Chapter 1

INTRODUCTION

Geodetic data collected since the 1990s from subduction zones have been interpreted using simple kinematic elastic dislocation models [Savage, 1983, 1995; Zweck *et al.*, 2002; Wang *et al.*, 2003; Chlieh *et al.*, 2008b]. However, over the last decade, a vast amount of geodetic data has become available from various subduction zones around the world, having not only good spatial coverage (using InSAR, see, e.g., Massonnet and Feigl [1998], Simons and Rosen [2007]), but also high temporal density and resolution (high-rate GPS, see for e.g., Larson et al. [2003]). Such dense datasets of velocity vectors provide an opportunity to explore more complex kinematic or quasi-dynamic mechanical models of the seismic cycle in subduction zones in order to estimate for instance, elastic thickness of the downgoing plate, or frictional properties on realistic 3D megathrust interface between the subducting and overriding plates. Eventually, such information will allow us to refine assessments of potential seismic hazard within different geographic regions of a plate boundary zone, thereby providing guidance on where to focus preventative measures (such as retrofitting buildings), and resources for emergency preparedness (such as evacuation plans and their facilitation).

Early theoretical attempts to model the kinematics of deformation during the entire seismic cycle were made in the late 1970s to mid 1980s, using 2D earth models having a single subduction interface embedded in (a) a fully elastic half-space [Savage, 1983], (b) an elastic layer (lithosphere) overlying a viscoelastic half-space (asthenosphere) [Thatcher and Rundle, 1979; Rundle, 1982; Thatcher and Rundle, 1984; Cohen, 1994], (c) an elastic layer (lithosphere) over a viscoelastic layer (asthenosphere), over an elastic half-space [Sato and Matsu'ura, 1988; Matsu'ura and Sato, 1989], or (d) a viscoelastic lithosphere over viscoelastic asthenosphere, over a viscoelastic half-space [Sato and Matsu'ura, 1992, 1993; Fukahata and Matsu'ura, 2006]. These models considered gravity and realistic subduction interface geometries. Models (a) and (b) assumed that there is no net accumulation of deformation in the overriding plate. Models (b) require two parameters that have high uncertainties in addition to those in elastic dislocation models – the asthenospheric viscosity, and the recurrence time for seismic events. The key result from models (b) was that the surface velocity field was much larger than the corresponding elastic field [model (a)] right after a megathrust event, and much smaller than the elastic field just before the subsequent event. So, the integrated velocity field during the interseismic period exactly cancels the coseismic displacements after each cycle, resulting in zero net deformation of the overriding plate. Including gravity diminishes the magnitude of viscoelastic deformation in the model, which reaches steady state faster than in the zero-gravity case [Rundle, 1982]. Because the more complex physics included in these models introduces additional parameters, they can fit the coseismic, postseismic, and at least in some cases, the interseismic deformation fields well [Thatcher and Rundle, 1984]. Models (c) and (d) predict a net accumulation of deformation in the overriding plate after each seismic cycle, owing to the steady state motion at the rate of plate convergence along the curved fault interface within the upper elastic lithosphere — this conclusion is unaffected by the inclusion of gravity. As we will see in Chapter 1, this so-called permanent deformation is very similar to that required to support elastic stresses resulting from bending of the subducting plate at the trench. Although this runaway surface deformation can be modulated by parameterizing accretion, erosion and sedimentation during the seismic cycle [Sato and Matsu'ura, 1993; Cohen, 1999], such complexity introduces many more free parameters having high uncertainties. Furthermore, Savage [1995] argued that the coseismic, postseismic and interseismic fields can be fit equally well (given the data uncertainties) with a modified elastic dislocation model having a fault patch downdip of the locked zone that slips only post-seismically — and that it is hard to demonstrate that asthenospheric relaxation contributed to interseismic deformation on the surface of the overriding plate.

Going beyond these semi-analytical approaches, finite-element-method (FEM)-based models also do not do better than dislocation models, given the current spatial resolution and uncertainty limits of geodetic data. Quasi-static models that are computationally more challenging, and are driven by dynamically consistent boundary conditions have also been developed. It is illustrative to consider two representative studies that model the two distinct types of plate compression zones — subduction and collision zones — using FEM models.

Williams and McCaffrey [2001] developed a 2D quasi-static, self-gravitating, purely elastic finite element model of the Cascadia subduction zone beneath Oregon and southwest Washington. The quasi-static deformation fields within the overriding plate are entirely determined by uniform and constant shear tractions along the locked subduction interface (a proxy for the effect of locking), as well as along its bottom surface (a proxy for upper mantle flow). Using that model, they attempted to constrain shear stresses acting along the fault interface and the bottom of the overriding plate using regional geodetic data. They compare the surface velocity and tilt-rate fields predicted by their preferred FEM with those of an equivalent elastic dislocation model having the same fault geometry, and find that both models fit the vertical velocities (at a single observation point) as well as the observed surface tilt rates equally well. Their main argument for preferring the FEM was its ability to better fit the location of the change in slope of the horizontal velocity profile, as well as a broad region of elevated horizontal velocities just beyond this slope change. As is well known, and also illustrated in Chapter 2, the location of the change in slope of the horizontal velocity profile predicted by an elastic dislocation model (with a locked zone extending all the way up to the trench) is sensitive to the abruptness of transition between zero to finite aseismic slip at the downdip end of the locked zone.

Vergne et al., [2001] compared the predictions of interseismic surface velocities and crustal stress concentrations from a realistic 2D finite element model of an intra-continental thrust fault — which is kinematically and dynamically similar to a subduction thrust interface — with an elastic dislocation model having the same fault geometry. The 2D finite element model incorporated a layered crust and mantle with temperature dependent rheology, topography, gravity, and surface processes, and fit all available constraints on interseismic and long-term surface displacements. Their main conclusion was that the dislocation model fit the data as well as the finite element model, including predictions of micro-seismicity near the bottom of the locked patch during the interseismic period.

So, unless complexities such as poro-elasticity, material heterogeneity, anisotropy, or inelastic bulk rheology are included in modeling the subduction zone [e.g., Masterlark, 2003], simple elastic dislocation models would do as well as FEM in fitting current geodetic data. It seems reasonable, therefore, that such models — which can be essentially described with only two parameters, the extent of the locked fault interface, and the plate geometry — have been widely used in modeling interseismic period geodetic data in subduction zones, and have been used to successfully fit geodetic observations using realistic plate interface geometries [Savage, 1983, 1995; Zweck *et al.*, 2002; Wang *et al.*, 2003; Chlieh *et al.*, 2008b].

Here, we want to understand late post-seismic and interseismic deformation in subduction zones, and as such, only consider a purely elastic crust (represented by the half-space for the purpose of computing the surface deformation field). We do not seek to model the complex dynamics of rupture nucleation, interaction between asperities, or rupture propagation [see, e.g., Rice, 1993; Lapusta and Rice, 2003; Kato, 2008; Perfettini and Ampuero, 2008]. We also do not model topographic evolution on time-scales longer than the interseismic since we use purely linear elastic bulk rheology that, by definition, cannot accumulate net long-term (geologic) deformation while keeping the stresses bounded. Instead, we pursue kinematic and quasi-dynamic approaches to modeling slip (and its evolution) on the fault over the seismic-cycle. Throughout this work, we assume that crustal deformation is localized along fault zones, and the bulk of the crust is rigid, and perfectly elastic. We therefore ignore any bulk relaxation processes in the crust owing to viscous or poro-elastic effects. While this assumption may not hold true over the geologic time-scale, over the span of several seismic cycles ($<10^4$ yrs) that we model here, it is reasonable to assume that crustal response to constant tectonic loading is elastic. This assumption is borne out by the ability of elastic dislocation models to fit much of the geodetic data collected over the past couple of decades. Further, elastic deformation fields can provide intuition about regions that could potentially experience long-term deformation, where it may be more appropriate to use non-linear rheologies. Therefore, within the context of elastic crustal deformation, we want to ask the following questions:

Why does the backslip model fit geodetic observations so well?

As a first step towards more complex models for interseismic deformation in subduction zones, we want to ask how the thickness of the downgoing plate influences the deformation field at the surface of the overriding plate. We also want to understand why the backslip model [Savage, 1983] works so well for interpreting interseismic geodetic data in subducton zones. In essence, how can one reconcile this half-space model with subduction of a downgoing plate? While standard textbooks discuss the elastic flexure of a subducting plate at the trench, the effect of this bending on overriding plate deformation has not been systematically analyzed so far. In Chapter 2, we introduce an elastic subducting plate model (ESPM), and compare its predictions with that of the backslip model (BSM), in order to address the above questions. The ESPM links elastic plate flexure processes to interseismic deformation, and helps clarify under what conditions the BSM is appropriate for fitting interseismic geodetic data at convergent margins. We show that the ESPM is identical to the BSM in the limiting case of zero plate thickness - thereby providing an alternative motivation for the BSM. The ESPM also provides a consistent convention for applying the BSM to any megathrust interface geometry. Even in the case of non-negligible plate thickness, the deformation field predicted by the ESPM reduces to that of the BSM if stresses related to plate flexure at the trench are released either continuously and completely at shallow depths during the interseismic period, or deep in the subduction zone (below ~100 km). However, if at least a portion of these stresses are not continuously released in the shallow portion of the subduction zone (via seismic or aseismic events), then the predicted surface velocities of these two models can differ significantly at horizontal distances from the trench equivalent to a few times the effective interseismic locking depth. We also suggest potential geographic areas where the subduction zone geometry is favorable for testing the ESPM in the near future

— especially as onshore geodetic coverage improves in these areas, and ocean-bottom geodetic measurements become available.

What are some practical surface observables that have immediate relevance to field-geologic studies or building intuition for numerical modeling?

Surface observables — especially, the location of the zero-crossing (hinge-line) and peak value of uplift rates — can be useful tool in determining the approximate location (to within a horizontal distance of 50 km) of the location of the downdip end of the locked portion of a megathrust interface. These two uplift-rate values are important because gradients in the surface deformation field (strains) are strongest ocean-ward of this region, and highest right above the downdip end of the locked patch. Geodetic or field-geologic observations can be taken more cost-effectively by choosing to sample at a higher spatial resolution in the zone of peak strains (using either land-based or ocean-bottom stations), and more sparsely farther away. If a reasonable estimate for slab dip can be obtained, then the location of this high-strain region can be narrowed down to a zone as narrow as 10–15 km. So, while the relationship between the hinge-line and downdip end of the locked zone is not necessary for geodetic inversions per se, such information can be very helpful in optimally collecting the data for these inversions. The relationships between surface observables and ranges of fault dip, as well as the effect of fault curvature and subducting plate thickness are discussed in Chapter 3. We show that irrespective of the fault geometry, the mean of the location of zero-vertical surface velocities, x_{hinge}, and the peak surface vertical velocities, x_{max}, gives a good approximation for the surface projection of the locked zone, x_{lock}, for both the BSM and the ESPM with shallow dipping plate interfaces (< 30°). However, in the presence of a transition zone, or a large plate thickness, x_{max} gives a more reliable estimate for x_{lock}, and hence, the extent of the locked zone. Therefore, the common notion that the location of the peak in vertical velocities (x_{max},) determines the extent of the locked megathrust (x_{lock}), is valid only if a transition zone is assumed downdip of the locked interface.

For a given subduction zone, what fraction of the current surface deformation field inferred from geodetic data can be explained by the stress-shadow effect of ruptures during the past century on known seismic asperities?

During the past decade, with the availability of high-resolution spatio-temporal geodetic data as well as strong-motion seismic data, the characteristic asperity model for the seismic cycle [e.g., Ruff, 1992] has been shown to apply to the Sumatra [Chlieh *et al.*, 2008b; Sieh *et al.*, 2008; Konca *et al.*, 2009], Kurile [Nanayama et al., 2003; Satake and Atwater, 2007], Chile [Cisternas et al., 2005; Satake and Atwater, 2007], and northeastern Japan [Tanioka et al., 1996; Nakayama and Takeo, 1997; Robinson and Cheung, 2003; Miyazaki et al., 2004; Miura et al., 2006; Umino et al., 2006] subduction zones. For the Japan trench, for instance, it is thought that the ruptures off Miyagi [Miura et al., 2006; Umino et al., 2006], Sanriku [Tanioka et al., 1996; Nakayama and Takeo, 1997], and Tokachi [Robinson and Cheung, 2003; Hamada and Suzuki, 2004; Miyazaki et al., 2004; Satake et al., 2006] occurred repeatedly over roughly the same region of the subduction megathrust. Owing to the fact that geodetic and seismic data resolution was much poorer during the earlier part of the last century — and good spatio-temporal coverage was lacking in most subduction zones excluding Japan even as recently as the 1990s — the exact details of coseismic slip distribution vary between each of these "repeating" sequence of ruptures. However, the picture that seems to be emerging is that, overall, coseismic slip tends to be restricted to only a small fraction of the shallow seismogenic megathrust interface — at "asperities" — while the rest of the interface slips aseismically during the postseismic or interseismic periods of the seismic cycle.

Inversions of geodetic data from interseismic periods, however, produce models that are locked (i.e., are modeled to have backslip) over spatially smooth and extensive region of the seismogenic megathrust [Bürgmann *et al.*, 2005; Suwa *et al.*, 2006; Chlieh *et al.*, 2008b], in contrast to the smaller discrete asperities estimated by the above earthquake source studies. Such smooth, broad regions may be a consequence of a lack of model resolution and the resulting need for regularization inherent to the use of onshore geodetic data. It is also possible that the inferred interseismically coupled regions are larger than the collective asperity sizes for known earthquakes due to an incomplete earthquake catalogue, and may imply the potential for large earthquakes in the future. Hence, the different levels of apparent coupling implied by interseismic and seismic-source inversions (Figure 1-1) have very different implications for regional seismic hazard.

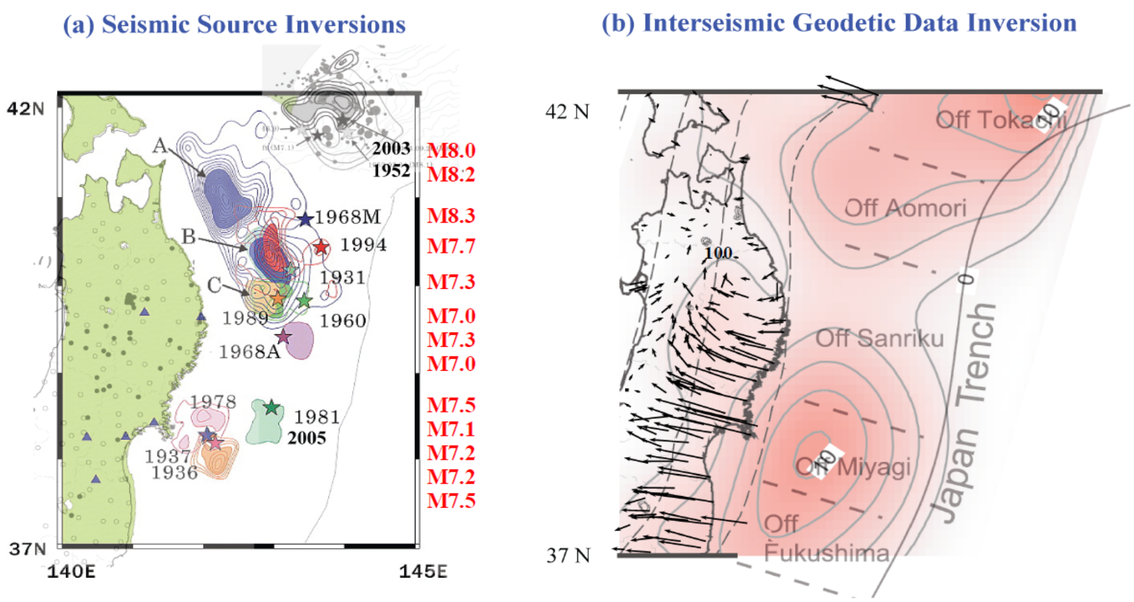

Figure 1-1. (a) Coseismic slip and (b) interseismic slip deficit ("backslip") estimates for the megathrust interface off northeastern Japan. Adapted from Yamanaka and Kikuchi [2003; , 2004] and Suwa et al. [2006].

Bürgmann et al., [2005] tested several asperity models for the Kamchatka subduction zone, but assumed that all areas outside the asperities were freely slipping — so they did not model slip evolution around the asperities. Recently, Hetland et al. [2010] and Hetland and Simons [2010] developed a 3D mechanical model of stress-dependent interseismic creep along the megathrust, considering frictional rheologies. Their mechanical "toy"-models predict that late in the seismic cycle, there are relatively smooth, long wavelength regions of very low slip-rates on the megathrust interface surrounding these asperities, owing to the "stress-shadow" effect of seismic ruptures. The effect of such "physical" smoothing on surface velocity predictions may be indistinguishable from the artificial smoothing produced by model regularization in inversions of interseismic geodetic data.

Here, assuming that (i) known asperities persist across multiple earthquake cycles, and (ii) ruptures are both time- and slip-predictable [see, e.g., Shimazaki and Nakata, 1980], we test the hypothesis that mechanical coupling on asperities inferred from the locations of past earthquakes alone is sufficient to explain currently available geodetic observations for Japan — or alternatively, that these data require additional regions of the Japan Trench megathrust to be coupled. Underlying our approach is the assumption that known asperities persist across multiple earthquake cycles. The modeling approach and setup are discussed in Chapter 4, and results presented in Chapter 5. The preliminary results presented here show that we can explain most of the horizontal interseismic GPS velocities in northern Japan, by assuming mechanical coupling only on the inferred asperities.

As a corollary to the last question, can the late post-seismic and interseismic response in models incorporating these asperities tell us something about the rheology of the megathrust interface over the seismic-cycle timescale?

Recently, several research groups have attempted to infer fault rheologies from inversions of post-seismic geodetic data at plate-boundary zones — for e.g., Sumatra [Hsu et al., 2006], California (Landers [Perfettini and Avouac, 2007], Parkfield [Johnson et al., 2005]), Taiwan [Perfettini and Avouac, 2004], and Japan (Tokachi-oki [Fukuda et al., 2009]). These have used either spring-slider type models [Perfettini and Avouac, 2004; Fukuda et al., 2009] or planar frictional faults made up of rectangular patches embedded in a half-space [quasi-static models, e.g., Johnson et al., 2005; Perfettini and Avouac, 2007]. The model introduced by Hetland et al. [2010] and Hetland and Simons [2010] (summarized in Sections 4.1 and 5.2) belongs to this class of quasi-static models. Spring-slider models have no spatial length-scale (or explicit fault geometry) associated with them, predictions using such models have only local applicability. On the other hand, fully heterogeneous fault frictional properties can be modeled by the latter class of models. There is another class of forward models that consider the dynamic evolution of stresses and slip on a frictional fault surface due to non-uniform rheology [e.g., Hori, 2006; Kato, 2008]. Currently, such quasi-dynamic models focus on simulating seismic ruptures only, and are not constrained by surface geodetic observations. In contrast, the quasi-static models mentioned above are designed to be constrained by geodetic observations and allow us to ask an important question from a forward modeling standpoint: what is the effect of lateral and depth variations in rheological parameters on predictions for afterslip and postseismic/ interseismic deformation? Another important question is the practicality of considering different rheologies (and therefore, different evolution time-scales) over different regions of the megathrust interface. For the preliminary results presented in Chapter 5, we only consider uniform rheological properties over the entire fault surface.

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