1 Methods and background

1.1 Introduction

The geological study of active faults requires a multidisciplinary understanding of several fields of earth science, including plate tectonics, structural geology, tectonic geomorphology, Quaternary stratigraphy, seismology, potential-field geophysics, geodesy, and Quaternary dating techniques. These have been described in a previous publication (Yeats *et al.*, 1997) but are summarized here, with more recent and more detailed references.

1.2 Tectonics

1.2.1 Introduction

The surface of the Earth is at two predominant levels, high-standing *continents*, with their mean land surface 840 m above sea level, and *ocean basins*, at a mean depth of 3700 m below sea level. The different levels are caused by different internal compositions of the Earth's crust. Continents are composed of silicic granitic rocks, composed of lighter minerals including orthoclase and plagioclase feldspar and quartz, while ocean basins are composed of basalt, made of denser minerals including pyroxene and olivine, in addition to plagioclase feldspar. The boundary between these two types of crust is relatively abrupt, forming continental slopes. Granitic crust stands high with respect to ocean crust because its lower density makes it buoyant, thereby supporting high topography. Continental crust is thicker than ocean crust such that the base of continental crust projects downward, and the thicker and deeper continental crust supports its elevation above sea level, just as an iceberg rises above the sea surface because of support from the part that is submerged.

Both continental and oceanic crust are separated from underlying *mantle*, composed of still-denser ultramafic rock composed predominantly of olivine and pyroxene, by the *Mohorovicic discontinuity* or *Moho*, first detected by seismic waves. At the surface and at shallow depths, rock stores elastic strain energy and deforms by brittle fracture, but under high temperature and pressure, rock may deform very slowly by hot creep. In the shallow crust, the increase in confining pressure with depth increases the strength of rock, but increasing temperature has the opposite effect and weakens rock. At depth, the decrease in strength due to higher temperature takes over from the increase in strength with increasing pressure. The depth and temperature where this takes place is the *brittle–ductile transition*. All crustal earthquakes are shallower than this transition. The strong crust and upper mantle comprise the *lithosphere*, which grades downward into the *asthenosphere*, weaker because of its higher temperature.

1.2.2 Plate tectonics

A map showing the worldwide distribution of earthquakes shows that they are not uniformly distributed but are concentrated in relatively narrow bands, an indication that deformation is concentrated in these bands. This observation, together with the observation that only the shallow shell of the crust has enough strength to generate earthquakes, leads to the conclusion that the Earth's surface consists of a series of *plates* that are moving with respect to one another. This conclusion has been confirmed geodetically, first based on radio telescopes (Very Long Baseline Interferometry) and more recently by the Global Positioning System (GPS). The most famous plate boundary is in the Atlantic Ocean, where Alfred Wegener and others proposed that the similarity of coastlines on opposite sides of the Atlantic were due to continental drift. Africa and South America were, indeed, adjacent to each other, but their separation was the result of formation of new ocean crust at the center of the Atlantic Ocean, along the Mid-Atlantic Ridge, so that the continents drifted apart as passive passengers on top of new ocean crust. Earthquakes are found at the ridge, not at the boundary between continent and ocean basin, which is called a *passive margin*.

As new ocean crust formed, the Earth's magnetic field underwent repeated reversals (magnetic north became south, and vice versa) that had been dated based on radiometrically dated volcanic rocks on the continents. These reversals were then mapped on the ocean floor using magnetometers, and the rate of formation of new oceanic crust leading to the separation of continents was then worked out. Newly formed oceanic crust is hotter and thus shallower due to its buoyancy, forming a ridge; as it cools, it subsides at a known rate. Three types of plate boundaries were identified, all marked by bands of earthquakes: (1) sea-floor spreading centers, where the rate of opening was determined based on symmetrical magnetic anomalies, (2) zones where the plates moved past each other, called *transform faults*, and (3) zones where the plates moved toward each other and one dove beneath the other, called *subduction zones*. Subduction zones contain earthquakes within a downgoing slab called *Wadati–Benioff (W–B) earthquakes* to depths as great as 700 km, with evidence for the slab based on higher transmission speeds of seismic waves (*seismic tomography*) at depth. Where the plate boundary is offshore, it may be marked by a narrow trench, with the greatest water depths in the oceans. Active volcanoes commonly are found in the upper plate, above the subduction zone. Subduction zones release about 90% of Earth's seismic strain energy, including most of the great earthquakes with $M > 8$ and all of them with $M \geq 9$.

Plate boundaries may be relatively sharp or they may be diffuse. For example, the boundary between the Pacific and North America plate is thousands of kilometers wide, extending from off the west coast of the United States to the edge of the Colorado Plateau. Long-term rates (10^5 years) of relative motion between adjacent plates have been worked out for all plate boundaries, and these rates may be compared with long-term rates on individual structures with the idea that rates on all the individual structures should equal the plate rate (*slip rate budget*). Shorter-term rates based on paleoseismology (10^4–10^3 years) or geodesy (10^2 years or shorter) may differ from the long-term rates.

Most subduction zones are flanked by mountain ranges that are higher than the continental averages, including the Andes and Himalaya. These ranges are high because they are supported by more-buoyant mountain roots that descend to greater depths than average continental crust; they are *isostatically compensated*. Some parts of ocean crust are shallower than crust formed simply by sea-floor spreading because they are covered by vast fields of basalt, called *large igneous provinces* (LIP). One of these, the Ontong Java Plateau, is colliding with the Solomon Islands, and another, comprising much of the Yakutat Microplate, is colliding with southern Alaska. Others form plateau basalts on the continents, including the Deccan Basalts of peninsular India and the Columbia River basalts of the Pacific Northwest. Because the offshore LIPs are thicker and shallower, they are more likely not to subduct beneath a continent but to collide with it, forming an *accreted terrane*.

The asthenosphere is roiled by currents, and some, called *hotspots*, rise toward the surface and form volcanoes on the sea floor. Because the plates above them are moving, the volcanoes form a linear chain, ending above the present position of the hotspot. Among the best known is the Yellowstone hotspot in the northwestern United States, which can be backtracked offshore, and the Hawaii–Emperor seamount chain, which includes active faults and earthquakes on the island of Hawaii, above the present position of the hotspot. Other anomalous features are topographic swells, particularly well displayed in Africa because of warping of a well-preserved Tertiary erosion surface. The centers of some of these swells have collapsed to form *rift valleys*, which may be incipient sea-floor spreading centers like the Red Sea.

A more detailed discussion of plate tectonics is provided by Condie (1997).

1.3 Structural geology

Structural geology is the study of deformation, including rigid-body translation, of the Earth's crust at all scales from mountain ranges to individual crystals in rock. Earthquakes are themselves a manifestation of rock deformation, most commonly an expression of displacement along a fault. Deformation results from *stress*, or force per unit area, and produces *strain*, or change of length, volume, or shape of a body with respect to its original length, volume, or shape. If a body is *elastic*, that is, strain is recoverable if the stress is removed, stress is proportional to strain with the proportionality constant the *modulus of elasticity*. One type of strain is a change in volume, where a body is subjected to stresses that are uniform in all directions, like the strain on a submarine beneath the surface. Liquid acting on the submarine changes its volume slightly, but this volume change is recovered when the submarine returns to the surface. Liquids have volume elasticity.

Rocks beneath the surface are subject to volume strain, but because the enclosing rock has strength, the stress is not equal in all directions, and the rock may deform, either plastically or by sudden release of strain. A stress acting on a surface can be resolved into a normal component σ acting perpendicular to the surface and a shear component τ acting parallel to the surface. The normal stresses acting on a body are maximum compressive stress, σ_1, minimum compressive stress, σ_3, and an intermediate stress, σ_2, at right angles to the other two. Shear stress, τ, on surfaces perpendicular to these principal stresses is zero. Shear stress acts on all other surfaces and is higher (τ_{max}) as differential stress ($\sigma_1 - \sigma_3$) increases.

For previously unfaulted rock without planes of weakness, each orientation of principal stresses results in two conjugate planes of high shear stress. However, these planes do not intersect at 45° to the principal stress σ_1 because the fault angle relative to principal stresses is also related to the coefficient of static friction, μ_s, such that the angle between one of the fault planes and σ_1 is closer to 30°. This relationship permits a classification of faults based on the orientation of the principal stresses: *normal faults* where σ_1 is vertical, *reverse faults* where σ_3 is vertical, and *strike-slip faults* where σ_2 is vertical. For strike-slip faults, the conjugate faults are vertical, one being left-lateral and one being right-lateral.

These stresses are compressional and act inward on a body. The body may also be porous and contain fluids that act outward, producing pore pressure. These pressures (P) may be hydrostatic, equal to the pressure of a column of fluid, reducing the compressive stress:

$$\mathbf{P}_{\text{compressive}} - \mathbf{P}_{\text{fluid}} = \mathbf{P}_{\text{effective}}$$

This principle is used to create fractures in rock to increase oil production: "hydrofracking." It also affects rock beneath a reservoir behind a recently constructed high dam, resulting in earthquakes, as described in this text at Oroville, California. The movement of low-angle thrust faults may be facilitated by fluid pressures greater than hydrostatic within fault zones.

A fault with a dip < 90° has a *hanging wall* that is directly above the fault and a *footwall* that is directly beneath it. For normal faults, the hanging wall moves down with respect to the footwall, resulting in horizontal extension. For reverse faults, the hanging wall moves upward relative to the footwall, resulting in horizontal shortening. In some cases, motion along the fault plane produces grooves called *slickenlines*. In well-exposed fault systems, slickenlines and fault dips, including those on minor faults, may be used to reconstruct the differential stress field existing during faulting. Some faults have moved by oblique-slip so that the slip consists of a strike-slip and a dip-slip component. If the slip vector is preserved along the fault plane, the angle between the slip vector (expressed as slickenlines) and the strike of the fault is the *rake*. For strike-slip faults, the slickenlines tend to be horizontal, and the rake is zero. In some cases where the coseismic slip is very large, as along the Himalayan front, the slip must be calculated from the vertical component of displacement and the dip of the fault.

Rocks may be folded rather than faulted, forming an *anticline* where the fold is upward or a *syncline* where the fold is downward. The *hinge* is that part of a folded surface where the *radius of curvature* (the radius of a circle whose arc most nearly matches fold curvature) is a minimum. The *axis* is a line within bedding that is parallel to the hinge. The surface connecting all the hinges is the *axial surface*, dividing the fold into *limbs*. *Symmetrical folds* are upright, with a vertical, generally planar axial surface. *Asymmetrical folds* have one limb steeper than the other, and the axial surface dips < 90°. *Overturned folds* have one limb overturned so that its dip is < 90°. The fold has *vergence* toward the steeper or overturned limb.

Drag folds show asymmetry appropriate to the sense of displacement on the adjacent fault; they respond to frictional drag on strata that are cut by the fault. *Fault-bend folds* are formed because underlying fault surfaces are not perfectly planar in the direction of slip. Movement across a non-planar fault surface should produce deformation of one or both blocks adjacent to the fault or cause the fault to break through one of the blocks to produce a straighter trace in the direction of slip. *Fault-propagation folds* emanate from the tip of a propagating fault such that separation on the fault decreases to zero as the fault tip is approached. Although fault-bend and fault-propagation folds are not earthquake sources, they provide information about the underlying *blind fault* that generates them. In some cases, the fold masks a steeply dipping reverse fault that extends downward into high-strength rocks at depth, whereas in other cases, the fault is low-angle and does not penetrate high-strength rocks. A *décollement fold* is a concentric fold overlying a zone beneath which there is no faulting. The length of a folded bed is compared with the horizontal distance across the folded structure to determine the amount of shortening. In many instances, several low-angle reverse faults (*thrust faults*) are overlain by décollement folds to form a stack of structures called a *fold-thrust belt*. An example of a seismogenic fold-thrust belt makes up the Western Foothills of Taiwan, source of the Chi-Chi earthquake of 1999.

For more detailed background information, two textbooks are suggested. Fossen (2010) is written from a European perspective, and Davis *et al*. (2012) is written from an American perspective. Both books emphasize the importance of field study of structures, even though the source of the mainshock of an earthquake is many kilometers beneath the Earth's surface.

1.4 Seismic waves

1.4.1 Introduction

Most earthquakes are caused by release of elastic strain accompanying sudden displacement on faults (Reid, 1910). The location of the earthquake within the Earth is its *hypocenter* or *focus*, and the point on the Earth's surface directly above the hypocenter is its *epicenter*. The ground shaking that results from this release of strain energy is recorded on *seismograms*, providing information about the earthquake process and the earth materials the seismic waves pass through. *Seismographs* were invented in the late nineteenth century, meaning that seismic waves can be studied only for earthquakes in the last century or so. In addition, improvement in instrumentation has led to more recent earthquakes yielding much more information than, say, information on the 1906 San Francisco earthquake. For example, the seismic information from the 11 March 2011 Tohoku-oki earthquake of M 9 is far superior to information on the 1964 Gulf of Alaska earthquake, for two reasons: (1) more accurate and more sophisticated recorders and (2) a much larger array of stations. Modern seismographs have three components, measuring motion up and down, east and west, and north and south, and also are able to measure earthquakes of vastly different frequencies. I have been working with Egill Hauksson of the California Institute of Technology and the seismic network for Los Angeles, which now permits very accurate location and depth determination and source parameters of even small earthquakes that was not possible even a few decades ago. In my view, modern seismograph networks are coming closer to imaging the geology directly!

Earthquake waves are of two general types: *longitudinal* or *compressional (P) waves*, in which particle motion is in the direction of propagation, and *transverse* or *shear (S) waves* in which particle motion is at right angles to the direction of propagation. Because liquids possess volume elasticity, P-waves can pass through liquids, but S-waves cannot. If transverse motion of shear waves takes place in a single plane, the wave is *polarized*. The distance between two successive crests of a wave is the *wavelength*. The maximum displacement of a particle from an equilibrium position as the wave passes through it is the *amplitude* of the wave. The length of time it takes one wavelength to pass a point is its *period*. The reciprocal of the period is the *frequency*. A wave passes out from a point source in the Earth as an expanding sphere, the surface of which contains the energy of the wave at any given time, with the energy per cm^2 of the spherical surface inversely proportional to the area of the surface, $4\pi r^2$. As a result, the strength of an earthquake wave decreases or *attenuates* away from the earthquake source. The amount of attenuation is related to the rigidity of the Earth materials the wave passes through. If the materials are not entirely elastic, seismic energy dissipates in the Earth, and the wave attenuates more rapidly

After an earthquake, the P-wave arrives at a seismograph first, since it is faster, and, because it is compressional, it can be transmitted into the air, making a sound. The S-wave is slower and arrives later. Both are called *body waves* because they travel through the Earth. A third type is called a *surface wave*, which is only observed close to the surface of the Earth, arriving generally after the S-wave. Complexity in the strain-release pattern at the source and in the Earth materials the waves pass through causes P-waves to be converted to S-waves, or S-waves to P-waves, adding further information about the earthquake and the materials earthquake waves pass through, as well as the internal structure of the Earth itself. In addition to attenuation, the speed of earthquake waves is affected by the strength of Earth materials the waves pass through, faster in cold Precambrian crust like that beneath continental shields and in subducting slabs. Slabs of subducting lithosphere are colder and more rigid than the enclosing asthenosphere, allowing the mapping of the slabs based on their higher speeds, a branch of seismology called *seismic tomography*, analogous to CT scan tomography in medicine. In addition, earthquake source parameters cause some earthquake waves to travel at slower speeds than others, some so slow that they are recorded geodetically rather than at a seismograph. In the Nankai, Cascadia, and Middle America subduction zones, slow earthquakes tend to predominate at greater depths along the subduction boundary than ordinary, highly damaging earthquakes, and are transitional between stick slip at the locked zone and stable sliding at still greater depths.

1.4.2 Orientation of fault plane based on earthquakes

Earthquakes accompany displacement on faults that are assumed to be oriented in a plane of high shear stress with respect to the maximum and minimum principal compressive stresses (σ_1 and σ_3, respectively, labeled P and T in the earthquake literature). Motion in the quadrants containing the P-axis is toward the earthquake source, which means that the earthquake wave moves away from the seismograph and reaches it as a dilatation. Earthquakes in the quadrants with the T-axis move away from the earthquake source, arriving at the seismograph as a compression, with a different first motion of the seismic wave. Because it is this motion toward or away from the seismograph that is recorded, with the *double-couple assumption*, a set of seismograms cannot tell which of the conjugate fault planes produced the earthquake, although the first motion can distinguish between other possible orientations or other possible senses of slip. Additional information is needed, such as a linear pattern of aftershocks or a mapped active fault. Non-double-couple patterns are possible, such as the pattern from an underground nuclear explosion, a gigantic landslide, or a curved fault plane.

Different types of sources take advantage of the relative amplitudes of seismic waves leaving the earthquake source in different directions, leading to *centroid moment tensor solutions* (see following section about seismic moment). Three-component seismograph stations measure the polarization of shear waves, with some orientations arriving before others, a phenomenon called *shear-wave splitting*. The first waves to arrive are in a plane at right angles to the minimum compressive stress and parallel to the maximum compressive stress. A dense array of seismographs can show the direction of propagation of a large earthquake along a fault surface (toward or away from the seismograph station), analogous to the Doppler effect of a train whistle moving toward or away from the listener.

1.4.3 Magnitude scales

The need for a measurement of earthquake size led Charles Richter in 1935 to design the first earthquake magnitude scale. His magnitude is the logarithm to the base 10 of the maximum seismic wave amplitude recorded 100 km from the epicenter on the Wood–Anderson seismograph then in common use. He recognized that his scale did not measure any fundamental physical parameters, and so he said it is simply a measure of earthquake size at its source, and is applicable to southern California, where he worked. Subsequent work showed that the increase in energy released is about 30-fold per unit of magnitude, and the increase differs for different magnitudes. The Wood–Anderson seismograph has a natural oscillation period of 0.8 s, and longer-period waves such as those arriving from great distances (*teleseisms*) are increasingly diminished on the record, so that the seismograph is better at characterizing local earthquakes. Accordingly, the Richter-designed scale is called *local magnitude*, or M_l.

An attempt to rectify the problem of using longer-period waves or measuring teleseisms was the *surface-wave magnitude scale*, M_s, measuring the largest wave amplitude in a surface wave train with a period of 20 s. The *body wave magnitude scale*, m_b, measures the maximum amplitude of teleseismic P-waves with a period of about 1 s. These scales were designed so that magnitudes of earthquakes of intermediate size would correspond roughly to $\mathrm{M_l}$. In this book, if the type of magnitude scale is not specified, the size is given as M.

For very large earthquakes, such as those in subduction zones, none of these scales is an accurate representation of the size of the earthquake. The Richter scale works well for earthquakes that are small enough that they can be considered as a point source, that is, the area of strain release is smaller than the wavelength of the seismic wave being used. But what about giant earthquakes where the rupture area of the fault is hundreds of kilometers long, and rupture must propagate along the fault surface? This led to the concept of *seismic moment* in which rupture along a fault involves equal and opposite forces that produce a couple. The *seismic moment* $M_o = \mu$ u A, where μ is the shear modulus of elasticity and u is the average slip over the rupturing segment of a fault with area A. M_o measures the energy radiated from the entire fault, not just an assumed point source, and it is independent of the frictional energy dissipated during faulting. The seismic moment may be measured from very long-period waves for which even a fault with a very large rupture area appears as a point source. It also may be measured without a seismograph, based on the geological evidence of surface rupture and average slip accompanying large earthquakes or the geodetic evidence of displacement using GPS. Hiroo Kanamori designed a *moment magnitude scale*, M_w, based on seismic moment that is capable of measuring superquakes with $M \geq 9$.

Intensity is a measure of the violence of earthquake shaking at a given site, calculated by the amount of damage done to structures, the degree to which the earthquake was felt by individuals, and the presence of secondary effects such as landslides, liquefaction, and ground cracking. To a first approximation, the zone of highest intensity is close to the epicenter of the earthquake, but unstable ground conditions can produce high intensities far from the epicenter. A map showing intensities resulting from an earthquake relies on reports on the intensity of shaking from a large number of locations. The USGS has a website called “Did You Feel It?” for people to call in their locations and their experiences during an earthquake. The scale most commonly used in Western countries is the *Modified Mercalli Intensity (MMI) scale*, from I to XII. Other scales have been designed in Japan and the former Soviet Union. The latter, the MSK scale, is widely used in central Asia. Different scales reflect differences in construction in central Asia, Japan, China, and Western countries.

In a project my company did in Afghanistan, we found that MSK and MMI are roughly equivalent at the isoseismal boundary between intensities VI and VII. The area of maximum intensity is called the *meizoseismal region*. The seismic intensity may be quantified by using a *strong-motion seismograph* to measure the *acceleration*, *velocity*, and *displacement* at the measuring site. The strong-motion seismograph does not operate continuously but is triggered by strong ground motion; its data are of greatest use to structural engineers in designing buildings against earthquake damage. In forecasting the probability of future earthquakes at a site, it is common practice to express this as *probability of exceedance* of an estimated acceleration or velocity over a selected future time period. Accelerations are commonly presented as percentages of *g*, which is the acceleration due to gravity. An acceleration exceeding 1 *g* was first observed after the 1897 Shillong earthquake in India, which caused rocks to be dislodged and even thrown into the air. Maximum velocity and displacement measured by strong-motion instruments are also important to engineers.

The intensity scale has the advantage that it can be used to estimate the magnitude of pre-instrumental earthquakes. The maximum intensity, I_o, is determined from examination of historical and archaeological records. If the source fault is known, then the damage at a site may be compared with the distance from the source fault by examining attenuation relations based on intensities along the same earthquake path from earthquakes that have occurred during the seismograph era.

For more details, refer to Bolt (2004).

1.5 Tectonic geodesy

1.5.1 Terrestrial geodesy

The precise measurement of distances, angles, and elevations on land has long been used in surveying property lines, but it was first used to measure deformation of the crust after earthquakes in Sumatra in 1892, India in 1897 and 1905, and California in 1906. Conventional techniques include *triangulation* (measurement of angles between survey markers), *trilateration* (measurement of line lengths), and *leveling* (measurement of vertical deformation and tilting). The earthquake in Sumatra struck while a triangulation survey was underway, and the surveyor found that the angles measured after the earthquake did not correspond with those measured before. The two Indian earthquakes were studied with the background of the high-resolution Great Trigonometric Survey of India of the nineteenth century, which meant that measurements taken more than a century ago can still be used today to measure deformation. The San Francisco earthquake of 1906 was studied by Harry Reid, along with the 1892 Sumatra earthquake, providing evidence for his *elastic rebound theory of earthquakes*.

The San Andreas fault has been surveyed repeatedly, starting in the 1970s, focusing on the San Francisco Bay region, with measurements related to the center of mass of the network. Displacements are generally parallel to strike, but there are extensional and rotational components. One subfield of geodesy is *near-field geodesy*, with closely spaced survey markers that are surveyed repeatedly. This was done for many years in New Zealand (fault-monitoring patterns) and is commonly done on active volcanic carapaces to monitor against future volcanic activity. Releveling of highways has been combined with repeated observations of tide gauges to monitor strain buildup toward a future Cascadia subduction zone earthquake.

1.5.2 Space geodesy

Very long baseline interferometry (VLBI) uses the radio signals from quasars as recorded at radio telescopes separated by a baseline vector AB, with A and B representing the position of the two telescopes, in some cases separated by ~10 000 km. The signals arrive as essentially planar wave fronts; as the Earth rotates, the signal arrives first at A, then at A and B simultaneously, then at B first. The time delay between the two radio telescopes is proportional to the baseline distance between them, the chord AB. Repeated measurements of AB resolved a slip-rate discrepancy between the Pacific–North America plate motion and slip rate on the San Andreas fault, leading to the discovery of a separate Sierran microplate.

Global positioning system (GPS) is based on a constellation of NAVSTAR satellites orbiting the Earth at an altitude of about 20 000 km. A position on the Earth can be located in three dimensions to an accuracy of generally several meters based on the distance of the position from three satellites with known orbits. *Relative* position can be determined with an uncertainty three orders of magnitude smaller than position location. Location is three dimensional, the space geodetic counterpart of ground-based trilateration and leveling, although measurements of relative vertical position are several times less accurate than those of horizontal position. Change of distance between source and receiver is measured rather than the distance. Measurements with GPS agree within 0.2 ppm of terrestrial measurements over baseline lengths of 10 to 40 km. The main sources of error are the precise orbital parameters of the satellite and uncertainties in path delays in the troposphere. Accurate data can be obtained with only a few hours of observations, which should be less in the future with improved techniques.

Early measurements were by *campaigns*, in which observation stations were reoccupied periodically. The success of GPS tectonic geodesy has led to the establishment of stations and arrays allowing *continuous observations*, permitting the detection of shorter-term transients. For example, local reversals in the direction of GPS-based vectors marking elastic strain buildup on the Cascadia subduction zone were determined to be slow earthquakes in the transition between the locked zone and the zone of stable sliding. The distribution of these transients within the array allowed the determination of equivalent moment magnitudes, even though none of the earthquakes was felt at the surface.

GPS measurements have been limited to land, where distances between a position and NAVSTAR satellites may be measured. However, a team from the Japan Coast Guard was able to install five sea-floor reference points off the coast of northeast Japan that could measure GPS in campaign mode and to take measurements prior to the 2011 Tohoku-oki earthquake. These reference stations were able to determine vertical and horizontal displacements accompanying that earthquake. Fortuitously, one reference point was at the epicenter.

A recent reference for greater detail is Thatcher (2009).

1.5.3 InSAR

Synthetic aperture radar (SAR) acquired by the ERS-1 satellite was used to map the displacement field of the 1992 Landers, California, earthquake. SAR measures the ground reflectivity and the distance (range) between the radar antenna and the ground. Images from an altitude of 785 km, pointed west at an angle 23° from the vertical, were acquired before and after the earthquake from similar orbits and under similar ground reflectivity. The two images were superimposed, canceling out the topographic differences except for the component of coseismic displacement that affected the range. The resulting image contains interferometric fringes that are a contour map of the changes of range relative to points far enough away that they are assumed to be unaffected by the earthquake. Each fringe is equivalent to 28 mm relative change in range, half the wavelength of the ERS SAR. The fringes are incoherent in areas of complex deformation and in areas of mountainous topography. In addition to measuring coseismic deformation, the method has been used to map buildup of strain along the southern San Andreas fault.

1.5.4 *In situ* stress

An important objective in tectonic geodesy is measurement of tectonic strain as a proxy for determining the orientation of the principal *in situ* compressive stresses. This could be