Chapter 1

General Introduction

1.1 Motivation

The island of Taiwan, situated at the plate boundary zone between the Eurasian and the Philippine Sea plates, is one of the best examples of young and rapidly-growing orogeny due to arc-continent collision [Chai, 1972; Wu, 1978; Biq, 1981; Suppe, 1981 and 1984; Davis et al., 1983; Tsai, 1986]. About 25~30% of the total plate convergence rate of 82 mm/yr [Yu et al., 1997] have been concentrated on the narrow arc-continent collision boundary, Longitudinal Valley (LV) in eastern Taiwan [Yu et al., 1990; Angelier et al., 1997; Lee et al., 1998; Angelier et al., 2000], which indicates that this boundary may pose a significant seismic threat. Since it has been undergoing active tectonic collision, the Longitudinal Valley fault (LVF) is characterized by active fault segments, high seismic activity, and rapid surface creep.

Creeping crustal faults often generate a number of microearthquakes, and less commonly, they may also produce large earthquakes that rupture the brittle crust. The LVF in eastern Taiwan characterized by such behavior has been known to undergo 1-3 cm/yr surface creep [Yu and Kuo, 2001], probably one of the most active creeping thrust faults known in the world. It gives an excellent opportunity for studying how a creeping fault can generate large earthquakes. However, due to limited geodetic coverage in this area, a well resolved picture of fault slip rates at depth has been lacking. Interferometric Synthetic Aperture Radar (InSAR) data provides better spatial sampling of the deformation field [e.g., Massonnet et al., 1993], but the nature of landcover and atmospheric condition of this area complicate the task of obtaining more detailed deformation information [Hsu and Bürgmann, 2006].

Study on repeating earthquakes (i.e., a group of events with nearly identical waveforms, locations, and magnitudes and thus represents a repeated rupture of the same patch of fault) has been proposed to infer fault slip rate at depth. However, it also has its limitation because of the poor station coverage. A new repeating sequence identification scheme for the region where the station coverage is spare and one-sided, therefore, is needed.

In this study, we first propose a relatively objective method to identify repeating
events in this particular region. Furthermore, using these results we aim to answer the following questions: 1) What are the spatial and temporal distributions of repeating earthquakes and what are their relationships to the distribution of seismicity and large earthquakes on the fault? 2) What is the distribution of deep creep rates that can be inferred from repeating quakes if they exist? 3) How does deep fault creep inferred from repeating earthquakes compare with deep creep determined geodetically? And 4), how can the repeating earthquake distribution, inferred rates, overall seismicity patterns, and large earthquakes distributions be used to improve our understanding of the earthquake potential in this unique tectonic environment? Because the repeating earthquakes observation offers advanced information for earthquake risk assessment and is of fundamental importance in understanding the earthquake cycle, it would be also important to know the nature of repeating earthquakes behavior. In this thesis we examine the diverse recurrence properties at different regions to understand what controls the recurrence interval of repeating events, what controls their regularity, and how the varying occurrence features of repeating earthquakes reflect the variation in fault zone properties.

1.2 Overview of repeating earthquakes studies

The observations of repetitive failures on the same fault patch, namely, repeating earthquake sequence (RES), have been observed in diverse tectonic environments such as Parkfield [e.g., Ellsworth, 1995; Nadeau et al., 1995], Calaveras fault [Templeton et al., 2007], and Hayward fault [Bürgmann et al., 2000] in California, North Anatolian fault in Turkey [Peng and Ben-Zion, 2005], Japan subduction zones [Matsuzawa et al., 2002; Uchida et al., 2003; Igarashi et al., 2003; Matsubara et al., 2005]. The relationship between repeating event’s size and recurrence interval can be used to infer critical parameters associated with a variety of fault zone processes, e.g., fault healing [Vidale et al., 1994; Marone et al., 1995; Peng et al., 2005], rate and state dependent friction [Dieterich, 1994], aftershock decay [Schaff et al., 1998], earthquake source parameters and properties [Marone et al., 1995; Nadeau and McEvilly, 1997; Nadeau and Johnson, 1998; Rubin et al., 1999; Waldhauser and Ellsworth, 2000; Schaff et al., 2002; Dreger et al., 2007; Uchida et al., 2007], aseismic transient event [Niu et al., 2003], subsurface changes of seismic wave propagation properties [Schaff and Beroza, 2004; Peng and Ben-Zion, 2005; Rubinstein et al., 2007], and fault slip rate variations at depth [Bürgmann et al., 2000; Nadeau and McEvilly, 1999 and 2004]. The major implications of repeating event observations are addressed below.
1.2.1 Deep slip rate estimate

Determination of deep fault slip rate using RESs has recently been developed and has been applied in diverse tectonic settings [Bürgmann et al., 2000; Nadeau and McEvilly, 1999; Matsuzawa et al., 2002; Uchida et al., 2003; Matsubara et al., 2005]. These repeating events are thought to fail repeatedly because they are continuously loaded by creep on the surrounding material [Vidale et al., 1994; Nadeau et al., 1995]. Given an assumption that the frequency of seismic slips on a fault patch is proportional to the tectonic loading rate due to the relative plate motion, one can determine the deep fault slip rate at each RES’s site using the recurrence intervals and seismic moments of RES [e.g., Nadeau and Johnson, 1998]. In other words, one can use the average seismic moment of a RES to obtain its slip, then divide the slip by recurrence interval between repeating events to give a slip rate. Unlike interseismic slip rates from inversion of GPS data, the RESs-derived slip rate is a direct measurement using RES’s seismic moment and recurrence interval, which can provide an independent measure for deep fault behavior.

Numbers of small RESs have been found in the creeping sections along the San Andreas fault [e.g., Bürgmann et al., 2000; Nadeau and McEvilly, 2004], where the temporal and spatial distribution of slip rate at depth can be used to compare with surface deformation and earthquake activity for a better understanding of seismogenic potential and earthquake hazard assessment.

1.2.2 Temporal variation in seismic wave propagation

Since source and path effects are common to repeating events in a RES, differences in their seismic wave characters can be attributed to changes in the characteristics of the medium. Using the RESs observations, local change in crustal properties associated with a major earthquake can lead to a remarkable change in recurrence rate [Schaff et al., 1998], seismic wave velocity [Schaff and Beroza, 2004], coda Q [Beroza et al., 1995; Chun et al., 2004], and waveform similarity [Baisch and Bokelmann, 2001]. Differences between the post-mainshock repeating events and pre-mainshock events in a RES can be used to compute such a change.

Baisch and Bokelmann [2001] show that the reduction in waveform similarity between pre- and post- mainshock (1989 Loma Prieta earthquake) repeating events is found at the stations close to the Loma Prieta rupture zone, where the temporal pattern of the reduction exhibits a recovery pattern. The spatio-temporal pattern in
waveform similarity change suggests a model of coseismically opened cracks acting as seismic scatterers with postseismic healing process. Post-mainshock repeating events reveal phase delays in the early S wave coda, which leads to velocity decrease of 1.5% for P waves and 3.5% for S waves. Following the mainshock, the decrease of velocity amplitude decays logarithmically in time with a postseismic recovery. In the Loma Prieta aftershock zone the RESs’ recurrence intervals is also found to follow Omori’s law that a large number of post-seismic event repeats occurred with extremely shortened recurrence intervals gradually increasing with time [Schaff et al., 1998].

1.2.3 Recurrence property

A key question in earthquake probability study is if the recurrence interval variability (or aperiodicity) is primarily due to inherent complexity of the earthquake cycle process or if stress interaction between faults influences event timing in a deterministic fashion. The recurrence intervals of repeating earthquake ruptures are found to be highly variable, where the irregular recurrence of observed repeating events is possible in response to nearby large earthquakes, change in the strain rate, or time-dependent variation in the frictional strength of the fault [e.g., Lay and Kanamori, 1980; Vidale et al., 1994; Nadeau et al., 1994; Ellsworth, 1995; Marone et al., 1995; Schaff et al., 1998; Igarashi et al., 2003; Uchida et al., 2003; Nadeau and McEvilly, 2004]. The irregular recurrence of observed repeating events is possible in response to nearby large earthquakes, change in the strain rate, or time-dependent variation in the frictional strength of the fault [e.g., Vidale et al., 1994; Nadeau et al., 1994; Ellsworth, 1995; Marone et al., 1995; Schaff et al., 1998; Igarashi et al., 2003; Uchida et al., 2003; Nadeau and McEvilly, 2004]. The aperiodic RESs usually accompany with relatively large variations in seismic moment than those of quasi-periodic sequences [e.g., Vidale et al., 1994; Ellsworth, 1995; Nadeau et al., 1995; Nadeau et al., 2004]. The observation is consistent with the simulation model for earthquake cycle by Zöller et al. [2005b], which shows that a narrow range of size scales leads to quasi-periodic recurrence, whereas a wide range of source sizes bring about the aperiodic sequences. Additionally, when the range of size scales increases, the recurrence property changes from quasi-periodic to temporal clustered. Kato [2004] presents a detailed numerical simulation of two interacting asperities within an otherwise creeping fault zone. His results indicate that the degree of interaction is a function of distance between neighboring events. These facts show that inter-RES
distances, RES’s size, and differences in average sizes of the events in a RES have impact on the recurrence behavior of earthquakes.

1.3 Geologic, geodetic background of eastern Taiwan

The LV of eastern Taiwan is a clear geological boundary between Eurasian plate and the Philippine Sea plate with a length of 160 km and an averaged width of 4 km (Fig. 1-1). It has been recognized as a major Late Cenozoic suture zone and an example of extreme shear concentration in an oblique collision zone [e.g., Angelier et al., 2000]. The convergent rate between the two plates is about 8.2 cm/yr in the azimuth of 310° as determined by the GPS observations [Yu et al., 1997]. The LV is an oblique suture between the Coastal Range (the northernmost segment of Luzon volcanic arc to the east) and the Central Range (the easternmost portion of the continental margin to the west) [e.g., Biq, 1972; Bowin et al., 1978; Angelier et al., 1986], where the Central Range with a variety of terranes is a part of the underthrust Eurasian continent, and Coastal Range is mainly composed of Neogene andesitic volcanic units and associated flyschoid and turbidite sediments that belongs to the Philippine Sea plate and represents a section of the Luzon arc that is being accreted onto the Eurasian continent [Ho, 1986].

The LVF, on the east side of the LV, is the surface expression of a major shear zone striking N20°E and dipping approximately 55° to the ESE with left-lateral strike-slip component [Angelier et al., 1997; Yu and Kuo, 2001]. There is a large gradient in deformation patterns along the LVF. In the south where the most active segment is located, it is characterized by a creeping behavior with a high slip rate of 3.2 cm/yr in a N38°W direction [Yu and Kuo, 2001] and high level of microseismicity [Kuochen et al., 2004]. Focal mechanisms of the background earthquakes indicate thrusting with small component of strike-slip faulting [Kao and Jian, 2001]. The middle portion of LVF has a surface slip rate of 2.2 cm/yr in a N43°W direction [Yu and Kuo, 2001] and sparse earthquake activity. This seismic gap is considered to be either creeping or locked. The northern portion has been characterized by a surface slip rate of 1.1 cm/yr in a N10°W direction [Yu and Kuo, 2001] and a relatively high frequency of large earthquakes with varying type of faulting, indicating a complex stress regime [e.g., Hu et al., 1996; Wu et al., 1997; Kao et al., 1998].
Fig. 1-1: Geodynamic framework around Taiwan (modified from Malavieille et al. [2002]). Vector of relative motion between Philippine Sea plate and Eurasia plate is shown by yellow arrow [Yu et al., 1997].
1.4 Seismologic background of eastern Taiwan

The earthquake records are complete from 1900 and separated into three catalogs [Tsai, 1986]. (1) 1900-1973 catalog: On-site analog seismic records began in 1900, including instrumental seismicity data from the Central Weather Bureau’s conventional seismographic network since 1973, instrumental seismicity from the World-Wide Standard Seismographic Network (WWSSN) since 1961, and various documentary descriptions. (2) 1973-1991 catalog: The operation of Taiwan Telemetered Seismic Network (TTSN) began in 1973 from Institute of Earth Sciences’ telemetered seismographic network and temporary local seismographic networks. (3) 1991-present catalog: TTSN stations were integrated into Central Weather Bureau Seismic Network (CWBSN) of a total of 75 stations.

Historic earthquakes of $M \geq 6$ in eastern Taiwan are shown in Fig. 1-2. Since 1900, earthquakes of $M \sim 7$ or larger occurred in 1908, 1919, 1937, 1938, 1951, and 1957. Among these major events, the most significant earthquake sequences occurred in 1951, called “1951 Hualien-Taitung earthquake sequence” (1951 H-T sequence), which caused several surface ruptures from the north of the LVF to the south. Including of 1951 sequence itself, eight $M \geq 7$ earthquakes occurred in the 50 years before the 1951 H-T sequence, yet only two $M \geq 7$ events occurred in the 50 years after 1951. Here we show the historic earthquakes since 1900 in mapview (Fig. 1-3A) and small earthquake occurrence rates in 1979-1999 (Fig. 1-3B). Note that the active small seismicity concentrates on the regions near Hualien, Fenglin, and Chengkung cities, as shown by yellow-to-red color in Fig. 1-3B, where these three areas had been experienced $M \geq 7$ earthquakes since 1900. This suggests that the creeping LVF characterized by a number of small earthquakes can generate large earthquakes.
Fig. 1-2: M ≥ 6 earthquake activities in eastern Taiwan from 1900 to 2004. Open circles indicate that more than one earthquake occurred that year. Star indicates the 1951 Hualien-Taitung earthquake sequence. Table on the right show a list of the M ≥ 6 events in the 1951 Hualien-Taitung earthquake sequence.

Fig. 1-3: Mapview of historic M6+ earthquakes and seismicity rate. (A) The historic earthquakes (M≥6) since 1900. Events are separated into three different periods, 1900-1951 (pre-1951 period), 1952-1990 (post-1951 period), and 1991-2002 (recent decades) shown by black, red, and blue open circles, respectively. (B) Seismicity rate for M ≥ 2.7 events during the period of 1979-1999 (20-yr prior to the September 21, 1999 Chi-Chi earthquake).
1.4.1 Fault segmentation from background seismicity rate

Fig. 1-4A shows that the 1900-2004 catalog is not uniform in its magnitude of completeness (Mc). The Mc for the 1900-1972 catalog (instrumental seismicity data from the Central Weather Bureau’s earlier seismographic network) is 4.75. The Mc for the 1973-1990 catalog from the Taiwan Telemetered Seismographic Network (TTSN) and the 1991-2004 catalog from the Central Weather Bureau Seismic Network (CWBSN) is 2.7 and 2.1 respectively. We obtain a representative pre-1951 $M \geq 4$ (i.e., $M \geq 4$) background rate by extrapolating from $M \geq 4$ rate through the frequency-magnitude distribution. We also calculate the $M \geq 4$ seismicity rate in the 20-year period before the 1999 Chi-Chi earthquake (1979-1999) for comparison.

Considering the large location uncertainties and low detection capability for pre-1951 earthquakes, we determine the background seismicity rate using a 10 km ×10 km grid. Taking into account only the earthquakes associated with the LVF (see the box in Fig. 1-4B), we plot the along-strike variation of $M \geq 4$ seismicity rate in Fig. 1-4C. The different study periods reveal a consistently high seismicity rate in the Hualien area. Fig. 1-4D shows a quantitative measurement of relative seismicity rates based on the maximum and minimum values in each segment window at different periods. In a given time period, the background rate for each segment is normalized by the Hualien rate. Therefore, the Hualien segment seismicity rates are normalized to be 1. The result of background seismicity rate computation offers useful information for further study in fault segmentation of the LVF.
1.4.2 Aftershock duration of M6 events

The interplate boundaries capable of generating large earthquakes have been selected for the most focused targets in seismic hazard assessment. It has been suggested that the areas of largest slip release correlate with high $b$-value regions, and unusually low $b$-value regions can be regarded as an indication of highly stressed patches in the fault [Wiemer and Katsumata, 1999; Schorlemmer and Wiemer, 2005]. Aftershock duration also offers a rich source of information about fault segmentation. The duration of the aftershock sequences seems to depend on the nature of the faults,
the focal depth, and the stress distribution on the fault. *Toda and Stein* [2002] examined the aftershock durations in Parkfield, California and found that the aftershock duration in the locked section is eight-fold longer than that in the creeping section. *Zöller et al.* [2005a] used a simulation model to show that the high ratio of creep coefficient leads to a fast aftershock decay rate.

The aftershock sequences in eastern Taiwan are selected if they are spatially close to the fault segment of interest and have a $M \geq 6$ mainshock (Fig. 1-5). Considering that the aftershock expansion pattern may be strongly associated with fault zone properties [*Tajima and Kanamori*, 1985], we determine the aftershock area using the events that occurred within one month after the mainshock (see the dashed ellipse in Fig. 1-5). In the Taitung segment there were no $M \geq 6$ earthquakes between 1973 and 2004, so we choose a $M 5$ mainshock instead. This $M 5$ mainshock occurred on 2003/12/18 and was likely triggered by the 2003/12/10 $M_L 6.4$ Chihshang earthquake that was used to estimate the aftershock duration of another segment (the Chihshang segment). However, the aftershock area of the $M 5$ Taitung event (dashed ellipse in Fig. 1-5D) is spatially isolated enough from the $M_L 6.4$ Chihshang aftershock (dashed ellipse in Fig. 1-5C), so that we may have an independent measurement of aftershock duration from the Chihshang segment. The selected aftershocks for different segments are then used to plot the earthquake numbers as a function of time (Fig. 1-6). With the estimated background rates prior to the mainshock, we can address the decay rate to the background level. As a result, the calculated durations at the Hualien, Yuli, Chihshang, and Taitung segments are 5.5 yr, 0.6-1.4 yr, 0.2 yr, and 0.9-3.3 yr respectively.

During the 1951 H-T sequence, the Yuli segment was characterized by the longest fault rupture and largest co-seismic surface deformation. During the inter-seismic period over the past two decades, the Yuli segment has experienced a similar rapid slip rate of 2-3 cm/yr, comparable to that of the creeping Chihshang section [*Yu and Kuo*, 2001]. However, unlike the abundant background seismicity in the Chihshang area, the Yuli area has been characterized by a much lower seismicity rate during the past few decades (see the 1979-1999 rate in Fig. 1-4). Whether strain energy accumulated in the Yuli segment has been released aseismically or is building up for the next big earthquake remains a subject of controversy.

The aftershock duration in the Yuli segment is calculated to be in a range of 0.6 to 1.4 yr. *Toda and Stein* [2002] studied the response of the Parkfield-Cholame section of the San Andreas fault to the 1983 Coalinga-Nuñez earthquake, which shows an apparent increase in aftershock duration from 0.6 yr in the creeping section to 5 yr in the locked section. They pointed out that the limited maximum size of earthquakes in the creeping section probably yields the small aftershock duration. In this study, the
$M_6$ mainshock on the creeping Chihshang fault is calculated to have an aftershock duration of less than one year, which is consistent with the short duration observed in the creeping section along the San Andreas fault. The Yuli segment, characterized by a 0.6 - 1.5 yr aftershock duration, may as well experience aseismic deformation during the past decade. The creeping behavior of the Yuli segment is supported by the inversion result from trilateration and leveling data [Yu et al., 1990; Yu and Kuo, 2001].

Fig. 1-5: Aftershock distribution of four major events. Dashed ellipses encircle the chosen aftershock zones that occurred one month following the mainshock for (A) the May 20, 1986 $M_{6.5}$ earthquake in Hualien, (B) the June 10, 2003 $M_{6.5}$ earthquake near Yuli, (C) the December 10, 2003 $M_{6.4}$ earthquake in Chihshang, and (D) the December 18, 2003 $M_{5.0}$ earthquake in Taitung.
1.4.3 Earthquake triggering of the 1951 Hualien-Taitung sequence

As the most destructive seismic episode ever known in eastern Taiwan, the 1951 H-T sequence consisted of sequential ruptures along four distinct fault segments [Gutenberg and Richter, 1954; Lee et al., 1978; Abe, 1981; Hsu, 1985; Cheng et al., 1996] as shown in Fig. 1-7. The $M_L$ 7.3 mainshock occurred near Hualien at 21:34 (GMT) on October 21, 1951 and created a surface rupture on the Meilun fault in the northern end of the LVF. Several hours after the mainshock, two other $M_L \geq 7$ events occurred in the same area (Events 2 and 5 in Fig.1-7). One month later, on November 24, two major earthquakes struck the middle part of the LVF. They occurred just 3 minutes apart and broke two different surface faults, the Chihshang fault and the Yuli fault. The first $M_L$ 6.0 event on the Chihshang fault was located ~100 km away from the mainshock and 5 km south of the second $M_L$ 7.3 event on the Yuli fault [Hsu, 1962; Cheng et al., 1996]. Thirty-six days after the mainshock, the last major earthquake occurred in the southern end of the LVF at Taitung, with $M_L$=6.0. Overall, the 1951
H-T sequence is characterized by a leaping behavior, that is, the first fault ruptured in the north end, the second ruptured 100-km away from the mainshock in the southern segment, the third ruptured in the middle segment, and the last occurred again in the southern end of the LVF [Taiwan Weather Bureau, 1952]. The sequential surface ruptures of the Hualien, Chihshang, Yuli segments, and the major events on the Taitung segment reported by Hsu [1962] are shown in Fig.1-7. Magnitude, occurrence time, and identify of each $M \geq 6$ events in this H-T sequence is shown in Table 1-1. Chen et al. [2008] seek to understand how segmented faults react to nearby large earthquakes, and why the subsequent events did not rupture the nearest fault first using the static Coulomb hypothesis and rate/state friction formulations.

![Fig. 1-7](image)

Fig. 1-7: Mapview of the 1951 Hualien-Taitung earthquake sequences. (a) Geodynamic framework of Taiwan. (b) Temporal and spatial distribution of major earthquakes (stars) during the 1951 H-T earthquake sequence in eastern Taiwan. Number in the star corresponds to the ID number listed in Table 1-1. White lines indicate the current active faults, and bold black lines indicate surface ruptures of the 1951 H-T earthquake sequence. Dashed line in the south indicates the non-ruptured fault segment during 1951 sequence. MF, Meilun fault; YF, Yuli fault; CF, Chihshang fault; LF, Luyeh fault.
Table 1-1 Major events ($M \geq 6$) in the 1951 H-T sequence

<table>
<thead>
<tr>
<th>ID in Fig. 1-7</th>
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<th>time</th>
<th>$M_L$</th>
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</table>

Calculations of static Coulomb stress change ($\Delta$CFF) and the stress evolution at successive rupture sites is illustrated by sequential plots of the Coulomb stress change in Fig. 1-8 [Chen et al., 2008]. Large earthquakes generally relieve stress along the rupture (red zones) and transfer the stress beyond the rupture tips and off the fault (blue zones). We first calculated the static stress changes caused by the $M_L$ 7.3 Hualien earthquake on Events 2 and 5 that clustered in the Hualien area (Fig. 1-8A-B). Fig. 1-8A shows increased Coulomb stress transfer of up to 1 bar at the location of the next event (Event 2). In addition, the events that occurred between the modeled event and next target earthquake (open red circles, hereafter called inter-event aftershocks) are mostly located in the stress increase zone. The combined effect of the mainshock and Event 2 on the fault plane of Event 5 is plotted in Fig. 1-8B. Event 5 and half of the inter-event aftershocks are distributed in the negative stress change regions. Considering the 6-8 km location uncertainty at depth for this offshore event [Cheng et al., 1997], Chen et al. [2008] examined how the stress transfer on the solved fault plane changes significantly with target depth and found that the stress change in the area between the fault planes of Events 2 and 5 to be very sensitive to the target depth. This suggests that depth uncertainty may perhaps explain why Event 5 occurs in the stress shadow in Fig. 1-8B.

The Coulomb stress change caused by the three $M$ 7+ Hualien earthquakes is examined at the hypocentral region of the $M_L$ 6.0 Chihshang earthquake (Fig. 1-8C). The Coulomb stress is raised by 3-20 bars near the edges of the Hualien rupture zone. Notice that the Chihshang rupture zone (denoted by a bold black line in Fig. 1-8C)
lies in a region with 0.14 bar ΔCFF, suggesting that the Hualien earthquakes may have brought the Chihshang fault closer to failure even though the stress magnitude is small. The inter-event aftershocks in Fig. 1-8C are mostly distributed in the zone of positive stress change on the modeled fault plane. It is noteworthy that three $M_{7+}$ Hualien events largely increase the stress in the middle to southern portions of the LVF (Fig. 1-8C), and that the Yuli rupture zone is also located in a positive ΔCFF zone ($\Delta$CFF = 0.2 bar). This brings into question some temporal features that conventional Coulomb stress hypothesis cannot explain alone. There is clearly a need, therefore, to further involve a temporal dimension in our computation.

Fig. 1-8D shows ΔCFF induced by the combined coseismic displacement of the three $M_{7+}$ Hualien events and the $M_{L}$ 6.0 Chihshang event. The ensuing 40-km rupture of the Yuli fault lies within a region of ΔCFF between 0.3-0.9 bars, and the epicenter of the $M_{L}$ 7.3 Yuli earthquake occurred in a region where ΔCFF is about 0.7 bar. Thus, the Yuli earthquake could have been caused by stress triggering. Fig. 1-8E shows cumulative stress changes computed from the combined coseismic displacements of the three $M_{7+}$ Hualien earthquakes, the $M_{L}$ 6.0 Chihshang earthquake, and the $M_{L}$ 7.3 Yuli earthquake. The Taitung segment (the Luyeh fault) lies within the region of ΔCFF of 0.1 bar, suggesting that the subsequent $M_{L}$ 6.0 Taitung earthquake was also located in one of the triggered zones. Consequently, regardless of the temporal behavior of triggering, a series of stress calculations indicates that static stress transfer appears to advance slip on the subsequent major earthquake segments in space.

However, the stress transfer alone cannot explain triggering across 100 km. With the rate/state stress transfer model, Chen et al. [2008] computed the temporal order of encouraged ruptures on different segments along the collision boundary. The rate/state friction model requires values for the stress change caused by the three $M_{7+}$ Hualien events, the aftershock duration $t_a$, the seismicity rate, and the combination of normal stress and constitutive constant $A\sigma$ on each receiver fault (i.e., the Yuli, Chihshang, and Taitung segments).
Fig. 1-8: Result of cumulative Coulomb stress changes caused by large earthquakes in the 1951 H-T earthquake sequence. The location uncertainties determined by Cheng et al. [1997] are denoted by crosses. Bold red lines denote the surface projections of fault models that are applied to calculate stress state. Bold black lines denote the next surface rupture. Dashed black lines are non-modeled fault planes. Red star shows hypocenter projected at the target depth for stress calculation. Open star shows hypocenter projected at the target depth for the next earthquake to rupture. Red open circle indicates earthquakes that occurred between two major events. Note that target depths are different between the panels according to the depth of the next major event. (A) ΔCFF at a depth of 5 km due to the co-seismic displacement of Event 1. Note that the 1 km source depth of Event 2 is too shallow and therefore manually changed to 5 km. Event 1 fault segment is indicated by a bold red line. The relocated epicenters of Event 1 and the next Event 2 are shown by a red star and open black star respectively. (B) ΔCFF at a depth of 18 km due to the co-seismic displacements of Event 1 and Event 2. (C) ΔCFF at a depth of 16 km due to the co-seismic displacements of Events 1, 2, and 5. (D) ΔCFF at a depth of 36 km due to the co-seismic displacements of Events 1, 2, 5, and $M_\text{L}$ 6.0 Chihshang earthquake. (E) ΔCFF at a depth of 49 km due to the co-seismic displacements of Events 1, 2, 5, $M_\text{L}$ 6.0 Chihshang earthquake, and $M_\text{L}$ 7.3 Yuli earthquake.
Using the estimated $M_{6+}$ earthquake rate derived from rate/state stress transfer model [Dieterich, 1994] and assumed time interval of 1-yr, Chen et al. [2008] computed the 1-yr earthquake probability in each 1 km$^2$ cell. Figure 1-9 A shows the 1-yr earthquake probability for $M_{6+}$ events. Following the mainshock and the two $M_{7}$ aftershocks in the Hualien area, the predicted $M_{6+}$ probability in the Chihshang segment is 2.2-3.1%, which is on average higher than that of Yuli segment (1.5-3.6%) and Taitung segment (0.7-0.9%) considering a wide range of $A\sigma$. Chihshang having the highest probability is consistent with the Chihshang rupture that occurred 34 days after the mainshock. After the Chihshang rupture, the $M_{6+}$ probability in the Yuli segment is predicted to be highest (4.9% in average), which is also consistent with the Yuli rupture that occurred 3 minutes later (Figure 1-9 B). This shows that the preferential triggering is observed when varying background seismicity rate, aftershock duration, and Coulomb stress change from different fault segments are resolved onto the probability model. The results show that 34 days following the major shocks in Hualien, the Chihshang segment had a higher $M_{6+}$ earthquake probability due to its significantly higher (at least an order of magnitude) background seismicity rate than the other two segments. After the Chihshang event, the rate/state model predicted a higher $M_{6+}$ earthquake probability in the Yuli segment, also matching the observation. In this case, the Yuli segment was triggered ahead of the Taitung segment because of its larger increase in Coulomb stress change.

Consequently, modeling of fault interactions and earthquake triggering during the 1951 $M_{L} 7.3$ H-T earthquake sequence allows us to understand the temporal patterns of distinct ruptures along the LVF in terms of static stress transfer and rate/state stress transfer models. In the 1951 H-T sequence, the sites of most triggered faults were located where the Coulomb stress was calculated to have increased by the previous event. The conventional static Coulomb stress change due to the $M_{L} 7.3$ Hualien earthquake, however, cannot explain the temporal triggering behavior where the first off-fault M6 aftershock did not occur at the closer fault segment ($\Delta$CFF for the Yuli fault ~0.2 bars) but jumped over a long distance (~100 km) and ruptured the Chihshang fault ($\Delta$CFF ~0.14 bars). Using the static Coulomb stress change coupled with the rate/state stress transfer is probably the correct path to explaining the leaping behavior of faulting along these four distinct faults. Our rate/state stress transfer modeling result implies that the physical properties associated with the high level of seismicity in the Chihshang segment are significantly important for the leaping triggering.
With measurements of the rate/state friction parameters among the different segments along the LVF, we are able to address fault segmentation as a function of time (e.g., along-strike variation of background seismicity rate), which is helpful for further earthquake probability studies in eastern Taiwan. From the short aftershock duration measurement in the Yuli segment, we also suggest that this seismically inactive segment is likely to experience seismic creeping instead of locked behavior with a larger earthquake potential.

Fig. 1-9: One year probability of $M$ 6+ earthquakes over the time period (A) between the mainshock and the Chihshang event and (B) between the Chihshang event and the Yuli event. The estimates for the Yuli, Chihshang, and Taitung segments are represented by blue, green, and red respectively. Vertical bars indicate the estimates using a wide range of $A_\sigma$ with the average denoted by filled squares. Vertical arrows indicate the occurrence of major events of interest in the 1951 H-T sequence.

1.5 A Thesis roadmap

This primary goal of this thesis is using repeating earthquake sequences (RESs) observation in eastern Taiwan to infer the fault behavior at depth and to make inferences about the nature of repeating earthquakes. To achieve this, it was necessary to develop method for solving the repeating earthquake identification problem at the study area where the seismic station coverage is sparse and one-sided. Chapter 2 details the identification problems in the study area and provides a relatively objective method for solving them. Chapter 2 is a part of the manuscript titled “Characteristic repeating earthquakes in an arc-continent collision boundary zone: The Chihshang fault of eastern Taiwan” that submitted to Earth and Planetary Science Letters on Feb. 2007. The co-author of this manuscript, Robert Nadeau, was the primary supervisor who provided the analytic expressions in the proposed relocation method and gave critical suggestions about the reliability of our identification scheme.
Using the new identification method, we are able to illustrate the characteristics of repeating earthquakes observation in space and time. In Chapter 3 we show that the RESs are found on the Chihshang fault of the southern LVF and in the Hualien RESs zone of the northern LVF, where the fault plane associated with the RESs is still under debate. The spatial distribution of RESs and their recurrence properties at these two areas are separately addressed in Chapter 3.1. Estimations of deep fault slip rates using RESs’ recurrence interval and seismic moment release with the measurement uncertainty are addressed in Chapter 3.2, which leads to the following section for spatiotemporal variation of deep slip rate in Chapter 3.3. In Chapter 3.4 we compare the two RESs clusters to discuss how the similarity and difference reflect the characteristic property of active fault zones, which follows a schematic fault model in Chapter 3.5. The content associated with the Chihshang fault zone is from the manuscript titled “Characteristic repeating earthquakes in an arc-continent collision boundary zone: The Chihshang fault of eastern Taiwan” submitted to Earth and Planetary Science Letters on Feb. 2007, and the content associated with the Hualien RESs zone is part of the manuscript submitted to Journal of Geophysical Research on Nov. 2007, titled “Variability of the repeating earthquakes behavior along the Longitudinal Valley fault zone of eastern Taiwan”. The co-authors of this manuscript, Ruey-Juin Rau and Jyr-Ching Hu, gave useful suggestions and valuable insight.

Chapter 4, which was partially published at Geophysical Research Letters titled “Towards a universal rule on the recurrence interval scaling of repeating earthquakes? ”, adds data from the Hualien RESs zone in a previously published work. The co-authors Robert Nadeau and Ruey-Juin Rau provided useful advice and encouragements. Using data from Robert Nadeau, Naoki Uchida, and some published materials, we discuss the magnitude and recurrence interval statistics from RESs and its implication in unify a wide variety of fault zone processes.

In Chapter 5, we summary the characteristics of RESs observed in eastern Taiwan and recommend the future works for potential of repeating earthquake study.