations the total post-seismic displacement recorded (between the date of each first solution and the 16 September) is thus not complete, but it is worth to compare them with the corresponding vectors predicted by the after-slip model (Fig. 2.13). The two sets of vectors have compatible directions. However, the magnitudes of the vectors required by the large after-slip below İzmit-Gölcük are about twice as much as the observed GPS vectors. This discrepancy is especially clear for stations around İzmit (HAMT; UCGT; BEST; MURT) but not for the station placed near the eastern end of the rupture (KOP1), where the two vectors are nearly coincident. Therefore, about half of the large after-slip determined with the SAR data below İzmit-Gölcük may have occurred during the first two days following the main shock. The rapidity of this large early after-slip would explain why it has not been incorporated in the models derived from the GPS data alone (Bürgmann et al., 2002a; Hearn et al., 2002). If this

line of reasoning is correct, then the maximum after-slip of about 2 m at 16-18 km depth near the hypocentre would have occurred at rates of up to 150-200 m/yr, significantly faster (two orders of magnitude) than deduced earlier (Bürgmann et al., 2002a). Nevertheless, the results found here are not inconsistent with the main inference from the GPS modelling, indicating that a significant albeit much smaller amount of after-slip has occurred below the Karadere fault segment, at the eastern end of the rupture (Bürgmann et al., 2002a; Hearn et al., 2002).

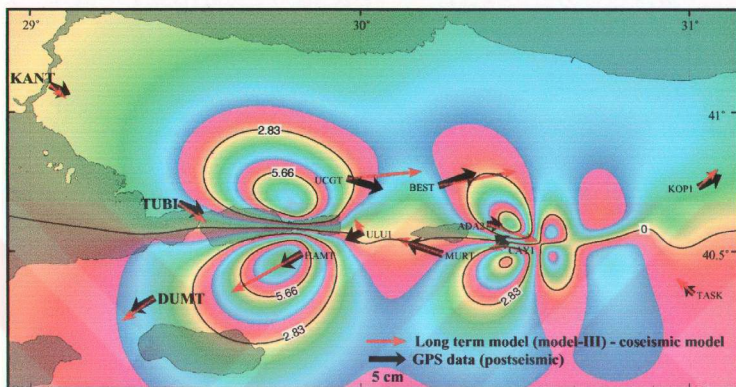


Figure 2.14 Postseismic deformation as seen at the earth's surface. The synthetic fringes represent the range change (in cm) corresponding to the after-slip model, as retrieved from the ERS1 data and depicted in Figure 2.13d. It seems improbable that such a set of fringes including symmetric paired lobes on both sides of the fault could be an atmospheric artefact. The corresponding horizontal displacement (in red) can be compared with the total displacement observed with GPS during the first 29 days after the mainshock (in black). The GPS data set on continuously recording stations shows rapidly decaying deformation but the record is not complete (Bürgmann et al., 2002a). The three stations in bold (KANT, TUBI and DUMT) are permanent stations already installed before the mainshock. The other stations were installed within few days following the mainshock and they do not include the deformation that may have occurred during the first 2-3 days after the mainshock. The large differences between the red and black arrows may be explained by very rapid, early after-slip reaching 1m that may have occurred in the region of Gölcük, around and below the hypocentre.

2.8 Implications Concerning the Long-Term Geological Record

The correlation between the coseismic slip at depth and measured slip along the surface rupture is good, but the coseismic slip is unevenly correlated with the long-term fault segmentation seen in the morphology. The coseismic slip distribution reproduces sharply the shape of the Sapanca-Akyazı and the Karadere segments along the eastern part of the rupture, but the boundaries between individual

segments are not visible in the slip distribution for this earthquake around the high-slip region around Gölcük. Brecciation mechanisms across fault jogs at segment boundaries may explain such features (Sibson, 1986). Then it seems possible to make a distinction between two different modes of rupture: An overloaded segment mode in Gölcük, which is capable of “erasing” the jogs at segment boundaries, and a critically-loaded segment mode, which prevents the segment boundaries from high coseismic slip so that its long-term shape is preserved. Both have been favourable to the propagation of the Izmit rupture over its 160-km length along strike. However, the rupture stopped at the eastern end of the Karadere segment and the Mw 7.2 Düzce earthquake ruptured the next individual segment, three months later, with a slip distribution comparable to that of the Sapanca-Akyazı segment (Akyüz et al., 2002). Thus, this latter segment also ruptured apparently under a critically-loaded mode. This shows that even under sufficient tectonic load, individual segments may or may not rupture in a concatenation of sub-events involving their neighbor segments. Conversely, it seems unlikely that the segment boundaries would have been enough to arrest the Izmit rupture inside the region of inferred large slip deficit and elastic overload, which may have existed in Gölcük before the earthquake. In other words, once triggered, the Izmit earthquake could not have been smaller than the size of the overloaded region around Gölcük. There the slip deficit had probably grown larger than the slip that any of the small individual fault segments visible at the surface could undergo alone, without having a high associated stress drop and producing high stress concentrations at the segment edges. The large coseismic slip in the overloaded region around Gölcük has also immediately triggered particularly large and fast after-slip in the velocity-strengthening region of the fault immediately below.

The particular conditions around Gölcük may have also influenced the rupture propagation. The very short S-P time (1.78 sec.), observed in a strong-motion station located beside the Sapanca-Akyazı segment 40 km east of the mainshock hypocenter, can be interpreted in two alternative ways: It could be the effect of either a supershear rupture propagation, or the triggering of an asperity by the P wave arrival from the hypocentre (Bouchon et al., 2001; Sekiguchi and Iwata, 2002). Both the inference of an overloaded region around the epicentral region and the observation of an extensional jog with less coseismic slip at the Sapanca Lake give support to the triggered asperity hypothesis, albeit without contradicting the supershear rupture propagation. The arguments above suggest that a heterogeneous slip and loading distribution along a large fault system like the North

Anatolian Fault may be determinant to allow propagation of large earthquake ruptures. For such a system, the notion of "characteristic earthquake" (Schwartz and Coppersmith, 1984) would apply only to the critically-loaded mode of rupture along individual segments. However, it will be very difficult to deduce from the surface slip distribution alone if contiguous segments with "characteristic ruptures" have ruptured together or not. In addition, the amount of coseismic slip on any segment will depend on variable degrees of slip deficit and load, or overload. Overloaded segments possibly undergo more slip than scaling laws would predict. These features are of concern to inferences of rupture length and moment magnitude for past earthquakes deduced from trenching. To describe distinct past events like the Izmit and the Düzce earthquakes would require relying upon the resolution of many measurements of slip along the fault trace and upon many well-resolved dates (provided that the events are separated by a reasonably long time interval).

Part of the heterogeneity in the loading along the NAF is likely to result from the fault segmentation, which may scale with the thickness of the seismogenic crust and may evolve as an effect of wear during progressive slip and fault growth (e.g., Scholz, 1987). However large stress heterogeneities (like in Gölcük) may also have grown-up and evolved from an uneven slip distribution during previous events. Thus the critical datum appears to be the distribution of slip deficit along the fault. For any segment along the NAF and at any time, the state of loading must integrate a complex slip history including sequences of earthquakes that probably never repeat the same way. The observations presented here give support to a variable slip model incorporating large earthquakes with variable magnitude and rupture length, which would result from unsteady segment-to-segment rupture propagation (from overloaded segments to critically loaded segments and vice versa).

2.9. Summary

Combining the SAR interferometry with the tectonic observations appears a powerful approach to resolve the first-order features of the slip distribution associated with the August 17, 1999 Izmit earthquake. In this work the fault geometry, the fault kinematics and the near-field deformation are resolved using particularly well-constrained tectonic observations collected in the field.

The SAR data set provide the best overall image of the surface deformation and appears to be the most appropriate to deduce an overall image of the static rupture at seismogenic depth. However, the possible occurrence of atmospheric artefacts may hinder a good solution. A critical analysis of the SAR data using a pair-wise

logic approach and independent meteorological data (from NOAA satellite images) is shown to allow to identify atmospheric artefacts and to remove them from subsequent modelling.

An overall forward modelling strategy, which combines a trial-and-error approach with a conventional inversion technique is used to calculate the slip distribution. Improving the data fits by steps seems more appropriate than uncontrolled inversion.

Slip is under-determined but the use of well-constrained measurements of slip at the earth's surface reduces considerably the range of possible solutions. The uncertainty in the slip estimates increases with depth. A good fit to the SAR data (RMS of less than 2 cm) corresponds to slip resolution of less than 0.5 m in the regions of the fault near the surface (12 km depth) and less than 1 m in the regions between 12 km and 24 km depth. However, the models cannot resolve slip of less than 1 m in regions of the fault at depth greater than 24 km.

The best fits to the SAR data define an inhomogeneous slip distribution with three main zones of higher slip along the fault and a total moment release of 2.3×10^{20} (Mw 7.6). The inhomogeneous slip distribution is robust and correlates well with the geometry of the fault segmentation, which is well defined from the morphology.

The Izmit earthquake rupture appears to have extended well into the eastern Sea of Marmara. The SAR data indicates that the Yalova-Hersek segment ruptured over 30-km length west of the Hersek peninsula, with slip tapering westwards from 2 m to zero and a moment release of 1.5×10^{19} Nm, equivalent to an event Mw 6.8. The western end of the rupture is located 40 km SSE from downtown Istanbul.

The ERS1 SAR interferogram -and thus the models- cover a "long-period" as they include the 29 days following the main shock and they may contain significant post-seismic deformation. Using the longer period model and the published GPS data describing the coseismic horizontal deformation (Reilinger et al., 2000) a slip model that represents better the coseismic slip alone is found. This model suggests that the moment release corresponding to the main shock is 1.9×10^{20} Nm (Mw 7.5), which is close to the seismological estimates.

The foregoing approach allows retrieving the early postseismic deformation that has been captured by the SAR data. The difference in moment release between the longer-period model and the seismological estimates appears explained by the occurrence of very fast aseismic after-slip reaching 2 m during the month following the main shock, within a zone of the fault located at 12-24 km depth, directly under

the zone of highest moment release in Gölcük. The Gölcük after-slip zone encompasses a region of the fault having substantially slipped coseismically, including the hypocentre.

Comparison of the after-slip retrieved from the SAR with the available GPS records of postseismic deformation (Bürgmann et al., 2002a) suggests that about half of the after-slip captured by the SAR data below Izmit-Gölcük may have occurred during the first two days following the main shock. The rapidity of the early after-slip would explain why it has not been incorporated in the models derived from the GPS data alone (Bürgmann et al., 2002a; Hearn et al., 2002). Accordingly, the maximum after-slip of 2 m at 16-18 km depth near the hypocentre would have started immediately after the mainshock at very fast rates (up to 1m in 2 days).

The correlation between the coseismic slip on the fault at depth and slip measured along the surface break is good. However, slip is unevenly correlated with the long-term fault segmentation: The Izmit earthquake slip distribution reproduces well the shape of some segments with a "characteristic rupture" (Sapanca-Akyazı and Karadere segments) but it has smoothed out the fault complexities at the boundaries between individual segments around the high-slip region of Izmit-Gölcük. This suggests that under the Gölcük region a large slip deficit and possibly a large elastic loading existed prior to the earthquake. This feature is consistent with the occurrence of both the rapid early after-slip under the overloaded region and the small events in this region in the years preceding the main shock.

The different correlation of the coseismic slip with the segment morphology also suggests a possible distinction between two different modes of rupture: A critically-loaded segment mode (in the Sapanca-Akyazı, the Karadere and the Düzce segments) and an overloaded segment mode (in Gölcük and surrounding segments). The heterogeneous coseismic slip and the state of loading on segments may result from a heterogeneous distribution of slip deficit accumulated during previous large earthquakes along the North Anatolian Fault.

3. ANALYSIS AND MODELLING OF ATMOSPHERIC ARTEFACTS IN THE COSEISMIC INTERFEROGRAMS OF THE AUGUST 17, 1999 IZMIT EARTHQUAKE

3.1 Introduction

Synthetic Aperture Radar interferometry (InSAR) is a powerful tool in mapping surface deformation with sub-centimetre accuracy and fine resolution over a wide area (Massonnet et al., 1993; Zekber et al., 1994; Peltzer & Rosen, 1995; Meyer et al., 1996; Wright et al., 1998; Bürgmann et al., 2000). There are however some limiting sources of error in interferometric SAR measurements. The accuracy of SAR measurements is influenced by several factors such as orbits, topography and the atmosphere (see Chapter 1). Of these, atmospheric effects are assumed to be one of the most limiting factors of differential interferometry as atmospheric artefacts are often difficult to recognize and correct (Zekber et al., 1997; Massonnet & Feigl, 1998; Delacourt et al., 1998; Rigo and Massonnet, 1999; Fruneau et al., 1998; Fruneau & Sarti, 2000; Beauducel et al., 2000; Feigl et al., 2002). Artefacts occur if atmospheric conditions such as pressure, temperature and, principally, water vapor, are not the same at the time of each acquisition of the two images from which an interferogram of phase difference is calculated (Tarayre and Massonnet, 1996; Hanssen et al., 2000). Spatial changes of % 20 in relative humidity can cause a phase delay of 10 cm (Zekber et al., 1997).

The atmospheric contributions on interferograms were first characterized and modeled in Mount Etna (Delacourt et al., 1998; Beauducel et al., 2000). Using more than 200 interferograms, Beauducel et al. (2000) finds that the atmosphere with a stratified structure might originate phase delay reaching as much as 2.4 fringes (6.7 cm). The biggest disadvantage of studying these artefacts in Etna is that both artefacts and ground deformation (i.e. deflection) give rise to fringes that are almost identical in shape, making rather difficult to differentiate atmospheric effects from surface deformation. In fact, one important characteristic of the atmospheric artefacts is that they are correlated to the topography and they mimic it (e.g. Massonnet & Feigl, 1998). Moreover, this phenomenon is the main source of error in measurements of subtle deformation such as subsidence, inter-seismic loading and

post-seismic deformation where strong topography is present (Lasserre et. al., 2001; Wright 2001b).

In order to highlight the possible confusion made between atmospheric artefacts and surface deformation the ERS1 co-seismic interferogram of the August 17 Izmit event is used. Atmospheric artefacts due to heterogeneous troposphere were identified in Chapter 2. Here in this Chapter, the atmospheric artefacts with topographic correlation that result from homogeneous changes in the lower troposphere are investigated only. This is achieved by subtracting the synthetic co-seismic interferogram from the observed data and by analyzing the residual signal versus the topography. However, as will be demonstrated such artefacts can be subtle to detect and thus might be misleading when interpreting small to large scale deformations seen in interferograms of displacement fields.

3.2 Identifying Atmospheric Artefacts

Leaving aside the main (fringe) features of the co-seismic interferogram discussed in Chapter 2, some small scale distortions in fringe pattern are present at two particular places, west of the Iznik lake and in the Mudurnu valley to the east where fringes are clearly deflected (white arrows in Figure 3.1a). These anomalies were previously interpreted as secondary tectonic motions on adjacent faults possibly induced by the main rupture (Armijo et al., 2000; Wright et al., 2001a; Feigl et al., 2002). After the Izmit earthquake, some open cracks were observed along the trace of the Mudurnu fault, which had ruptured in 1967, but no sign of an earthquake surface break was reported in the Geyve basin area. Wright et al. (2001a) have modelled these features of the interferograms with a variety of fault kinematics and have preferred models that surprisingly involve left-lateral strike slip (opposite to the known sense of slip on those faults), which would have been triggered by the main shock. In a similar way, Feigl et al. (2002) have chosen to model deformation in the Mudurnu valley with right-lateral slip but in the Geyve area with left-lateral slip.

However, although less pronounced the same kind of patterns in fringes can also be seen elsewhere in the interferogram (particularly to the north of the epicenter). For example, a similar bending is present south east of the Lake Iznik. As shown in Figure 3.2a, bending and deflection of fringes appear to coincide with the mountain ridges and valleys (dashed lines) and the fringes hug (or mimic) the topography like contour lines. The artefacts cannot be derived from errors in the digital elevation model used to remove the topographic contribution to the interferogram since the

sensitivity of the interferogram to topography is very low due to high altitude of ambiguity (over 3300 m in the image center).

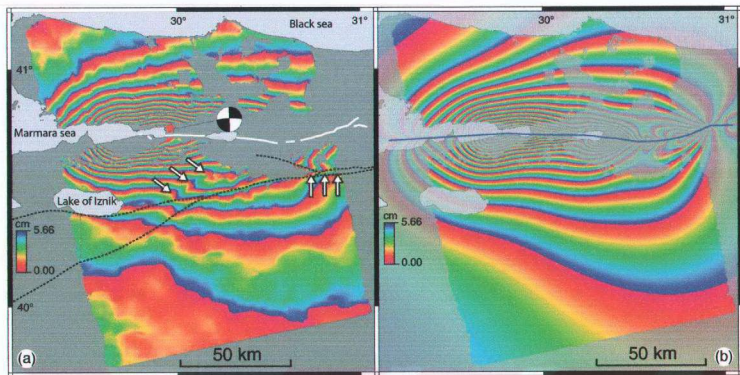


Figure 3.1 (a) ERS1 Interferogram showing surface deformation due to the August 17, 1999 Izmit earthquake (same as Fig. 2.3a). White lines show the mapped surface rupture (from Barka et al., 2002). Dashed lines are known active faults. White arrows indicate the location of some anomalies in the fringe pattern west of Izmit lake and along the Mudurnu valley. Star indicates the epicenter of the main shock with the focal mechanism solution from the Harvard CMT. (b) Synthetic interferogram (Model III, Fig. 2.8a) calculated using dislocations on rectangular faults in homogenous half space. Each fringe indicates 5.66 cm of surface displacement along the radar line of sight. Blue line is the surface trace of the model fault.

3.3 Evidencing Atmospheric Artefacts

Because the fringes arising from atmospheric artefacts are interfered with the ones due to co-seismic surface deformation, artefacts are not completely revealed and thus can not be calibrated very well. This is because the co-seismic signal is big with respect to the atmospheric one, which is not the case in other studies (Rigo & Massonnet, 1999). Therefore, the phase gradient due to the co-seismic surface displacement needs to be removed. This is done by subtracting the best model (Fig. 3.1b) from the interferogram (Fig. 3.1a), obtaining a residual interferogram shown together with topographic contours in Figure 3.2b.

The residual interferogram shows the misfit between the data and model that arises mainly from inaccurate modelling, large aftershocks, postseismic deformation and atmospheric artefacts. Inadequate modelling is apparent especially close to the fault between lake Sapanca and Gölcük where about 3 fringes of high gradient are seen. This is probably due to the complexities in the surface rupture geometry and coseismic slip distribution. The long wavelength (about one fringe) residual signal in

the northern side of the fault could also be due to the discrepancy between the real and modelled fault parameters or/and atmospheric artefacts that are not correlated with topography since the topography is fairly flat there. Such artefacts can result from horizontal heterogeneity of the atmosphere due to turbulent mixing (Hanssen, 2001). Residuals to the north of Izmit can be partly attributed to the small-scale surface deformation (~ 2 cm) caused by a large aftershock (13 September 1999, Mw=5.9) and detected by GPS observations (Ergintav et al., 2002). Because the co-seismic interferogram spans a period of about a month following the Izmit event, some of the residuals might be partly due to post seismic deformation that could not be taken into account in modelling such as pore-elastic rebound (Peltzer et al., 1996).

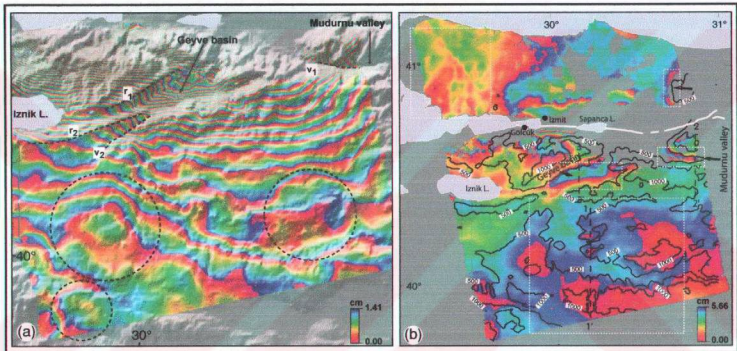


Figure 3.2 Evidence for atmospheric artefacts correlated with topographic elevation. (a) Coseismic interferogram shaded with digital elevation model. Dashed circles show some of the hills where residual signal hug the topography like contour lines. Dashed lines indicate the crest lines of mountain ranges (r) and valley axis (v) along which fringes are deflected abruptly. (b) Residual interferogram obtained by subtracting the model (Fig. 3.1a) from the data (Fig. 3.1b). Black lines are topographic elevation contours with an interval of 500 m. White dashed boxes show the locations where residual phases corresponding the same elevation are calculated to construct profiles shown in Figure 3.4b. White lines indicate the mapped surface ruptured.

Most of the residual signal appears to be due to atmospheric artefacts correlated with topography, which becomes more evident in regions with strong topographic variation (generally over 500 m of elevation) (Figures 3.2 and 3.3). For example, the correlation between the residual signal and topographic elevation is robust in the southernmost part of the interferogram where the residual phase amounts to about

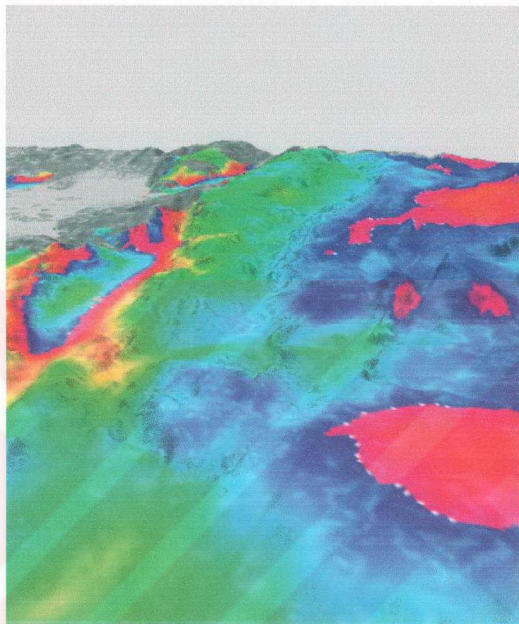


Figure 3.3 A 3D view of residual interferogram (Fig. 3.2b) draped over a digital elevation model. Note the correlation between the residual signal and topographic elevation (see also Fig. 3.2b).

4 cm up to 1300 m of elevation (Fig. 3.4b). There the residual signal mimics the topography very well. This can also be seen in Figure 3.4a (profile 1-1'), in which remarkable resemblance between profiles of residual signal and topography is illustrated. Residual signal in either side of the Mudurnu valley (profile 2-2') is nearly symmetric forming a trough whose axis coincides with the axis of the Mudurnu valley that is controlled by an active fault zone (Fig. 3.2b). In a similar way, the residuals in the Geyve area (profile 3-3') reveal the elliptical shape of the Geyve basin very well (Fig. 3.2b). The artefacts in the northern side of the surface rupture are relatively less pronounced since the topography is rather flat there. Nevertheless, the topographic correlation is evident to the north of the eastern

termination of the fault rupture where elevation exceeds 500 m (shown with an arrow in Fig. 3.2b).

3.4 Modelling Artefacts with a DEM

The delay in the radar signal is caused by an integration over the refractive index of the propagation medium along the line of sight. Horizontal and vertical heterogeneities in refractive index are influenced by the spatial distribution of water vapor, pressure, temperature, liquid water, and electron content (Hanssen et al., 2000). If the change in atmospheric conditions is spatially heterogeneous, modelling and removing the resulting artefacts can be only achieved by estimating propagation delay of electromagnetic waves using meteorological data set collected by a network of meteo-stations and/or GPS observations (zenith delay) over the image frame (Delacourt et al., 1998; Fruneau et al., 1998; Fruneau and Sarti, 2000, Hoeven et al., 2002; Webley et al., 2002). If, however, the change is horizontally homogeneous in the lower troposphere then phase difference will be directly proportional to topographic elevation (Fig. 3c), giving rise to fringes that look like topographic contours in the interferogram (Figures 2b and 3c). In such a case, the elevation difference required to create one fringe depends on the amount of phase shift due to different atmospheric conditions, not on the orbital baseline between the two images. Once calibrated, such artefacts can be modelled using a digital elevation model.

In areas where the correlation between topography and residuals appears to be strong, residual signals were measured in boxes (shown with dashed rectangles in Fig. 3.2b) and phases of the same elevations were averaged (Fig. 3.4b). As expected, the signal delay is not the same everywhere and varies with elevation indicating that the change in the atmosphere was not quite uniform, neither horizontally nor vertically in the scale of the image frame. While the signal delay is highest in the Mudurnu area where the residuals reaches over 2 fringes (~ 6 cm) up to 1100 m (Fig. 3.4b) it is less elsewhere with the same elevation. Nevertheless, although the phase-elevation ratio varies from one place to other, it appears to decrease with increasing elevation everywhere in a curvilinear manner. In the modelling, only the vertical variations are taken into account, the horizontal variations being ignored. A multi-layer atmospheric model as in Mount Etna can be used (Beauducel et al., 2000). However, in this study a curve of reverse exponential function is preferred. The function defined as $Phase = a * (1 - \exp^{-(b * Elevation)})$

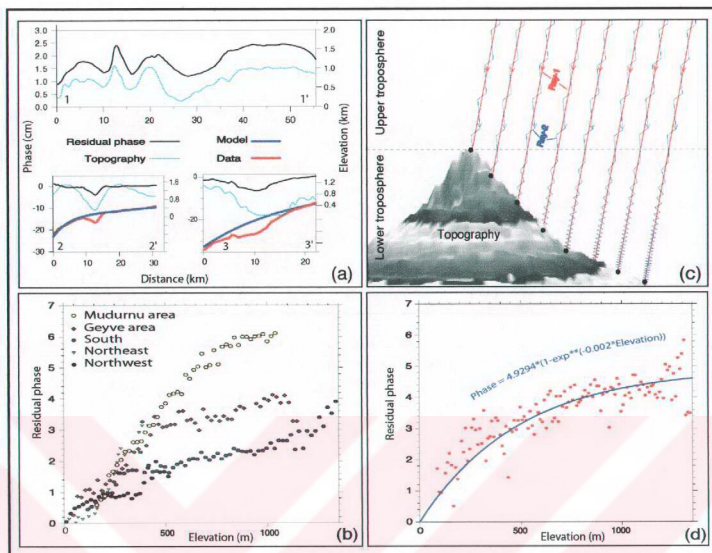


Figure 3.4 (a) Profiles of residual range-change and topographic elevation along the lines shown in Fig. 3.2b (1-1', 2-2' and 3-3'). (b) Plot showing the averaged residuals versus topographic elevation calculated at different sites within the interferogram. (c) Sketch illustrating the effect of horizontally homogenous changes in the atmosphere (modified after Massonnet and Feigl, 1998, fig. 7). It is supposed that after the acquisition of the first image (rays 1) humidity in the troposphere increases with decreasing elevation (rays 2). This will create a relative phase delay (illustrated as zigzags in rays 2) directly proportional to topographic elevation, giving rise to fringes that look like topographic contours in the interferogram (e.g. see Fig. 3.2b). Horizontally homogeneous phase delay above topography (here defined as upper troposphere) will not be detected. (d) Profile of averaged residual phase versus elevation calculated in the area south of the fault rupture. The black curve derived from a reverse exponential equation shows the relationship between topographic elevation and range change (delay) used in modelling the artefacts in this study.

depends on two parameters: a and b . To obtain an overall optimum model these two parameters were obtained by fitting a curve to a profile of elevation versus averaged residual signal for the entire southern part of the interferogram (black line in Figure 3.4d). The chosen curve thus gives a smooth model with a minimal misfit due to local variations in the phase-elevation ratio. Because the relationship between residual phase and topography is generally curvilinear, the misfit occurs where the gradient of residual phase varies significantly from that of the model, in particular, at lower elevations (mostly below 500 m). The resulting phase-elevation model

produced using a digital elevation model (DEM) is shown in Figure 3.5a. The DEM was filtered in order to avoid high frequency undulations in topography.

3.5 Correction of the Observed and Synthetic Interferograms

The interferogram obtained after subtracting the artefact model from the data is shown in Figure 3.5c. As seen in this figure, deflection of fringes in the Mudurnu and Geyve regions, as well as elsewhere in the southern parts of the data, is almost removed. Corrections can be improved locally. For example, deflected fringes across the Mudurnu valley can be further straighten if the model curve is fitted to the phase-topography profile measured there (Fig. 3.5f-2). In addition, “a” and “b” values of the reverse exponential function can be interpolated. The tests conducted show that such models do not improve the correction significantly as stated above the misfit becomes apparent where the gradient of the model curve differs considerably from that of phase-topography profiles. Therefore a simple, horizontally uniform model is kept also because of the subjectivity and complexities in interpolation of “a” and “b” values of the exponential function.

In order to clearly illustrate the effects and extend of the artefacts, the artefact model is added to the synthetic coseismic interferogram (Fig. 3.5e). As seen in Figure 3.5, resemblance of fringe pattern between the data and corrected models is striking; noise, bending and deflections seen in the entire data being almost reproduced.

It follows that the deflection of the fringes, which was previously interpreted as triggered slip in those regions (Armijo et al., 1999; Wright et al., 2001a,b; Feigl et al., 2002), is most likely reproduced by atmospheric artefacts. This is also supported by the fact that the shape of the residual signal in the Mudurnu valley cannot be produced by tectonic motions on the Mudurnu fault because the signal shows range increase on either sides of the valley along which the Mudurnu fault zone lies, rather than increase on one side and decrease on the other as one would expect from a tectonic slip on a strike slip fault. Secondary faulting may partly explains these anomalies but, it is unnecessary (Fig. 3.6). Apart from these secondary features, a large-scale correction is revealed in north easternmost side of the interferogram.

Fringes trend E-W in there (dashed circle in Fig. 3.5b), but after the correction (Fig. 3.5c) their trend becomes SE-NW closing around the eastern end of the surface rupture as predicted by synthetic model (Fig. 3.5d). When the possibility of atmospheric artefacts is not considered such an E-W trend of fringes implies that either the easternmost fault segment (i.e. Karadere segment) dips to the north (e.g.

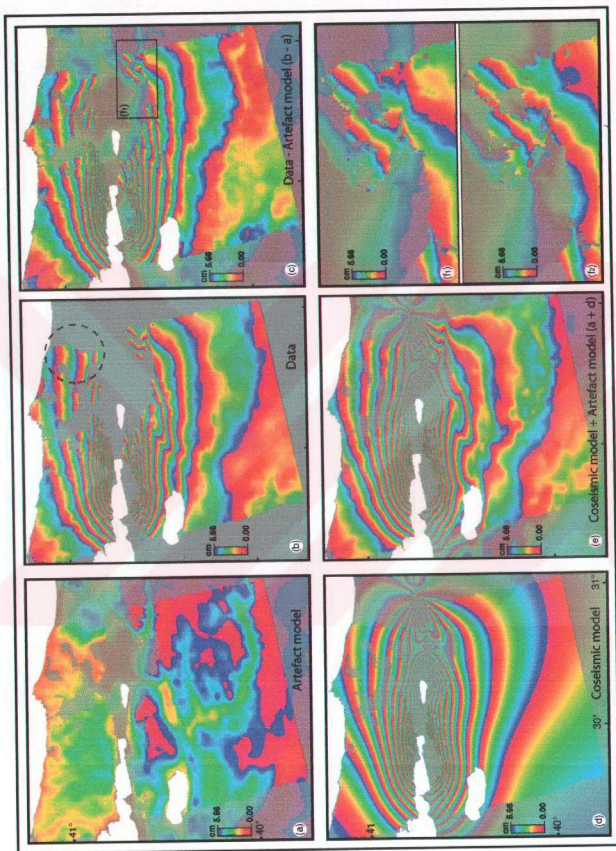


Figure 3.5 Modelling artefacts and correction of the observed data and the coseismic model. (a) Phase-elevation model (atmospheric artefacts), (b) Observed interferogram. (c) Correction of data by subtracting atmospheric artefacts. (d) Synthetic coseismic interferogram. (e) Correction of the displacement model by adding the atmospheric artefacts. (f) Correction of data in the Mudumu valley. (f1) blown up from c. (f2) Improving the correction using another artefact model with the two parameters of the exponential function, "a" and "b" are changed respectively from 4.9294 and 0.0020 to 7.2189 and 0.0023.

Wright et al., 2001a) or the fault rupture continues further east than it is observed in the field (Barka et al., 2002). Therefore, when modelling a large-scale deformation this phenomenon may also give rise to incorrect solutions.

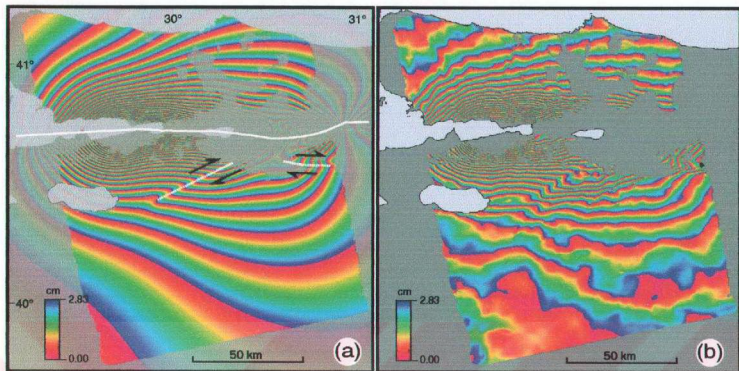


Figure 3.6 Modelling deflected fringes in the regions of Geyve and Mudurnu by elastic dislocations on secondary faults. For clarity, each color cycle represents here 2.8 cm of range change in the line of satellite sight (fringe frequency is twice as much as in previous figures). The observed fringes in the ERS1 interferogram are illustrated in the right panel (b) to facilitate comparison. (a) shows the synthetic fringes are obtained by adding right-lateral motion on two secondary faults to the model III (short white lines). The fault model requires the Mudurnu fault with 60°NNE dip, 10 km of length and 20 cm of right-slip between 1-15 km depth ($M_0=1 \times 10^{16}$ Nm; $M_w=5.8$). The NW Geyve fault is vertical with rupture 27 km long and 22 cm of right-lateral slip between 1-5 km depth ($M_0=4 \times 10^{17}$ Nm; $M_w=5.7$).

3.6 Discussion and Summary

The ratio between phase and topography is not linear, but decreases exponentially with increasing elevation from sea level to the top of the mountains. This is in good agreement with findings of a recent study in Etna reported by Webley et al. (2002) who investigated water vapor-based signal delay using GPS measurements along with SAR data although phase gradient in Izmit (reaching up to 6 cm of phase delay) is steeper than that found in Etna (4.35 ± 0.6) by Webley et al. (2002).

Despite of horizontal variations in atmosphere, as illustrated in Figure 3.4, a simple model (even a linear model) may account for most of the artefacts in an interferogram and answer whether or not a signal or an anomaly in the fringe pattern could be possibly related to atmospheric artefacts even though it exists in other interferograms. In this case, for example, the same anomalies with slight differences

are also present in the ERS2 co-seismic interferogram (Fig. 2.3b) and thus discriminating deformation from artefacts is not possible using pair-wise logic (i.e. cross checking of interferograms, Massonnet and Feigl, 1995b). It is, therefore, worthwhile to make a simple model of atmospheric artefacts before analyses of surface deformation and remove them from interferograms when possible. This is especially crucial when studying subtle deformation such as interseismic loading, post-seismic deformation and subsidence where strong topography is present.

4. KINEMATICS OF THE NOVEMBER 12, 1999 (Mw=7.2) DÜZCE EARTHQUAKE DEDUCED FROM SAR INTERFEROMETRY AND GPS MEASUREMENTS

4.1 Introduction

The Düzce area was struck once more on 12 November 1999 by a large earthquake, 87 days after the devastating 17 August 1999 Izmit (Mw=7.4) earthquake (Fig. 4.1) that occurred to the west in the Sea of Marmara region. The two adjoining earthquakes produced a 200-km-long surface rupture along the North Anatolian fault. The Düzce earthquake (Mw=7.2) is the latest strong earthquake in a sequence of westward migrating earthquakes since the 1939 great Erzincan earthquake (Mw=8.0). This sequence has been interpreted as triggering due to Coulomb stress transfer (Stein et al., 1997; Nalbant et al., 1998). In this context, the Düzce earthquake was expected by Barka (1999) who, after the Izmit earthquake, defined the Düzce area as a potential seismic gap taking in to account that the Düzce fault was the only segment of the North Anatolian fault that did not brake in this sequence. In addition to static stress interactions, possible effect of transient postseismic deformation of the Izmit earthquake on the Düzce event was explored using GPS data (Hearn et al., 2002). Detail analysis of the Coulomb stress triggering due to the Izmit event is discussed in Chapter 5 using the accurate fault parameters found here.

The most interesting feature of this earthquake is related to its rupture geometry, which appears to contradict with what is commonly observed in strike-slip earthquakes and tectonic field observations. First, compared to its magnitude, the length of the surface rupture mapped in the field was rather short (~35 km). Second, according to seismic and some geodetic observations (Ayhan et al., 2001; Bürgmann et al., 2002), the fault plane dips to the north at a very low angle ranging from 54° to 64° despite of its predominant strike-slip motion. In order to constrain the rupture geometry and deduce subsurface slip distribution we use SAR interferometry. An ERS coseismic interferograms is calculated and modelled using the same inversion technique described in Chapter 2. In addition, a set of coseismic GPS data published by Ayhan et al. (2001b) is also used along with the InSAR data. Taking into account the field observations and well located aftershock distribution,

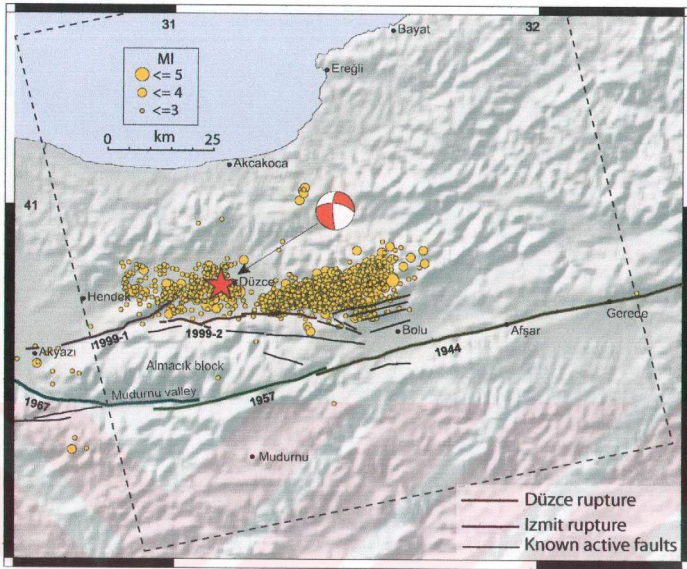


Figure 4.1 The 1999 and previous earthquake's breaks and neighboring active faults. Breaks of the Izmit (17/8/99) and Düzce (12/11/99) events are highlighted in purple and red, respectively. Star denotes the epicenter of main shock of the Düzce earthquake with the focal mechanism solution from Harvard CMT. Yellow circles are aftershocks recorded between 12/11/99 and 20/11/99 by the TÜBİTAK-MAM network (Özalaybey et al., 2000). The background DEM image is from GTOPO30. Dashed box shows the location of the ERS Interferograms (Frame 812, Track 114).

the two data sets are inverted separately and jointly for subsurface slip. Two models with different fault geometry are obtained and compared with previously published solutions for the same earthquake (Ayhan et al., 2001; Bürgmann et al., 2002b).

4.2 Tectonic Background and Field Observations of the Fault Rupture

The North Anatolian fault zone, expressed as a well defined single trace in the east, splays into two major branches west of Bolu (Fig. 4.1). The southern branch runs through the Lake Abant and Mudurnu Valley. The Düzce fault splays from the southern branch of the NAF in a complex, 10- to 20-kilometer-wide right step-over. Several intervening short faults are located within the step-over and serve to accommodate the transfer of slip between the northern (i.e. Düzce fault) and southern branches. To the west near Gölyaka, there is a sharp change ($> 30^\circ$) in

the strike of the NAF from E-W to NE-SW. This NE-SW trending portion of the fault is the Karadere segment that ruptured in the August 17, 1999 Izmit event. The southern branch was completely broken by The 1944 Bolu-Gerede, 1957 Abant, and 1967 Adapazarı earthquakes. In a similar way in the east, the two branches merge in a deformation zone between Akyazı and Sapanca, bordering the Almacık block. As stated in Chapter 2.2, with the 1999 earthquake sequence this block is now surrounded almost completely by recent fault breaks (Fig. 4.1).

Detail field observations describing rupture geometry and slip distribution were reported by Akyüz et al. (2002). Here, a brief summary relevant to this study is given. The rupture in general trends E-W and is approximately 35 km of long from Kaynaşlı in the east to Gölyaka in the west. Because of some political reasons, field observations were not allowed around the eastern termination of the rupture where a highway tunnel was under construction and thus the length of the rupture is only approximate. Official reports later acknowledged that the tunnel was partly destroyed by the earthquake, suggesting that the fault rupture probably continues through the tunnel reaching at least 40 km of long.

Abundance of man made features, such as roads and fences that were offset by the fault rupture made it possible to measure the displacements accurately and very frequently. The rupture is confined in a generally narrow deformation zone (5-50 m), and it can be divided into 4 sub segments considering the slip distribution and its geometry (Fig. 4.2). While the three segments in the east show almost pure right-lateral slip reaching up to 5.5 m in the middle one, the westernmost segment experienced an oblique slip with a (down-to-the-north) normal slip component reaching up to 3.5 m. The distribution of slip along the rupture is fairly simple with high slip occurring in the middle and tapering off symmetrically towards the both edges. Eastern segments are separated by two right step-overs of about one km wide, within which thrusting and left lateral displacements arising from block rotations were observed. The westernmost segment can be considered as an oblique transfer fault that connects the NE-SW trending Karadere fault to the E-W trending Düzce fault. This change in the fault geometry leads to formation of a releasing step-over zone, which explains the present day morphology in the Gölyaka area where the flat Düzce basin is bordered by the north-facing, steep fault scarps with triangular facets and the Lake of Eften in the immediate hanging-wall. It is very likely that this step-over may have acted as a geometric barrier to propagation of

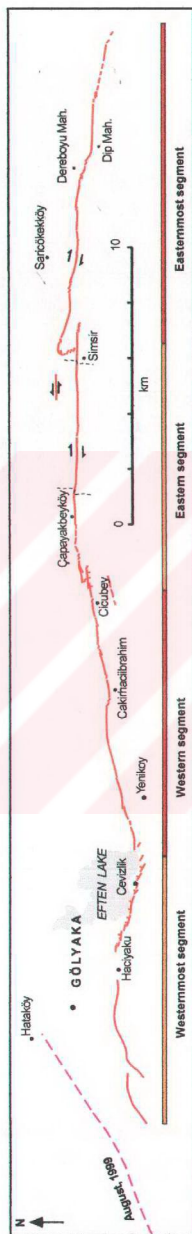


Figure 4.2 Surface rupture of the November 12, 1999 Düzce earthquake (from Akyüz et al., 2002).

August 17 Izmit earthquake (c.f. Barka and Kadinsky-Cade, 1988). After the Izmit earthquake, surface cracks were observed along the westernmost segment, which is interpreted as sympathetic slip on an adjacent fault (Hartleb et al., 2002).

Focal mechanism solutions of the Düzce earthquake indicate that the fault dips to the north at an angle ranging between 54 and 65 with a rake between 158 and 184 (Table 4.1). The fact that the majority of the aftershocks are located on the northern side of the fault and that the earthquake epicenter is located in the alluvial plain about 5-8 km north of the surface rupture also confirm the inference that the fault has a significant northward dip.

On the contrary to the seismological observations, except along the westernmost segment, no indication of such a low angle fault dip and significant rake were observed in the field. The north-dipping fault geometry also contradicts with previous studies (Şengör et al., 1985; Armijo et al., 1999), in which a south dipping transpressional fault is inferred. In the light of new results found in this study these contradictory views on the geometry of the Düzce fault (Şengör et al., 1985; Ayhan et al., 2001; Bürgmann et al., 2002b) will be addressed in Chapter 4.5.

Table 4.1. Parameters of the 1999 Düzce earthquake compiled from the following sources: UGGS: USA Geological Survey, HVD: Harvard University, CSEM: European-Mediterranean Seismological Center, OBN: Obninsk Seismological Observatory, Obninsk, Russia, GFZ: GeoForschungsZentrum Postdam (Bock *et al.* 2000), ERD: Earthquake Research Department (Disaster Affairs, Ankara), TUBI: TUBITAK-MAM (Özalaybey, pers. comm. 2002) .

Source	E°	N°	Depth (km)	Mw	Strike °N	Dip°	Rake°	Mo (Nm)
USGS	31.161	40.758	19.0	7.1	269	73	177	5.6 10 ¹⁹
HVD	31.250	40.930	18.0	7.2	268	54	167	6.7 10 ¹⁹
CSEM	31.161	40.758	15.0	7.0	282	74	170	4.1 10 ¹⁹
OBN	40.758	31.161	15.0	7.3	260	53	175	9.1 10 ¹⁹
ERD	31.21	40.79	11.0	7.2	-	-	-	-
GFZ	-	-	-	7.1	264	64	184	4.6 10 ¹⁹
TUBI	31.16	40.83	9.6	-	-	55	-	-

4.3 InSAR Data

4.3.1 Data processing

Three ERS images were requested and processed, one about a month before the earthquake and one tandem pair ten days after. The images were acquired at the ascending pass of the satellite (Fig. 4.1). Two coseismic interferograms were calculated using the ROI_PAC software package developed jointly by the Jet Propulsion Laboratory of NASA and the California Institute of Technology (Rosen et

al., 1996). To increase the coherence the interferograms are processed at 2 by 10 looks, resulting in a pixel size of 40x40 m (20x2-4x10). A 30 second posting DEM (USGS GTOPO30) is simulated to remove the topographic contribution from the interferograms.

One of the interferograms (which is not shown here) has a very low coherence (pair 2, see Table 4.2). This low level of coherence can be attributed to the small altitude of ambiguity, h_a of the pair (37 m, in the scene center). In other words, decorrelation occurs because the digital elevation model used to remove the topographic contribution to the interferogram is rather rough. The altitude of ambiguity of the second interferogram is relatively high, about 165 m in the center of the scene and thus coherence is better than the first interferogram (Fig. 4.3). Nevertheless, artefacts resulting from errors in the DEM appear as high frequency short wavelength signals in the areas with rugged topography, particularly southeast of the fault rupture (Fig. 4.4c).

Table 4.2. Details of the image pairs used for constructing coseismic interferograms and their corresponding values of altitude of ambiguity (h_a). B_⊥ represents perpendicular baseline component of orbital separation.

Pair	Track	Frame	Orbit-1	Date-1	Orbit-2	Date-2	B _⊥ (m)	h_a (m)	T. Baseline (day)
1	114	812	23014	14-Sep-1999	43689	22-Nov-1999	67	165	69
2	114	812	23014	14-Sep-1999	24016	23-Nov-1999	258	37	70

4.3.2 Data analysis and interpretation

The coseismic interferogram is shown in Figure 4.3 with wrapped fringes, each representing half a wavelength (i.e. 2.83 cm) range change along the radar line of sight. If we assume that all the displacement is pure right lateral slip then each fringe represents 7.24 cm of horizontal displacement. Because the interferogram spans a period of about a month before and 10 days after the Düzce earthquake, it may also contain some transient postseismic deformation following the Izmit and Düzce events.

Correlation is completely lost in the flat Düzce plain probably owing to a combination of high gradient deformation close to the rupture, and change in water content and in vegetation cover due to high agricultural activity. It is also generally inadequate in regions of rugged topography, particularly around the Almacık block in the south and in the northeast of the fault rupture. Correlation is preserved mainly in the southern

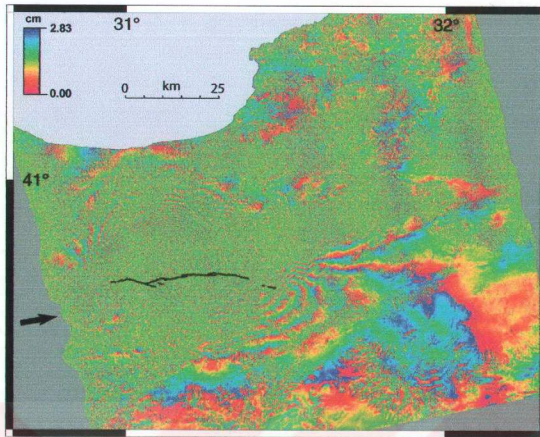


Figure 4.3 Coseismic interferogram of the November 12, 1999 Düzce earthquake. Each fringe represents 2.83 cm of range change along the radar line of sight. Surface rupture is shown with black lines. Radar look direction is indicated by a black arrow.

parts of the interferogram, but the concave fringes in the north-northwest of the surface rupture can still be seen. In the south, seven continuous fringes representing 19.81 cm of range change along the radar line of sight can be counted, where as in the north at least 15 fringes are counted.

The surface rupture, as in the case of the 1999 Izmit event, trends roughly East-West, sub parallel to the radar look direction (black arrow in Figure 4.3). Therefore, a dominant strike-slip movement on a steep, near vertical fault is expected to produce fringes that are nearly symmetric about the fault rupture. However, this is not the case; while the fringes in the south trend mainly E-W with low gradient, those in the north have a circular shape with high gradient. This apparent asymmetry confirms the seismological observations that suggest that the fault dips to the north and the displacement is not pure dextral-slip.

Because the topography is quite strong, especially in the south (Fig. 4.4a), it is possible that, as in the case of Izmit, atmospheric artefacts correlated with topography might exist in the interferogram. Therefore, assuming a constant

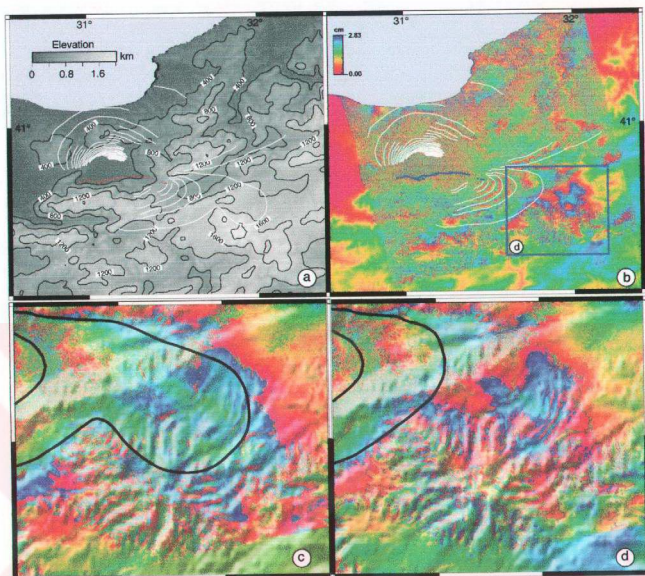


Figure 4.4 Correction of atmospheric artefacts correlated with topographic elevation. (a) Digital elevation model of the study area. Fringes in the observed interferogram are manually digitized and shown with white lines. Topography is stronger in the southern side of the surface rupture (red lines) than in the northern side and thus artefacts should not be significant in the north. (b) Calculating an artefact model (background image) and subtracting it from the observed interferogram (Fig. 4.3). The artefacts model is calculated using the digital elevation model (a) assuming a constant correlation rate of 1.44×10^{-6} , that is, half a fringe (1.44 cm) per 1000 m elevation. White lines indicate digitized fringes after correction. (c) The observed interferogram in the inset box shown in b. (d) Interferogram after correction blown out from b. Note that the cusp in the first fringe is removed after correction. Short wavelength signal (noise) is most likely due to errors in the digital elevation model used to remove the topographic contribution to the interferogram.

phase-elevation ratio several artifact models are calculated. These models are then subtracted from the observed data and the change in the fringe pattern is interpreted against the possible topographic correlation. A resulting interferogram is shown in Figure 4.4b. As seen in this interferogram the concave-westward cusp in the southernmost fringe (Figures 4.3 and 4.4c) appears to be associated with atmospheric artefacts. The fringes in the north are mainly located in the flat topography (Fig. 4.4a) and thus no significant change occurs when the artefact model is subtracted from the data.

4.4 Modelling with Elastic Dislocations

Because of its low coherence, the interferogram was not unwrapped. Instead, the visible fringes in the corrected interferogram were digitized and sampled (Fig. 4.4b). The sampled SAR data (1315 measurements) and three components of the coseismic GPS measurements (52x3 sample) are modelled using rectangular dislocations in an elastic, homogeneous, and isotropic half-space (Okada, 1985) as described in Chapter 2.4.

Fault geometry plays a key role in determining the subsurface slip when modelling geodetic data with elastic dislocations. The better the fault geometry is constrained the more accurate the solution that is found through inversions and forward modelling. Surface observations thus provide an essential piece of information and reduce considerably the range of possible solutions. The Inversions can allow one to estimate a simple fault geometry only. Complicated fault ruptures such as those involving multiple faults with varying dip and strike are difficult to estimate without additional information such as tectonics field observations and seismicity. First, a simple north-dipping fault is used to model the geodetic data and then a complicated fault geometry is sought.

4.4.1 Modelling with single-fault-rupture geometry

In the first stage of modelling, ignoring the complexities of the rupture geometry, a simple, continuous model fault is placed roughly along the surface rupture. The first-order model is obtained with a vertically uniform slip on 12 fault patches of 4-km-long along strike. Trending nearly in the E-W (N88°E) direction, the fault is 48-km long and 18-km wide.

To determine an optimum fault dip the InSAR and the GPS data are inverted separately and jointly using the same linear inversion algorithm described in Chapter 2.4. Keeping all the parameters fixed, the geodetic data set is inverted for uniform

slip on each patch with dips varying from 30° to 90° (Figure 4.5). As shown in Figure 4.5, the GPS data require a steeper dip angle (54°) than the InSAR data (67°). If the two data sets are jointly inverted weighting them equally, the optimum dip angle for the two data sets is found to be 62° . An overall optimum rake is also sought using the same approach above and found to be 10° . This ranges of dip angle and the rake are consistent with the independent seismological evidence (Table 4.1).

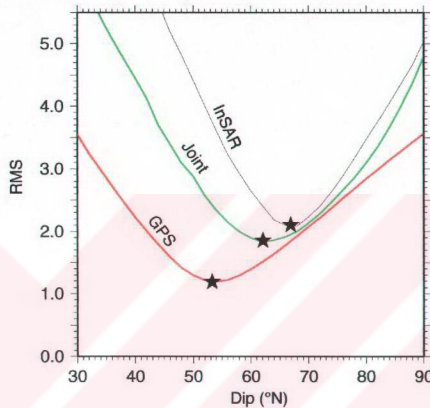


Figure 4.5 Plots of RMS versus fault dip for the Düzce rupture derived from separate and joint inversion of the geodetic data set using uniform slip model along depth. The GPS data require less, about 14° - 15° , fault dip than do the InSAR data. Optimal dip for joint data set is about 62° . Model with a 62° -north-dipping-fault geometry is used as starting solution for the variable slip inversions (see the text).

Therefore, both the seismological and geodetic observations suggest that the Düzce fault should dip to the north. Profiles of precisely located aftershocks (Özalaybey et al., 2000) taken roughly perpendicular to the fault are shown in Figure 4.6 with the 62° -northward-dipping model fault used in the inversions. The model fault appears to be well delineated by the aftershocks in the west. However, the distribution of aftershocks to the east, although rather clustered, suggests a steeper fault dip and possibly a more complicated rupture geometry that will be investigated later in the next section.

In the second step, the fault is further discretized into 4.5-km-wide patches along dip (48 patches 12 by 4 elements) for distributed slip inversions. The uniform slip model with a 62° dip is used as a starting solution in all the inversions. The uppermost

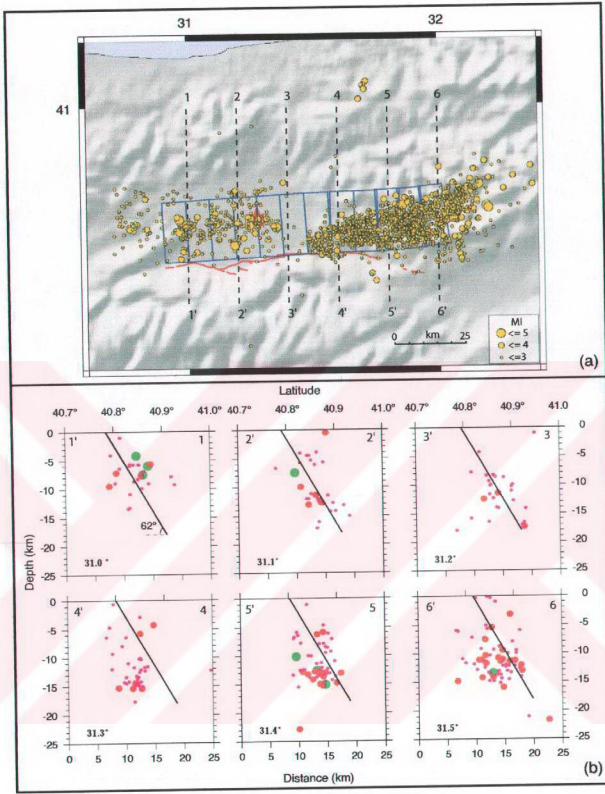


Figure 4.6 (a) Aftershocks recorded between 12/11/99 and 20/11/99 by TÜBİTAK-MAM (Özalaybey et al., 2000). Red lines show the surface rupture of the 1999 Düzce event. Blue rectangles are fault patches from surface to 18 km of depth with 62° N dip projected to the surface. (b) Cross sections showing the aftershock distribution at depth (locations are shown with dashed line in a) with the model fault used.

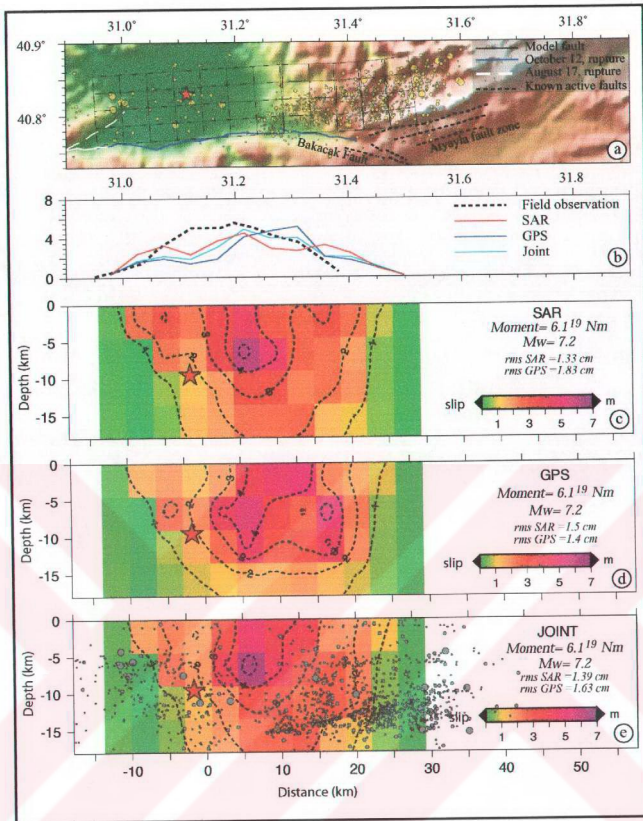


Figure 4.7 The Düzce earthquake fault trace, the coseismic surface slip and the modelled slip distribution at depth with single-fault-rupture geometry. (a) Shaded topographic map with simplified fault trace and the model fault projected to the surface. The model fault has 48 patches (4 km x 4.5 km) and dips 62° to the North. Locations of mainshock (red star) and of aftershocks (yellow circles) as in Fig. 4.6. (b) Surface right-lateral slip projected along the fault trace observed in the field is shown with a black dashed line. For comparison, surface slip predicted by SAR, GPS and joint inversions are also shown. Red star represents main shock hypocenter with the aftershocks shown with gray circles. (c) The InSAR model with a view from south to north. (d) The GPS model (data from Ayhan et al., 2001). (e) The slip distribution derived from joint inversion of InSAR and GPS data. Scalar moment (M_0) Magnitude (Mw) and RMS to the ERS and GPS data are indicated for each model.

patches are however constrained by the field observations because the resolution of the InSAR and the GPS data is relatively less in the near field. Because the starting model explains most of the InSAR and the GPS data very well (rms=2.1 cm), the inversions are stable and give similar solutions (Fig. 4.7).

While all the models fit fairly well to the InSAR data in the northern side of the fault, the fit in the south is not very good (Fig. 4.8). This poor fit can be attributed to atmospheric artefacts and the simple geometry of the model fault. Nevertheless, overall fit is within the uncertainty level of the geodetic data as the models explain

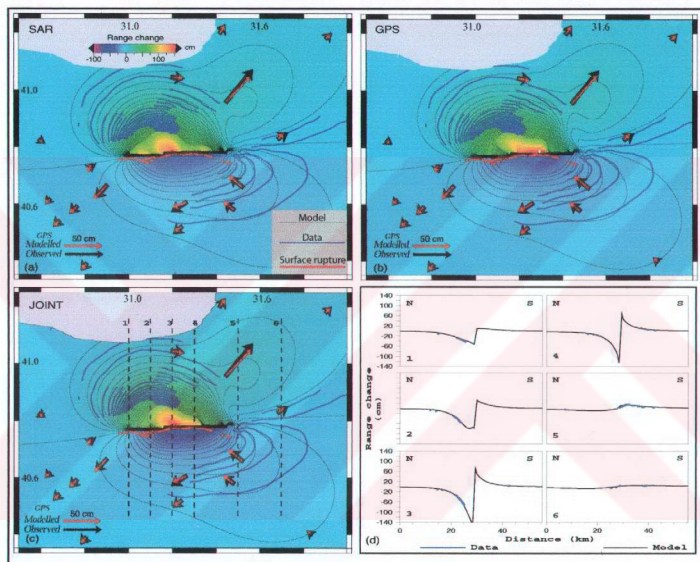


Figure 4.8 Comparison of the observed data and models with single-fault-rupture geometry. (a) SAR model. Background color map shows the modelled range change. Model fringes (thin gray lines) are contoured at every 2.83 cm to facilitate comparison with the observed fringes (thick blue lines). Black and read arrows show horizontal component of coseismic GPS vectors and modelled vectors, respectively. (b) GPS model. (c) Joint inversion model. (d) Profiles of observed (red) and modelled (black) range changes across the fault along the lines shown in c. See figure 4.7 for corresponding model of slip distribution at depth.

the InSAR data with an rms error of around half a fringe (1.41 cm). Good fit between the models and the observations is illustrated with some selected profiles of range change in Figure 4.8d. Some discrepancies between the observed GPS vectors

(black) and the vectors predicted by GPS model (red) are apparent, which is most likely due to the complexities in the rupture geometry and the slip distribution that are missing in the model parameterization.

Figure 4.7 shows the strike-slip distributions inverted from the InSAR data, the GPS data, and the joint data set. The overall pattern of slip distribution in all models is very similar. Both the GPS and the InSAR data suggest a similar shape and a similar amount of slip, and hence the same geodetic moments. Unlike the one found for the Izmit rupture (Fig. 2.6), the slip distribution predicted by the two data sets for the Düzce fault is fairly simple, coseismic slip reaching to a maximum around the fault center and gradually diminishing towards the edges and towards the deeper parts of the fault rupture. The highest coseismic slip predicted by the both data sets occurs within the upper 15 km of the seismogenic crust, and it reaches up to 6.5 m. The GPS model predicts less slip in the west, but more slip in the east compared to the InSAR model. The slip distribution obtained by the joint inversion simply lies in between the two models found from the separate inversions. The geodetic moment derived from the models is 6.1×10^{19} Nm (equivalent of $M_w=7.2$) and is comparable with seismological estimations that range between 4.5×10^{19} and 6.6×10^{19} Nm (Table 4.1).

4.4.2 Modelling with multiple-fault-rupture geometry

In the second stage of modelling, a rupture with more complicated geometry consistent with the geodetic, seismic and tectonic observations is sought. It is clear from the seismic and geodetic observations that at least some portion of the fault rupture must dip to the north at some depth. As stated earlier, the westernmost segment shows such a dip. However, models with faults dipping only along the westernmost segment do not satisfactorily explain the geodetic data. Therefore, a longer fault with a northward dip is required.

Westward propagation of the North Anatolian fault is known to have been guided by older deformation zones (Şengör and Yılmaz, 1981; Şengör et al., 1985; Elmas and Yiğitbaş, 2001). In other words, the present day trace of the North Anatolian fault coincides with the preexistent zones of weakness. According to Elmas and Yiğitbaş (2001), the North Anatolian fault (including the Düzce segment), in northwestern Turkey, forms the contact between two distinct tectonic units; the Sakarya block and the Western Pontite block. This zone is thought to have accommodated the subduction of the Sakarya continent underneath the Pontites during the Mid Jurassic time, and left-lateral motion between the two tectonic units later in the Late Cretaceous time (Elmas and Yiğitbaş, 2001). Therefore, it is plausible that the

Düzce fault is in fact vertical but, cuts an old north-dipping structure at depth. In such a case the Düzce earthquake might have been nucleated on a preexistent, north-dipping fault plane that could be an old thrust fault.

Accordingly, first a rupture geometry consisting of two intersecting faults is formed, and then to what extent such a rupture geometry can explain the geodetic data is explored using the same inversion procedure described above. The rupture is assumed to have been nucleated on a 50° -north-dipping plane and propagated upwards on the uppermost 6-8 km portion of the vertical Düzce fault. The two fault planes do not probably strike parallel to each other. However, because construction of a rupture geometry using rectangular patches of different strike and dip results in gaps and overlaps between the rectangular faults, the two faults are assumed to be parallel.

Two different fault geometries with an intersection depth of 6 and 8 km are shown in Figure 4.9. The geodetic observations are well met by the synthetic data derived from these model faults (Fig. 4.10). The models match with the InSAR data as well as do the previous models with the single-fault-rupture geometry (rms=1.4 cm). Overall, the observed GPS vectors are correctly reproduced by the two models (rms=2.1-2.3 cm), with the exception of a few stations (see Fig. 4.10). Although the predicted slip distributions are roughly similar to those obtained from single-fault-rupture geometry inversions, some important differences are present. Both the single-fault-rupture geometry models (MOD-I) and the multiple-fault-rupture geometry models (MOD-II) suggest a similar pattern of slip distribution along the western half of the fault. However, to the east the coseismic slip deduced from the MOD-II type models is less than that deduced from the MOD-I type models. The magnitude of slip in the MOD-II type models is much higher on the deeper parts of the rupture. The overall amount of slip and thus geodetic moment is slightly higher (%3-5) in the MOD-II type models (Fig. 4.9).

4.5 Discussion and Summary

The available geodetic data set can be explained by the models with either single- or multiple-fault-rupture geometry within the uncertainty level of the geodetic data set. Inadequate coherence of the InSAR data and lack of GPS points particularly in the near field allow models with both types of rupture geometry to fit the geodetic data within the acceptable limit of error.

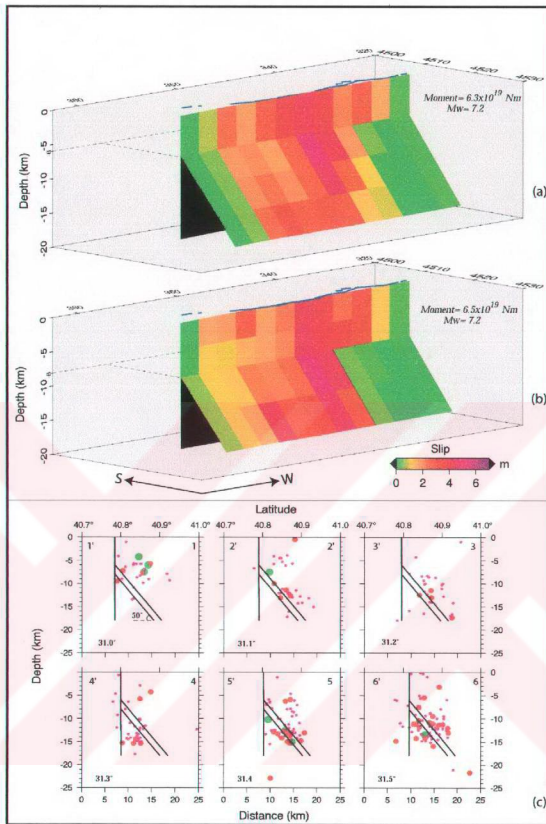


Figure 4.9 3D perspective views of the models with multiple-fault-rupture geometry with slip distribution on two intersecting fault planes (one vertical one inclined). View is from NE towards SW. The intersection depth is 6 and 8 km in the first (a) and second (b) model, respectively. Black surfaces are the areas of no slip on the vertical faults. Blue lines are the surface rupture. (c) Cross sections showing the faults both in (a) and (b) in relation to one month aftershock distribution at depth (locations of profiles are as in Fig. 4.6a).

Therefore, a better spatial coverage of geodetic observations is required for a better constrain on the fault geometry. Nevertheless, the MOD-II type models are preferred considering that the Düzce rupture with a predominant strike-slip trends parallel to the strike of the almost pure strike-slip, vertical fault rupture of the Izmit earthquake

to the west, and that the Düzce fault forms part of a major strike-slip plate boundary between the Anatolian block and the Eurasian plate. Accommodating a significant horizontal motion in the long term via a fault with a significant fault dip (as in the MOD-I type models) is mechanically difficult to explain. To some extent, this inference is also consistent with that of Şengör et al. (1985), in which a near-vertical south-dipping Düzce fault is suggested. Bürgmann et al. (2002b) inverted the same InSAR and GPS data jointly for subsurface slip using a single-fault-rupture geometry with 54° north dip and found a slip distribution similar to that found by Ayhan et al. (2001) who made use of GPS data only. Their fault geometry and the distribution and magnitude of slip are also in good agreement with the MOD-I types models found in this study. As they use single-fault-rupture geometry only, they have concluded that the geodetic data rule out a near-vertical geometry of the Düzce fault. However, as illustrated above, a composite fault rupture encompassing a vertical Düzce fault can too explain the geodetic observations reasonably well.

Although overall fit of the MOD-II type models to the geodetic measurements is reasonable, some discrepancies remain. Inadequate fit can be attributed to simple parameterization of the fault rupture, simplifying assumptions in calculations of surface deformation (homogeneous crust, no topography, elastic properties of rocks, etc.). For example, the strike of the model fault is rather inconsistent with the strike of the easternmost segment of the surface rupture. A good fit could not be obtained with model faults trending parallel to the strike of easternmost segment, suggesting that the rupture probably bifurcates to the east, and thus the composite nature of the rupture is more complicated than it is assumed in the models.

All the models indicate that the two geodetic data sets are in general consistent with each other. This observation in turn suggests that the contribution of the postseismic deformation (about one month) to the InSAR data is insignificant following the earthquake. Therefore, on the contrary to the 17 August 1999 Izmit event, a fast dynamic after-slip does not seem to have occurred following the Düzce earthquake. The differences between the GPS and InSAR derived solutions can be ascribed to different spatial coverage of the two data sets, possible orbital and atmospheric artefacts in the InSAR data and unrecognized errors in the GPS data.

Both the GPS and InSAR data suggest a longer (~15 km) fault rupture to the east than the one mapped in the field. This explains why the magnitude of the Düzce event is surprisingly high considering its short rupture length observed in the field.

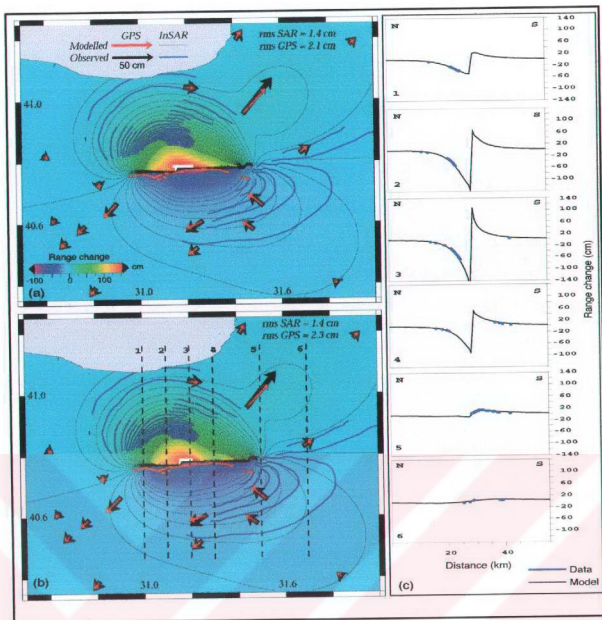


Figure 4.10 Comparison of the observed data and the models with a multiple-fault-rupture geometry derived from joint inversion of the GPS and InSAR data. (a) Model with two faults intersecting at 6 km of depth. Background color map shows the modelled range change. Model fringes (thin gray lines) are contoured at every 2.83 cm to facilitate comparison with the observed fringes (thick blue lines). Black and red arrows show horizontal component of coseismic GPS vectors and modelled vectors, respectively. (b) Model with two faults intersecting at 8 km of depth. (c) Observed (blue) and modelled (black) profiles of range change across the fault along the lines shown in b. See Figure 4.9 for corresponding model of slip distribution at depth.

5. COULOMB STRESS INTERACTIONS AND THE 1999 MARMARA EARTHQUAKES

5.1 Introduction

The August 17 ($M_w=7.4$), Izmit and October 12 ($M_w=7.2$), Düzce earthquakes occurred in eastern Marmara, causing an extensive destruction in a heavily industrialized and populated region of Turkey. The Izmit earthquake was not a surprise because the site has long been identified as a seismic gap (Toksöz et al., 1979). Taking into account the space-time migration of the earthquakes along the North Anatolian fault (NAF) in the 19th and 20th century, Toksöz et al. (1979) pointed out that the portion of the NAF (29° - 30° E) in the Izmit bay area posed a seismic hazard associated with an earthquake of magnitude 6 or greater. Recently before the Izmit earthquake, the progressive failure of the NAF particularly during the last century has also been interpreted in terms of Coulomb stress interaction (i.e. triggering due to increase in Coulomb stress) (Stein et al., 1997; Nalbant et al., 1998). Coulomb analysis of the westward migrating earthquake sequence since the 1939 Erzincan event (Fig. 5.1) (Stein et al., 1997), and analysis of historical earthquakes in the Marmara region (Nalbant et al., 1998) too showed that the gulf of Izmit was subject to an earthquake danger. As mentioned in Chapter 4.1, the Düzce earthquake was too expected by Barka (1999) who, after the Izmit earthquake, taking into account the earthquake sequence of the 20th century and its slip distribution along the NAF around the Almacık block (Figures 4.1 and 5.1), concluded that, not only the area to the western end of the Izmit rupture, but also the Düzce fault to the east might break in the near future.

In recent years, the analysis of Coulomb stress changes due to coseismic dislocation has been widely applied to investigate the variation in failure stresses on known faults (Harris and Simpson, 1992; Stein et al., 1992, 1994, 1997; King et al., 1994, 2001; Hubert et al., 1996; Harris, 1998; Nalbant et al., 1998; Hubert-Ferrari et al., 2000; King and Cocco, 2000). These studies show that earthquakes cause static stress changes on neighboring faults that may delay, hasten or trigger subsequent earthquakes. Therefore, determination of stress changes is important in seismic hazard assessments.

After the 1999 earthquakes, one or two earthquakes as great or greater than the Izmit earthquake ($M_w=7.4$) are now expected to occur within the submarine fault system that extends west of the Izmit fault under the Sea of Marmara, adjacent to Istanbul (Barka, 1999; Hubert-Ferrari et al., 2000; Parsons et al., 2000; Ambraseys, 2001; Atakan et al., 2002; King et al., 2001). Coulomb analysis of the 1999 Marmara earthquakes has been previously made by several workers (Parsons et al., 2000; Hubert-Ferrari et al., 2000; Pinar et al., 2001; Papadimitriou et al., 2001). However, different workers have used different fault parameters, and hence have found varying results. In this study, the fault geometry and the slip distribution used in the Coulomb stress calculations are those obtained directly from the InSAR and GPS modelling. As they explain the geodetic data very accurately (Chapters 2 and 4), maps of stress changes determined using them are thought to better represent the actual stress distribution. Here, the Coulomb stress changes due to four large earthquakes are calculated to map the static stress distribution on both the Izmit rupture plane and its surrounding regions prior to the Izmit event. Then, the stress changes caused by the Izmit event on the Düzce rupture are investigated.

5.2. Calculation of Static Stress Changes

When an earthquake occurs, it changes the state of stress on the nearby faults. In order to estimate the state of stress, the Coulomb failure stress is calculated using elastic dislocations on rectangular planes in a homogeneous and isotropic half space following Okada (1985, 1992). Change in the Coulomb stress $\Delta\sigma_f$ is given by

$$\Delta\sigma_f = \Delta\tau - \mu' \Delta\sigma_n \quad (5.1)$$

where $\Delta\tau$ is the change in shear stress (positive in the direction of slip) and $\Delta\sigma_n$ is change in the effective normal stress (positive in compression) on target faults. μ' is the effective coefficient of friction with range 0.0-0.8 (King et al., 1994). Here, the effective friction coefficient in all the calculations is assumed to be 0.4. A μ' of 0.4 minimizes the calculation error caused by the uncertainty in μ' to ± 25 percent (King et al., 1994). Failure is facilitated on specified or optimally oriented fault when the Coulomb failure stress, σ_f , rises. Unless specified the optimal fault orientation is defined by the given regional stress field (Anderson 1951).

The accuracy of Coulomb stress change due to an earthquake depends mainly on the accuracy of the source parameters of that earthquake (i.e. location and geometry of fault rupture, amount and sense of slip distribution). The more accurate the source parameters the more reliable results and thus interpretations can be

made. A reliable estimate of the slip distribution and fault geometry is therefore very important for stress transfer calculations. Small differences in slip distribution and fault geometry can lead to significant perturbations in the Coulomb failure stress. Further details of the technique can be found in King et al. (1994).

5.3. Coulomb Stress Field prior to the 1999 Izmit Earthquake

The Coulomb stress field caused by four large earthquakes that had occurred on the northern branch of the North Anatolian fault in northwestern Turkey prior to the 1999 Izmit earthquake is shown in Figure. 5.1a. Stress increase due to continues loading of the NAF (secular stress) is not taken into account because the history of the previous earthquakes on all the faults is not known. The total stress accumulation cannot be deduced because which historical earthquake broke which portion of the fault system particular in the Sea of Marmara region is not known very well. The four earthquakes are the $M_s=7.4$ 1912 Ganos, $M_s=7.3$ 1944 Gereede, $M_s=7$ 1957 Abant and $M_s=7.1$ 1967 Mudurnu earthquakes. Some smaller events, such as the 1935 west Marmara ($M_s=6.4$), 1943 Hendek ($M_s=6.4$), and 1963 east Marmara ($M_s=6.4$) earthquakes, are not taken into account because their fault parameters (in particular the location) are poorly known and their contribution to the Izmit earthquake is thought to be insignificant. One of these small earthquakes, however, is thought by King et al. (2001) to control the propagation of the Izmit rupture. As shown in Chapter 2, the Izmit rupture terminated about 30 km west of Hersek. According to King et al. (2001), the reason why the rupture stopped there, is that the western termination of the rupture is located in a stress shadow induced by the $M_s=6.4$ 1963 east Marmara earthquake. However, exact location of this event is a matter of debate (Nalbant et al., 1998). Seismic focal mechanism solutions indicate that it is a normal faulting event, but as the location of the event could not be resolved very well it is not known whether the event occurred on the north-dipping or south-dipping boundary fault of the Çınarcık basin. All the modelled earthquakes produced surface break and were mapped in the field (Ergin, 1969; Ambraseys and Zatopek, 1969; Barka, 1996; Ambraseys and Jackson, 1998; Altunel et al., 2000). Thus, their locations and surface slip distributions are well known. Moment magnitudes calculated from the source parameters used in Coulomb modelling are consistent with the seismological estimates.

The Coulomb stress distribution on optimally oriented strike-slip faults shown in Figure 5.1 indicates that the faults in the Sea of Marmara region are stressed by the previous earthquakes at two locations: the Izmit region in the east and the Ganos

region in the west. Stress increase in the epicentral area of the upcoming 1999 Izmit earthquake is about 0.3 bar. On the other hand, the Düzce area is located in a stress shadow.

The trend and type of the optimum faults are set indirectly by accepting a regional stress field in which the maximum and minimum stresses are horizontal with a compression axis (150 bars) trending N30°W. This definition results in a set of two conjugate strike-slip faults of optimal orientation at each calculation point (Figure 5.1b). One set of faults trends E-W consistent with the current tectonic regime. Therefore, the stress change along the fault shown in Figure 5.1 is not the real representation of the stress change along the entire fault as the fault strike deviates from its general E-W trend. Consequently, instead of calculating the Coulomb stress change on optimally oriented faults, stress changes resolved on the Izmit and Düzce ruptures themselves are calculated. Stress change at the center of each fault patch used in InSAR modelling is calculated and then all the values found are interpolated. The advantage of this method is that the spatial distribution of the resolved stress on the entire rupture can be visualized and thus variation of stress can be seen in 2-D (Fig. 5.2).

The shear stress imposed on the 1999 Izmit and Düzce ruptures due to the previous earthquakes (i.e. 1912, 1944, 1957 and 1967) is shown in Figure 5.2b (for simplicity the Düzce rupture is assumed to be vertical here). The sense and the magnitude of the shear stress on the 1999 rupture surface varies both along strike and depth (Fig. 5.2b). This is because the geometric relationship between the previous ruptures and the 1999 rupture changes from place to place. As the strike of the 1999 rupture and its location with respect to the previous ruptures change a variety of fault kinematics is promoted by the block motion induced by the previous earthquakes (Fig. 5.2a). Because the 1967 Adapazarı rupture strikes at an angle to the Izmit rupture and its western termination is located in the vicinity of Sapanca town, mostly left-lateral strike-slip with normal and reverse component is encouraged to east of Sapanca, whereas only right-lateral strike-slip with reverse component is promoted to the west Sapanca (Fig. 5.2a,b,c). While the shear stress being right-lateral (promote failure) or left-lateral (inhibit failure) depends on whether the 1999 ruptures are located to the west or to the east of the western end of the 1967 rupture, the shear stress being reverse or normal and hence the normal stress being compressive (clamping effect [inhibit failure]) or dilatational (unclamping effect [promote failure]) depends on

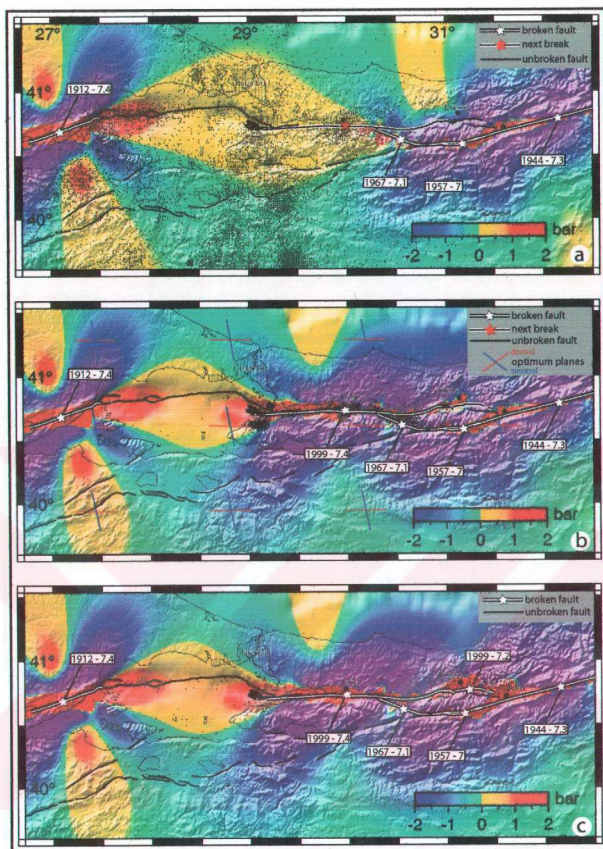


Figure 5.1 Coulomb stress field on optimally oriented vertical strike-slip faults (~E-W for right-lateral faults) at 10 km depth with a friction coefficient value of 0.4 and a compression axis at an azimuth of 330°. (a) Coulomb stress loading due to four large earthquake that occurred prior to the August 17, 1999 Izmit earthquake along the northern branch of the North Anatolian fault in the sea of Marmara region. Stress due to the secular loading is not taken in to account. Coulomb stress before the Izmit earthquake is enhanced in the Izmit and the western Marmara sea regions. Most of the seismic activity prior to the Izmit earthquake between 1967-1999 is located in regions of enhanced stress. (b) Change in Coulomb stress field after the Izmit earthquake. The event results in Coulomb stress increase of over two bars on the faults around the eastern end western terminations of the Izmit rupture. Located in the stress shadow before the Izmit event (a) the Düzce area is now situated in a region of enhanced stress prior to the November 12, 1999 Düzce event. (c) Coulomb stress field at present after the Düzce earthquake.

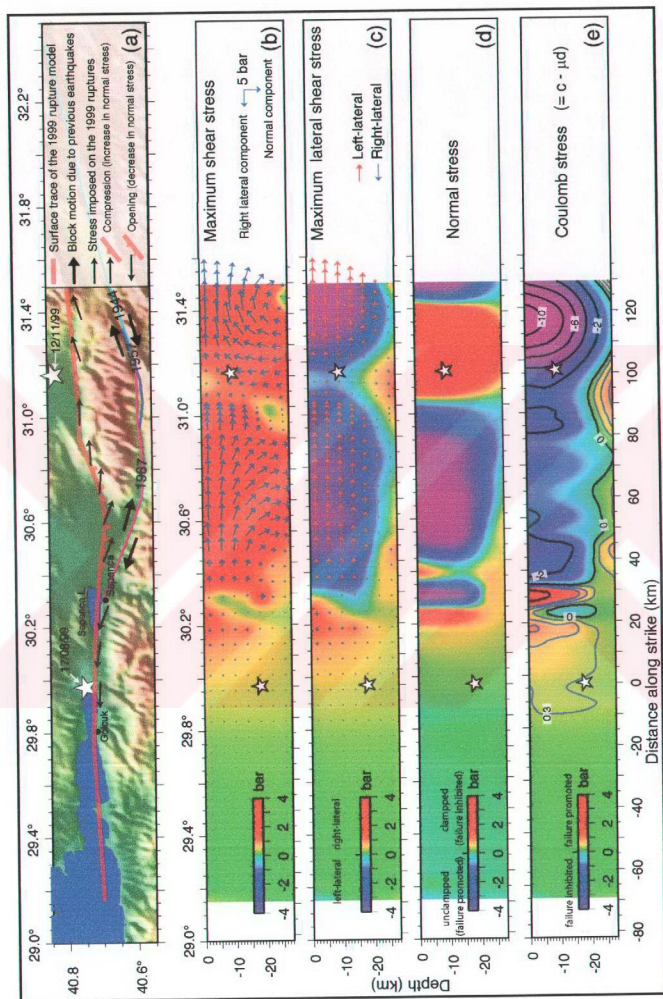


Figure 5.2 Resolved stress on the 1999 ruptures due to previous earthquakes (a) Shaded topographic map with 1999 model faults and previous fault breaks. (b) Maximum shear stress. Arrows indicate the sense and magnitude of the shear stress. (c) Lateral components of the maximum shear stresses. (d) Normal stress. (e) Coulomb stress.

the angle between the 1999 ruptures and the previous ones (Figures 5.2a, 2b and 2d). The distribution of the Coulomb stress can be divided into two distinct parts: (1) the west Sapanca section, on which stress is entirely increased and (2) the east Sapanca section, on which stress is mostly reduced (Figure 5.2e). A lobe of high stress increase reaching to 2.5 bars occurs in the vicinity of Sapanca, which is mostly due to the edge effect. Except this part, the increase in Coulomb stress along the rupture is quite low (< 1 bar). It is about 0.3 bar around the Izmit hypocenter. As mentioned above, the effective coefficient of friction, μ' , is assumed to be 0.4 in all the Coulomb stress calculations. Thus, if the calculations are made using different values of μ' , as illustrated in Figure 5.3, the distribution and the amount of the Coulomb stress change along the rupture surface will differ. When μ' is increased the Coulomb stress around the hypocenter decreases. On the other hand, with increasing values of μ' the Coulomb stress change along the Karadere segment and south of Adapazarı gets higher and higher and becomes positive (red) as a result of high normal stress decrease there (Fig. 5.3). On the contrary to the Karadere segment, with increasing values of μ' the Coulomb stress decrease along the Düzce rupture gets lower and lower, revealing the segmentation of NAF due to difference between the strike of Karadere and Düzce faults (Figure 5.3). Therefore, the clear difference between the Karadere and Düzce segments revealed in the distribution of the Coulomb stress change with increasing μ' may have been one of the factors that prevented the Düzce fault from breaking simultaneously with the Izmit earthquake.

With any value of coefficient of friction, it is clear that the Izmit earthquake nucleated in an area of enhance stress (Figure 5.3). If μ' is assumed to be 0.4, its rupture propagated also towards the east into the stress shadow. And thus, If that is a reasonable value for μ' , then the stress shadow in this region did not stop the rupture (although it might have hindered it). Propagation of the earthquake ruptures into the stress shadow is not paradoxical and can occur as observed in the case of the 2000 Hector Mine earthquake after the 1992 Landers event (Pollitz and Sacks, 2002; Fred and Lin, 2001).

Calculation of secular stress loading based on the modelling of interseismic GPS measurements (McClusky et al., 2000) shows that stress accumulation along the northern branch of the NAF is 0.37 bar per year (Fig. 5.4), in consistent with King et

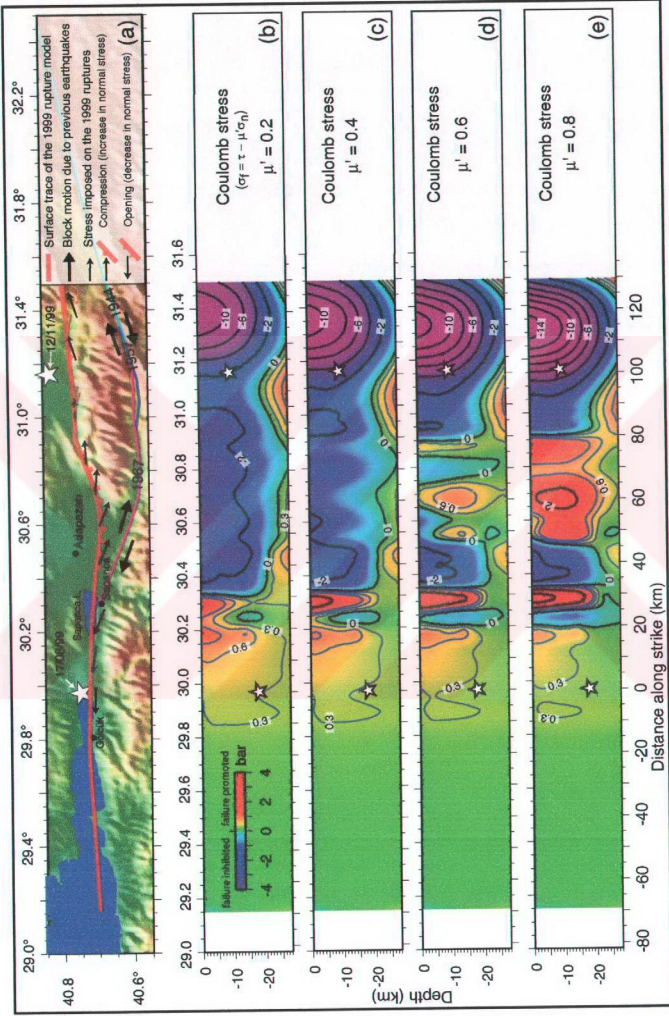


Figure 5.3 Coulomb stress resolved on the 1999 ruptures due to previous earthquakes using different values for the effective coefficient of friction, μ' .

al. (2001) who suggest 0.4 bar increase per year. Accordingly, ~ 0.3 bar of stress increase induced by the previous events at the hypocenter of the Izmit event is loaded by the continuous plate motion in about a year. Thus, the Izmit earthquake is weakly promoted by the previous earthquakes. One of the reasons for this low stress increase is the relationship between the Izmit and 1967 ruptures. Because the trend of the 1967 rupture is NW-SE direction (Fig. 5.1a), the Coulomb stress increase caused by this event is not significant on the E-W Izmit rupture. In addition, the slip on this rupture is assumed to be pure strike-slip, in consistent with the field observations. If, however, the event is in fact associated with some oblique normal component, then the amount of the Coulomb stress increase will be higher.

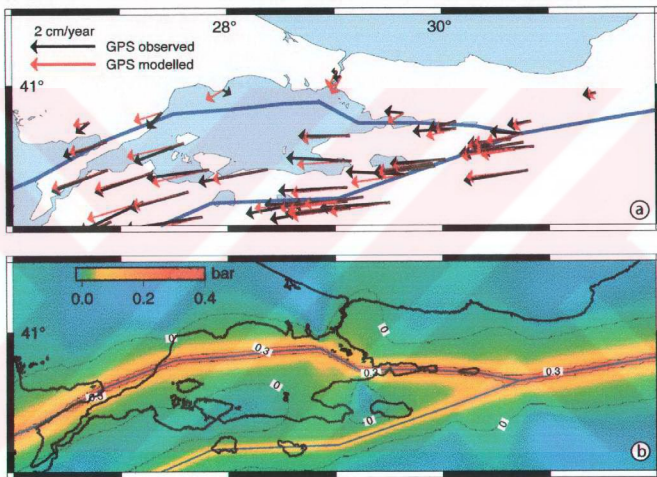


Figure 5.4 Stress accumulation induced by loading due to continuous plate motions in the Sea of Marmara region. (a) Modelling GPS measurements (McClusky et al., 2000) showing the motion of the Anatolian block relative to the Eurasian plate. Blue lines are the North Anatolian faults that bounds the two plates. Faults are assumed to be locked at 15 km depth, below which plate motion is continuous at a slip rate of 3 cm per year on the northern branch and about 0.8 cm per year on the southern branch of the NAF. (b) Annual stress loading at 10 km of depth derived from the model in (a), which is about 0.37 bar per year on the northern strand.

5.4 Coulomb Stress Changes Induced by the Izmit Event and its Effect on the Düzce Earthquake

As shown in Figure 5.5, there have been several Coulomb models of the Izmit earthquake published by different researchers. Although they are roughly the same,

there are some significant differences in the distribution and amount of static stress changes between the models. The differences arise mainly from the different rupture geometry and slip distribution of the earthquake used in the calculations. As a result, seismic hazard analysis based on each model will be different. Therefore, here the fault parameters of the Izmit earthquake used in the Coulomb stress calculations come directly from the modelling results of the geodetic data discussed in Chapters 2 and 4. As they explain the InSAR and GPS observations within the resolution uncertainties of the geodetic data set, maps of the Coulomb stress changes using such sources are more reliable.

The distribution of the Coulomb stress changes on a variety of optimum faults calculated in this study are shown in Figures 5.5a-c. The aftershocks are mostly located in regions of stress increase on optimally oriented strike-slip faults (Fig. 5.a), and thus they are most likely to be triggered due to the Coulomb stress transfer. It is interesting to see that although the Coulomb stress was increased by well over 5 bars in the Düzce region, seismic activity is very low there prior to the Düzce earthquake. The distribution of the cumulative stress changes due to the previous shows that the Düzce area is located in a stress shadow prior to the Izmit earthquake (Fig. 5.1a). However, the Izmit earthquake caused the Düzce fault to get out of the stress shadow (Fig. 5.1b) promoting the failure of the Düzce fault (Figures 5.1a and 5.1b).

As discussed in Chapter 4, the Düzce rupture is most probably associated with several fault breaks with different dips (see Fig. 4.9b). Therefore, it is better to reveal the Coulomb stress change resolved on the Düzce rupture. This time, the fault is divided into smaller patches (2x2 km) to 18 km of depth and the changes in the normal, maximum shear and Coulomb stresses due to all the previous events including the Izmit earthquake are then calculated for each patch (Fig. 5.6). The shear stress on the Düzce rupture caused by the 1944, 1957 and 1967 events is left-lateral (inhibit failure) and the normal stress that is about 2 bars around the hypocenter of the upcoming Düzce earthquake is mostly compressive (inhibit failure). Therefore, these events do not promote any earthquake on this fault as it is a right-lateral fault and the normal stress is high (Fig. 5.6a). The Coulomb stress decrease due to the previous events around the hypocenter of the upcoming Düzce earthquake is 1-2 bars before the 1999 Izmit earthquake. On the contrary to the previous events, the Izmit earthquake imposes right-lateral shear on the Düzce rupture and increases the Coulomb stress around the Düzce hypocenter 4-6 bars (Fig. 5.6b). Therefore, the Izmit earthquake gets the Düzce fault out of the stress shadow and promotes the Düzce earthquakes (Fig. 5.6c).

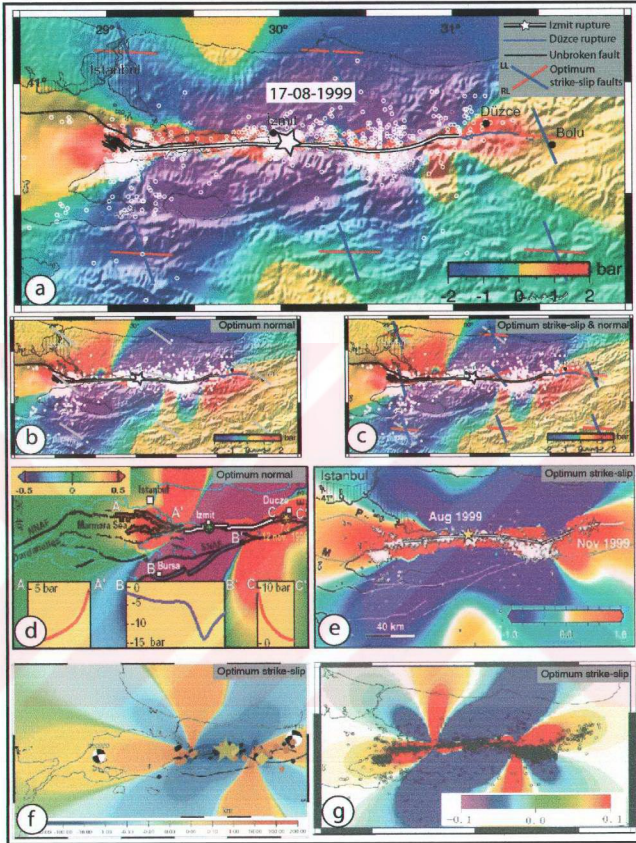


Figure 5.5 Stress changes associated with the Izmit earthquake. (a), (b) and (c) are stress changes on optimally oriented strike-slip, normal and the two together, respectively (this study). Fault parameters used in the calculation of the stress distribution are from InSAR modelling (see Fig. 2.6f). White circles are the Izmit aftershocks before the Düzce earthquake (from Kandilli observatory). (d), (e), (f) and (g), are previously published maps of stress changes due to the Izmit earthquake (b: Hubert-Ferrari et al., [2000], c: Parsons et al., [2000], d: Papadimitriou et al., [2001], e: Pinar et al., [2001]). The differences between all the models of Coulomb stress change result mainly from the different source parameters used in the calculations.

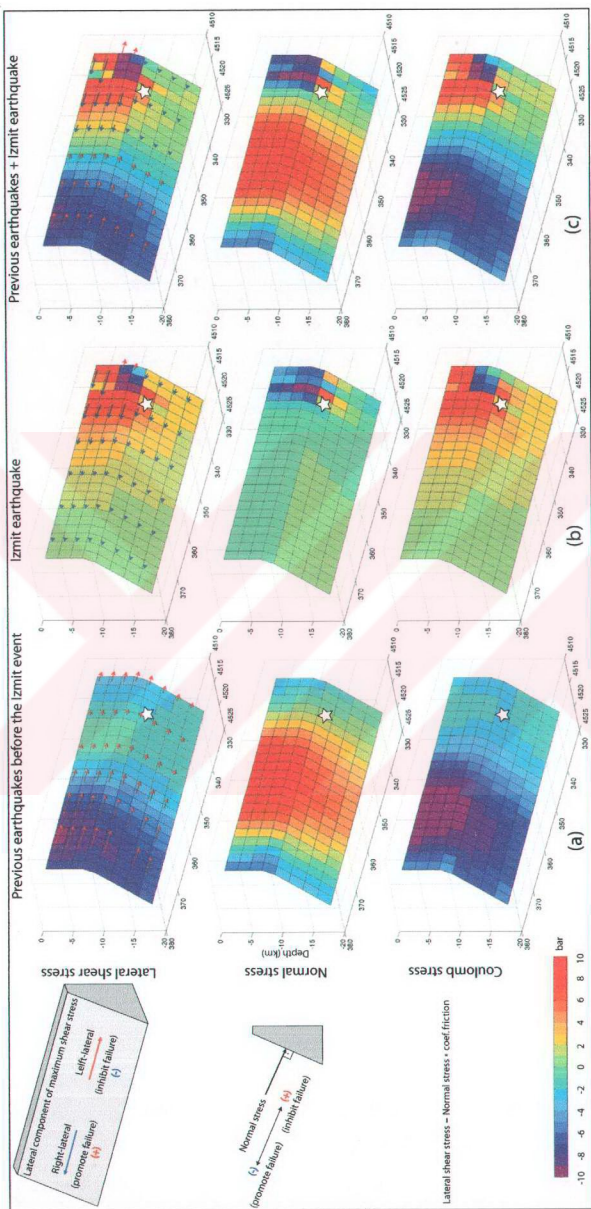


Figure 5.6 3D perspective views (from NW to SE) of the Düzce rupture (see Figure 4.9) with resolved stresses induced by the previous events prior to the Düzce earthquake (a). The 1999 Izmit earthquake increases the Coulomb stress around the hypocenter by 3-6 bars (b), which gets the hypocenter of the Düzce earthquake out of the stress shadow induced by the previous events (c).

5.5 Coulomb Stress Changes after the Düzce Earthquake

As shown in Figures 5.1b and 5.6, the 1999 Düzce earthquake occurred in a region of enhanced Coulomb stress caused by the Izmit earthquake. Using the source parameters deduced from geodetic data (the one with a multiple-fault-rupture geometry [see Fig. 4.9b]), the Coulomb stress changes caused by the Düzce earthquake are calculated and shown in Figure 5.6. Owing to its complex rupture geometry, the Düzce earthquake caused a stress increase over two bars around the epicentral area of the earthquake, inducing high seismic activity there. The earthquake also results in high stress increase around the eastern and western terminations of the Düzce rupture. When the stress changes due to the previous earthquakes are taken into account, the areas to the west of the Düzce rupture is located in a stress shadow (Fig. 5.1c). However, the area of enhanced stress to the east of the rupture still remains because of the 12-15 km-long gap between the 1944 rupture and the eastern termination of the Düzce rupture (Fig. 5.7). As discussed in Chapter 4.1, the Düzce fault splays from the southern branch of the NAF in a complex step-over, within which several intervening short faults accommodate the transfer of slip between the northern (i.e. Düzce fault) and southern branches (Fig. 5.7). Detail paleoseismological studies of Barka et al. (2001) suggest that this area should not be considered as a potential seismic gap that could produce events larger than magnitude 6.

To reveal the distribution of the Coulomb stress change on the main Çınarcık fault (i.e. the northern boundary fault of the Çınarcık basin), first a 34-km-long fault with 2x2 km patches to the depth of 20 km (170 patches in total) is formed and then the Coulomb stress change on each fault patch due to the Izmit earthquake is calculated (Fig. 5.8). The location and the geometry (dipping 85° SW) of the fault are constrained with the high-resolution bathymetry data and deep seismic profiles (Le Pichon et al., 2001; Armijo et al., 2002; Singh et al., 2002). As shown in Figure 5.8, the Coulomb stress change is maximum (over 5 bars) close to the end of the Izmit rupture and decrease westwards and downwards. The shear stress induced by the previous earthquakes (mostly Izmit event) on the main Çınarcık fault is dominantly strike-slip with minor normal component. Therefore, the strike-slip aftershocks that have occurred along this fault (Özalaybey et al., 2002) were most likely triggered due to transient stress changes induced by the Izmit earthquake and thus, they do not necessarily confirm long term kinematics of this fault.

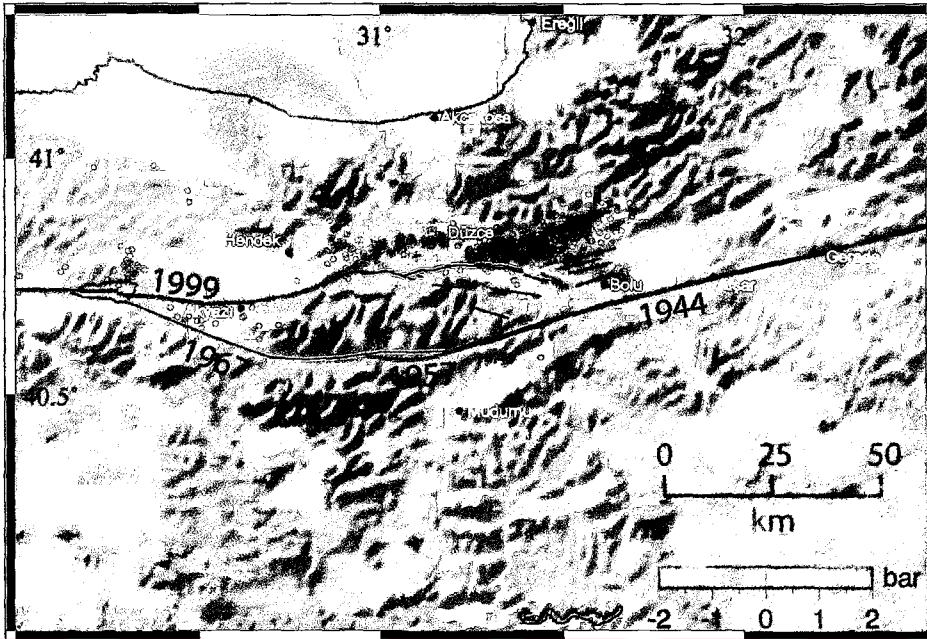


Figure 5.7 Coulomb stress changes on optimally oriented strike-slip faults due to the November 12, 1999 Düzce earthquake. Aftershocks recorded between 12-11-1999 and 20-11-1999 by TÜBİTAK-MAM (Özalaybey et al., 2000) are shown with black circles. Fault geometry and slip distribution used in the stress calculation is the one deduced from InSAR and GPS modeling in which a multiple-fault-rupture geometry is used (see Figure 4.9b).

5.6 Discussion and Summary

At present, the Sea of Marmara region is located in an area of enhanced stress increase due to the large earthquakes ($M_s > 7$) since 1912 Ganos event (Fig. 5.1c). The 1912 and 1999 events, in particular, increased the static stress over 5 bars on the submarine fault system in the east and west, respectively. A stress increase of 5 bars corresponds to an increase normally accumulated in about 12 years by secular loading due to the continuous plate motion. In other words, the previous earthquakes brought forward the next earthquake in the Sea of Marmara by 12 years. The faults in this region therefore pose a serious seismic hazard particularly for Istanbul where over 12 million people live.

Although it is under debate (Le Pichon et al., 2001), detail studies of Armijo et al. (2002) based on the high resolution bathymetry data and deep seismic profiles suggest that the NAF in the Sea of Marmara is fragmented into three segments. If this is assumed to be the case and one or two segments simultaneously may break, then the question is whether the future earthquake will occur in the eastern or

western Marmara region. Considering the westward migration of earthquakes since the 1939 Erzincan event, one can suggest that the earthquake will likely occur in the eastern Marmara region. However, history of the large earthquake that occurred in the Sea of Marmara before 1912 must be better known to answer this question with confidence. Historical records about the past earthquakes give clues about which earthquake broke which fault (Ambraseys and Finkel 1991; Ambraseys 2001), but are inadequate. Submarine studies on the Marmara faults by a CNRS-TUBITAK (French-Turkish) cooperation are aimed to reveal the history of the earthquake in this region.

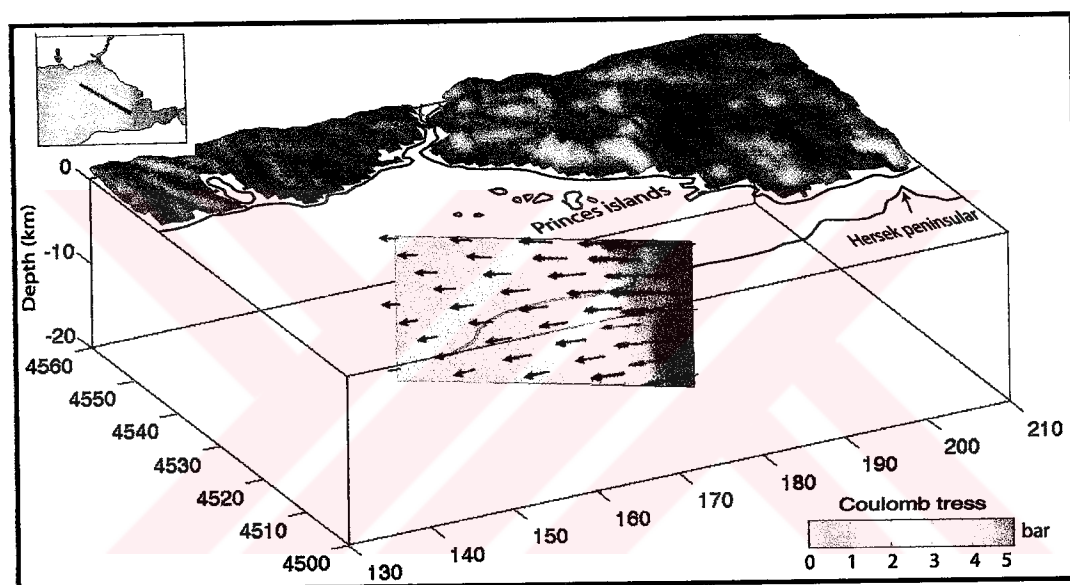


Figure 5.8 A 3D perspective view of the main Çınarcık fault with the Coulomb stress change resolved on it from the 1999 Izmit earthquake (coordinates are in km with UTM projection). Black arrows indicate the direction and magnitude of the maximum shear stress resolved on the fault plane, which shows oblique loading. The inset shows the map view of the fault with coastlines. The fault is 34 km of long, 20 km wide and dip 85° to the south.

As shown in Figure 5.5, several workers have modelled the Coulomb stress changes caused by the Izmit earthquake, leading however rather different results. This is mainly because source parameters of the event used in these Coulomb stress calculations are quite different. Accordingly, in Coulomb based seismic hazard studies, if available, well-constrained source parameters should be used. Location of aftershocks after an earthquake can be better predicted by using better source parameters of that earthquake. However, Coulomb models that predict the location of aftershocks fairly well do not necessarily suggest that they represent the actual distribution of stress change. For example, dismissing the segment west of

Hersek deduced from geodetic data (Reilinger et al., 2000; Armijo et al., 2000; Wright et al., 2001a), Pinar et al. (2001) prefer a much shorter coseismic rupture for the Izmit earthquake because their Coulomb model with such a short rupture predicts the aftershocks, particularly those around Yalova, better than the model with a longer fault rupture that continues west of Hersek. When discussing the location of aftershocks in relation to the Coulomb stress changes, what is commonly forgotten to be taken into account is the kinematics of the aftershocks. The distribution of the Coulomb stress changes on different type of optimal faults (i.e. strike-slip, normal and thrust) will be different than one another. Therefore, if for example the changes in Coulomb stress are calculated for optimally oriented strike-slip faults, there may not be any relation between the changes of stress and the location of the normal faulting aftershocks. For example, focal mechanism solutions of the aftershocks in the Yalova region show that most of them are purely normal faulting events (Özalaybey et al., 2000; Örgülü and Aktar 2001; Pinar et al., 2001). Thus, in order to suggest that these earthquakes were triggered by Coulomb stress transfer or not, one has to calculate the Coulomb stress changes on optimally oriented normal faults, not optimally oriented strike-slip faults as did by Pinar et al. (2001)

Thus, care must be taken when evaluating the fault parameters of an earthquake on the basis of the correlation between the aftershock distribution and the Coulomb stress changes. After all, not all the aftershocks are directly related to the Coulomb stress increase as some other factors such as dynamic stresses, fluid movements, viscoelastic relaxation of the lower-crust or upper-mantle, after-slip and complex fault systems probably play a role in the locations of after shocks. The Yalova cluster appears to be a good example of induced seismic activity unrelated to static stress increase. The cluster falls mostly in the stress shadow when the Coulomb stress changes is calculated on optimally oriented normal faults (Fig. 5.5b). Therefore, seismicity in Yalova could not be related to Coulomb stress increase. This, of course, assumes that the source parameters obtained through InSAR and GPS inversions are correct. Dynamic stress triggering is thought to be a possible explanation to this seismic activity there (Özalaybey per. com., 2002).

Both in the Izmit and the Düzce cases, there is no correlation between the distribution of the Coulomb stress changes and the coseismic slip or, between the maximum Coulomb stress increase and the location of the hypocenter. It is clear that the adjoining Izmit earthquake definitely promoted the 12 November Düzce earthquake by raising the static stress on the Düzce rupture over 5 bars but,

coseismic stress changes alone cannot satisfactorily explain the 3-month delay between the Düzce and Izmit earthquakes. In addition to coseismic stress change, Hearn et al. (2002) found that postseismic deformation following the Izmit earthquake contributed substantially to the Coulomb stress change on the Düzce rupture.



6. CONCLUSIONS

Because each Chapter includes its own summary of conclusions, here an overall summary of the findings and the general conclusions is going to be given.

Although the possible occurrence of atmospheric artefacts may obstruct a good solution, a critical analysis of the SAR data using a pair-wise logic approach and independent meteorological data can allow one to identify atmospheric artefacts and to remove them from interferograms. InSAR appears to be the most appropriate to deduce an overall image of the static rupture at seismogenic depth because of its unique capability of high spatial coverage with high resolution on the contrary to other measurements such as the GPS measurements that sample discrete observation points with generally no comparable spatial coverage.

Combining SAR interferometry with the well-constrained surface observations, such as the fault geometry, the fault kinematics and the near-field deformation appears a powerful approach to resolve the first-order features of the slip distribution associated with earthquakes. As solution to a data set of geodetic observations is nonunique, this approach reduces the range of possible solutions that can adequately explain the geodetic observations. An overall forward modelling strategy that combines a trial-and-error approach with a conventional inversion technique appears to be more appropriate than uncontrolled blind inversions.

Fault geometry plays a key role in determining the subsurface slip when modelling geodetic data with elastic dislocations. The better the fault geometry is constrained the more accurate the solution that is found through inversions and forward modelling. The Inversions can allow one to estimate a simple fault geometry only. Complicated fault ruptures such as those involving multiple faults with varying dip and strike are difficult to estimate without additional information such as tectonics field observations and seismicity. As shown in Chapter 4, the geodetic data set can be explained reasonably well with a rupture of complex geometry. Such rupture geometry is more plausible than the one with north-dipping single fault that is difficult to explain when the regional tectonics is considered.

The Izmit earthquake rupture appears to have extended well into the eastern Sea of Marmara to 40 km SSE from downtown Istanbul. The SAR data indicates that the

Yalova-Hersek segment ruptured over 30-km length west of the Hersek peninsula, with slip tapering westwards from 2 m to zero. The best fits to the SAR data define an inhomogeneous slip distribution with three main zones of higher slip along the fault. The inhomogeneous slip distribution is robust and correlates well with the geometry of the fault segmentation, which is well defined from the morphology. It appears that a dynamic aseismic after-slip reaching 2 m during the month following the main shock occurred within a zone of the fault located at 12-24 km depth, directly under the zone of highest moment release in Gölcük. Comparison of the after-slip retrieved from the SAR with the available GPS records of postseismic deformation suggests that about half of the after-slip captured by the SAR data below Izmit-Gölcük may have occurred during the first two days following the main shock. The correlation between the coseismic slip on the fault at depth and slip measured along the surface break is good. However, slip is unevenly correlated with the long-term fault segmentation: The Izmit earthquake slip distribution reproduces well the shape of some segments with a "characteristic rupture" (Sapanca-Akyazı and Karadere segments) but it has smoothed out the fault complexities at the boundaries between individual segments around the high-slip region of Izmit-Gölcük. This suggests that under the Gölcük region a large slip deficit and possibly a large elastic loading existed prior to the earthquake. This feature is consistent with the occurrence of both the rapid early after-slip under the overloaded region and the small events in this region in the years preceding the main shock. The different correlation of the coseismic slip with the segment morphology also suggests a possible distinction between two different modes of rupture: A critically-loaded segment mode (in the Sapanca-Akyazı, the Karadere and the Düzce segments) and an overloaded segment mode (in Gölcük and surrounding segments). The heterogeneous coseismic slip and the state of loading on segments may result from a heterogeneous distribution of slip deficit accumulated during previous large earthquakes along the North Anatolian Fault.

Analysis of the coseismic interferograms of the 17 August 1999 Izmit earthquake, reveals that the interferograms include atmospheric artefacts correlated with topography. The phase-elevation ratio decreases with increasing elevation, reaching up to 6 cm of relative phase delay. Although the phase-elevation ratio also varies laterally, a simple, horizontally uniform model of atmospheric artefacts is calculated using a digital elevation model. Correction of the observed interferograms using this model reveals that some of the anomalies in the fringe pattern, which were previously interpreted as triggered slip, are also associated with atmospheric

artefacts. The model not only explains the wide spread noise, deflection and bending in fringes but also reveals some large scale artefacts previously undetected. This study shows that a simple model may account for most of the artefacts in an interferogram and answer whether or not a signal or an anomaly in the fringe pattern could be possibly related to atmospheric artefacts even though it exists in other interferograms. It is, therefore, worthwhile to make a simple model of atmospheric artefacts before analyses of surface deformation and remove them from interferograms when possible. This is especially crucial when studying subtle deformation such as interseismic loading, post-seismic deformation and subsidence where strong topography is present.

Geodetic measurements of the coseismic displacements due to the Düzce earthquake suggest a longer (~15 km) fault rupture to the east than the one mapped in the field. This explains why the magnitude of the Düzce event is surprisingly high considering its short rupture length observed in the field. The available geodetic data set can be explained by the models with either single- or multiple-fault-rupture geometry within the uncertainty level of the geodetic data set. Inadequate coherence of the InSAR data and lack of GPS points particularly in the near field allow models with both types of rupture geometry to fit the geodetic data within the acceptable limit of error. However, the models with multiple-fault-rupture geometry are favored taking into account that the Düzce rupture with a predominant strike-slip trends parallel to the strike of the almost pure strike-slip, vertical fault rupture of the Izmit earthquake to the west, and that the Düzce fault forms part of a major strike-slip plate boundary between the Anatolian block and the Eurasian plate. The composite nature of the rupture is thought to be more complicated than it is assumed in the models. The overall pattern of slip distribution in all the models is very similar. Both the GPS and the InSAR data suggest a similar shape and a similar amount of slip, and hence the same geodetic moments. Unlike the one found for the Izmit rupture, the slip distribution predicted by the two data sets for the Düzce fault is fairly simple, coseismic slip reaching to a maximum around the fault center and gradually diminishing towards the edges and towards the deeper parts of the fault rupture. The highest coseismic slip predicted by the both data sets occurs within the upper 15 km of the seismogenic crust, and it reaches up to 6.5 m. The geodetic moment derived from the models varies between 6.1×10^{19} and 6.5×10^{19} Nm ($M_w=7.2$) and is comparable with seismological estimations that range between 4.5×10^{19} and 6.6×10^{19} Nm. Although previous studies suggest that the Düzce rupture must dip

strongly to the north, this study illustrates that composite fault rupture that comprises a vertical Düzce fault can too explain the geodetic observations reasonably well.

An accurate set of source parameters of an earthquake is the key to obtain accurate Coulomb stress changes caused by that earthquake. The more accurate the source parameters the more reliable the results are. Small differences in slip distribution and fault geometry can lead to significant perturbations in the Coulomb failure stress. Seismic hazard analyses based on Coulomb models of the same earthquake with different source parameters will be different. A reliable estimate of the slip distribution and fault geometry is therefore very important for stress transfer calculations. Accordingly, Coulomb models based on the fault parameters and the slip distributions that are deduced from geodetic and seismic observations are more reliable. Coulomb models that predict the location of aftershocks fairly well do not necessarily suggest that they represent the actual distribution of stress change. After all, not all the aftershocks are directly related to the Coulomb stress increase as some other factors such as fluid movements, viscoelastic relaxation, after-slip and complex fault systems probably play role in the locations of after shocks. Coulomb modelling of the large earthquakes and the 1999 seismic sequence indicates that the Sea of Marmara region is located in an area of enhanced stress increase. The 1912 and 1999 events, in particular, increased the static stress over 5 bars on the submarine fault system in the east and west, respectively. A stress increase of 5 bars corresponds to an increase normally accumulated in about 12 years by secular loading due to the continuous plate motion. In other words, the previous earthquakes brought forward the next earthquake in the Sea of Marmara by 12 years. The faults in this region therefore pose a serious seismic hazard particularly for Istanbul. If one or two segments simultaneously break, then the question is whether the future earthquake will occur in the eastern or western Marmara. History of the large earthquake that occurred in the Sea of Marmara before 1912 must be better known in order to answer this question with confidence. Historical records about the past earthquakes give clues about which earthquake broke which fault, but are inadequate. This study indicates that 12 October Düzce earthquake is most likely to have been triggered by the adjoining Izmit earthquake that raised the static stress resolved on the Düzce fault over 3 bars.

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14. YÜKSEKÖĞRETİM KURULU
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Tez Sahibinin Numarası: 205972007

Adı, Soyadı: Ziyaeddin ÇAKIR

Tezin Kendisi ☒

Tezin Türkçe Özeti ☒

Tezin İngilizce Özeti ☒

Tez Veri Giriş Formu ☒

Tez Öneri Formu ☒

Tez Çalışma Raporu/Raporları ☒

Kontrol Edilme Tarihi: 28.03.2023

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