

The Earthquake Cycle of the Manyi Fault, Tibet

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Trinity, 2013

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Declaration

The contents of this thesis are all my own work, except where otherwise stated. The views and opinions expressed herein are mine and not necessarily those of any other person or body unless so attributed.

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To Mum and Dad; I hope I have made you proud.

Abstract

The Earthquake Cycle of the Manyi Fault, Tibet

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Trinity Term, 2013

This thesis focuses on the Manyi Fault in Northern Tibet which experienced a M_w 7.6 earthquake in 1997. The remoteness of the area limits the feasibility of measurements in the field, however the climate makes it ideal to study by remote sensing, specifically Interferometric Synthetic Aperture Radar (InSAR). The mechanics of the earthquake have been well documented however there are still numerous questions about the other stages of the earthquake cycle (postseismic and interseismic) across the fault. Previous studies of the postseismic motion across the Manyi Fault using four years of ERS SAR data show the deformation can be explained by either viscoelastic relaxation of a Standard Linear Solid body with a viscosity of 4×10^{18} Pa s or afterslip. We use the ERS timeseries and ratemaps formed from a network of ENVISAT SAR scenes from 2003-2010 to analyse the postseismic deformation. We create a series of afterslip models based on rate-and-state frictional laws, along with series of viscoelastic models with various rheologies (Maxwell and Burgers). Our results show that an afterslip model fits the data slightly better than a Burgers rheology but not within resolvable errors. A range of afterslip models fit the data well, with frictional parameters ranging from 8×10^{-4} to 2×10^{-3} and a preseismic slip rate of 8 to 20 mm/yr. The best-fitting Burgers rheology has a Kelvin element viscosity of 4×10^{18} Pa s and Maxwell element viscosity of 6×10^{19} Pa s. We analyse the interseismic InSAR signal observed before the 1997 earthquake using ERS data from 1992-1997 to find that the Manyi Fault was accumulating strain at 3 ± 2 mm/yr. We also find the seismic locking depth was 22 ± 15 km which correlates with the maximum depth of slip during the earthquake. We show there is no significant deformation across the fault to the north of the Manyi Fault which may be an extension to the Kunlun fault. We discuss an analytical 2D thin viscous channel model from literature that has been shown to match the data in this thesis. We show that, once errors are properly accounted for, their model cannot explain both the post and preseismic datasets.

Extended Abstract

The Earthquake Cycle of Manyi Fault, Tibet

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Trinity Term, 2013

The earthquake cycle is a vital tool for explaining continental deformation and estimating the seismic hazard of faults. The earthquake cycle states that strain is built up during the interseismic motion due to the relative motion of plates. This accumulated strain is then released during the coseismic phase. In the immediate aftermath of earthquake the residual stress field can be dissipated in what is known as postseismic motion. This transient motion decays with time until eventually the major deformation is from the interseismic motion.

The Manyi Earthquake

The Manyi Earthquake was one of the largest earthquakes to occur on the Tibetan Plateau in the modern age. The fault lies to the west of the Kunlun fault, one of the major faults on the plateau. The area around the event is particularly remote making any field observation of the event impossible. The area however is ideal for InSAR as there is little vegetation and topographic change across the fault and there is little variation in atmospheric water vapour. Previous studies using ERS data from the

period 1997-2001 that modelled the postseismic deformation of the Manyi Earthquake determined there are two possible causes for the observed surface deformation in the area: afterslip and viscoelastic relaxation. The ambiguity may be resolvable by using more data from later stages of the postseismic cycle which is the aim of this thesis.

Extending the Timeseries

Unfortunately a malfunction with one of the gyroscope on the ERS satellite in 2001 meant SAR data were no longer suitable for interferometry. The launch of the ENVISAT satellite in 2003 provided an opportunity to continue the timeseries into the decade after the event. I process the data from the main track covering the Manyi Earthquake to see if any deformation is still observable. As the ENVISAT data is so much later after the earthquake, the signal in any one interferogram is significantly smaller than the signal observed using the ERS data (assuming the same temporal baseline). I explore a series of techniques to improve the signal-to-noise ratio of the ENVISAT data including stacking, chaining and using a network approach. A network approach allows us to form three ratemaps covering 5.5-8 years, 10-11 and 11-13 years after the event that clearly show the motion is still occurring. I also analyse the data to check there is no topographically correlated signal in each ratemap.

Pre-1997 Interseismic Slip Rate Estimation

Postseismic motions decay with time after the earthquake, eventually to the point that the largest signal is the interseismic strain accumulation across the fault. As the observed signal in the ENVISAT data is significant smaller than in the ERS data it is possible that the postseismic signal has faded to the point it has returned to the interseismic state. No estimate of the interseismic slip rate exists in the vicinity of the Manyi Fault so I use the ERS from before the 1997 event to estimate one. As

the signal is again small we use a network technique to improve the signal-to-noise ratio of the data. I model the strain accumulation using the established Savage and Burford Model and compare that to the pre-1997 ERS data to determine an interseismic rate is 3 ± 2 mm/yr with a locking depth of 22 ± 15 km. I also model the strain accumulated across a fault to the north of the Manyi Fault and find that there is no significant slip across it in the observation period. Both of these results have important implications for the validity of block models in the area which, up until the publication of results from this study, assumed slip of 10 mm/yr across one of the faults. Assuming the interseismic slip rate is constant between earthquake cycles I also show that the interseismic rate is insufficient to explain the deformation within the ENVISAT data so it must be due to a postseismic deformation mechanism.

Postseismic Deformation Modelling

Afterslip Modelling

Previous studies of the postseismic deformation showed that afterslip was a viable mechanism to reproduce the observed ERS deformation pattern. The models were formed from inversion of the observed data to calculate the temporal evolution of slip on an extension of the fault with depth. In this thesis I use a different technique to model the afterslip. I use the hot-friction approximation of the rate-and-state frictional laws to create a series of models with a various preseismic slip rate and frictional parameters. I then perform an parameter-space search to find the best-fitting models to the ERS data. Once I have narrowed the range of acceptable models I then compare the data to the ENVISAT data. I find that there is a trade-off in the best-fitting model between the preseismic rate and the frictional parameters however the models with the lower values ($(a - b) = 1-2 \times 10^{-3}$ and V_0 of 10-20 mm/yr) tend to be the most stable with increasing time. These values are in line with those postseismic studies of other

events around the world.

Viscoelastic Relaxation

A significant stress change in the Earth can cause viscoelastic flow. In the previous study of the Manyi postseismic signal the observed deformation could be modelled by the viscoelastic relaxation of a Standard Linear Solid (SLS) body beneath the an elastic lid. SLS bodies however do not reproduce the observed deformation pattern observed in the ENVISAT data. I create a series of models with different rheologies (Maxwell and Burgers) and earth-models (layered and half-space) to ascertain the best-fitting model that fits the data best over the entirety of both datasets. I find the best-fitting model is a Burgers body with a Kelvin and Maxwell element viscosity of 4×10^{18} and 6×10^{19} Pa s. The behaviour of this model is similar to that of a SLS for the ERS data however is a significantly better match to the ENVISAT data.

Concluding Remarks

Comparison of the two models and their misfits show there is not a significantly preferable model. The afterslip model has a slightly lower RMS misfit however the improvement is not resolvable once errors in the datasets are accounted for.

Both these models however may be mimicking the behaviour of a more complex rheology such as a power-law. The best-fitting Burgers model at each time point changes with increasing time. The best-fitting solution shown here a good approximation of these best-fitting models at each time as the errors do not allow us to resolve the variation with time. This evolution of the characteristic timescale of deformation is typical behaviour of a body with a power-law rheology.

I compare the results of this study to other results in the literature. given the data available at the time of study I find our results agree with those in literature.

One study attempts to reconcile the preseismic data determined in this study and the postseismic ERS data across the Manyi Fault with a 2D thin viscous channel model. I show that when all the postseismic data are included and the errors in the preseismic are accounted for then the proposed thin viscous channel is unsuitable.

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As the Beatles said it's a long and winding road and I would not have got to the end without help from a large number of people.

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Chapter 1

Introduction

Earthquakes have never been far from the news headlines in the 21st Century such as the M_w 9.2 Boxing Day 2004 Earthquake in Sumatra and the M_w 9.0 Tohoku earthquake. Earthquakes themselves are not a 21st Century problem, they have been responsible for some of the worst natural disasters in history, such as the January 23, 1556 Earthquake in China which killed nearly 1 million people. They are an increasingly pressing issue however as the number of deaths due to earthquakes has risen dramatically over the last thousand years (*Jackson*, 2006). This change in death rate is attributed not to an increase in seismic events but to an increase in population, especially in developing countries. Fatal earthquakes are generally within the continental interiors (Figure 1.1) as people living in the Alpine-Himalayan belt settle in close proximity to faults which provide resources essential to early settlers, such as water. With knowledge of where (and ultimately when) earthquakes occurs we could reduce the number of fatalities but first we must improve our understanding of how the Earth deforms.

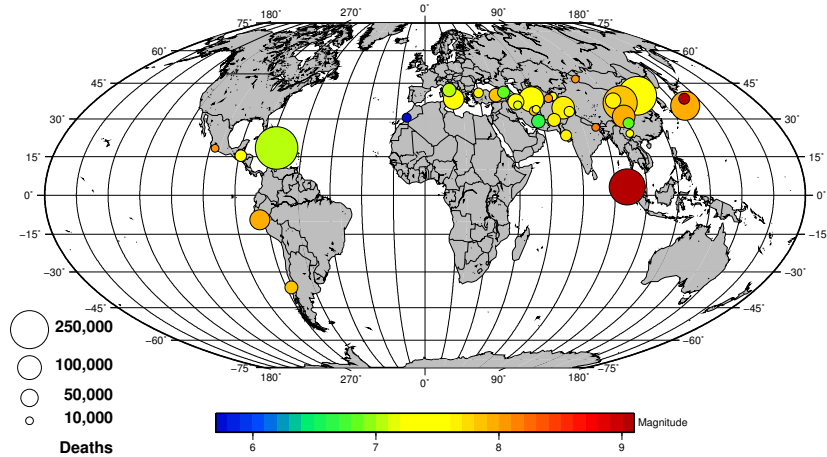


Figure 1.1: Map of the global distribution of earthquakes with death tolls over 10,000. Circles are scaled according to number of deaths and colour according to the magnitude of the event. Map is an update of *Elliott* (2009) and data are taken from http://earthquake.usgs.gov/earthquakes/world/world_deaths.php

1.1 Continental Tectonics

The concept of plate tectonics saw the Earth's lithosphere mapped into a series of discrete rigid plates separated by narrow boundaries that are approximated as a fault. This could explain the occurrence of earthquakes at all boundaries where one side of the fault was oceanic in nature. The traditional theory of plate tectonics could not explain the presence of deforming zones over 1000 km in extent that occur when continents collide (*Molnar*, 1988).

To explain continental deformation two classes of models have been suggested. One model breaks up continental collision zones into a series of smaller micro-plates which can be as small as 100 km across. This model, sometimes referred to as a block model, is comparable with traditional plate tectonics only on a smaller scale. The extent of blocks and slip rates across them are taken from measurements of surface motion (*Meade and Hager*, 2005). The second model is a continuum model, which treats the continental lithosphere as a viscous medium that can flow, where the flow is driven by internal buoyancy and other external forces (*England and McKenzie*, 1982). As the size of blocks in a block model decreases they more closely approximate a continuum model (Figure 1.2). Any deformation at depth is filtered through an elastic lid so we

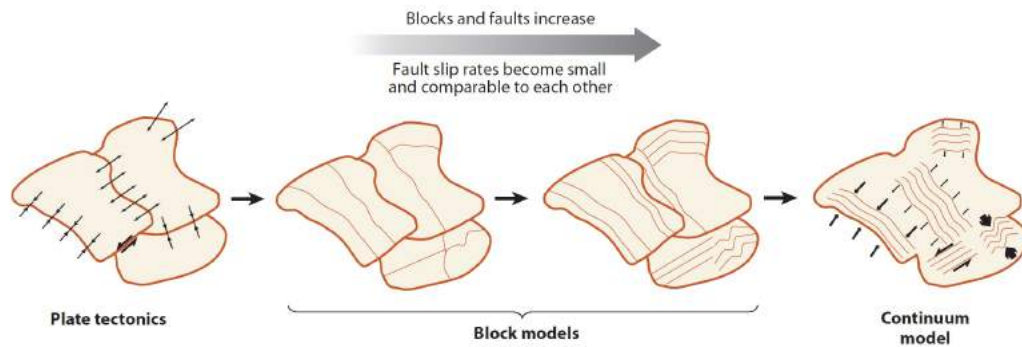


Figure 1.2: Image showing the transition from global plate kinematics of block models to continuum driven models. Fault slip rates become comparable to each other over the course of the transition. Bold lines denote major block boundary faults. Thin lines are faults separating smaller blocks. Reproduced from *Thatcher* (2009).

can only observe long wavelength (with respect to the lid thickness) deformation at the surface which makes block and continuum models increasingly difficult to differentiate using surface measurements of deformation. One other critical difference between the two models is the strength of the lithosphere required for each model. In block models the upper crust is stronger than the lower crust/mantle; in continuum models deformation is driven by flow in the lower lithosphere, which is stronger than the upper crust. Insights into the strength profile of the lithosphere can be gained from postseismic deformation, one of the stages of the earthquake cycle.

1.2 The Earthquake Cycle

The concept of the earthquake cycle was first proposed by H.F. Reid following observations made before and after the San Francisco Earthquake of 1906 (*Reid*, 1910). By comparing triangulation surveys from before and after the earthquake, he modelled an earthquake as a sudden relaxation of slowly built up elastic strain caused by continuous and long-term motions on either side of the fault (Figure 1.3).

Technological advances allowed more accurate observations of the cycle that found failings in Reid's elastic model. The original model required a homogeneous elastic half-space, which failed to take into account how the properties of rocks vary with depth.

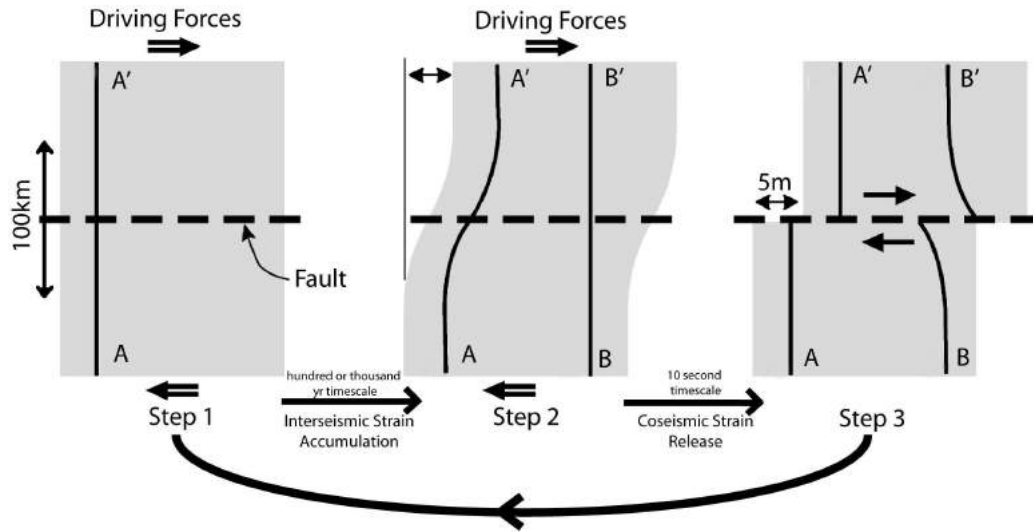


Figure 1.3: Reid's Earthquake Cycle from *Wright* (2002). Step 1 is an aerial view of a theoretical fault immediately after an earthquake. Note that A-A' is straight. Step 2 is immediately before the next earthquake and shows the deformation pattern as interseismic strain accumulates across the fault. Note that A-A' is now curved and B-B' is straight just before the strain is released. Step 3 shows the deformation just after the coseismic rupture has finished. A-A' is now straight again but offset and B-B' is now curved with a displacement that decays with distance from the fault; representing the deformation field from the coseismic rupture.

Reid's model cannot explain why deformation in between events is localised around faults nor why there is a large postseismic deformation signal after the earthquake. It is now accepted that, for continental earthquakes at least, there are three phases to the earthquake cycle (*Wright*, 2002). The first phase is the interseismic phase. This involves the accumulation of strain across faults at a steady rate. This strain is released during an earthquake during the next phase of the cycle known as coseismic phase. The coseismic phase is generally short in nature, lasting seconds to minutes. The third phase is the postseismic phase where transient deformation relieves stress increases caused by the earthquake through a variety of inelastic means. The postseismic phase eventually decays back to the interseismic phase and the cycle repeats. We discuss each of these phases in more detail in the following sections.

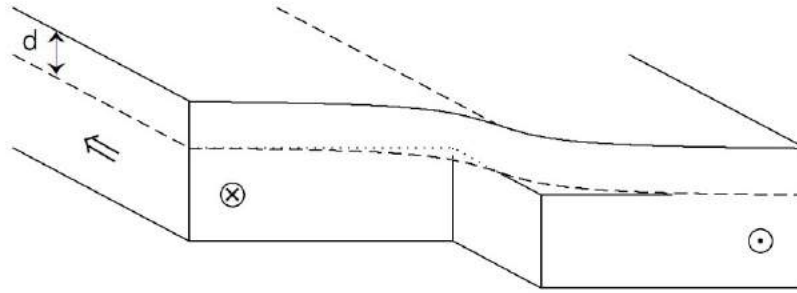


Figure 1.4: Schematic view of the Savage and Burford model (*Savage and Burford, 1973*). The top layer down to depth d is locked and does not move, however below d the fault stably slides at a given rate s that is dependent upon far-field forces.

1.2.1 Interseismic Deformation

Interseismic strain is caused by distant forces resulting from relative plate motions that deform an area around a locked fault. The motions were those first described after the 1906 San Francisco Earthquake by *Reid* (1910). One of the first models of interseismic strain accumulation was devised by *Savage and Burford* (1973) by analysing triangulation data across the San Andreas Fault. Their model describes a fault as a buried infinitely long screw dislocation in an elastic half-space (Figure 1.4). Lateral motion occurs along the dislocation at a rate s below a depth d . Between the surface and depth d the fault is considered locked due to friction, hence d is often referred to as the locking depth of the fault. As the fault is infinitely long there is no variation in strain parallel to the fault, however the model gives a variation in elastic strain with perpendicular distance away from the fault (y) given by Equation 1.1;

$$y = \frac{s}{\pi} \tan^{-1} \left(\frac{x}{d} \right) \quad (1.1)$$

In this model the majority of the deformation is localised near the fault (as can be seen on profile A-A' in Step 2 of Figure 1.3). The proposed model is very simple as shear is localised onto one plane at depth. *Prescott and Nur* (1981) proposed a more complex model where the shear is instead distributed across a finite width region. Such additional details however are not generally required to model interseismic strain as any

deformation at depth is filtered through the locked elastic layer above the dislocation. This causes the two models to be indistinguishable apart from in the most extreme situations.

Faults are not perfectly elastic as the Savage and Burford model assumes. *Savage and Prescott* (1978) incorporate viscoelasticity in the half-space and find that there is again little discernible difference in the surface motions. Recent work by *Hetland and Hager* (2006) and *Johnson et al.* (2007) have extended this viscoelastic model and found similar results once the postseismic transients have decayed. These factors mean that the simple Savage and Burford model is still the most common model used to constrain interseismic deformation and it is the model used in Chapter 5 of this thesis to model the strain accumulation across the Manyi Fault prior to the 1997 event.

1.2.2 Coseismic Deformation

Eventually the build up of strain below the locked fault plane causes the stress on the fault plane to be too great to be restrained by friction and discrete slip across the fault plane occurs. As this slip accelerates the ground rapidly, it emits seismic radiation that can be detected on seismometers all over the world. It also permanently displaces the ground around the fault, which can be measured using GPS and InSAR. The source of the earthquake can be studied using each of these techniques, all of which have their own merits. Seismology measures the radiation emitted by the earthquake that can be detected at remote locations spread around the world. This provides excellent temporal coverage as the data can be recorded continuously after the earthquake. However problems with seismology arise from finding the precise locations of the event as its position must be triangulated from positions far from the event. InSAR provides images of the deformation at the surface at high spatial resolution. This allows for more precise locations of the rupture to be made. InSAR is limited however as it has poorer temporal resolution which is limited by the satellite trajectory around the earth, which can mean a delay of several days to weeks until appropriate data can be collected for any given event. GPS can provide very high precision continuous measurements of the

deformation of fixed points around an event. This gives very precise measurements over a continuous time period, but GPS is limited in spatial density of points. Although GPS receivers could be used as seismometers, current technological limitations mean this is not possible in all but the most specific of locations (*Wright et al.*, 2012). Efforts are ongoing to try and merge the techniques but further discussion is beyond the scope of this thesis.

Studies of the coseismic phase of the earthquake cycle generally attempt to find the source parameters of the earthquake, i.e. the strike, dip, rake and spatial extents of the fault planes that ruptured during the event. These parameters can be used for many things such as stress modelling to assess the seismic hazard of future earthquakes near the fault. The distribution of slip with depth can be used to gauge the thickness of the seismogenic layer.

1.2.3 Combining Interseismic and Coseismic Studies

During the interseismic phase, slip occurs at a steady rate across a fault (or region of finite width) below a given depth, with the top portion remaining locked. During an earthquake, the fault slips, releasing the accumulated strain in the previously locked section. In a perfect world the sum of the interseismic strain accumulated and the coseismic strain release should be the same, resulting in no net strain over one cycle, and leading to a finite offset of the blocks either side of the fault (Figure 1.3c line A-A'). Such a cyclical behaviour suggests that it may be possible to model the timing of earthquake ruptures with a series of models (*Shimazaki and Nakata*, 1980) (Figure 1.5). In all models the interseismic slip is assumed to accumulate linearly over time and/or the initial and final stress (τ_1, τ_2) control the timing and magnitude of the coseismic slip.

The first model is the periodic model: we assume the earthquake stress drop and the maximum sustainable stress (and by definition the initial stress) are the same for every event. Such a model produces periodic earthquakes the time and size of which can be predicted using the interseismic strain rate and the time and size of the last event. In

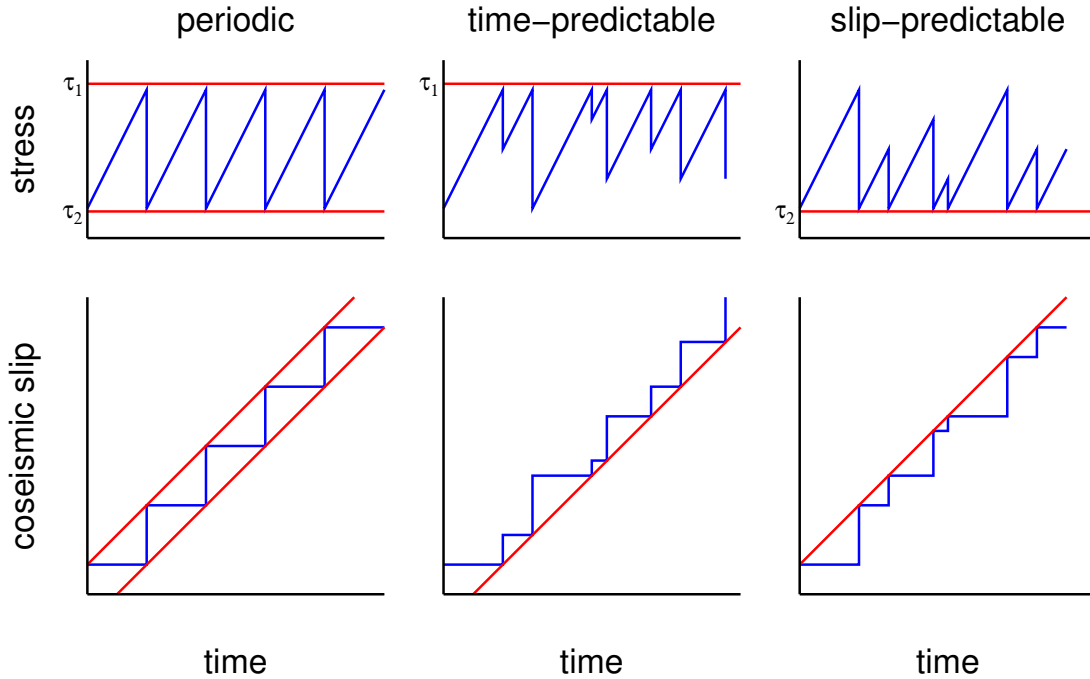


Figure 1.5: Three models of earthquake recurrence known as periodic, time predictable and slip predictable. All these models assume stress is accumulated across faults at a constant rate and that the stress change is proportional to the slip. The other assumptions that are made for each model are described in the text. Modified from *Shimazaki and Nakata (1980)*.

the second model, the time-predictable model, the maximum stress is assumed to be constant but the stress drop is not. In this model the timing and size of events can vary between events. If the stress drop of the last event is known the time until the next event can be predicted. In the final, slip-predictable model the initial stress is assumed constant. This means any event will reduce the stress down to the same level after each event. In such a model the size of the next event can be predicted but not when it will happen. In practice none of these models are applicable to real faults as neither the initial or final stress is the same for any given fault. We can therefore say that earthquake predictability is well beyond our current understanding. One (but by no means the only) reason the models described above do not work is that there is another stress dissipating mechanism in the earthquake cycle that they do not account for: the postseismic deformation.

1.2.4 Postseismic Deformation

In the immediate aftermath of an earthquake the relative displacements across the fault continue to increase. This motion tends to be quite rapid at first and can reach displacement rates of cm/yr. Over time the motion decays in amplitude, eventually returning to the interseismic rate. This motion is not due to aftershocks as the motion is observable around events that have few aftershocks and also is continuous and not episodic. There are three major mechanisms that cause this postseismic deformation: poroelastic rebound, afterslip and viscoelastic relaxation. This section discusses briefly each mechanism briefly and describes how each is modelled.

Poroelastic Rebound

Fluids, most commonly water, occur frequently in the Earth's crust. These fluids can influence the response of the surface after an earthquake as the crust moves more like a fluid-saturated poroelastic medium (*Roeloffs, 1988*) than an elastic one. After an earthquake, fluids in the crust initially resist the coseismic stress changes. However over time, as the fluids flow from regions of high to low pressure, the stress decreases. The surface deformation due to poroelastic rebound can be calculated by looking at the difference between surface displacements in undrained (coseismic) and drained (post-seismic) conditions. It is worth noting that there is no way to constrain the temporal evolution of poroelastic rebound. For most earthquakes it is eliminated as a feasible mechanism as the spatial pattern of modelled deformation poorly matches the observed deformation. Poroelastic deformation signals have however been the major contributing postseismic signal in studies of the South Iceland (*Jónsson et al., 2003*) and 1992 Landers (*Pollitz et al., 2000*) earthquakes (Table 1.1).

Afterslip

Afterslip is frequently invoked as the principle postseismic deformation mechanism in postseismic studies (Table 1.1). In this thesis we define afterslip as brittle creep, where creep occurs in the brittle domain as a result of frictional sliding and sub-

critical crack growth occurring on the coseismic fault plane (*Scholz, 1968; Cruden, 1970; Amitrano and Helmstetter, 2006*). The deformation (slip) rate for brittle creep is an exponential function of the driving stress (for the purposes of postseismic studies this is the coseismic stress change) as opposed to a power-law function generally associated with ductile flow. Afterslip is highly dependent upon the friction across a fault plane. As temperature increases with depth the frictional properties also change with depth (*Marone, 1998*). In shallow areas faults exhibit rate-weakening friction where slip is energetically favourable, leading to stick-slip behaviour that can generate earthquakes. Deeper down the frictional regime changes to velocity strengthening where any rupture is quickly suppressed. The transition from velocity weakening to velocity strengthening is generally assumed to occur around 250°C (*Blanpied et al., 1995*). The transition is also affected by other factors such as lithology and fluids. We will discuss more about frictional regimes in Chapter 6 when we model the afterslip deformation of the Manyi Earthquake.

Afterslip can be modelled kinematically or dynamically. The kinematic method assumes that afterslip is described by slip on an extended (both with depth and laterally) fault plane and inverts the observed deformation data to determine a slip distribution. The best-fitting slip distribution is then forward modelled to produce a deformation map that can be compared to the data. The correlation between coseismic and post-seismic slip on the fault plane can also be compared. As afterslip is driven by coseismic stress increases, regions of high afterslip are adjacent to regions of high coseismic slip as these are the locations where the coseismic shear stress increases the most. They generally do not overlap significantly as the high coseismic slip regions experience a stress decrease which does not cause afterslip. The other way afterslip is modelled (*Perfettini and Avouac, 2007*) is to use frictional laws to model the temporal evolution of afterslip (as discussed in Chapter 6). The slip distribution calculated can be converted into deformation maps, which can be compared to observed deformation fields. Afterslip has been observed following a number of events including the 1999 Izmit Earthquake (*Hearn et al., 2009*), 1999 Chi-Chi Earthquake (*Perfettini and Avouac, 2004; Hsu et al.,*

2002; *Savage et al.*, 2005) and 2003 Tokachi-oki earthquake (*Hsu et al.*, 2006).

Viscoelastic Relaxation

The Earth is a viscoelastic body. Glacial rebound and even the oblate spheroid shape of the Earth attest to the fact. Plate tectonics describes the motion of strong continental lithospheric plates that overlay a weak convective asthenosphere. The Earth's lithosphere is however more complex than a conducting skin. To determine the strength profile of the lithosphere we can use many avenues of study, one of which is postseismic viscoelastic deformation. Viscoelastic relaxation is dependent upon the rheology of the lithosphere. Rheology is the study of the flow of all forms of matter (*Reiner*, 1964). Food analogies are often used to describe the rheological and strength profiles of the Earth's continental lithosphere (Figure 1.6). The first analogy is the 'Jelly Sandwich' where a weak lower crust is sandwiched between a strong upper crust and a strong upper mantle (*Burov and Watts*, 2006). In this model much of the long term strength of the lithosphere generally lies in the (lithospheric) upper mantle. The second model is termed the 'Creme Brulee' model. In this model the strength of the lithosphere lies in the crust, and the upper mantle is very weak due to the high temperatures (*Jackson*, 2002). The last model is termed the 'Banana Split' model. In this model there is little strength in the lithosphere along plate boundaries as they have been weakened due to fault processes (*Zoback et al.*, 1987). It is stipulated these areas of weakness could exist at all depths.

Large earthquakes cause a stress change in the crust and upper mantle that can induce flow that is detectable at the surface. The flow, and hence the surface deformation, due to viscoelastic relaxation is modelled using numerical techniques. Numerical codes take a specific user-defined Earth model with specified rheologies and solves equations of static equilibrium in the Earth (as discussed in Chapter 2). Solutions to these equations can be evaluated at the surface to yield the surface deformation field at given times, which can then be compared to the observed deformation data. Small earthquakes do not produce an appreciable stress change below the seismic layer which is

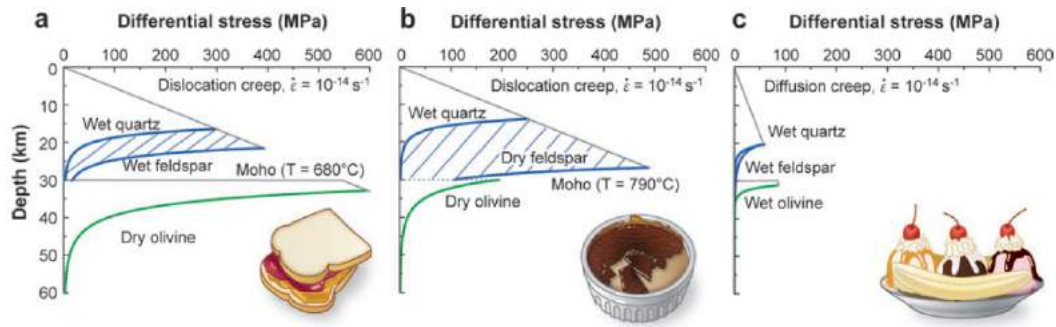


Figure 1.6: Schematic view of models of continental lithospheric strength envelopes. They all show an initial increase in strength with depth (and pressure) following Byerlee’s law, however the two models on the left have a higher coefficient of friction due to a lower pore pressure. Model a) shows a low strength lower crust due to the presence of water. Model b) has a strong lower crust but weak upper mantle. In this model the strength resides in the dry, brittle crust. Model c) considers faults as weak due to strain weakening and feedback processes. Taken from *Bürgmann and Dresen (2008)*.

assumed to be elastic and so do not exhibit viscoelastic postseismic deformation. This restricts the number of earthquakes that can be studied for viscoelastic relaxation. Viscoelastic relaxation has been observed following the $M_w 7.9$ Denali earthquake, amongst others that are shown in Table 1.1.

Previous Postseismic Studies

Table 1.1 details postseismic studies of continental events. The majority of studies took place in western US, with the other events taking place in the Alpine-Himalayan Belt. One thing to note from Table 1.1 is that often different studies of the same earthquake can derive different mechanisms for the postseismic deformation. There is however a trend to observe predominantly viscoelastic relaxation as the period of observation increases.

It is noticeable that there is a large increase in studies in the 21st Century. This is largely due to an increase in available postseismic data. A large portion of these data comes from InSAR, a remote sensing technique that allows us to study millimetre scale deformation over large areas and long periods of time. The next part of this chapter briefly describes what the technique of InSAR is and how to create an interferogram.

Event	Year	Publications	Mechanism	Obs. Period
America				
Hebgen Lake	1959	<i>Reilinger</i> (1986)	VR*	23 years
		<i>Nishimura and Thatcher</i> (2003)	VR	27 years
Loma Prieta	1989	<i>Pollitz et al.</i> (1998)	VR and AS	5 years
Landers	1992	<i>Massonnet et al.</i> (1996)	AS [†]	3 years
		<i>Savage and Svarc</i> (1997)	AS [†]	3.4 years
		<i>Deng et al.</i> (1998)	VR	3.4 years
		<i>Peltzer et al.</i> (1998)	AS [†] and PR	4 years
		<i>Fialko</i> (2004)	AS [†] and PR	7 years
		<i>Perfettini and Avouac</i> (2007)	AS [†]	6 years
		<i>Deng et al.</i> (1999)	AS [†]	2 years
Northridge	1994	<i>Donnellan et al.</i> (2002)	VR*	2.9 years
Hector Mine	1999	<i>Pollitz</i> (2003)	VR	2.5 years
Parkfield	2004	<i>Johanson et al.</i> (2006)	AS [†]	4 months
		<i>Freed</i> (2007)	AS	2 years
		<i>Barbot et al.</i> (2009)	AS	3 years
		<i>Bruhat et al.</i> (2011)	AS and VR	6 years
El Mayor Cucapah	2012	<i>Pollitz et al.</i> (2012)	AS and VR	1.5 years
Eurasia				
Manyi, Tibet	1997	<i>Ryder et al.</i> (2007)	AS and VR	3 years
		<i>Yamasaki and Houseman</i> (2012)	VR*	3 years
Izmit, Turkey	1999	<i>Reilinger et al.</i> (2000)	AS [†]	2.5 months
		<i>Hearn et al.</i> (2002)	AS	2.5 months
		<i>Burgmann et al.</i> (2002a)	AS [†]	3 months
		<i>Hearn et al.</i> (2009)	AS and VR	4 years
		<i>Wang et al.</i> (2009b)	AS and VR	6 years
		<i>Ergintav et al.</i> (2009)	AS and VR	6 years
Kokoxilli, Tibet	2001	<i>Diao et al.</i> (2011)	AS and VR	7 years
		<i>Ryder et al.</i> (2011)	VR	5 years
Gorny-Altai, Siberia	2003	<i>Barbot et al.</i> (2008)	AS	3 years
Nima-Gaze, Tibet	2008	<i>Ryder et al.</i> (2010)	AS	9 months

Table 1.1: Studies of Postseismic Deformation following continental earthquakes. Studies that focus on postseismic signals following earthquakes near subduction zones such as Denali (*Biggs et al.*, 2007) and Chi-Chi (*Hsu et al.*, 2006) have been omitted due to their tectonic setting. In the above table AS signifies afterslip is the predominant mechanism, VR viscoelastic relaxation and PR poroelastic rebound. † and * denote that afterslip and viscoelastic relaxation were the only mechanism tested respectively.

1.3 InSAR

Interferometric synthetic aperture radar, commonly known as InSAR, is a satellite based geodetic technique capable of measuring millimetre-scale surface deformation over large areas (<100 kms) at a fine spatial resolution of 1-10s of metres. The first InSAR capable satellite was the European Space Agency's ERS-1 satellite launched in 1991. The first earthquake to be studied using InSAR was the 1992 Landers Earthquake in California (*Massonnet et al.*, 1993). Since then both the number of SAR capable satellites, and the natural and man-made phenomena studied by InSAR has grown rapidly (*Massonnet and Feigl*, 1995; *Bürgmann et al.*, 2000; *Wright*, 2002). InSAR is now used to study phenomena such as the movement of glaciers, landslides, subsidence caused by mining and oil extraction and millimetre-scale crustal motions from faults and volcanoes. A list of InSAR capable satellites is shown in Table 1.2. All the data used in this thesis comes from either the ERS1/2 satellites or ENVISAT.

Detailed reviews of synthetic aperture radar and interferometry are given by *Bamler and Hartl* (1998); *Massonnet and Feigl* (1995); *Rosen et al.* (2000); *Hanssen* (2001) so we only provide a very brief introduction to what InSAR measures and how it can be used to form deformation maps of an area. We also do not discuss the software used to process interferograms from the raw data and instead refer the reader to *Rosen et al.* (2004). We start with a description of the SAR images made by the satellite and show how they can be used to make an interferogram.

InSAR Satellites

Satellite Platform	Space Agency	Availability	Band	Wavelength cm	Incidence Angle °	Repeat Time days
ERS-1	ESA	1992–2000	C	5.66	23	35
ERS-2	ESA	1995–2011 ¹	C	5.66	23	35
ENVISAT	ESA	2002–2012 ²	C	5.62	15–45	35
TERRASAR-X	DLR	2007–present	X	3.10	44	11
RADARSAT2	CSA	2008–present	C	5.66	20–49	24
ALOS	JAXA	2006–2011	L	23.6	10–51	46
SENTINEL 1/2	ESA	2013 ³ –?	C	5.55	20–45	12 ⁴

Table 1.2: Characteristics of InSAR-capable satellites for the European Space Agency (ESA), Canadian Space Agency (CSA), German Space Agency (DLR) and Japanese Aerospace Exploration Agency (JAXA) used to study earthquakes. 1: A gyroscope failure in 2001 made the majority of the data from ERS-2 unsuitable for InSAR. 2: The lack of fuel on the satellite resulted in an orbital wobble, the wobble was adjusted such that only data at 40°N is suitable for InSAR from 2011 onwards. 3: Currently planned launch date. 4: Upon the launch of Sentinel-2 in 2015 (subject to change) this will reduce to 6 days.

1.4 SAR Images

The SAR antenna emits a beam of microwave radiation from ≈ 800 km above the Earth onto a region of the Earth’s surface where it backscatters. It then records the phase and the magnitude of the backscattered signal. The first product produced during InSAR data processing is a single look complex (SLC) image. This image comprises of two bands: 1) the intensity and 2) the phase of the returned signal. An example of an SLC is shown in Figure 1.7. The geometry of the image is not geographic but rather that of the satellite orbit. Pixel dimensions vary in the azimuth and range of the satellite. Topographic features appear to be one sided due to the off-nadir viewing geometry. Each pixel in an SLC is stored as a complex number, $Z = |z|e^{i\phi}$ where z is the magnitude and ϕ is the phase of the returned signal.

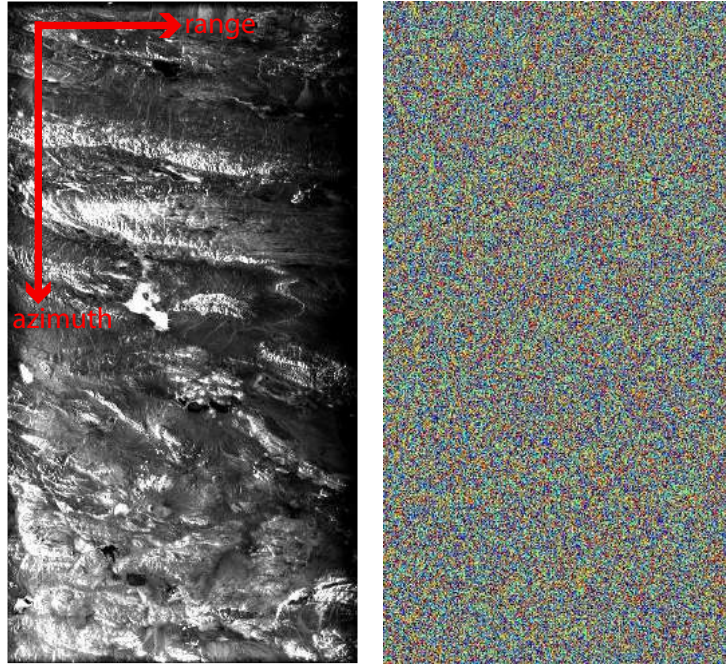


Figure 1.7: The two layers of an SLC image; amplitude (left) and phase (right). The image was taken with the ENVISAT satellite on April 8th 2008 over descending track 305. A vertical stretch of five is due to the difference in pixel size between azimuth (4 m) and range (20 m). The range extent is ~ 100 km. The phase is random, white noise. The black patches near the bottom of the magnitude image are lakes which do not backscatter strongly as most of the radiation is reflected. The satellite look geometry causes the image to be flipped vertically such that east is the left of the image and west right. As track 305 is descending the top of the image corresponds to north.

1.5 Making an Interferogram

Interferograms describe the path change between an early (master) and late (slave) SAR acquisition. An interferogram is formed by differencing the measured phase of co-registered pixels in two SAR images. The amplitude of the return can be used to spatially co-register each SAR image. The difference in phase between two adjacent pixels in each SAR image is deterministic but unpredictable as it is based on the configuration of the scatterers on the ground in the region represented by the pixel. The collections of scatterers introduce an arbitrary phase change upon scattering, ϕ_{pixel} . This makes the spatial distribution of phase across many pixels appear to be random.

An interferogram contains a phase signal that is affected by a variety of phenomena. The observed phase change for each pixel, $\Delta\phi$, can be written as a sum of contributions

from different effects as

$$\Delta\phi = \Delta\phi_{pixel} + \Delta\phi_{orb} + \Delta\phi_{topo} + \Delta\phi_{atmos} + \Delta\phi_{def} \quad (1.2)$$

Each different component in the above equation is defined and how it is removed or accounted for is described below;

1.5.1 Pixel Errors, $\Delta\phi_{pixel}$

The interaction of the microwave radiation and scatterers within a pixel affect the phase change of the radiation upon backscattering. Although we are not interested in the absolute value of the phase change we are concerned whether the value changes between the master and slave acquisitions. If the scattering properties of a region of pixels do change between the two SAR acquisitions the area becomes decorrelated as $\Delta\phi_{pixel} \neq 0$, and it appears as random white noise in the interferogram and the region is said to be decorrelated. One of the major ways decorrelation occurs is from changes in vegetation or agricultural changes such as ploughing fields. As this problem becomes more influential with a longer time span between master and slave images it is referred to as temporal decorrelation. Snow and erosion can also cause decorrelation if the surface, and its scattering properties, change dramatically. This problem is a major limiting factor in the use of InSAR in areas with widespread agriculture or forests e.g Northern Europe and the Indonesian Rainforest. There is little that can be done to extract a signal from incoherent areas of interferograms. We can however minimise the problem by using longer wavelength microwave radiation such as the L-band ALOS satellite, or we can attempt to use short temporal baselines to minimise the chance that the scattering properties of the ground have changed.

Another major source of decorrelation is geometric and caused when the difference in path length between adjacent pixels is greater than the radar wavelength. This is a common cause of incoherence near fault ruptures where there are large deformation gradients. Although this reduces the data near the fault, it is not a major problem as

often the near-field must be masked out anyway when modelling earthquakes to add stability when inverting for the fault geometry. In some cases the areas of decorrelation can be used to map emerging features such as newly emplaced lavas (*Rosen et al.*, 1996; *Zebker et al.*, 1996).

1.5.2 Orbital Position, $\Delta\phi_{orb}$

Repeat pass interferometry requires images that are taken on different days over the same region of the Earth. Tidal and gravitational forces along with solar radiation and atmospheric drag can cause perturbations to satellite orbits (*Ziebart et al.*, 2005). Fuel constraints on-board the satellite restrict the orbital corrections of the perturbations that can be made to a 2 km width tube (for ENVISAT). The baseline separation of two antennae (Figure 1.8) causes a difference in path length r . In an interferogram such a difference manifests itself in the form of a phase ramp. These ramps can be largely removed using our knowledge of the satellite's orbital position. Such positions however are only known to 5-7 cm radially and 10-15 cm across track (*Scharroo and Visser*, 1998). These residual errors can leave remnant, small magnitude phase ramps in the interferograms that can be removed empirically, although due to possible trade-offs with long wavelength deformation (*Wright et al.*, 2004; *Biggs et al.*, 2007), we shall consider these effects further in Chapter 3.

1.5.3 Topography, $\Delta\phi_{topo}$

Surface topography along with an orbital baseline difference causes a perspective change that results in fringes of path difference that correlate with topography (Figure 1.9). The height change that causes a phase change of 2π (one fringe) is given by the height of ambiguity h_a , defined as

$$h_a = \frac{r\lambda\sin\eta}{2B_{\perp}} \approx \frac{10,000}{B_{\perp}} \text{ for ENVISAT} \quad (1.3)$$

where η is the look angle of the satellite (23° for ENVISAT), λ is the wavelength of

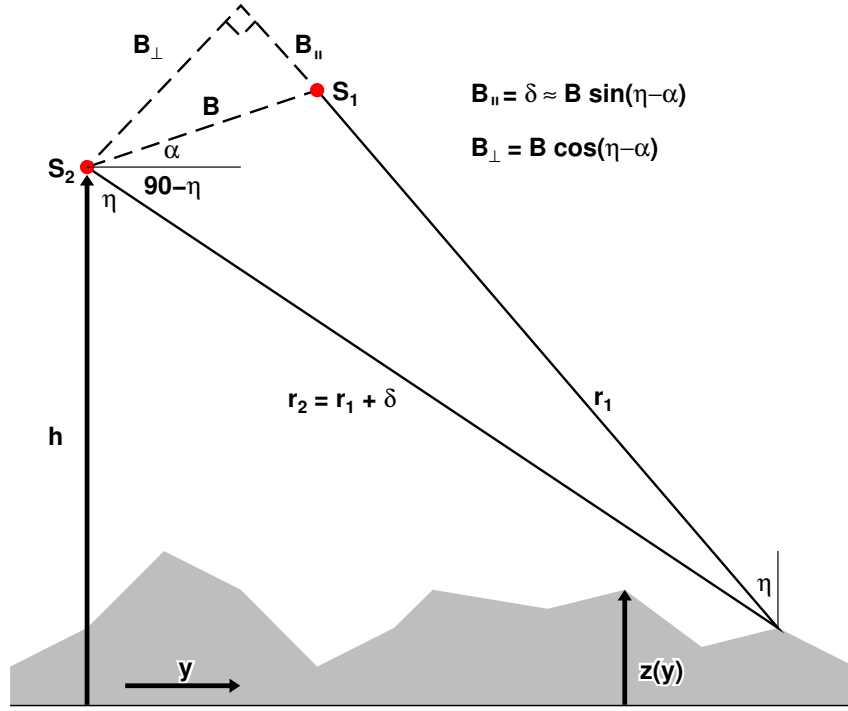


Figure 1.8: The position of the satellite at each acquisition is denoted by $S_1, 2$, with the two positions separated by the baseline distance B . We can resolve B into perpendicular $B_{\perp}$ and parallel $B_{\parallel}$ components. The satellite height and look angle, η and h are labelled, the values of η for each satellite can be found in table 1.2. The parallel baseline has no real influence on an interferogram as it is constant across each pixel and thus can only cause an offset which is removed by redefining 0 in the interferograms. From *Elliott (2009)*

the incident radiation (5.6 cm for ENVISAT) and r is the range between satellite and the target. For an ENVISAT interferogram with a typical 100 m baseline we expect a fringe for every 100 m of topography. To remove these topographic fringes from an interferogram we can use a digital elevation model (DEM) to generate a correction. *Farr and Kobrick (2000)* used SAR instruments of known baseline separation on-board a space shuttle to produce interferograms of topographic fringes which were used to produce one of the most complete and accurate global digital elevation models of the last decade: the SRTM DEM.

1.5.4 Atmosphere, $\Delta\phi_{pixel}$

The radar signal must travel through the atmosphere between the satellite and ground (Figure 1.8). The presence of gases and water vapour in the atmosphere alter its

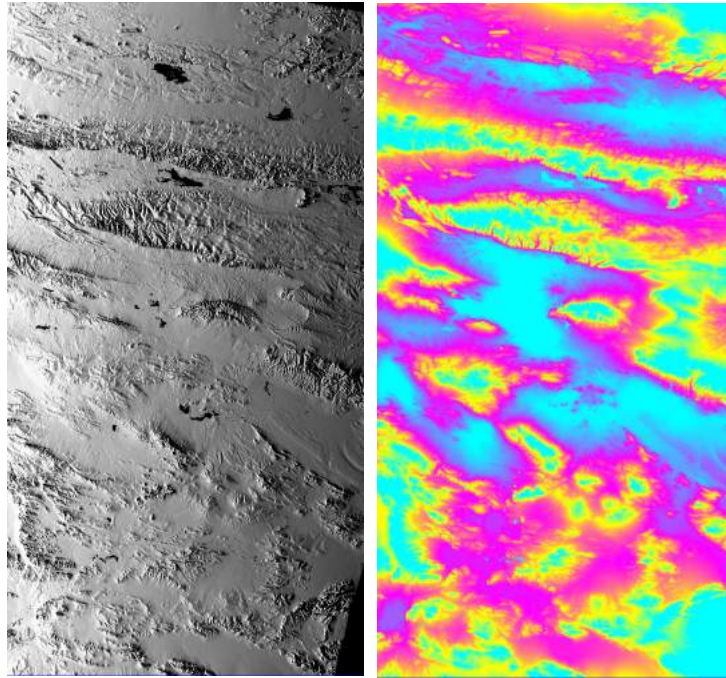


Figure 1.9: The topographic phase correction for ENVISAT interferogram 031007-040713. The interferogram is over track 305, and covers the area of the 1997 Manyi Earthquake. The left image shows the shaded topography from the 90m SRTM DEM. The image on the right shows the topographic phase correction in wrapped fringes. As the baseline of the interferogram is about 40 m each fringe represents 200 m of topographic change.

refractivity and this in turn affects the wavelength of the microwave radiation as it travels through it. The wavelength change causes a phase difference over the whole travel path. The spatio-temporal distribution of water vapour is highly variable and not very well modelled. The largest limitation in the accuracy of InSAR measurements is from water vapour as the influence of atmospheric water vapour can be very significant in given circumstances. For coseismic InSAR studies the LOS deformation signal is often a metre or more, so the effect of water vapour, which is typically of the order of centimetres, is only small. However, when the deformation is only small (a few centimetres as in the case of post and interseismic) it may be of the same magnitude as any atmospheric signal.

There is also a tendency for the spatial distribution of water vapour to correlate with topography. If the topographic change across an interferogram is small it has negligible systematic influence but when there are large topographic gradients it is

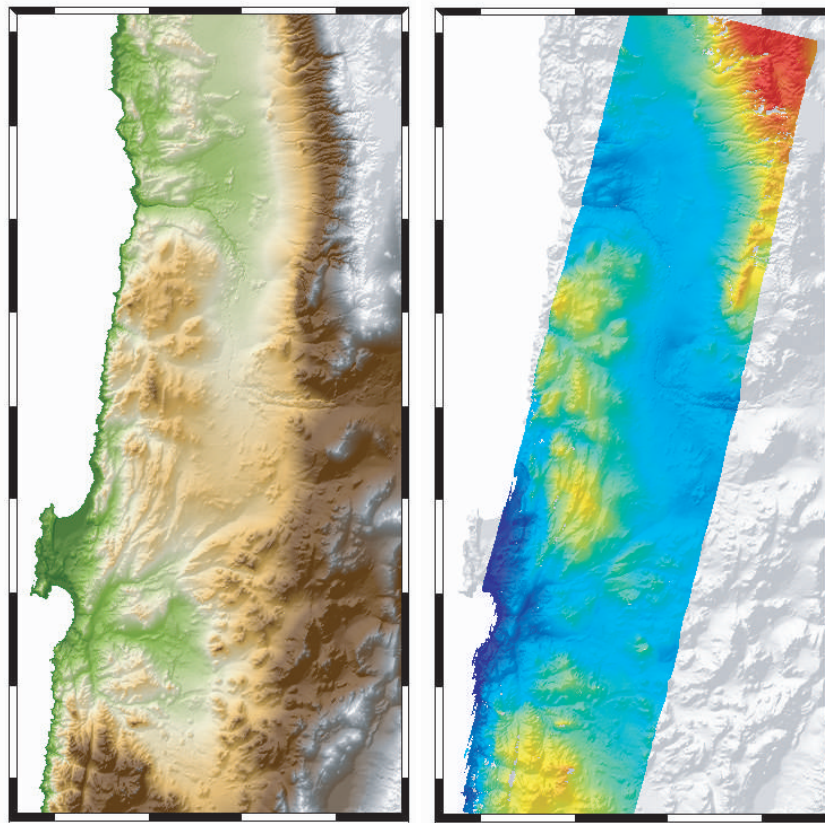


Figure 1.10: The left panel shows the shaded SRTM DEM of the area affected by the 2007 Tocopilla Earthquake. The right side shows an interferogram of LOS range change overlaying the topographic shading for ease of reference. The red patch in the north east corner corresponds to the start of the Andes mountains shown in the DEM (white). The low topography correlates with the dark blue region along the coast in the south of the images.

possible that water vapour can adversely affect tectonic models. Figure 1.10 shows a postseismic interferogram of the area affected by the 2007 Tocopilla earthquake. In the interferogram there is a clear 'signal' that correlates well with the topography. Any postseismic studies would require this atmospheric signal be removed before the postseismic deformation could be measured. *Elliott et al.* (2008) study the interseismic strain accumulation across the Altyn Tagh fault, one of the major faults that bound the Tibetan Plateau, which has over 4km of relief change across it. They show that when atmospheric water vapour is not corrected for, the slip rate across the fault is 35-44 mm/yr; this value is significantly higher than the 10 and 27 mm/yr given by other geodetic studies (*Wallace et al.*, 2004) and Quaternary dating (*Mériaux et al.*, 2004)

respectively. If the data are corrected for water vapour delays using topographically correlated maps of atmospheric water vapour, the estimate reduces to 11 ± 5 mm/yr, a similar value to that obtained (*Mériaux et al.*, 2004).

1.6 Overview of Thesis

This thesis looks at the application of InSAR to the earthquake cycle. The target area for this thesis is the Manyi Fault in Northern Tibet, which is interesting for a number of reasons

- The 1997 event on the fault has been well studied and a detailed analysis of the coseismic rupture already exists in the literature (*Funning et al.*, 2007a)
- The initial postseismic response of the event has been analysed and the interpretation was inconclusive (*Ryder et al.*, 2007)
- There is a significant ENVISAT dataset acquired over the fault region that could help distinguish between the ambiguity between the models used to interpret the initial postseismic study
- There is significant interest in the rheological properties of the Tibetan Plateau from those who aim to model the deformation of the Plateau.
- The excellent coherence generally observed on the Plateau permits the construction of datasets where even millimetre scale deformation may be observable.

Chapter 2 introduces the theory behind deformation modelling. It starts with a description of the basic equations of equilibrium within the Earth, progressing onto a description of elastic dislocation modelling that is used as the basis for modelling postseismic afterslip and coseismic slip. We then discuss how the correspondence principle can be used to turn elastic equations of deformation into viscoelastic ones. We then give a very brief description of how numerical codes are formulated, focusing on

the PSCMP/PSGRN code which can be used to model the viscoelastic deformation of earthquakes.

Chapter 3 starts with a review of Tibetan tectonics and then proceeds to a description of the Manyi Earthquake, detailing previous work that has been written on the event. We then discuss how we can extend the previous postseismic work and include analysis of ENVISAT data acquired 5.5-13 years after the 1997 event. We review network techniques which can be used to improve the signal to noise ratio of the ENVISAT data. We show results from applying such a network approach to the postseismic ENVISAT data to create three distinct ratemaps showing the rate of line-of-sight deformation that shows a similar spatial shape to the postseismic signal observed in the initial postseismic study.

Chapter 4 is a study of the interseismic deformation immediately preceding the 1997 event using data collected between 1992-1997 by the ERS satellites. This chapter applies the principles introduced in Chapter 3 to form ratemaps. We then analyse the data to determine the slip rate across the Manyi fault and a fault to the North which may be an extension of the Kunlun Fault. We also briefly discuss the implications for the tectonics of Tibet and also how the result found is affected by viscoelastic processes in the vicinity of the fault.

Chapter 5 presents results from dynamic afterslip modelling of the Manyi Earthquake. The modelling is based on rate-and-state friction laws using the hot friction approximation to derive the slip on the coseismic fault plane. These measurements of slip are then converted into displacement maps at the surface. We then compare the data (from previous studies and also Chapter 3) to a series of models with different frictional parameters to ascertain the best-fitting model to the data. Lastly we discuss the feasibility of the models and whether they can explain the observed deformation over very long timescales.

Chapter 6 presents the results of the viscoelastic modelling of the Manyi Earthquakes. We create a series of viscoelastic models with various geometries and rheologies and attempt to match the models to the data. We then analyse the limits of resolution

in the InSAR data by comparing the range of best-fitting models.

Chapter 7 compares the best-fitting afterslip and viscoelastic models and the implications for the Manyi Fault. We also discuss how the best-fitting models may be approximating a non-linear rheology. The results of this study are compared with previous studies in the literature. We use the data from this thesis to analyse a 2D model which (some authors show) explains the interseismic and postseismic data used in this thesis simultaneously. We finish the chapter with some discussion of future work that could be performed to determine postseismic deformation.

Chapter 2

Modelling the Deformation of the Earth's Surface

This thesis uses the geodetic technique of InSAR to study displacement of the surface of the Earth as a result of earthquakes, and uses these observations to understand some of the physical processes occurring in the Earth. In this chapter we introduce the concepts of stress, strain, dislocations and static equilibrium and how they can be used to calculate the deformation of increasingly complex elastic bodies. We then discuss how we can incorporate viscoelasticity into deformation model to model viscoelastic behaviour culminating in a discussion about the numerical viscoelastic modelling code PSGRN/PSCMP.

2.1 Deformation and Strain

When the distance between two points on the surface of the Earth changes we can make one of two assumptions. The first is that each point lies on a rigid body. A rigid body is comprised of a series of points which move together without a change in shape of the the body. in this case a change in distance between two points is due to relative (rigid body) motion between the bodies on which each point lies. The alternative assumption is that both points lie on the same body but the body is no longer rigid and and the

shape of the body changes, i.e it deforms. We can describe the difference between the two situations mathematically by considering the displacement $\mathbf{u}$ of a point at $\mathbf{x}$ relative to an arbitrary origin $\mathbf{x}_0$ (Figure 2.1).

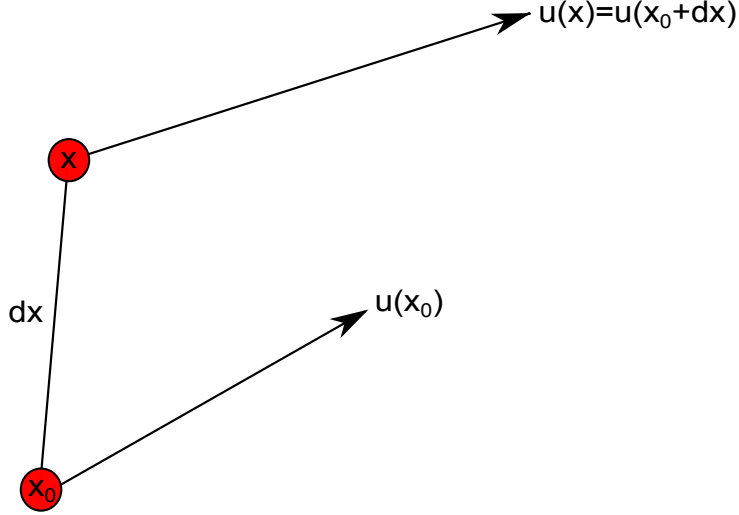


Figure 2.1: Deformation of a line segment of length $d\mathbf{x}$. The figure shows the displacement vectors $\mathbf{u}$ at two points $\mathbf{x}_0$ and $\mathbf{x}$ that are a distance $d\mathbf{x}$ apart.

If we take the first two terms in the Taylor's series expansion of the displacement of $\mathbf{x}$ about $\mathbf{x}_0$,

$$u_i(\mathbf{x}) = u_i(\mathbf{x}_0) + \left. \frac{\partial u_i}{\partial x_j} \right|_{x_0} dx_j + \dots \quad i = 1, 2, 3 \quad (2.1)$$

The first term, $u_i(\mathbf{x}_0)$, represents a rigid body translation as all points in the vicinity of $\mathbf{x}_0$ share the same displacement. The second term, consisting of all the differentials of u_i with respect to x_j the 9-component displacement displacement gradient tensor $\mathbf{L}$ or L_{ij} ,

$$\begin{bmatrix} \frac{\partial u_1}{\partial x_1} & \frac{\partial u_1}{\partial x_2} & \frac{\partial u_1}{\partial x_3} \\ \frac{\partial u_2}{\partial x_1} & \frac{\partial u_2}{\partial x_2} & \frac{\partial u_2}{\partial x_3} \\ \frac{\partial u_3}{\partial x_1} & \frac{\partial u_3}{\partial x_2} & \frac{\partial u_3}{\partial x_3} \end{bmatrix} \quad (2.2)$$

We can split L_{ij} into symmetric and anti-symmetric parts. The symmetric part is known as the (infinitesimal) strain tensor, ϵ_{ij} , and has six independent components,

$$\epsilon_{ij} = \frac{1}{2} \left(\frac{\partial u_i}{\partial u_j} + \frac{\partial u_j}{\partial u_i} \right) \quad (2.3)$$

The anti-symmetric part, which has three independent components, is given by,

$$\omega_{ij} = \frac{1}{2} \left(\frac{\partial u_i}{\partial u_j} - \frac{\partial u_j}{\partial u_i} \right) \quad (2.4)$$

Substituting Equations 2.4 and 2.3 into Equation 2.1 we can find that any displacement $\mathbf{u}$ is due to three components,

$$u_i(\mathbf{x}) = \underbrace{u_i(\mathbf{x}_0)}_{\text{rigid body translation}} + \underbrace{\omega_{ij} dx_j}_{\text{rigid body rotation}} + \underbrace{\epsilon_{ij} dx_j}_{\text{strain}} \quad (2.5)$$

The first two of these three components describe the motion of rigid bodies. The remaining component, strain, describes the change in shape of a body.

2.1.1 Theory of Elasticity

To develop physically realistic models of tectonic processes we must determine how the forces within the earth (stresses) and deformation (strain) relate to one another. The seismogenic layer, which resides in the upper part of the crust, behaves elastically between earthquakes. Perfectly elastic media deform (strain) instantaneously when a stress is applied. The strains will remain as long as the stress is maintained. Once the stress is removed the media will return to its original state. Hooke's Law is used to describe the behaviour of an elastic medium when stress is applied to it, in general the linear relation between stress and strain can be written as Equation 2.6,

$$\sigma_{ij} = C_{ijkl} \epsilon_{kl} \quad (2.6)$$

where σ is the stress, ϵ is the strain and C is the elasticity tensor. The strain is symmetric so we can state C_{ijkl} must be symmetric in k and l , and as the stress tensor is symmetric it must also be symmetric in i and j . Consideration of the strain energy function (*Ranalli, 1995*) allows us to also state that $C_{ijkl} = C_{klij}$. If we limit

our consideration to materials that behave isotropically, i.e those with elastic properties that do not vary according to the direction, Equation 2.6 reduces to,

$$\sigma_{ij} = \lambda \epsilon_{kk} \delta_{ij} + 2\mu \epsilon_{ij} \quad (2.7)$$

where λ and μ are the Lamé parameters and δ_{ij} is the Kronecker delta (*Malvern, 1969*). The first parameter λ has no simple physical interpretation, but it serves to simplify the stiffness matrix in Hooke's law. The second parameter μ is the shear modulus and relates shear stress to shear strain. The two parameters together constitute a parametrisation of the elastic moduli for homogeneous isotropic media, and are thus related to the other elastic moduli such as the bulk modulus $\kappa = (\lambda + 2/3\mu)$, which relates the mean normal stress $\sigma_{kk}/3$ to the volumetric strain ϵ_{kk} .

2.1.2 Deformation in Elastic Media

To describe deformation due to faults in elastic media we start by thinking of a fault as a dislocation. Volterra first described dislocations in 1905 as imposed displacement discontinuities within a crystal matrix. Dislocations come in two major types: screw and edge (Figure 2.2). Edge dislocations are defects caused from the introduction of an extra plane of atoms midway through the crystal. Such dislocations are used to model normal and thrust faults. Screw dislocations are more difficult to describe but can be thought of as a cut partway through a crystal and one side of the cut moving parallel to the base of the cut. We consider strike-slip faults as equivalent to screw dislocations in this chapter.

Betti-Rayleigh Reciprocity Theorem

To describe the deformation field due to dislocations we ultimately want to relate the forces and displacements at dislocations to those at distant observation points. To do so we use the Betti-Rayleigh reciprocal theorem described below following *Sokolnikoff* (1983).

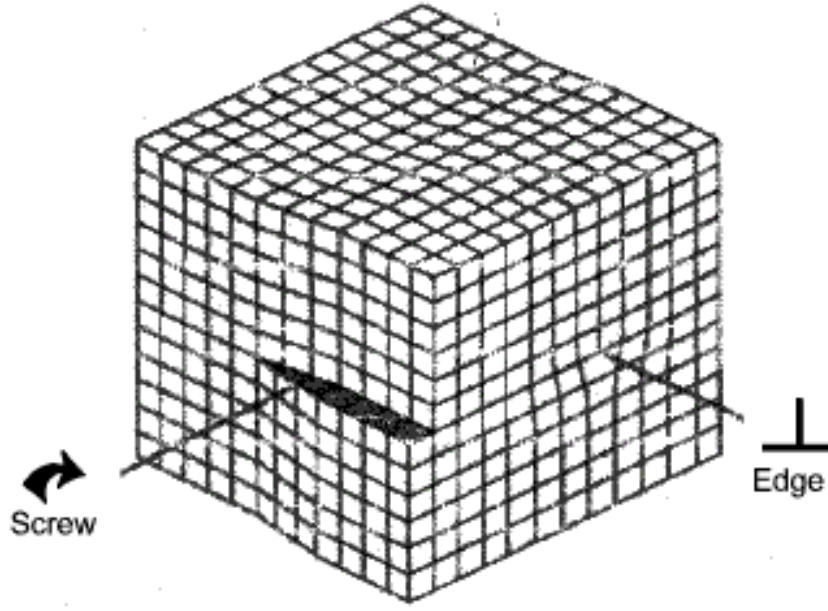


Figure 2.2: The two types of dislocations in crystal lattices; screw and edge. They are used to model the deformation due to slip on a plane (an earthquake). The sense of motion with respect to the dislocation, known as the Burger's Vector, determines which type is most appropriate. From *Ranalli* (1995).

Consider two sets of body forces and surface tractions (f_i^1, T_i^1 and f_i^2, T_i^2) acting through the same elastic medium (Figure 2.3). The work done (W) by the forces and tractions in the first state acting through the displacements in the second is given by Equation 2.8,

$$\int_A T_i^{(1)} u_i^{(2)} dA + \int_V f_i^{(1)} u_i^{(2)} dV = W \quad (2.8)$$

The equation of equilibrium for an elastic medium subject to a point force, f_i is also,

$$f_i + \frac{\partial \sigma_{ij}}{\partial x_j} = \rho \frac{\partial^2 u_i}{\partial t^2} = 0 \quad (2.9)$$

As we are only concerned with static displacements and not seismic wave propagation, the time derivative term is set to zero.

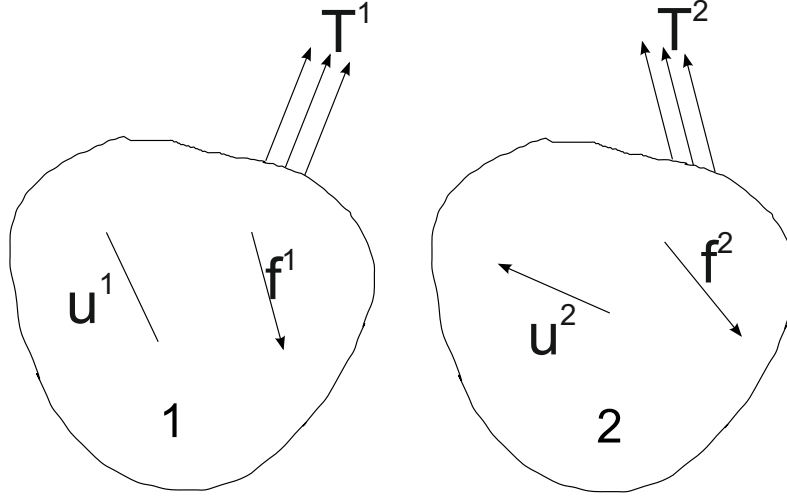


Figure 2.3: Two sets of body forces, tractions and displacements acting on different areas of the same elastic body. The Betti-Rayleigh Reciprocity Theorem states the work done by the forces and tractions in the first state acting through the displacement in the second must equal the work done by the forces and tractions in the second state acting through the displacements in the first.

Substituting Equation 2.9 into Equation 2.8 yields,

$$\int_A T_i^{(1)} u_i^{(2)} dA + \int_V f_i^{(1)} u_i^{(2)} dV = \int_A T_i^{(1)} u_i^{(2)} dA - \int_V \frac{\partial \sigma_{ij}}{\partial x_j} u_i^{(2)} dV \quad (2.10)$$

Using the divergence theorem on the first term on the right hand side and the definition of traction, $T_i = \sigma_{ij} n_j$ yields,

$$\begin{aligned} \int_A T_i^{(1)} u_i^{(2)} dA + \int_V f_i^{(1)} u_i^{(2)} dV &= \int_V \frac{\partial \sigma_{ij}^{(1)} u_i^{(2)}}{\partial x_j} dV - \int_V \frac{\partial \sigma_{ij}^{(1)}}{\partial x_j} u_i^{(2)} dV \\ &= \int_V \sigma_{ij}^{(1)} \frac{\partial u_i^{(2)}}{\partial x_j} dV \end{aligned} \quad (2.11)$$

substituting in Hooke's law (Equation 2.6) we find,

$$\int_A T_i^{(1)} u_i^{(2)} dA + \int_V f_i^{(1)} u_i^{(2)} dV = \int_V C_{ijkl} \epsilon_{kl} \frac{\partial u_i^{(2)}}{\partial x_j} dV \quad (2.12)$$

The symmetry of the elasticity tensor shows that the above equation is symmetric with respect to the first and second state, in other words they can be interchanged.

This means the work done by the forces and tractions in the first state acting through the displacements in the second is the same as the work done by the forces and tractions in the second state acting through the displacements in the first. This is commonly referred to as the reciprocity theorem Equation 2.13,

$$\int_A T_i^{(1)} u_i^{(2)} dA + \int_V f_i^{(1)} u_i^{(2)} dV = \int_A T_i^{(2)} u_i^{(1)} dA + \int_V f_i^{(2)} u_i^{(1)} dV \quad (2.13)$$

Volterra's Formula

We now use the reciprocity theorem to derive Volterra's Formula (*Steketee*, 1958). Let the first state be a dislocation in a body with displacements u_i^d and tractions T_i^d acting on the fault surface Σ . Let the second state be a point source f^p which has displacements u_i^p and tractions T_i^p acting upon Σ , evaluated at a generic point, $\boldsymbol{\xi}$. In this instance the reciprocity theorem allows us to write,

$$\underbrace{\int_V f_i^p u_i^d dV}_1 + \underbrace{\int_\Sigma u_i^d \sigma_{ij}^p n_j dA}_2 = \underbrace{\int_V f_i^d u_i^p dV}_3 + \underbrace{\int_\Sigma u_i^p \sigma_{ij}^d n_j dA}_4 \quad (2.14)$$

There are no body forces associated with the dislocation so term 3 equals 0. Tractions and displacements caused by a point force in an elastic medium are continuous everywhere which means that term 4 is also 0. The displacements caused by the dislocation are also continuous everywhere apart from on the surface of the dislocation itself. If we consider the point force f_i^p to be a Dirac delta function at $\mathbf{x}$ acting in the k direction described as $\delta_{ik}\delta(\mathbf{x} - \boldsymbol{\xi})$, term 1 in Equation 2.14 becomes,

$$\int_V f_i^p u_i^d dV = \int_V u_i^d(\boldsymbol{\xi}) \delta_{ik} \delta(\mathbf{x} - \boldsymbol{\xi}) dV(\boldsymbol{\xi}) = u_k^d(\mathbf{x}) \quad (2.15)$$

The fault displacements u_i^d are discontinuous across the boundary of the dislocation whereas the stress is continuous. The integral involves displacements on both sides of Σ . If we separate the integral into two halves, one on each side of the dislocation with a unit normal pointing in opposite directions and displacements of u_i^{d+} and u_i^{d-} , we can write the integral in term 2 as,

$$\begin{aligned}
\int_{\Sigma} u_i^d \sigma_{ij}^p n_j dA(\boldsymbol{\xi}) &= \int_{\Sigma} u_i^{d-} \sigma_{ij}^p n_j dA(\boldsymbol{\xi}) - \int_{\Sigma} u_i^{d+} \sigma_{ij}^p n_j dA(\boldsymbol{\xi}) \\
&= - \int_{\Sigma} (u_i^{d+} - u_i^{d-}) \sigma_{ij}^p n_j dA(\boldsymbol{\xi}) \\
&= - \int_{\Sigma} s_i \sigma_{ij}^p n_j dA(\boldsymbol{\xi})
\end{aligned} \tag{2.16}$$

where s_i corresponds to the slip vector, the difference in displacement between the two sides of the dislocation (assuming no tensile component of displacement). We can combine Equations 2.14, 2.15, and 2.16 into Volterra's Formula,

$$u_k(\mathbf{x}) = \int_{\Sigma} s_i(\boldsymbol{\xi}) \sigma_{ij}^k(\boldsymbol{\xi}, \mathbf{x}) n_j d\Sigma(\boldsymbol{\xi}) \tag{2.17}$$

where $\sigma_{ij}^k(\boldsymbol{\xi}, \mathbf{x})$ is the stress at $\boldsymbol{\xi}$ due to a point force in the k direction located at $\mathbf{x}$. This stress is often written in terms of the derivatives of a Green's tensor, $g_i^k(\mathbf{x}, \boldsymbol{\xi})$. The components of the Green's tensor describe the displacement at $\mathbf{x}$ in the i^{th} direction due to a unit force acting in the k direction located at $\boldsymbol{\xi}$ (*Okada* (1985, 1992); *Segall* (2010)). The common form of Volterra's formula that is suitable for an isotropic medium is then given by,

$$u_k(\mathbf{x}) = \int_{\Sigma} s_i(\boldsymbol{\xi}) \left[\mu \left(\frac{\partial g_k^i}{\partial \xi_j} + \frac{\partial g_j^i}{\partial \xi_k} \right) + \lambda \delta_{ij} \frac{\partial g_k^m}{\partial \xi_m} \right] n_j d\Sigma(\boldsymbol{\xi}) \tag{2.18}$$

Volterra's formula relates the displacement at an arbitrary observation point caused by a dislocation to stresses over the dislocation caused by a point force at the observation point. To determine the displacement at an observation point we can calculate the Green's function for any given source and geometry of the problem. These can be substituted into Equation 2.18 and integrated across the whole dislocation surface, Σ . Extra terms may be required to satisfy any boundary conditions in the given geometry.

Application of Volterra's Formula

Having derived Volterra's formula we can now use it to derive the displacements due to a screw dislocation, the most appropriate dislocation to model an infinitely-long,

vertical strike-slip fault. We first solve the equilibrium equations in a whole-space, then consider the solutions in a half-space and how to satisfy the necessary boundary conditions at the surface. The geometry of the half-space problem is shown in Figure 2.4. .

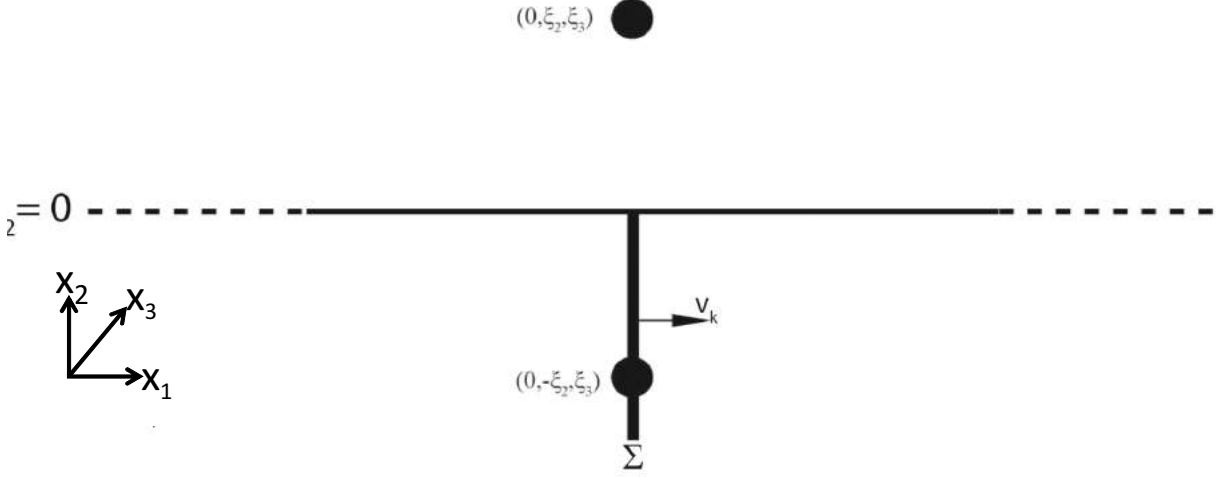


Figure 2.4: The Method of Images. The surface of the Earth is $x_2 = 0$ and the fault plane is represented by the surface Σ . x_3 is out of the plane of the page. The solution for the image source is obtained from the whole-space solution by adding another point source at the position of the image.

Initially we consider a screw dislocation within a uniform elastic whole-space across which slip, s_3 , occurs in the x_3 direction across Σ . To simplify the problem, rather than finding the Green's function for a point source and integrating over fault width and infinite length, we consider displacements and stresses in an arbitrary (x_1 - x_2) plane through a line source at arbitrary depth, and integrate over fault depth alone. For a line source where all the force act in the x_3 direction the only non-zero displacements occur in the x_3 direction which allows us to neglect other components of displacement. The Green's function $g_k(\boldsymbol{\xi}, \mathbf{x})$ is the solution to the equation of static equilibrium (Equation 2.9) at $\mathbf{x}$ due to a force at $\boldsymbol{\xi}$. For this particular problem the body force $\mathbf{f}$ is a line force in the x_3 direction, allowing us to state,

$$\frac{\partial \sigma_{ij}}{\partial x_j} + \delta(\mathbf{r}) = 0 \quad (2.19)$$

where $\mathbf{r}$ is the position vector with respect to the line force. Due to the symmetry of

the problem we have only two non-zero stress components,

$$\sigma_{13} = \mu \frac{\partial g_3}{\partial x_1}, \quad \sigma_{23} = \mu \frac{\partial g_3}{\partial x_2} \quad (2.20)$$

Substituting Equation 2.20 into Equation 2.19 gives

$$\frac{\partial}{\partial x_1} \left(\frac{\partial g_3}{\partial x_1} \right) + \frac{\partial}{\partial x_2} \left(\frac{\partial g_3}{\partial x_2} \right) = \frac{-\delta(\mathbf{r})}{\mu} = \nabla^2 g_3 \quad (2.21)$$

Rewriting Equation 2.21 in cylindrical co-ordinates centred on the line source gives,

$$\frac{1}{r} \frac{\partial}{\partial r} \left(r \frac{\partial g_3}{\partial r} \right) = \frac{-\delta(\mathbf{r})}{\mu} \quad (2.22)$$

The solution to Equation 2.22 is,

$$\begin{aligned} g_3 &= -\frac{\log(r)}{2\pi\mu} \\ &= -\frac{1}{4\pi\mu} \log((\xi_1 - x_1)^2 + (\xi_2 - x_2)^2) \end{aligned} \quad (2.23)$$

Equation 2.23 is the solution for the whole-space problem, however the geometry of Figure 2.4 involves a half-space and thus a boundary condition that the stress σ_{23} is zero at the surface ($x_2 = 0$). In order to satisfy this boundary condition we use the method of images to add another “image” line source, this time located at $(0, \xi_2)$. The Green’s function for this source is of the same form as that for the Equation 2.23, just with ξ_2 replaced with $-\xi_2$. The Green’s function for the half-space problem G_3 is sum of the Green’s functions of each source,

$$G_3 = -\frac{1}{4\pi\mu} \log((\xi_1 - x_1)^2 + (\xi_2 - x_2)^2) - \frac{1}{4\pi\mu} \log((\xi_1 - x_1)^2 + ((-\xi_2 - x_2)^2)) \quad (2.24)$$

The derivative of the above equation with respect to ξ_1 at $\xi_1 = 0$ is,

$$\frac{\partial G_3}{\partial \xi_1} = \frac{1}{2\pi\mu} \left(\frac{x_1}{x_1^2 + (\xi_2 - x_2)^2} + \frac{x_1}{x_1^2 + (-\xi_2 - x_2)^2} \right)_{\xi_1=0} \quad (2.25)$$

We can substitute Equation 2.25 into Volterra's Formula (Equation 2.18) but we can first simplify Volterra's formula by considering the geometry of the problem. From Figure 2.4 we define $s_i = s_3$, therefore $i = 3$. The normal is in the x_1 direction therefore $j = 1$ in n_j . As $i \neq j$ we can neglect the term containing δ_{ij} . There is also no component of force in the x_1 direction so the second term in Equation 2.18 is zero. Volterra's formula is then reduced to,

$$u_3(\mathbf{x}) = \int_{\Sigma} s_3 \mu \frac{\partial G_3}{\partial \xi_1} d\Sigma \quad (2.26)$$

Substituting Equation 2.25 into 2.26, assumeing s is constant with depth and integrating over the depth of faulting, to the surface $\xi_2 = -d \rightarrow 0$ gives

$$\begin{aligned} u_3(\mathbf{x}) &= \int_{-d}^0 \frac{s}{\pi} \left(\frac{x_1}{x_1^2 + (\xi_2 - x_2)^2} + \frac{x_1}{x_1^2 + (-\xi_2 - x_2)^2} \right) d\xi_2 \\ &= \frac{s}{\pi} \left[\tan^{-1} \left(\frac{-d - x_2}{x_1} + \frac{-d + x_2}{x_1} \right) \right] \end{aligned} \quad (2.27)$$

Evaluating 2.27 at $x_2 = 0$ yields the widely used expression for the coseismic displacement of an infinitely long strike-slip fault rupturing from a depth d to the surface (*Savage and Burford, 1973*)

$$u(x) = \frac{s}{\pi} \tan^{-1} \left(\frac{d}{x} \right) \quad (2.28)$$

Altering the limits of integration from $-d$ and 0 to $-\infty$ and $-d$ leads to the equation for interseismic strain accumulation,

$$u(x) = \frac{s}{\pi} \tan^{-1} \left(\frac{x}{d} \right) \quad (2.29)$$

Elastic Layers over an Elastic Half-space

Thus far only homogeneous media have been considered. This is unrealistic as the rheology of the Earth changes with depth. If the Earth is modelled as a series of elastic layers overlaying an elastic half-space we can again use the theory of images to

describe the deformation expected. Adding layers increases the number of boundary conditions as the strain must be continuous across each of them. Increasing the number of boundary conditions that need matching requires an infinite number of images to satisfy them as by adding one image to satisfy a lower boundary condition an additional one is then needed to rebalance the upper one and so on. For an infinitely long strike-slip fault with slip of s down to depth d embedded in an elastic layer of rigidity μ_1 and thickness H ($d < H$), lying over a half space with rigidity μ_2 the displacements due to a slip of s are described by Equation 2.30, (*Rybicki, 1971*).

$$u = \frac{s}{\pi} \left(\tan^{-1} \left(\frac{x}{d} \right) + \sum_{n=1}^{\infty} \left[\frac{\mu_1 - \mu_2}{\mu_1 + \mu_2} \right]^n \cdot \left[\tan^{-1} \left(\frac{x}{2nH + d} \right) - \tan^{-1} \left(\frac{x}{2nH - d} \right) \right] \right) \quad (2.30)$$

This result is equivalent to an infinite sequence of sources in a uniform half-space. The uppermost source is the same as that in the elastic layered case. Below d however there is a series of sources, extending from $(2nH - d)$ to $(2nH + d)$, with a normalised amplitude corresponding to the $[\mu_1 - \mu_2] / [\mu_1 + \mu_2]^n$ term in Equation 2.30. The equation becomes increasingly more complex for increasing number of layers, so much so that for all but the geometry used in Equation 2.30 they are not feasible to state analytically.

2.2 Viscoelasticity

Elastic dislocation theory can be used to describe the displacements we see as an elastic response to the coseismic moment release in an earthquake but it cannot account for the continuing displacements observed immediately after the earthquake with rates that generally decay over time. One of the major time-dependent processes that can cause this kind of motion is viscoelastic relaxation of the ductile lower crust/mantle in response to the stresses applied during the earthquake.

In elastic deformation models, springs are used to represent the elastic behaviour of materials. Springs obey Hooke's Law (Equation 2.6) and will strain instantaneously

when a stress is applied and any deformation is recovered immediately upon removal of the stress. There is no time dependence during the static elastic deformation of springs, as shown by the equations in the previous section of this chapter. In viscous deformation models, dashpots are used to describe the viscous behaviour of materials. The deformation of a viscous medium is time dependent, and when subject to a stress, it will not strain instantly like a spring, but instead the strain will build up over time, on a timescale that is dependent upon the viscosity. Once the stress is removed no strain is recovered so any deformation is permanent. The equations of motion describing the motion of a spring and dashpot are given by Equations 2.31 and 2.32 respectively

$$\sigma = \mu\epsilon \quad - \quad \text{Spring} \quad (2.31)$$

$$\sigma = \eta\dot{\epsilon} \quad - \quad \text{Dashpot} \quad (2.32)$$

We should note a factor of 2 is sometimes present in definitions of the above equations. To describe viscoelastic behaviour we construct mechanical analogues using combinations of dashpots and springs where the components are connected in series and/or parallel. We can describe the time dependent (or independent) motion of each individual component mathematically so we can describe the behaviour of each combination. Some commonly used combinations and their equations of motion are shown in Figure 2.5.

These different combinations describe how different rheological materials behave, both when they are stressed and once stress is removed. The simplest of these models are a dashpot and spring connected in series, known as a Maxwell rheology and a dashpot and spring connected in parallel known as a Kelvin rheology. The behaviour of these bodies is easy to determine by considering the behaviour of the component parts.

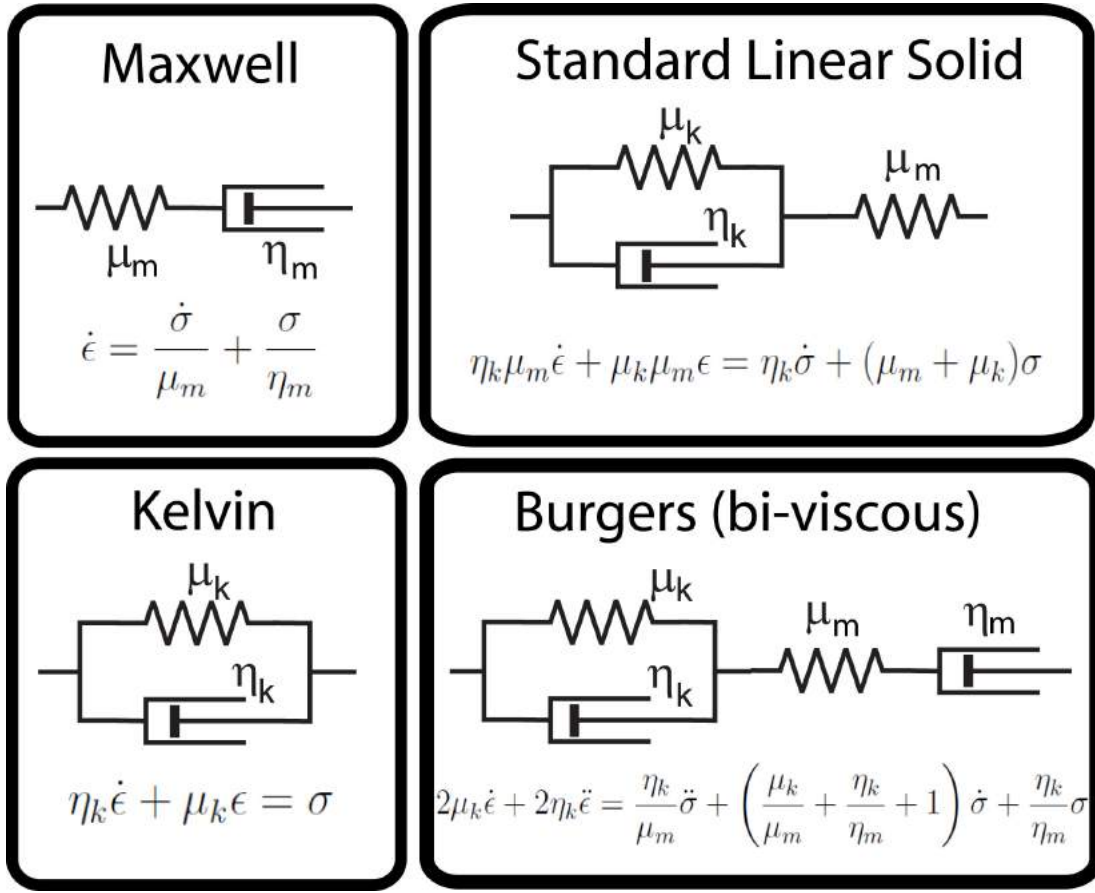


Figure 2.5: Mechanical Analogies to linear viscous rheologies. They contain springs (with shear modulus μ) and dashpots (with viscosity η) to represent the elastic and viscous behaviour respectively. The subscripts m and k refer to the parameters for the elements in the Maxwell and Kelvin elements. It should be noted that a Standard Linear and Maxwell are special cases of a Burgers rheology. The equations in each frame show the constitutive equation for that rheology, some of which are derived in the text.

Maxwell Rheology

When elements are connected in series the stress is the same across both elements but the strain is given by the sum of the strains across each part. Differentiating Equation 2.31 with respect to time and combining it with Equation 2.32 gives the equation for a Maxwell body,

$$\dot{\epsilon} = \frac{\dot{\sigma}}{\mu_M} + \frac{\sigma}{\eta_M} \quad (2.33)$$

where μ_M and η_M are the rigidity and viscosity of the spring and dashpot respectively.

When Equation 2.33 is integrated from $t = 0 \rightarrow t$ with a constant stress, σ_0 applied at $t=0$, assumed we get,

$$\epsilon = \frac{\sigma_0}{\mu_M} + \frac{\sigma_0}{\eta_M} t \quad (2.34)$$

showing that a Maxwell body exhibits instantaneous elastic strain followed by a linear steady state deformation. If we look at the behaviour of a Maxwell body when $\dot{\epsilon} = 0$ we find the solution for stress is,

$$\sigma = \sigma_0 e^{-\frac{\mu_M}{\eta_M} t} \quad (2.35)$$

Equation 2.35 shows Maxwell bodies exhibit an exponential stress relaxation according to the timescale $\tau_M = \eta_M/\mu_M$, which is often called the Maxwell time. It shows that Maxwell bodies have no long-term strength and all the long-term deformation is permanent.

Kelvin Rheology

If we describe the motion of a spring and dashpot connected in parallel (Kelvin) the total stress is the sum of the stress over each component and the strain is the same across both elements. For a Kelvin element we can use Equations 2.31 and 2.32 to write,

$$\sigma = \mu_k \epsilon + \eta_k \dot{\epsilon} \quad (2.36)$$

where μ_k and η_k are the rigidity and viscosity of the spring and dashpot in the Kelvin element respectively. Rearranging Equation 2.36 and multiplying by an integrating factor $e^{\frac{\mu_k}{\eta_k} t}$ we obtain,

$$\dot{\epsilon} e^{\frac{\mu_k}{\eta_k} t} + \frac{\mu_k}{\eta_k} \epsilon e^{\frac{\mu_k}{\eta_k} t} = \frac{\sigma}{\eta_k} e^{\frac{\mu_k}{\eta_k} t} \quad (2.37)$$

which is equivalent to,

$$\frac{d}{dt} \left(\epsilon e^{\frac{\mu_k}{\eta_k} t} \right) = \frac{\sigma}{\eta_k} e^{\frac{\mu_k}{\eta_k} t} \quad (2.38)$$

and yields, upon integration, and applying the boundary condition $\epsilon = 0$ at $t = 0$ for constant stress σ_0 ,

$$\epsilon = \frac{\sigma_0}{\mu_k} \left(1 - e^{-\frac{\mu_k}{\eta_k} t} \right) \quad (2.39)$$

The above equation shows that strain approaches some asymptotic value σ_0/μ_k as $t \rightarrow \infty$. When the load is removed $\sigma = 0$ and integration of Equation 2.36 leads to,

$$\epsilon = \epsilon_0 e^{-\frac{\mu_k}{\eta_k} t} \quad (2.40)$$

A rheology equivalent to a Kelvin element is an unrealistic approximation for the rheology of the Earth. Seismic waves propagate through the Earth so it must, on temporal scales of seconds and minutes behave elastically. A Kelvin rheology is incapable of producing this behaviour.

Burger's Rheology

We can use the results derived above to for describe a more complicated type of linear rheology known as a Burger's body. This combines a Maxwell and Kelvin element in series. It has four material parameters, μ_k, η_k, μ_M and η_M , the rigidities and viscosities of the Kelvin and Maxwell elements. We know that the total strain ϵ across the Burgers body is the sum of strains across the Kelvin ϵ_k and Maxwell ϵ_M elements i.e $\epsilon = \epsilon_k + \epsilon_M$. We can differentiate Equation 2.36 to find,

$$\dot{\sigma} = \mu_k (\dot{\epsilon} - \dot{\epsilon}_M) + \eta_k (\ddot{\epsilon} - \ddot{\epsilon}_M) \quad (2.41)$$

substituting Equation 2.33 into Equation 2.41 and rearranging we find,

$$\eta_k \ddot{\epsilon} + \mu_k \dot{\epsilon} = \frac{\eta_k}{\mu_M} \ddot{\sigma} + \left(\frac{\eta_k}{\eta_M} + \frac{\mu_k}{\mu_M} + 1 \right) \dot{\sigma} + \frac{\mu_k}{\eta_M} \sigma \quad (2.42)$$

A Burgers body has instantaneous elastic behaviour followed by transient creep and then long term viscous behaviour so is a realistic model to approximate the Earth, although any physical reason why it should be so is still unclear. It often matches the observed postseismic deformation, e.g (*Ryder et al.*, 2011), which is unsurprising as the increase in number of dashpots allows an increase in the number of timescales that describe the deformation.

The behaviour of a Kelvin, Maxwell and Burgers model when a constant stress is applied and then removed is summarised in Figure 2.6.

2.2.1 Modelling Viscoelastic Deformation

The relationship between stress and strain in viscoelastic materials is more complicated than the simple stress/strain relationship in elastic materials. To model viscoelastic deformation we need to incorporate the viscoelasticity into the the elastic models described earlier in this chapter. To do this we utilise the correspondence principle described in the next section.

Correspondence Principle

Constitutive relations of simple linear rheologies can be described using a general constitutive relations equation (specific versions of this equation are shown for the rheologies in Figure 2.5).

$$\left(a_0 + a_1 \frac{\partial}{\partial t} + a_2 \frac{\partial^2}{\partial t^2}\right) \sigma_{ij} = \left(b_0 + b_1 \frac{\partial}{\partial t} + b_2 \frac{\partial^2}{\partial t^2}\right) \epsilon_{ij} \quad (2.43)$$

The linear viscoelastic constitutive relation above is a set of linear operators applied to the stress and a series of linear operators applied to the strain. The correspondence principle allows us to construct the viscoelastic equations of deformation if the elastic case is known. The key point of the correspondence principle is recognising the similarity in form between the Laplace transformed version of the constitutive relations for a viscoelastic medium and the constitutive relation of an elastic medium.

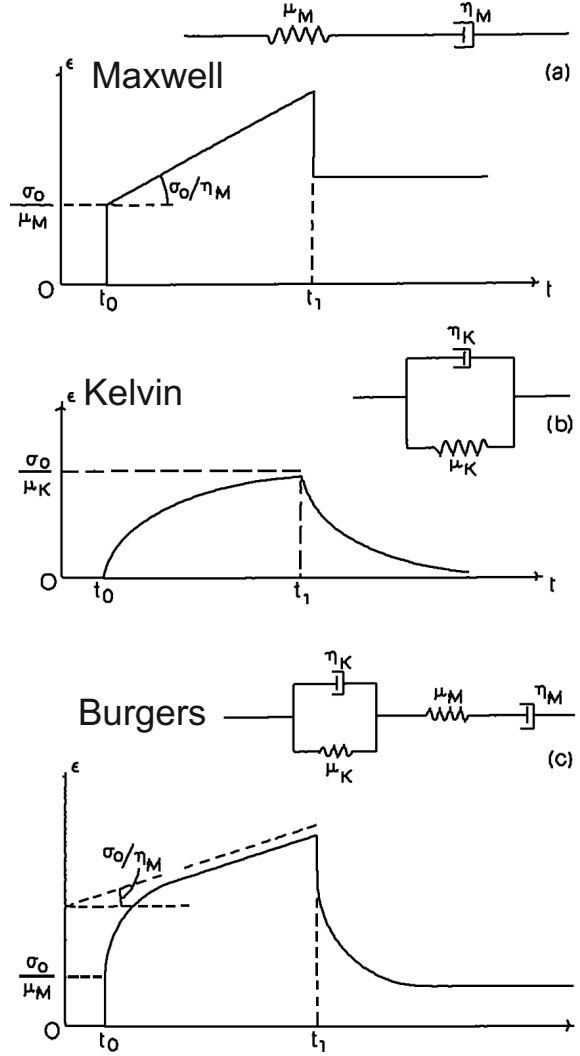


Figure 2.6: The strain response over time of Maxwell (a), Kelvin (b) and Burgers (c) rheologies when a constant stress, σ_0 , is applied from t_0 and t_1 and then removed at t_1 . Adapted from *Ranalli* (1995)

We demonstrate this below by comparing a simple Maxwell viscoelastic constitutive relation (rewritten here from Equation 2.33) to a purely elastic one, (rewritten from Equation 2.31). Note we have dropped the M subscript for clarity.

$$\dot{\epsilon} = \frac{1}{\mu} \dot{\sigma} + \frac{1}{\eta} \sigma \quad \text{and} \quad \epsilon = \frac{1}{\mu} \sigma \quad (2.44)$$

The Laplace transform of a function $f(t)$ is defined by,

$$\bar{f}(s) = \int_0^\infty f(t)e^{-st}dt \quad (2.45)$$

where s is the Laplace transform variable. We also know the relation between the Laplace transform of a function and it's derivative is,

$$\bar{\dot{f}} = s\bar{f} - f(0) \quad (2.46)$$

The Laplace transform of Equation 2.44 then becomes,

$$\begin{aligned} s\bar{\epsilon}(s) - \epsilon(0) &= \frac{1}{\eta}\bar{\sigma}(s) + \frac{s}{\mu}\bar{\sigma}(s) - \frac{1}{\mu}\sigma(0) \\ \bar{\epsilon}(s) &= \left(\frac{\mu + s\eta}{s\mu\eta} \right) \bar{\sigma}(s) \\ &= \frac{1}{\bar{\mu}(s)} \bar{\sigma}(s) \end{aligned} \quad (2.47)$$

where,

$$\bar{\mu}(s) = \frac{s\mu\eta}{\mu + s\eta} \quad (2.48)$$

We have removed $\sigma(0)$ and $\epsilon(0)$ as $\sigma(0) = \mu\epsilon(0)$. We can see that the constitutive law for a viscoelastic material in the Laplace transform domain (Equation 2.47) exactly parallels that from Hooke's law (Equation 2.31). Since time does not enter into the static equilibrium equations it follows that the Laplace transformed viscoelastic problem is equivalent to an elastic problem. This is known as the correspondence principle. It allows us to take a solution to the elastic equation of motion, replace μ with $\bar{\mu}(s)$ and the slip time function (normally a Heavyside function - i.e 0 at $t < 0$, 1 at $t > 0$) with it's Laplace transform, and then transform the equation from the s -domain to the time domain using the reverse Laplace transform to obtain the final time-dependant viscoelastic solutions.

An Analytical Solution for an infinitely long strike-slip fault

Nur and Mavko (1974) then *Savage and Prescott* (1978) use the correspondence princi-

ple to extend the model described by Equation 2.30 of an elastic layer over viscoelastic half-space rather than over an elastic one. They start with the elastic equation of deformation for an infinitely-long, pure strike-slip fault given by Equation 2.30, defining the slip as ΔU and then, using the correspondence principle substitute $\bar{\mu}(s)$ (Equation 2.48) for μ_2 in Equation 2.30 and ΔU by it's Laplace transform $\Delta U/s$.

Taking the inverse Laplace Transform of the resulting viscoelastic equation, the displacements for an infinite strike-slip fault to depth d in an elastic layer of width H overlying a viscoelastic half-space can be written as Equation 2.49 (from *Cohen* (1999)),

$$\Delta U = \frac{1}{2} - \frac{1}{\pi} \tan^{-1} \left(\frac{x}{d} \right) + \sum_{n=1}^{\infty} V(x, t, n) \quad (2.49)$$

where $V(x, t, n) = G(x, n) \times T(t, n)$,

$$G(x, n) = \frac{1}{\pi} \tan^{-1} \left[\frac{2 \frac{x}{d}}{(2n \frac{H}{d})^2 - 1^2 + (\frac{x}{d})^2} \right] \quad (2.50)$$

$$T(t, n) = \frac{1}{(n-1)!} \int_0^{t/2\tau} r^{n-1} e^{-r} dr \quad (2.51)$$

$T(t, n)$ is the incomplete gamma function. The $n = 1$ term is,

$$T(t, 1) = 1 - e^{-t/2\tau} \quad (2.52)$$

and subsequent terms can be derived using a recursion relationship such that,

$$T(t, n+1) = T(t, n) - \frac{r^n e^{-r}}{n!} \quad (2.53)$$

where $r = t/2\tau$ and $\tau = \mu/\eta$

The solutions from Equation 2.49 are shown in Figure 2.7 at various times. These solutions however only apply to an infinitely-long strike-slip fault. To describe real faults the solutions must describe the deformation from faults of finite length and also of varying fault geometries. Unfortunately apart from in very specific cases, such as

the one above, the inverse Laplace transforms for these faults are nearly impossible to derive analytically so must be solved by numerical means.

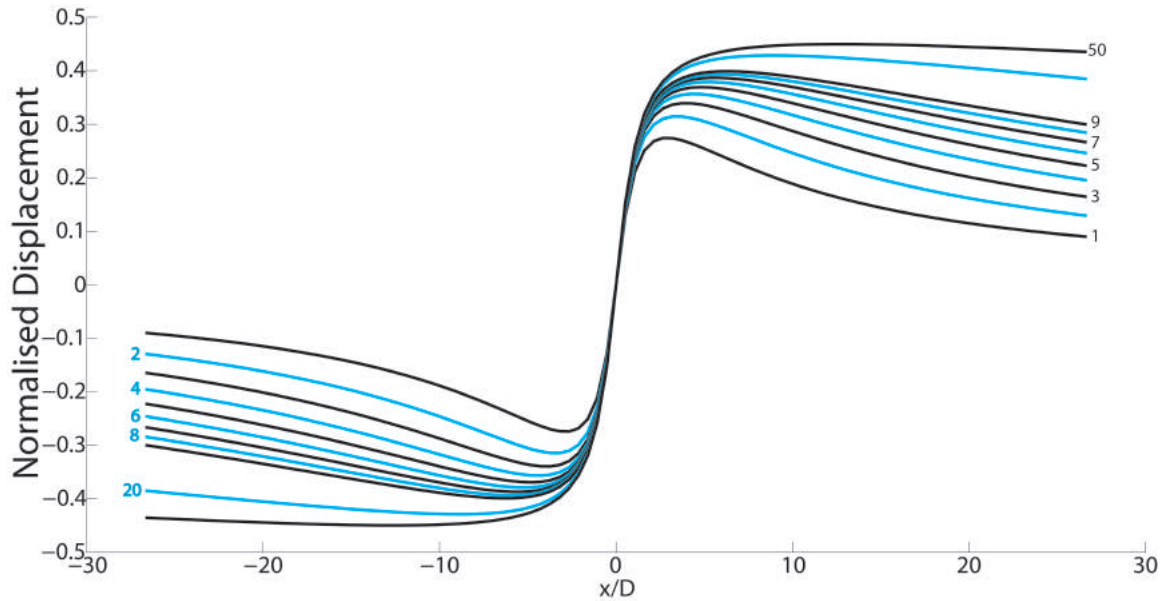


Figure 2.7: The horizontal surface displacements due to postseismic relaxation following an earthquake. The plot is from an analytical solution of an infinitely long pure strike-slip faulting event as given by Equation 2.49. The numbers next to the lines indicate the time (in Maxwell times (Equation 2.35)) of the deformation. The x-axis is normalised according to the depth of the earthquake (which is assumed to rupture the whole elastic layer), and the y-axis is normalised to a metre of slip during the event

2.3 Numerical Modelling of Viscoelastic Deformation

We have seen how it is possible to model the deformation due to dislocations in elastic media and also how it is possible to convert elastic solutions into viscoelastic ones using the correspondence principle. All the examples considered so far have had relatively simple geometry of idealised infinitely long vertical faults. Although we have considered the solutions for finite depth faults in a homogeneous layer over a homogeneous and uniform half-space however in reality the Earth's properties vary with depth. Therefore to describe the observed motions of the surface of the Earth we require a more complex model that accounts for heterogeneity of the Earth's rheological parameters. One

method we can use is the method of images as used to obtain Equation 2.30 and thus Equation 2.49. This becomes very complicated however when attempting to calculate 3-D deformation problems. A more common approach is to use propagator matrices.

2.3.1 Propagator Matrix Methods

First proposed by *Thomson* (1950) and *Haskell* (1953), propagator matrices are widely used in seismology to compute wave-fields in layered media (*Aki and Richards*, 2002). The partial differential equations of equilibrium (Equation 2.9) and Hooke's Law (Equation 2.7) can be used to eliminate stress components that do not exert tractions on horizontal boundaries. The remaining stress components are then reconstructed given the known displacements and stresses. The resulting partial differential equations can be reduced into a series of ordinary differential equations (in the vertical coordinate) by Fourier transforming in both horizontal directions. A simple example, called the anti-plane problem (*Ward*, 1985), is shown below for a vertical fault using the geometry from Figure 2.4.

2.3.2 Propagator Matrix for the Anti-plane problem

For the anti-plane problem we are only concerned with how displacements and stress in the x_3 direction (parallel to the fault) vary on the x_{12} plane. Applying Equations 2.3, 2.7, and 2.9 to the anti-plane problem we can state,

$$\begin{aligned}\frac{\partial \sigma_{23}}{\partial x_2} &= -\frac{\partial \sigma_{13}}{\partial x_1} \\ &= -\mu \frac{\partial^2 u_3}{\partial x_2^2}\end{aligned}\tag{2.54}$$

The equations only apply away from the dislocation as the force term is neglected for simplicity. A more detailed treatment including the force term can be seen in *Segall* (2010). We can also state,

$$\frac{\partial u_3}{\partial x_2} = \frac{1}{\mu} \sigma_{23}\tag{2.55}$$

These two equations can be written in matrix form as,

$$\frac{\partial}{\partial x_2} \begin{bmatrix} u_3 \\ \sigma_{23} \end{bmatrix} = \begin{bmatrix} 0 & \frac{1}{\mu} \\ -\mu \frac{\partial^2}{\partial x_1^2} & 0 \end{bmatrix} \begin{bmatrix} u_3 \\ \sigma_{23} \end{bmatrix} \quad (2.56)$$

Taking the Fourier transform of both sides of the above equation with respect to x_1 (and replacing x_2 with z) gives,

$$\frac{\partial}{\partial z} \begin{bmatrix} u_3 \\ \sigma_{23} \end{bmatrix} = \begin{bmatrix} 0 & \frac{1}{\mu} \\ \mu k^2 & 0 \end{bmatrix} \begin{bmatrix} u_3 \\ \sigma_{23} \end{bmatrix} \quad (2.57)$$

The above equation is of the general form of a propagator matrix equation,

$$\frac{d\mathbf{v}(z)}{dz} = \mathbf{A}\mathbf{v} \quad (2.58)$$

where $\mathbf{v}$ is the length n stress-displacement vector and $\mathbf{A}$ is an $n \times n$ matrix that can be found using the above Fourier or Hankel transform method. A propagator matrix, $\mathbf{P}$ is defined as a matrix which propagates stresses and displacements from a depth at which they are known z_0 , to a depth z . The propagator matrix that corresponds to $\mathbf{A}$, assuming that $\mathbf{A}$ is constant between z and z_0 , is defined as

$$\begin{aligned} \mathbf{P}(z, z_0) &= \mathbf{I} + \int_{z_0}^z \mathbf{A}(\zeta_1) d\zeta_1 + \int_{z_0}^z \mathbf{A}(\zeta_1) \left[\int_{z_0}^{\zeta_1} \mathbf{A}(\zeta_2) d\zeta_2 \right] d\zeta_1 + \dots \\ &= \mathbf{I} + \mathbf{A} \int_{z_0}^z d\zeta_1 + \mathbf{A}\mathbf{A} \int_{z_0}^z \left[\int_{z_0}^{\zeta_1} d\zeta_2 \right] d\zeta_1 + \dots \\ &= e^{\mathbf{A}(z-z_0)} \end{aligned} \quad (2.59)$$

Using the equations of static equilibrium and Hooke's law we can determine the matrix $\mathbf{A}$ and therefore we can find our propagator matrix $\mathbf{P}$. Using Equation 2.57 we can form the propagator matrix for the anti-plane problem:

$$\mathbf{P}(z, z_0) = \begin{bmatrix} 1 + \frac{(z-z_0)^2}{2} k^2 + \frac{(z-z_0)^4}{4!} k^4 \dots & (z-z_0) \frac{1}{\mu} + \frac{(z-z_0)^3}{3!} k^3 \dots \\ \mu(z-z_0) + \mu \frac{(z-z_0)^3}{3!} k^4 \dots & 1 + \frac{(z-z_0)^2}{2} k^2 + \frac{(z-z_0)^4}{4!} k^4 \dots \end{bmatrix} \quad (2.60)$$

$$\mathbf{P}(z, z_0) = \begin{bmatrix} \cosh(z - z_0)k & \frac{1}{\mu k} \sinh(z - z_0)k \\ (\mu k) \sinh(z - z_0)k & \cosh(z - z_0)k \end{bmatrix} \quad (2.61)$$

We now have a propagator matrix to allow us to calculate the displacement-stress vector at any point but first we need to define the starting point z_0 . We choose the z_0 to be the shallowest point at which the problem is uniform (there are no boundaries or sources below it). This means $\mathbf{v}(z)$ must satisfy Equation 2.58 for $z > z_0$. The solutions to Equation 2.58 are given by the eigenvalues and eigenvectors of $\mathbf{A}$ (*Strang*, 1976). The $n \times n$ matrix $\mathbf{A}$ has n eigenvalues λ_i , each corresponding to an eigenvector Ψ_i and a coefficient c_i which give the solution,

$$\mathbf{v}(z) = c_1 \Psi_1 e^{\lambda_1 z} + c_2 \Psi_2 e^{\lambda_2 z} + \dots + c_n \Psi_n e^{\lambda_n z} \quad (2.62)$$

For the anti-plane problem in Equation 2.57, the eigenvalues are $+k, -k$ and the corresponding eigenvectors $[1/\mu, k]^T$ and $[1/\mu, -k]^T$. We can therefore write the homogeneous solution to the anti-plane problem,

$$\mathbf{v}(z) = C_1(k) \begin{bmatrix} \frac{1}{\mu} \\ k \end{bmatrix} e^{kz} + C_2(k) \begin{bmatrix} \frac{1}{\mu} \\ -k \end{bmatrix} e^{-kz} \quad (2.63)$$

As both u_3 and σ_{23} decay to 0 as $z \rightarrow \infty$ then the solution must also decay, hence $C_1 = 0$. We can then write Equation 2.63 as,

$$\mathbf{v}(z) = \begin{bmatrix} u_3 \\ \sigma_{23} \end{bmatrix} = C(k) \begin{bmatrix} \frac{1}{\mu} \\ -|k| \end{bmatrix} e^{-|k|z} \quad (2.64)$$

where C is determined by the boundary condition that tractions vanish at the surface. By incorporating the force terms which we previously neglected (*Segall*, 2010) we can use the solutions above to reproduce the result derived earlier for the deformation due to an infinitely long strike-slip fault (Equation 2.28).

The above example of anti-plane strain are used above for their simplicity and symmetry. In order to have a framework useful for real faults (i.e 3D) we require a

solution for plane strain, that is displacements that occur in the x_1, x_2 plane. This problem is significantly more complex as there are now two independent equilibrium equations (Equations 2.65 and 2.66) that have to be solved, however the techniques for doing so are the same as those described above (*Segall, 2010*).

$$\mu \nabla^2 u_1 + (\lambda + \mu) \frac{\partial u_{k,k}}{\partial x_1} + f_1 = 0 \quad (2.65)$$

$$\mu \nabla^2 u_2 + (\lambda + \mu) \frac{\partial u_{k,k}}{\partial x_2} + f_2 = 0 \quad (2.66)$$

The plane and anti-plane solutions were combined into one propagator solution where the displacement fields are described in spherical vector harmonics by *Ben-Meneham and Singh (1968)*; *Wason and Singh (1972)*. The solutions were transformed into radial and transverse coordinate systems to yield a 6×6 $\mathbf{A}$ matrix. These were then decoupled into a 2×2 and 4×4 matrix, often referred to as the SH and P-SV systems by seismologists as they are essentially the solutions for the anti-plane and plane strain problems we have already described.

Elastic to Viscoelastic

We now have a framework for numerically solving equations in an elastic medium. The correspondence principle is used to form the equations of a viscoelastic problem by substituting the elastic shear modulus by the Laplace transformed equivalent that describes a viscoelastic rheology. The viscoelastic equation of deformation can then be reconstructed using an inverse Laplace transform. Such techniques were used by *Barker (1976)* and *Rundle (1978)* to consider the effects of viscoelastic layers within the Earth.

Up until now we have considered forces caused by dislocations only; we have neglected gravity. For many cases in tectonics, gravitational stresses are negligible in comparison to the elastic stresses. Once the spatial scales get very large however we can not neglect the gravitational terms in the equilibrium equations if the displacements

do significant work against gravity. If we consider viscoelasticity, the influence of gravitational effects becomes even more pronounced as the lack of long-term strength mean the assumption that gravitational forces have negligible effect becomes increasingly less valid with time.

2.3.3 Gravitational Effects

Love (1911) was the first to describe the governing equations for infinitesimal static deformation in an elastic, self-gravitating, hydro-statically prestressed, spherical Earth using the following equations,

$$\frac{\partial \sigma_{ij}}{\partial x_j} + \rho \frac{\partial}{\partial x_i} (\psi + u_k g_k) - g_i \frac{\partial}{\partial x_k} \rho u_k = 0 \quad (2.67)$$

$$\frac{\partial^2 \psi}{\partial_i \partial_i} - 4\pi G \frac{\partial}{\partial x_i} (\rho u_i) = 0 \quad (2.68)$$

where σ is the stress tensor, ρ is the density, $\mathbf{u}$ is the displacement vector, ψ is the gravitational potential, $\mathbf{g}$ is the acceleration due to gravity and G the gravitational constant.

A series of papers by *Rundle* (1980, 1981, 1982) applied the spherical Earth equations to a flat Earth by approximating the gravity as its surface value and treating it as an external body force. He also show that the self gravitating term in Equation 2.67, $\rho \nabla \psi$, is negligible for coseismic and postseismic deformation, hence the problem became known as the reduced problem shown below,

$$\frac{\partial \sigma_{ij}}{\partial x_j} + \rho g \frac{\partial u_z}{\partial x_i} - g_i \left(\rho \frac{\partial u_n}{\partial x_n} + \frac{d\rho}{dz} u_z \right) = 0 \quad (2.69)$$

The above equation can be solved using propagator matrix methods. The results for the anti-plane problem do not change as it does not contain either vertical displacements nor body forces. The plane strain problem is modified to result in an $\mathbf{A}$ matrix of (from *Segall* (2010)),

$$\begin{bmatrix} 0 & -k & \frac{1}{\mu} & 0 \\ \frac{-k\lambda}{\lambda+2\mu} & 0 & 0 & \frac{1}{\lambda+2\mu} \\ \frac{4\mu k^2(\lambda+\mu)}{\lambda+2\mu} & k\rho g & 0 & \frac{k\lambda}{\lambda+2\mu} \\ k\rho g & 0 & -k & 0 \end{bmatrix} \quad (2.70)$$

The four solutions to the above matrix are,

$$c_1^\pm \psi_1^\pm e^{\pm m_1 z} \quad \text{and} \quad c_2^\pm \psi_2^\pm e^{\pm m_2 z} \quad (2.71)$$

where the four eigenvalues $\pm m_{1,2}$ are,

$$\begin{aligned} \pm m_1 &= \pm k \sqrt{1 + \frac{k_g}{k}} \\ \pm m_2 &= \pm k \sqrt{1 - \frac{k_g}{k}} \end{aligned} \quad (2.72)$$

where $k_g = \frac{\rho g}{\sqrt{\mu(\lambda + 2\mu)}}$

To satisfy the boundary condition when $z \rightarrow \infty$ we know that $\psi_{1,2}^+ = 0$. The solution can now be written as,

$$\mathbf{v}(k, z) = c_1^- \psi_1^- e^{-m_1 z} + c_2^- \psi_2^- e^{-m_2 z} \quad (2.73)$$

The above solution is feasible for any $k > k_g$ as there are two constants to satisfy two surface boundary conditions (on for each side of the layer). Problems arise however when $0 \leq k \leq k_g$. In this instance the boundary value problem has no solution as the only one of the solutions in Equation 2.72 is regular and two are required (Wang, 2005). Generally for elastic problems k_g is too small to ever cause instabilities, but in viscoelastic problems the effective shear modulus tends to zero for long times which causes k_g to become very large.

Adam-Williamson Condition

To overcome this numerical problem we can use the Adam-Williamson condition. This requires a density gradient as a result of the initial hydrostatic equilibrium (*Wang*, 2005).

$$\frac{d\rho}{dz} = \frac{\rho^2 g}{\kappa} \quad (2.74)$$

Where κ is defined as the bulk modulus. It is derived using the chain rule shown below,

$$\frac{d\rho}{dz} = \frac{d\rho}{dp} \frac{dp}{dz} = \frac{\rho}{\kappa} \frac{dp}{dr} = \frac{\rho^2 g}{\kappa} \quad (2.75)$$

Using an Earth model with an Adam-Williamson condition Equation 2.69 reduces to,

$$\frac{\partial \sigma_{ij}}{\partial x_j} + \rho g \frac{\partial u_z}{\partial x_i} - g_i \left(\rho \frac{\partial u_n}{\partial x_n} + \frac{\rho^2 g}{\kappa} u_z \right) = 0 \quad (2.76)$$

where the bulk modulus, κ , is assumed constant).

The solutions of this problem are similar to those of Equation 2.71 but the eigenvalues are now,

$$\pm m_1 = \pm \sqrt{k^2 + \frac{2}{3} \left(k_g^2 + k_g \sqrt{k_g^2 - k^2} \right)} \quad (2.77)$$

$$\pm m_2 = \pm \sqrt{k^2 + \frac{2}{3} \left(k_g^2 - k_g \sqrt{k_g^2 - k^2} \right)} \quad (2.78)$$

where,

$$k_g = \frac{\rho g}{\sqrt{\frac{4}{3} \kappa (\lambda + 2\mu)}} \quad (2.79)$$

By introducing an Adam-Williamson condition $\pm m_{1,2}$ remains finite for any value of k meaning even if $\mu = 0$ there are sufficient solutions to match the boundary conditions, thus the numerical instability for long timescale deformation is avoided.

PSGRN/PSCMP - a Flat Earth postseismic modelling program

Some of the range of numerical techniques to determine postseismic motions use the starting point of the equations of static deformation highlighted above, and use the correspondence principle to obtain the viscoelastic solution in the Laplace transformed domain. Two types of numerical models have become popular, one writing the solutions in terms of normal modes using a spherical earth, the other using a flat earth model and with solutions in Fourier spatial domain. Throughout this thesis we use a flat Earth model which is detailed below.

Wang (2003) presented new numerical codes for modelling the coseismic response of the Earth to earthquakes called EDGRN and EDCMP. In 2006 these codes were extended to include models of postseismic deformation, the new code called PSGRN and PSCMP (*Wang et al.*, 2006). These new numerical codes model the Earth as a multi-layered viscoelastic gravitational half-space, with the rheologies defined by the user. The codes use an Adam-Williamson condition to avoid numerical instabilities when solving the eigenvalues of the boundary value problem. The correspondence principle is used to turn the elastic solutions into viscoelastic ones in the Laplace domain. Once the wavenumber and Laplace domain problems have been solved inverse transforms (performed numerically) allow the spatial solutions to be determined. These solutions take the form of time varying Green's function, evaluated at specific radial source distances. Linear superposition of the Green's functions allow the modelling of any combination of finite source dislocations.

The numerical code is split into two programs, PSGRN and PSCMP. PSGRN is the most computationally intensive as it is responsible for preparing the Green's functions required to describe the response of the model. The response to four types of dislocation sources are calculated. These are dip-slip, strike-slip, inflation and compensated linear vertical dipole. Each Green's function containing information about 11 different variables: three displacement, six stress, and two tilt components. Although the time taken to run the PSGRN code is generally far greater than PSCMP, once the Green's functions have been calculated they can be used for any fault model as long as the

earth model is kept the same. PSCMP uses the Green's function output of PSGRN to calculate the desired output (in this thesis the output is the surface displacements). The input to PSCMP is a discrete number of rectangular fault planes with parameters defined by the user. The plane is then split up into patches, with the same spatial resolution as the Green's functions, and a point dislocation is placed at the centre of each patch. The deformation is then calculated at any observation point by integrating the deformation due to each point source on the fault plane.

2.4 Numerical Experiments

2.4.1 Strike Slip Faulting - Comparison of Numerical Codes to Analytical Solution

To test the reliability of the PSGRN/PSCMP code, the displacements from an 'infinitely long' strike-slip fault were modelled and compared to the analytical solution from *Cohen* (1999). In both models a viscoelastic half-space (in this case $\eta=1 \times 10^{19}$ Pa s) was overlain by a rigid 15 km elastic layer (in both cases $\mu=3 \times 10^{10}$ Pa). The numerical code requires a finite length fault whereas the analytical solution only applies for an infinitely long fault. In order to approximate an infinitely long fault in the numerical code a fault length of 4500 km was used, 300 times larger than the thickness of the elastic layer and profiles were taken across the centre of the fault. Figure 2.8 shows the deformation at two time intervals (5 and 100 Maxwell times) using both methods.

2.4.2 Normal Faulting

As opposed to strike-slip faulting, there are no realistic analytical solutions for the displacements along normal faults. The analytical solutions that exist apply to infinitely long faults with depth, a situation that is impossible to replicate in the numerical code PSGRN. In order to test the accuracy of the code when dealing with vertical displacements from normal faults the coseismic displacements were compared to those from a

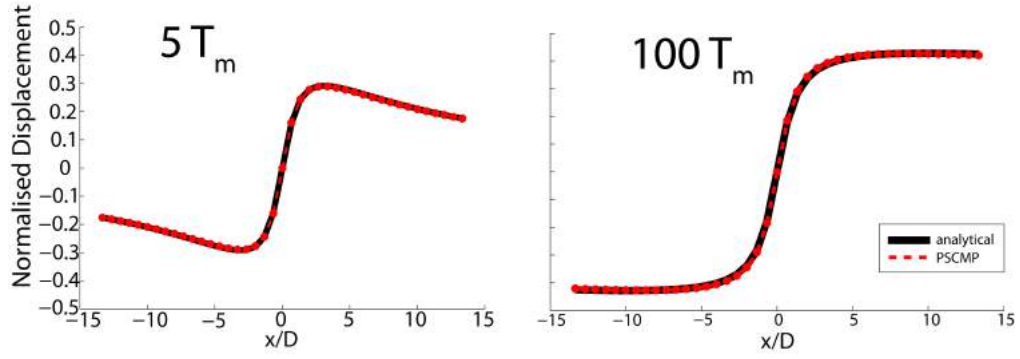


Figure 2.8: The Displacement Pattern given by the Numerical Codes PSGRN and PSCMP and the analytical formulation given by *Cohen* (1999). The red line and dots are the output from the numerical code and the solid black line the output from the Cohen analytical formulation. Both models use the same rheological structure. The horizontal scale is normalised to thickness of the elastic layer (D) and the vertical scale normalised with respect to the coseismic slip

previously established code *oksar* (*Funning et al.*, 2007b). A 250 km long fault was assumed to rupture the whole elastic layer (of depth 15 km) at a dip of 45° .

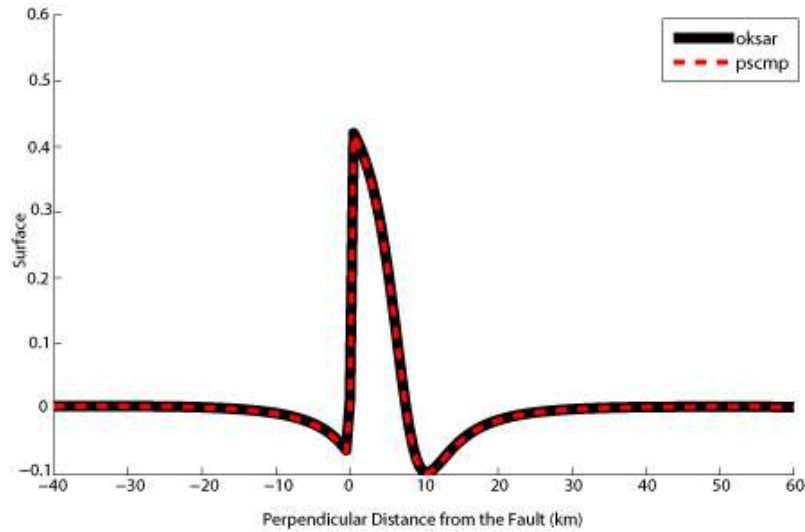


Figure 2.9: The Vertical Displacement Pattern given by the numerical codes PSGRN and PSCMP and the semi-analytical formulation given by *oksar*. The red line is the output from the numerical code and the solid black line the output from *oksar*. The horizontal scale is normalised to thickness of the elastic layer (D) and the vertical scale normalised with respect to the coseismic slip. The fault dips to right and intercepts the surface at 0. Positive displacement is down

Chapter 3

The Manyi Earthquake

3.1 Introduction

The Tibetan Plateau (Figure 3.1) is one of the most highly studied regions of the world for continental tectonics and is the area of interest for this thesis. This chapter provides an overview of the tectonics of the Tibetan Plateau, with particular focus on the strike slip faulting that takes place within it. We then focus on one earthquake in particular, the 8th November 1997 Manyi Earthquake, presenting the results of previous InSAR studies of the deformation field from the earthquake itself and during the three years after it. We then describe techniques that can be used to study small scale (10^{-3} - 10^{-2} m) deformation using InSAR, and apply the techniques to a dataset collected by the ENVISAT satellite (2003-2009). We use the resulting deformation maps around the location of the 1997 Manyi Earthquake, Tibet to determine whether the area around the Manyi Earthquake is still deforming.

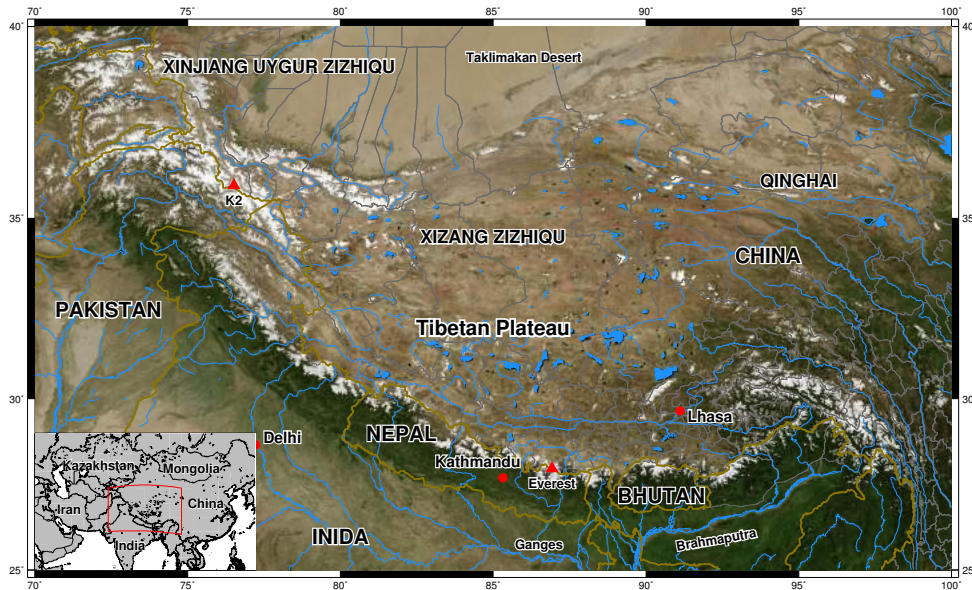


Figure 3.1: Geographic borders and locations of features in the region of Tibet. National boundaries are marked in brown. Major rivers and lakes are marked in blue. Inset map shows region of Asia under study and surrounding countries.

3.2 The Tibetan Plateau

The Tibetan Plateau is a result of the collision between India and Eurasia about 50 million years ago. This event was one of many continental accretions onto the Siberian Shield. As a result, the lithology of the Tibetan lithosphere comprises of a series of different age and compositional units which are, generally, east-west trending. *Molnar and Stock* (2009) reconstruct the relative positions of points in NE and NW India and show a relative convergence rate change of 118 mm/yr to 83 mm/yr, and 109 mm/yr to 59 mm/yr respectively at around 50 million years ago. They also find a reduction in the convergence rate of 40% around 10-20 million years ago, which is attributed to the removal of mantle lithosphere concurrent with the onset of tectonic activity on the plateau. A review of the evolution of the Tibetan Plateau is provided by *Royden et al.* (2008).

Since the first collision about 50 million years ago it is estimated that 2400 km of northward convergence has occurred between India and Asia (*Dewey et al.*, 1988). This has resulted in a doubling of the crustal thickness from 35 to between 60-80 km (*Mooney et al.*, 1998; *Bassin et al.*, 2000). The current elevation of the plateau is about 5 km

and despite covering a 3,000,000 km² area, has remarkably little topographic variation (*Fielding et al.*, 1994).

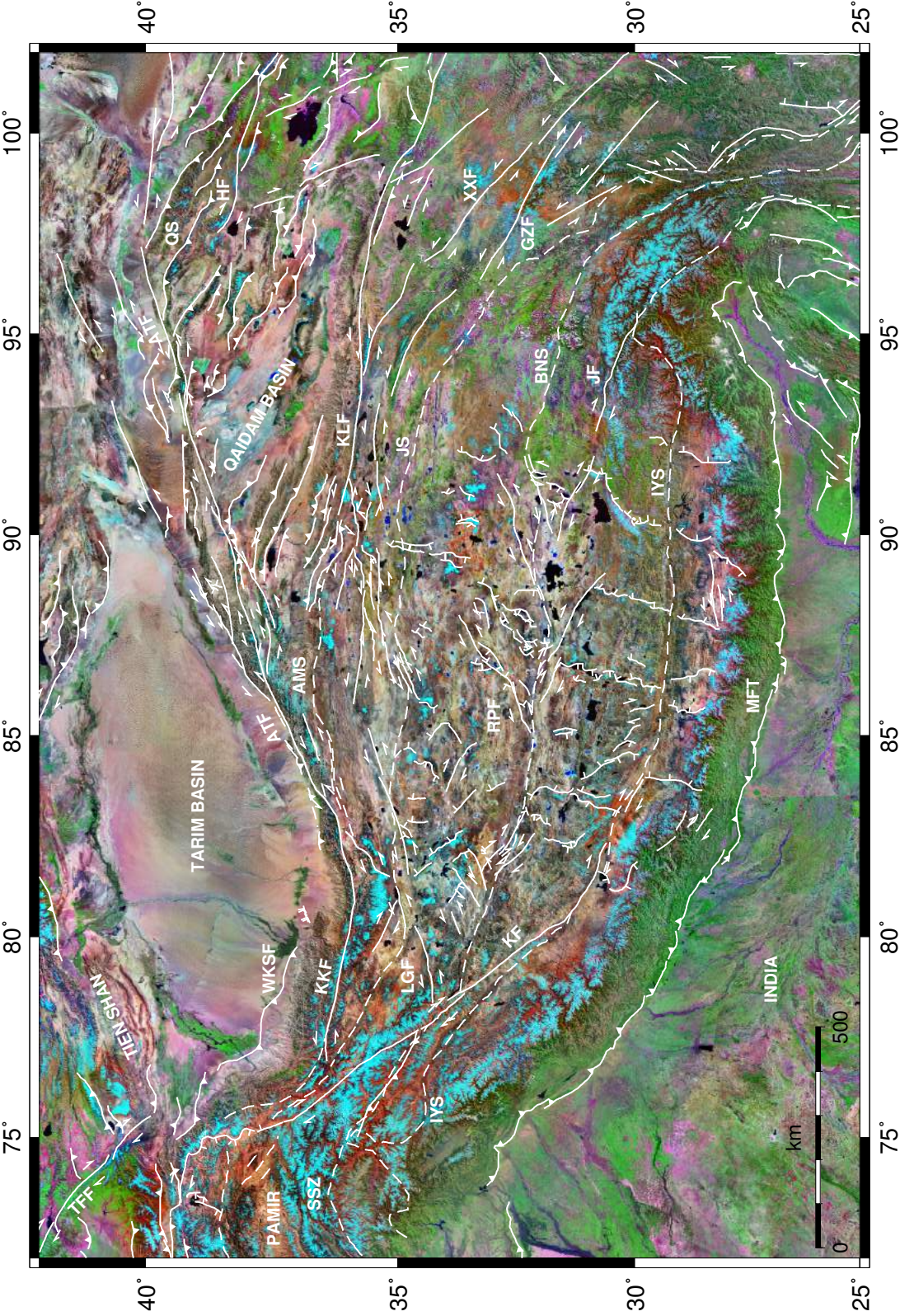
3.2.1 Active Faulting in Tibet

The dynamics of the Tibetan Plateau has been difficult to unravel as it is a highly complex continental collision zone, with numerous, active faults spread across 1000s of kilometres exhibiting any of the three different (normal, thrust and strike-slip) faulting mechanisms (Figure 3.2). There are however some underlying trends as to the distribution of each type of faulting and their individual contributions to the convergence are described below.

Thrust Faulting

Thrust faults are mainly confined to the southern edge and parts of the northern edge of the Plateau (Figure 3.2). A series of low-angle, north-dipping thrust faults are present in the south and have been active through the India-Asia collision (*Yin and Harrison*, 2000), including the Main Central Thrust and Main Boundary Thrust. These faults are considered part of Main Himalayan Thrust decollement. GPS velocities (*Larson et al.*, 1999) show that there is at least 20 ± 3 mm/yr of convergence between India and the southern part of Tibet, 80% of which is accommodated in a 50 km zone around the main thrust faults. This convergence produces up to 2 m metres of potential slip every 120 years along the locked portions of the thrust faults. *Bilham et al.* (2001) study historical records of earthquakes around this locked zone, finding a large portion of this accumulated strain is not accounted for in the record, highlighting the potential risk of a major event in the area that could be worse than the M_w 7.6 2005 Kashmir Earthquake which killed around 80,000 people (*Pathier et al.*, 2006).

The thrust faults along the northern edge of the plateau are mainly split into two groups. The first is the series of NE-SW striking faults that lie next to the Tarim Basin and Tien Shan which currently accommodates 20 mm/yr of convergence (*Abdrakhmatov et al.*, 1996), the whole range having accommodated 200 ± 50 km of shortening and



experienced a number of historical earthquakes (*Avouac et al.*, 1993). The second group lies in the the Qaidam Basin to the NE of the Plateau, where a series of NE-SW trending thrust faults make up the Nan Shan thrust belt at the eastern termination of the Altyn Tagh fault. 150-200 kilometres of crustal shortening has occurred in this region over the last 10 million years (*Meyer et al.*, 1998; *Tapponnier et al.*, 2001). The constraints on timing and interaction of this region with the Altyn Tagh has important implications for the timing of the growth of the Plateau itself.

Normal Faulting

Although normal faulting occurs throughout the Plateau, it is confined to areas above 4.5km in elevation (*Elliott et al.*, 2010) and commonly occur in large north/south striking graben structures (Figures 3.2 and 3.3). The majority of those that do not conform to this pattern lie near the eastern syntaxis and Sichuan Basin and strike E-W. *Copley and McKenzie* (2007) propose that these are the result of extension caused by an elevation change which increases the down slope velocity gradient, assuming the flow is driven by topographic pressure gradients.

The north-south trending graben structures were first mapped by *Tapponnier and*

Figure 3.2 (preceding page): Major faults of the Tibetan Plateau and surrounding region overlain on a mosaic Landsat image. The Main Frontal Thrust (MFT) defines the southern boundary of the Himalaya and accommodates a large proportion of N-S convergence of India and Eurasia. The strike-slip faults are in the north of the plateau and, generally left-lateral and east-west-striking, with the exception of the right-lateral NW-SE Karakoram Fault (KF) to the west of the plateau. The Altyn Tagh Fault (ATF) is the longest of all the strike-slip faults, trending ENE for 1500 km and separating the Tarim Basin to the north from the plateau in the south. A series of smaller conjugate strike-slip faults run in an east-west belt across the central plateau. A series of north-south trending grabens occur in the south of the plateau and become less obvious in the north. Faults are from *Taylor and Yin* (2009) which is compiled from *Armijo et al.* (1986, 1989); *Taylor et al.* (2003); *Taylor and Peltzer* (2006). KEY: ATF - Altyn Tagh Fault, GZF - Ganzi Fault, HF - Haiyuan Fault, JF - Jiali Fault, KF - Karakoram Fault, KKF - Karakax Fault, KLF - Kunlun Fault, LGF - Longmu-Gozha Fault, MFT - Main Frontal Thrust, QS - Qilan Shan, RPF - Riganpei Co Fault, TTF - Talas-Fergana Fault, WKSF - Western Kunlun Shan Fault, XXF - Xianshuihe Fault. AMS - Anyimaqen-Kunlun-Muztagh suture zone, BNS - Bangong Nujiang suture zone, IYS - Indus Yalu suture zone, JS - Jinsha suture zone, SSZ - Shyok suture zone.

Molnar (1977); *Molnar and Tapponnier* (1978) using seismology and remote sensing. Later observation made in the field (*Tapponnier et al.*, 1981; *Armijo et al.*, 1986) showed that these normal faults were a result of E-W extension that post-dates the compressional structures in the area (*Mercier et al.*, 1987). Although early work mainly focused on the most obvious rifts in southern Tibet (*Armijo et al.*, 1986) recent work by *Yin et al.* (1999) suggest that similar features exist in Northern Tibet with slip rates of 2mm/yr, comparable to the slip rates of 1-4mm/yr observed in the south.

The presence of extensional features, in what is predominantly an area of convergence, adds another layer of complexity to models of Tibetan deformation. The explanation of east-west extension usually invokes gravitational forces causing orogenic collapse (*Dewey*, 1988; *England and Houseman*, 1989; *Armijo et al.*, 1986, 1989; *Royden et al.*, 1997; *McCaffrey and Nabelek*, 1998). The details of each different model and how they account for the east-west extension is beyond the scope of this thesis but a good overview is given in *Elliott et al.* (2010)

Strike-Slip Faulting

Strike-slip faults on the Tibetan Plateau and their associated slip-rates are integral to an explanation of Tibetan dynamics (*Molnar and Tapponnier*, 1975; *England and McKenzie*, 1982). Although strike-slip faults exist throughout the plateau, there are six major strike-slip faults on the Tibetan Plateau which have been studied extensively (Figure 3.2, *Searle et al.* (2011)).

- The Altyn Tagh Fault - left lateral fault that bounds the northern edge of the Plateau from the Tarim basin.
- The Karakoram Fault - right lateral fault to the west of the Plateau.
- The Xianshuihe Fault - left lateral fault in Eastern Tibet.
- The Jiali Fault - right lateral fault to the north of the eastern syntaxis.
- The Haiyuan Fault - left lateral fault in North-East Tibet.

- The Kunlun Fault - left lateral fault in Northern Tibet south of the Qaidam Basin.

The role of strike-slip faults in accommodating N-S shortening and E-W extension is critical in the modelling of deformation in Tibet with some models proposing high slip rates across the few major faults (block models) and others requiring small amounts of slip to be distributed over many faults. One of the major differences between these models is the implied slip rates on the major faults. The slip rates have been studied using a variety of techniques such as InSAR eg (*Wright et al.*, 2004; *Taylor and Peltzer*, 2006; *Elliott et al.*, 2008), GPS e.g (*Taylor and Yin*, 2009) and Quaternary dating e.g (*Mériaux et al.*, 2005; *Gold et al.*, 2009). For the Altyn Tagh Fault alone there have been over 20 estimates of the slip rate along it (see *Elliott et al.* (2008) and references therein), ranging from 0 to 45mm/yr, although most recent results tend to cluster around values of 10-20 mm/yr.

Of the major faults in Tibet one is of particular interest in this thesis. The left lateral Kunlun Fault has produced two large recent earthquakes, the most recent being the 2001 M_w 7.8 Kokoxili Earthquake which ruptured 350km of the fault (*Lasserre et al.*, 2005) and the earliest the 1997 M_w 7.6 Manyi Earthquake that occurred 500km to the west. It is this earthquake that is the focus of this and the following three chapters.

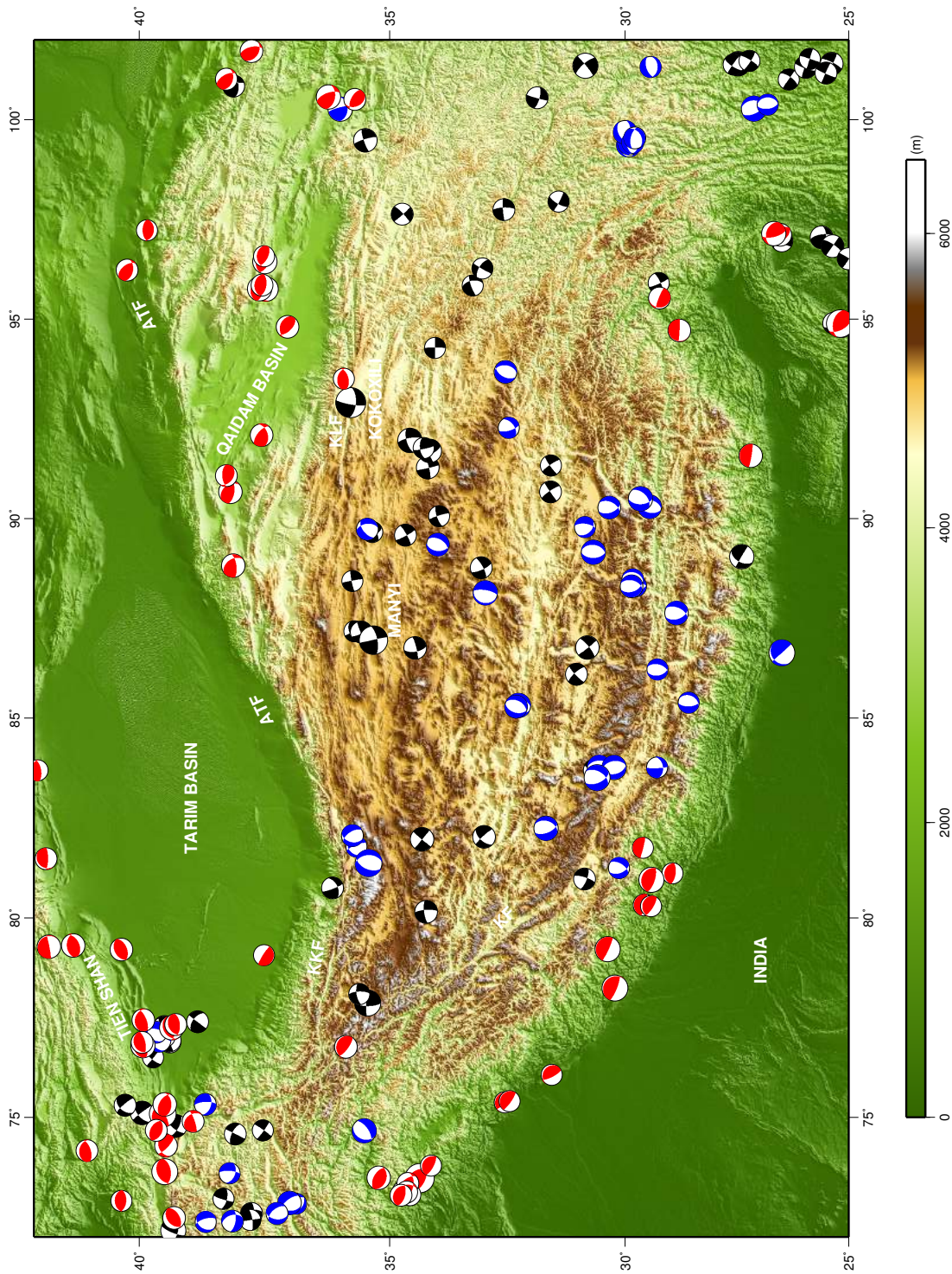


Figure 3.3 (preceding page): M_w 5.5 and greater earthquakes (1976–2011) from the Global CMT catalogue identified as normal (blue), thrust (red) or strike-slip (black) mechanisms of the Tibetan Plateau and surrounding region overlain on a mosaic Landsat image. Labels as those in Figure 3.2. Thrust faulting occurs at low elevations along the edges of the plateau: at the Himalayan front in the south, the NE next to the Qaidam Basin and the NW near the Tien Shan. Strike-slip earthquake mechanisms generally occur on nominally east-west striking faults. The major strike-slip earthquakes of Manyi (1997) and Kokoxili (2001) are highlighted. Normal faulting earthquakes are constrained to the high regions of the plateau (*Elliott et al.*, 2010) and are distributed across the plateau. They are mostly north-south striking and indicate east-west extension. There is little seismicity in the Indian shield and Tarim Basin.

3.3 The 8th November 1997 Earthquake

The M_w 7.6 Manyi Earthquake occurred on 8th November 1997. Due to the remoteness of the area (Figure 3.3) affected there were no reported fatalities despite its relatively high magnitude. This remoteness, however, limits our ability to observe the fault in the field, and we must rely upon remote sensing techniques to study the event. An early InSAR study by *Peltzer et al.* (1999) showed that there was significant asymmetry in the deformation pattern. *Peltzer et al.* (1999) used non-linear elasticity to explain the deformation pattern, however a later study by *Funning et al.* (2007b) showed that such unusual material properties were not required if the fault geometry was more complex.

Funning et al. (2007b) used a series of geodetic techniques to approximate the surface trace of the fault as 11 straight fault segments. A series of elastic dislocation models (*Okada*, 1992) were then used to determine the best-fitting geometry of the fault when compared to the observed interferograms (Figure 3.5). To account for the asymmetry in the deformation pattern a change in dip was introduced along the fault trace. Having determined the geometry of the fault, the fault planes were split into a series of 4x4 km fault patches and displacement data were inverted for slip and rake on each patch (Figure 3.6).

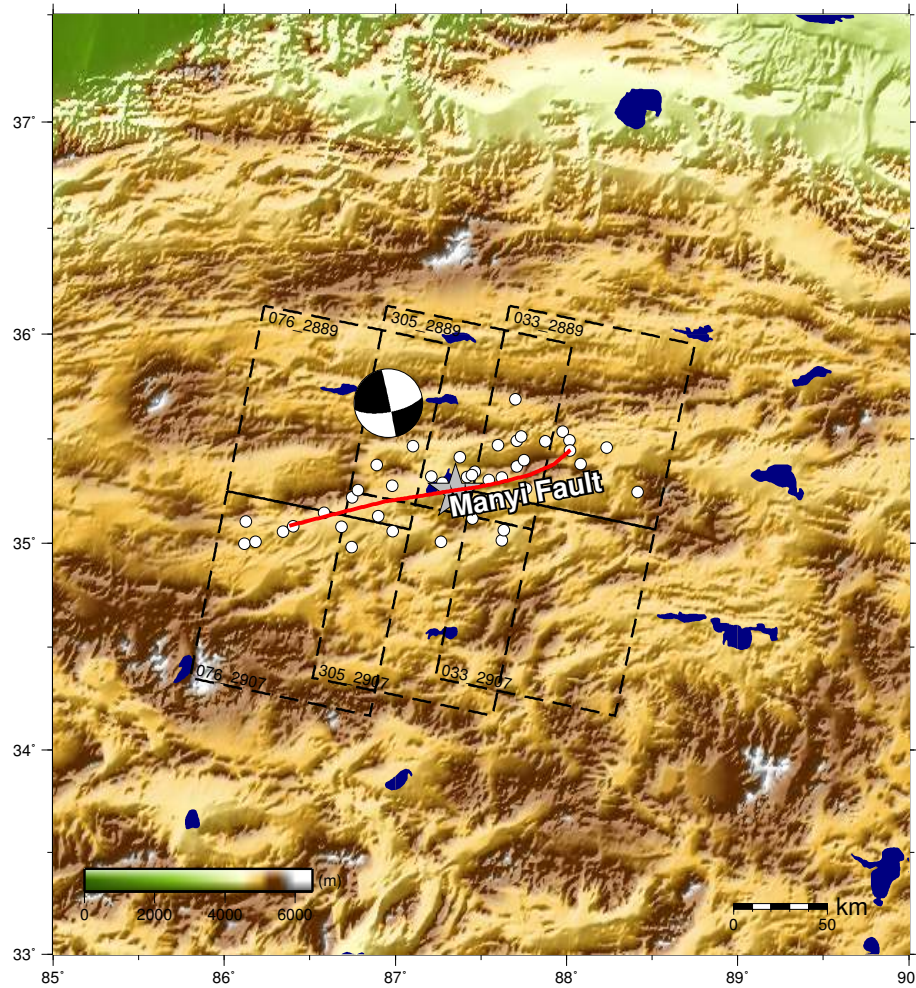


Figure 3.4: SRTM Topography of the Manyi Area (*Farr and Kobrick, 2000*), and seismicity from November 1997 to December 2001 (from *Funning et al., 2007b*). The grey star shows the hypocentral location of the mainshock from *Steck et al. (2001)*. The white circles are the relocated aftershock locations from *Funning et al. (2007b)*. The focal mechanism is taken from *Velasco et al. (2000)*. The black frames show the location of the two frames on three SAR tracks that provided the data used in the interferograms shown in figure 3.5. The red line shows the surface location of the fault that ruptured in the earthquake as determined by *Funning et al. (2007b)* and shown in Figure 3.6.

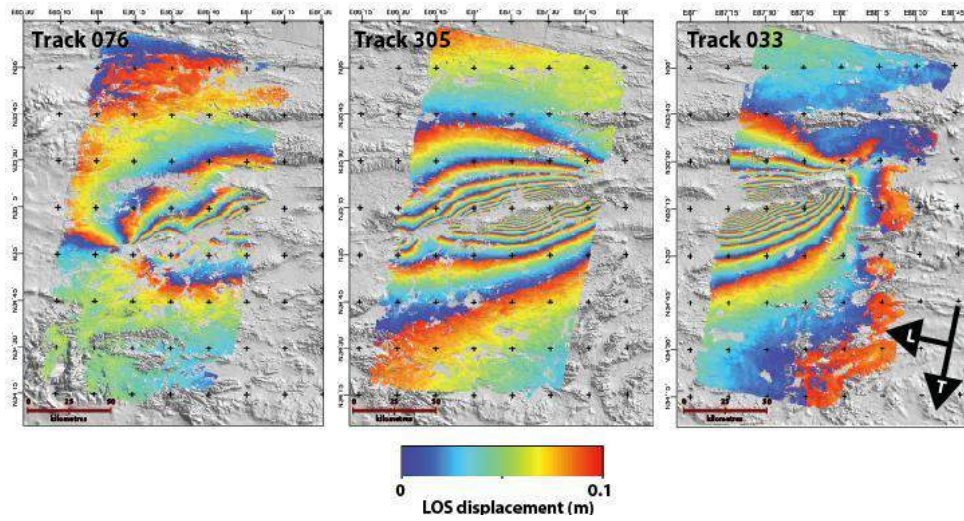


Figure 3.5: The three coseismic interferograms used by *Funning et al.* (2007b) to determine the geometry of the fault plane of 1997 event. Each interferogram shows the line of sight displacement of the ground and has been wrapped to 10cm intervals. Each interferogram is overlain on the DEM of the area. T and L denote the along-track and LoS direction of the satellite respectively. The peak LoS displacement is 2.1m which if purely horizontal displacement occurred on the fault, is equivalent to about 8m of coseismic slip. The pattern of deformation is not the same on the north and south sides of the fault. The gaps in the interferograms are areas of incoherence; there is a large region of incoherence around the fault as the gradient of deformation was too large in this area. Reproduced from *Funning et al.* (2007b).

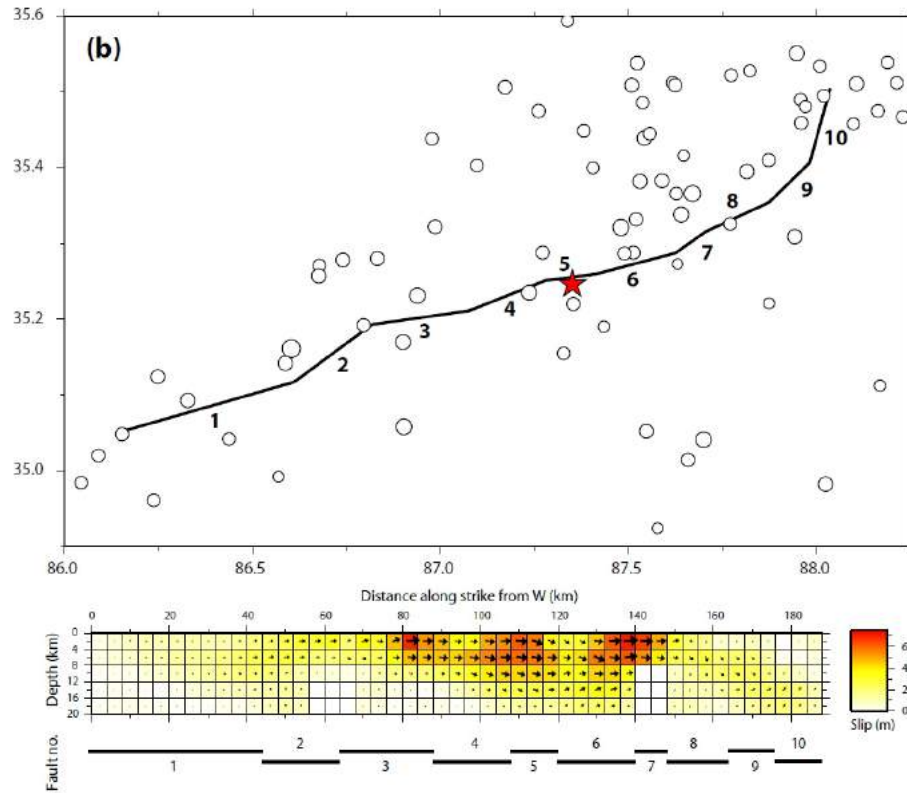


Figure 3.6: The surface location of the Manyi Fault and the slip distribution on the fault (*Funning et al.*, 2007b). The white circles denote the aftershocks and the red star the epicentre (equivalent to the same features in Figure 3.4). The majority of the slip occurs within the top 12 km and highest slip patches are near the location of the hypocentre. The arrows on the fault patches demote the rake of the slip at that point. Reproduced from *Funning et al.* (2007b).

3.4 Previous Postseismic Studies

Ryder et al. (2007) analysed the ERS data collected from 16th November 1997 (9 days after the earthquake) until 19 September 2001 (978 days after the earthquake) over tracks 33, 76 and 305. These interferograms were used to form a smooth time series that investigated the early postseismic motion around the fault. Using this time series, along with the stress changes caused by the fault model from *Funning et al.* (2007b), *Ryder et al.* (2007) determine which of the three major postseismic mechanisms (viscoelastic relaxation, afterslip and poroelastic rebound) describe the motion best (Figure 3.7). The poroelastic rebound model could not match the spatial distribution of deformation so could be disregarded immediately. To model the afterslip, they inverted the time series to determine the best-fitting slip distribution at each time point. The viscoelastic modelling tested a series of rheological models, each with a different layered structure and/or rheology. They conclude that the deformation pattern can be described by either afterslip on the coseismic plane or viscoelastic relaxation of a medium with an elastic 15 km thick layer overlying a Standard Linear Solid (SLS) half-space with viscosity 4×10^{18} Pa (Figure 3.7). Although each mechanism reproduced the observed deformation field well, both mechanisms have issues that need addressing before any definitive conclusion can be made.

Firstly the afterslip is purely kinematic. As the distribution of slip was determined from the data it is unsurprising that it matches the deformation pattern. The slip distribution, and more importantly the temporal evolution of the slip, was not governed by any physical principles, it was the best-fitting, in a least-squares sense, solution to the data at each epoch subject to the conditions of slip on planar extensions of the coseismic rupture plane. To determine whether or not the deformation is aseismic afterslip the temporal evolution of slip must obey the laws that are thought to govern frictional afterslip.

Secondly, there is a complication with the best-fitting viscoelastic solution as an SLS has a long term shear strength. This is physically unrealistic as the region below the elastic lid behaves viscously over very long timescales (*Bürgmann and Dresen, 2008*).

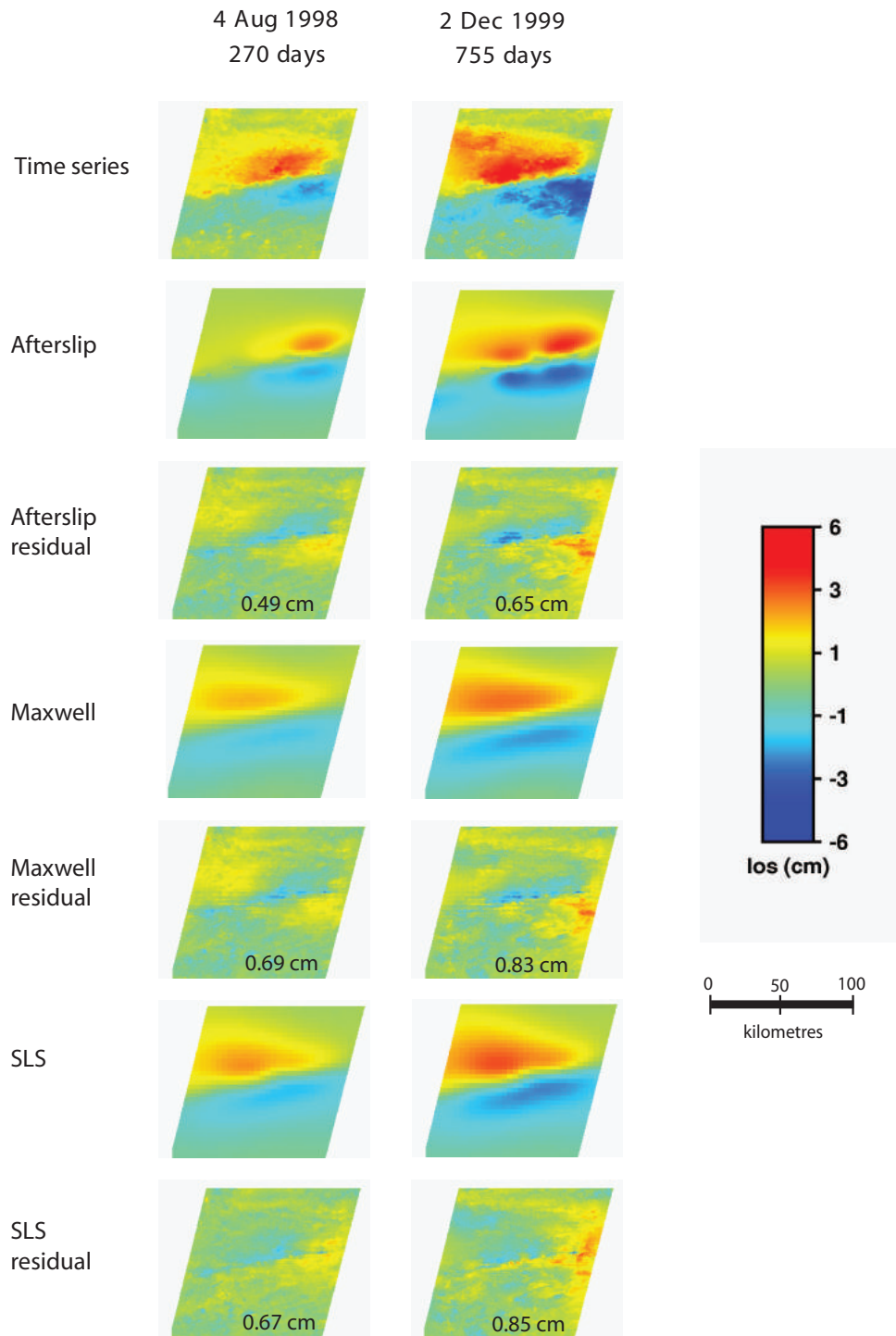


Figure 3.7: Time series observations and modelling images from track 305 from two points in the initial time series (270 and 755 days after the earthquake). Residuals are shown on the same scales as the deformation. The values correspond to the RMS misfit. The viscoelastic synthetics are computed using the best-fitting viscosity for each date rather than the best-fitting overall viscosity. Reproduced from *Ryder et al. (2007)*.

A more realistic rheology would be a Maxwell solid, but the study by *Ryder et al.* (2007) shows this model would not match the data unless the Maxwell time changed over the course of postseismic relaxation. If each epoch was taken separately it was possible to model the deformation using a medium in which the viscosity increases with time. Such a behaviour could be observed by either a more complex Burgers rheology, spatially variable viscosity or even a non-linear rheology. *Ryder et al.* (2011) suggest a Burgers rheology describes the postseismic deformation of the 2001 Kokoxili earthquake which occurred approximately 200 km east of the 1997 Manyi event.

3.5 Continuing the Timeline

Many of the first postseismic deformation studies (*Pollitz et al.*, 2000; *Reilinger et al.*, 2000; *Hearn et al.*, 2002; *Burgmann et al.*, 2002b; *Jónsson et al.*, 2003; *Pollitz*, 2005) focused on the initial postseismic response that lasted a few months to years beyond the event (see Chapter 1). Postseismic studies rarely extended beyond this timescale as the deformation rates become so small that any deformation is difficult to resolve. Recently technological advances as well as a dramatic increase in the data archive of both GPS and InSAR have resulted in several studies of postseismic deformation (*Hearn et al.*, 2009; *Barbot et al.*, 2009) extending the observations to greater than a decade after the event. These studies have shown that, by continuing the period of observation, we can get a lot more reliable and accurate picture about the response of the Earth to an earthquake.

Figure 3.8 shows how the expected deformation rate for the Manyi postseismic signal decays between the ERS era (1997-2001) and the start of the ENVISAT era in 2003 using the best-fitting viscoelastic model of *Ryder et al.* (2007). Both models are calculated using the PSGRN/PSCMP viscoelastic modelling code (see Chapter 2) and the data are from *Ryder et al.* (2007). The profile is taken perpendicular to the fault trace of *Funning et al.* (2007b).

Figure 3.9 shows how the signal goes from being easily observable in a single ERS

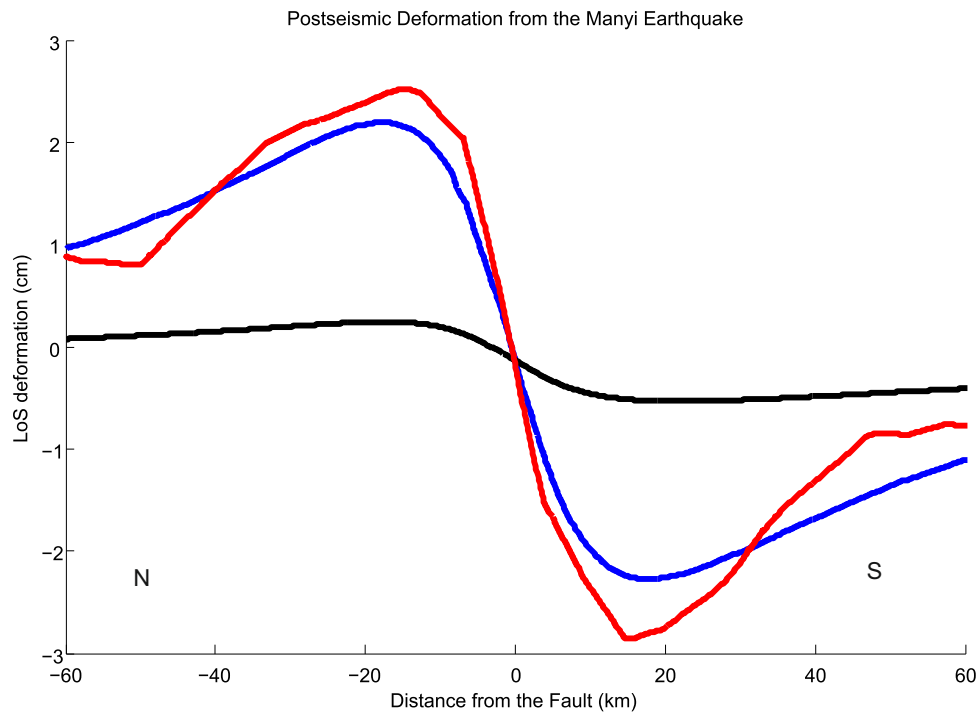


Figure 3.8: LoS Deformation profile of the preferred SLS model with viscosity 4×10^{18} taken across the Manyi Fault. The red lines denotes the data for the first 305 days and the blue line shows the best-fitting model. The black line denotes the deformation that is expected over a 305 day period that starts 6 years after the event. The LoS vector is taken from the geometry of track 305.

interferogram initially after the earthquake to hidden behind the noise in the later ENVISAT interferogram. The magnitude of expected line of sight displacement rate in the ENVISAT era is less than 0.5 cm/yr. This is problematic as such rates are similar in magnitude to the level of noise in any given interferogram. Therefore any study that uses the ENVISAT era data has to increase the signal-to-noise ratio. The next section of this chapter describes the ENVISAT data available in the Manyi Area and various methods to improve the signal-to-noise ratio of the ENVISAT interferograms.

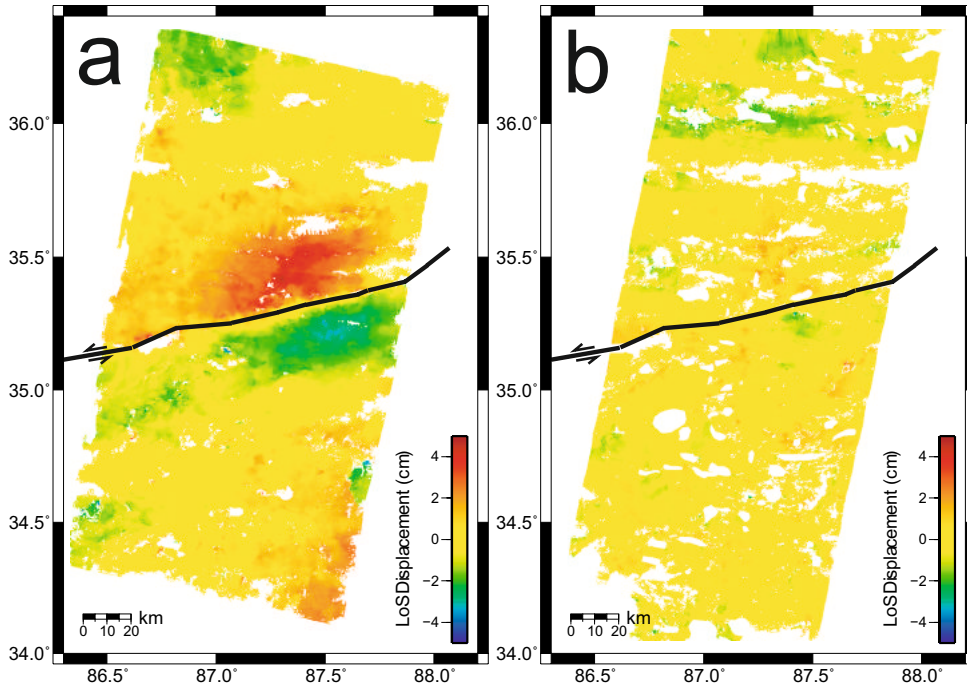


Figure 3.9: a) An interferogram covering the the period 971202-980908 (YYMMDD convention) and b) an interferogram covering the period 031007-040713 (both have a temporal baseline of 305 days). Both interferograms are from track 305 and have had a linear baseline correction applied during the processing. The signal is obvious in the earlier interferogram but in the later one it is only hinted at.

3.6 Extracting the Deformation Signal

We process all the available ENVISAT ASAR data (Table 3.1) over the Manyi area from track 305. Using these data, we construct all the possible interferograms with a baseline of less than 500 m (Figure 3.10). All interferograms are processed using the JPL/Caltech ROI_pac software (*Rosen et al.*, 2004) with initial orbital baseline corrections using Delft precise orbits (*Scharroo and Visser*, 1998). All the interferograms were filtered (and then refiltered as appropriate) using a power spectrum filter (*Goldstein and Werner*, 1998) and unwrapped using a branch-cut algorithm (*Goldstein et al.*, 1988). Each interferogram was duplicated, one with no further orbital baseline correction and the other with a residual orbital ramp removed from the final interferogram by fitting a linear plane to the far field data (with respect to the fault) and removing it from the whole interferogram.

Scene No.	Date yyyy-mm-dd	Date (decimal)	Days
1	2003-04-15	2003.29	1985
2	2003-10-07	2003.77	2160
3	2003-11-11	2003.86	2195
4	2003-12-16	2003.96	2230
5	2004-02-24	2004.15	2300
6	2004-03-30	2004.25	2335
7	2004-06-08	2004.44	2405
8	2004-07-13	2004.53	2440
9	2004-10-26	2004.82	2545
10	2004-11-30	2004.91	2580
11	2005-02-08	2005.1	2650
12	2005-06-28	2005.49	2790
13	2005-08-02	2005.59	2825
14	2005-09-06	2005.68	2860
15	2005-10-11	2005.78	2895
16	2005-12-20	2005.97	2965
17	2006-01-24	2006.06	3000
18	2007-02-13	2007.12	3385
19	2007-03-20	2007.22	3420
20	2007-11-20	2007.89	3665
21	2008-01-29	2008.08	3735
22	2008-03-04	2008.17	3770
23	2008-04-08	2008.27	3805
24	2008-05-13	2008.37	3840
25	2008-06-17	2008.46	3875
26	2008-07-22	2008.56	3910
27	2008-08-26	2008.65	3945
28	2008-09-30	2008.75	3980
29	2008-11-04	2008.84	4015
30	2008-12-09	2008.94	4050
31	2009-02-17	2009.13	4120
32	2009-03-24	2009.23	4155
33	2009-04-28	2009.32	4190
34	2009-06-02	2009.42	4225
35	2009-07-07	2009.52	4260
36	2009-08-11	2009.61	4295
37	2009-09-15	2009.7	4330
38	2009-10-20	2009.8	4365
39	2009-12-29	2009.99	4435
40	2010-03-09	2010.19	4505
41	2010-08-31	2010.28	4540

Table 3.1: Date of all the SAR acquisitions on track 305 over the Manyi Fault. Note there are data available from other dates that has not been included because it does not cover both sides of the fault. The days column describes the number of days since the 1997 earthquake.

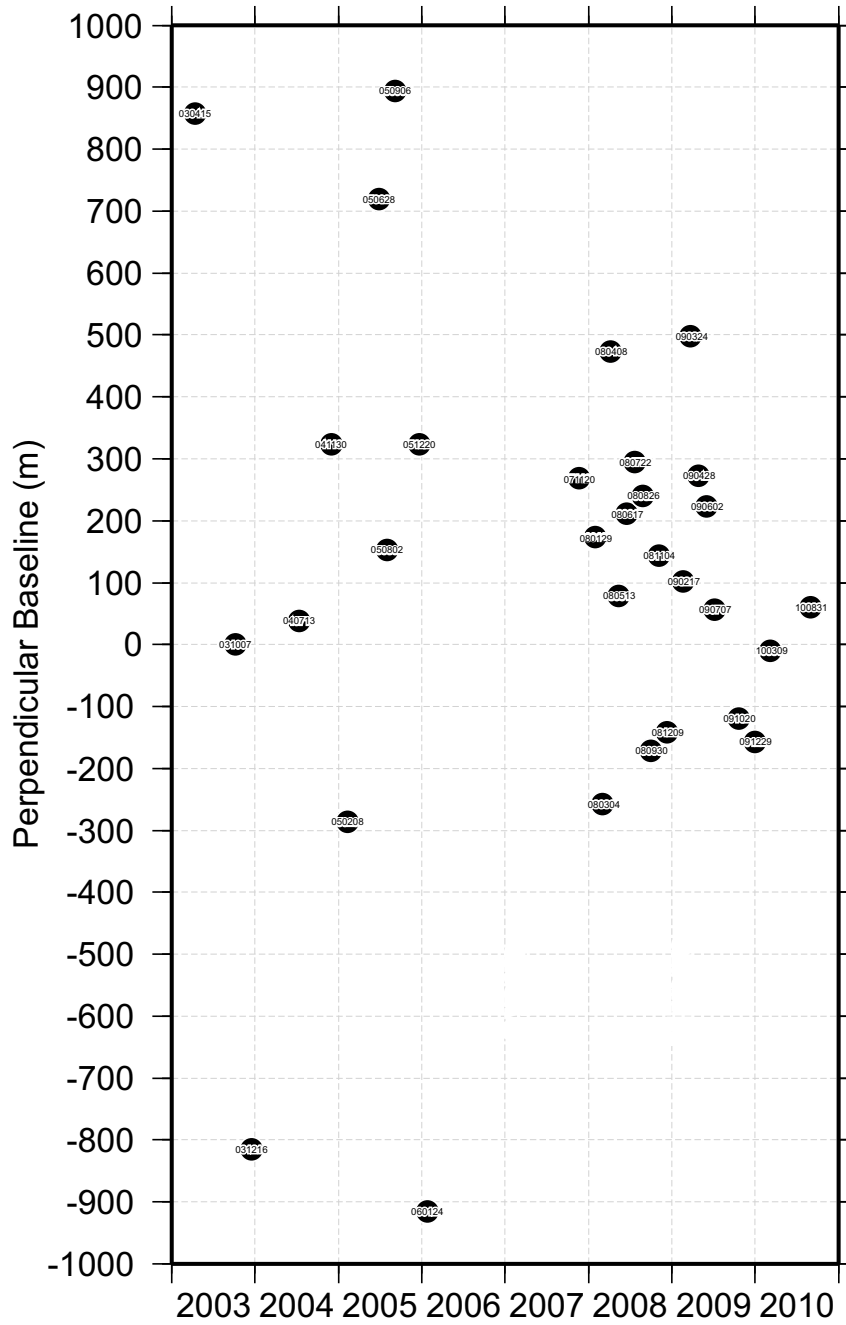


Figure 3.10: The dates of the SAR acquisitions over track 305. Each circle represents one SAR image and the perpendicular baseline values are taken relative to 2003-10-07. The improved steering of the satellite is evident in the tighter distribution of baselines after 2007

3.6.1 Unwrapping Interferograms

Interferograms contain measurements of phase changes of the backscattered microwave radiation and as such all the pixels have phase values between 0 and 2π . These values corresponds to a LOS path difference of $\lambda/2$, which for ENVISAT is 2.8 cm. It is

only half the wavelength as the radiation must travel to the ground and back. A raw interferogram is referred to as wrapped as it only gives deformation between 0 and 2.8 cm. To measure how much the ground has moved we need to convert these bounded measurements by integrating the phase gradient to determine a continuous range of values in a process known as unwrapping. The branch-cut algorithm used to unwrap wrapped interferograms (*Goldstein et al.*, 1988) is based on the principle that any closed path integral should equal zero. Point around which the integral is non-zero (residues) are found and those with adjacent positive and negative residues are joined by a cut and not integrated across. An example of a wrapped and unwrapped coseismic interferogram from the 2007 Tocopilla Earthquake is shown in Figure 3.11.

Generally the unwrapping algorithm correctly unwraps the majority of the interferogram but occasionally it makes errors. For interferograms with a large signal-to-noise ratio these unwrapping errors are usually insignificant and are often ignored, however when studying small scale deformation (< cm scale), unwrapping errors can often mask the signal. Figure 3.12 shows an example of an unwrapping error for a postseismic Manyi interferogram and the same interferogram after it has been corrected. An unwrapping error appears as a jump of $2n\pi$ which can (generally) be spotted by visual inspection of the interferogram. Unwrapping errors can also be found using temporal phase closure loops. Phase closure loops work on the principle that, regardless of temporal path, all the sources of phase change are conservative, i.e

$$\phi_{ln} - \phi_{lm} - \phi_{mn} = 0 \quad (3.1)$$

where ϕ_{ln} is the phase contribution from interferogram ln which is constructed from acquisitions l and n and so on. Unwrapping errors do not behave conservatively, violating Equation 3.1, so by adding (or subtracting) a closed network of interferograms we can identify unwrapping errors. Ideally, techniques would exist for automatic identification of such errors however, in practice, this has not been possible. Instead correction of unwrapping errors has to be performed manually by inspecting each interferogram (or set of interferograms) in turn using data manipulation software. Manual correc-

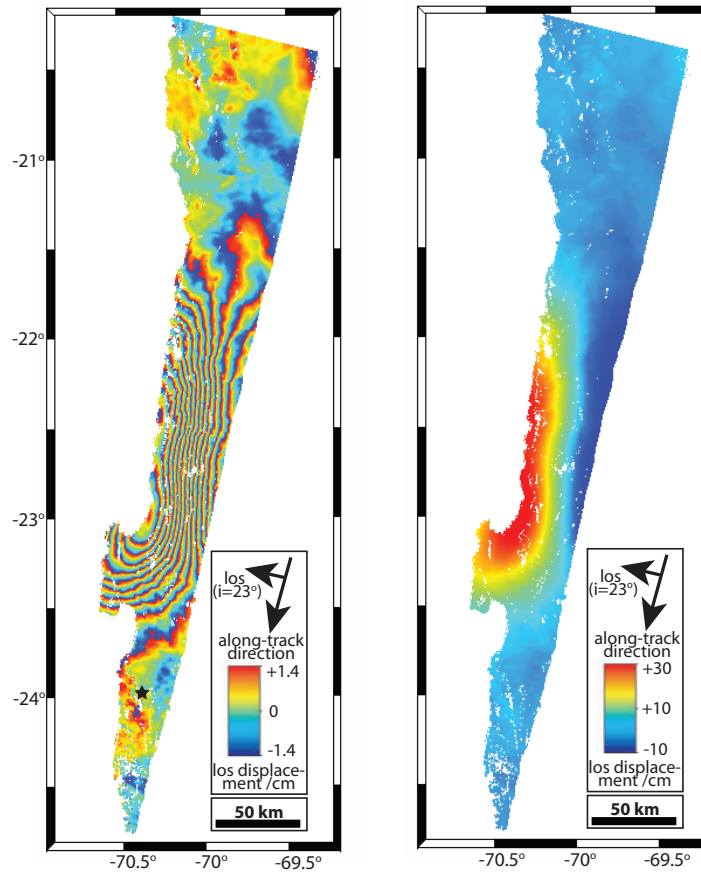


Figure 3.11: The track 368 interferogram showing the LOS deformation due to the 2007 Tocopilla earthquake. One can manually count the fringes on the wrapped interferogram (left) from a zero point which is marked by a star. Counting the fringes as we move directly north towards the centre we go through 10 fringes. 10 fringes corresponds to about 30 cm LOS deformation, which is the peak deformation observed in the unwrapped interferogram (right).

tions are very time intensive as they often occur on or near areas of incoherence, where defining the boundaries of the error can be difficult as the filtering process smears the edges. To avoid ambiguity, the edges of the errors are often masked out. Unwrapping errors under 10 km in spatial extent were removed rather than corrected.

Once the unwrapping errors have been eliminated, we can often see if there is any observable deformation signal. If the signal is not obvious we adopt a variety of techniques to combine interferograms and increase the signal-to-noise ratio. We also remove residual orbital phase ramps.

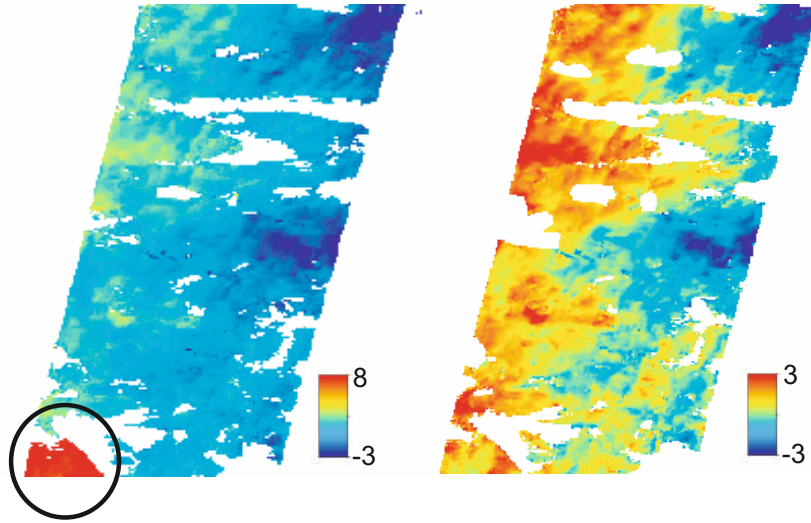


Figure 3.12: The left image is the interferogram with the unwrapping error, the right is the corrected image. The error saturates the rest of the interferogram so any signal there is difficult to see. Subtracting 2π from the region changes the upper bound of the colour scale such that the signal (in this case a tilt) can now be seen easily.

3.6.2 Simple Stacking

Simple stacking is the simplest method of combining interferograms. It assumes that the signal in the interferograms is temporally correlated, as opposed to the atmospheric noise which is random at any given time. The sum of N interferograms with equal duration has a signal N times larger than a single interferogram, but the noise is only $\sqrt{N}$ times larger assuming that the noise is random i.e uncorrelated. The signal to noise ratio is therefore improved by a factor of $\sqrt{N}$. Such methods require interferograms that are completely independent and share no master or slave dates. This method has been used successfully to detect and measure interseismic strain accumulation in various places around the world including Turkey (*Wright et al.*, 2001; *Walters et al.*, 2009), California (*Peltzer et al.*, 2001; *Fialko*, 2006) and Tibet (*Wright et al.*, 2004). These studies have been carried out in areas of relatively good coherence, and only pixels which are coherent in all interferograms are used. They are also regions where there was a large archive of SAR data collected over a long period of time.

We apply the simple stacking technique to the data collected by ENVISAT on track 305 (Figure 3.13). We use the interferograms that have had a second baseline correction

during the simple stacking process. The interferograms used are highlighted in Figure 3.13a and the data split into three periods after the earthquake: 5.5-8 years (black), 10-11 years (red) and 11-13 years (blue). The LOS velocity maps for each period are shown in Figure 3.13b, c and d respectively.

Figure 3.13 clearly shows how the noise is reduced more for the earliest stack (Figure 3.13b) than the latest (Figure 3.13d) as the timespan represented in the stack is larger. As we require each epoch to be used only once, we are limited in the choices of interferograms that can be stacked. This is a particular problem for the two later stacks as there are fewer interferograms with long temporal baselines and thus a higher signal-to-noise ratio.

3.6.3 Chain Stacking

Each interferogram is constructed using two acquisitions, the master and slave acquisition. Hence, the atmospheric and orbital effect in an interferogram contain contributions from atmospheric and orbital terms in the master and slave images. The simple stacking method can be extended by constructing a chain of interferograms, such that the master of one interferogram is the slave of the next. The orbital and atmospheric contributions from acquisitions in the middle of the chain cancel leaving only the contributions from the first master and last slave (Equation 3.5).

$$\phi_{lm} + \phi_{mn} = \left[\frac{2\pi}{\lambda} r_{los}(t_l - t_m) + \phi_{orb,l} - \phi_{orb,m} + \phi_{atm,l} - \phi_{atm,m} \right] + \quad (3.2)$$

$$\left[\frac{2\pi}{\lambda} r_{los}(t_m - t_n) + \phi_{orb,m} - \phi_{orb,n} + \phi_{atm,m} - \phi_{atm,n} \right] \quad (3.3)$$

$$= \frac{2\pi}{\lambda} (t_l - t_n) + \phi_{orb,l} - \phi_{orb,n} + \phi_{atm,l} - \phi_{atm,n} \quad (3.4)$$

$$= \phi_{ln} \quad (3.5)$$

The chain can be made of interferograms with short temporal baselines which do not suffer so much from temporal decorrelation so the coherence will be better than a single interferogram covering the same time-span. The resulting stack is still dom-

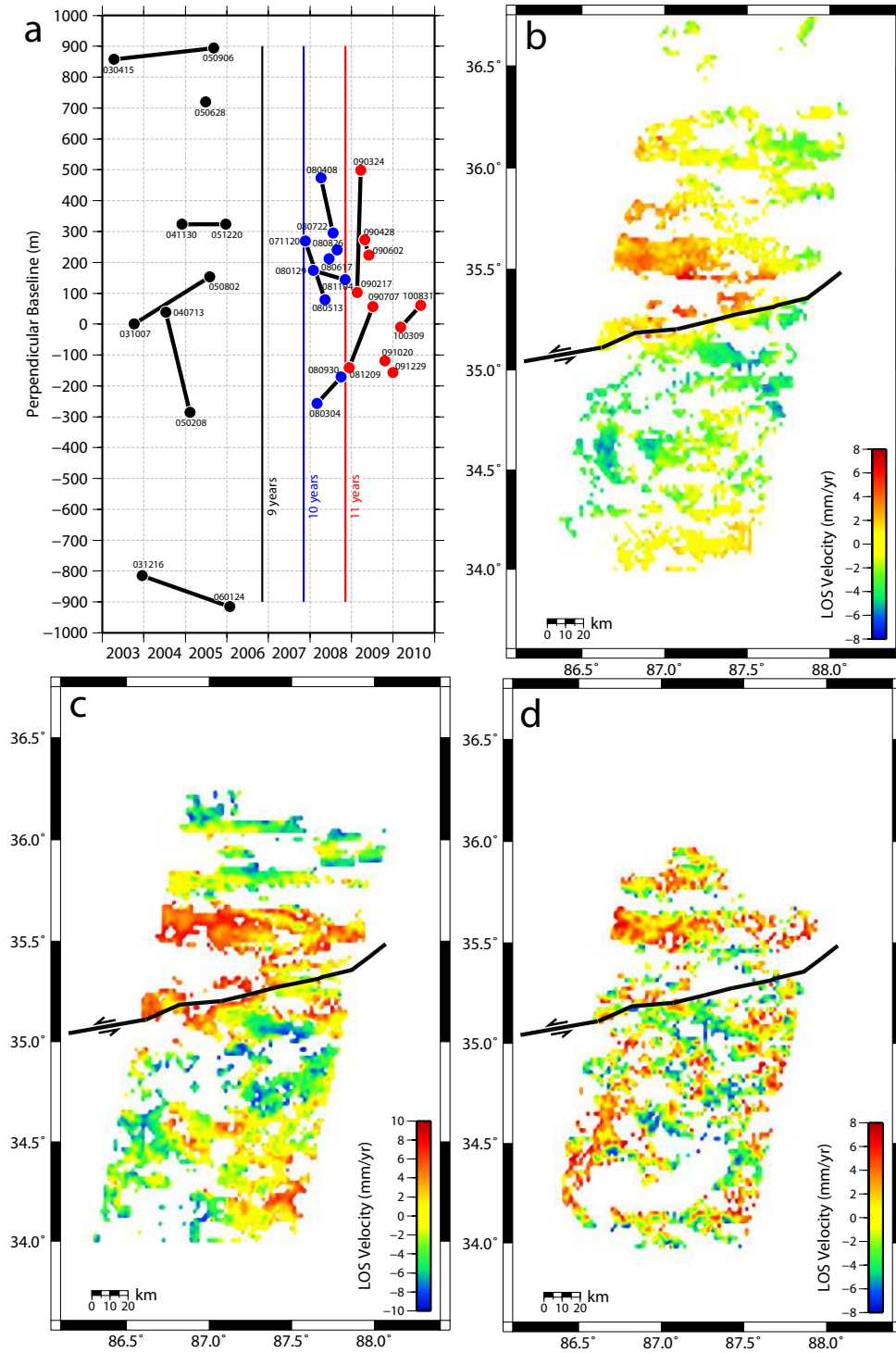


Figure 3.13: a) The baseline-time plot of the postseismic SAR acquisitions from track 305. The lines indicate the interferograms used in the each of the stacks, with the colour indicating which stack each epoch belongs to. The solid vertical lines indicate the time after the earthquake. b) The stack showing velocity between 5.5 and 8 years after Manyi. The cumulative timespan of the stack is 7.94 years. We do not use epoch 050628 as there is no other epoch to which an interferogram could be made that is independent. c) The stack showing velocity between 10 and 11 years after Manyi. The cumulative timespan of the stack is 2.29 years. d) The stack showing velocity between 11 and 13 years after Manyi. The cumulative timespan of the stack is 1.55 years. We do not include any interferogram containing epoch 091229 as the data only covers part of the area and drastically reduces the number of coherent pixels in the stack.

inated however, by atmospheric and orbital contributions from first master and last slave. The velocities are calculated by converting the phase to LOS displacement and dividing by the time between the two end epochs of the chain. Figure 3.14 shows chain stacking of the Manyi postseismic interferograms with the second baseline correction. This technique is not optimal for this area as only short chains (temporally) can be formed and these chains require the use of interferograms with comparatively large perpendicular baselines which reduce the coherence. As the temporal baselines of the stacks are so small also the signal-to-noise ratio remains small and it is still dominated by the atmospheric noise from the epochs at the ends of the chain.

3.6.4 Ratemaps

Stacking methods utilise pixels that maintain coherence in every interferogram in the stack. However pixels that are coherent in some, but not all, interferograms still contain useful information about displacement rate. In order to compare pixels that are coherent in different numbers of interferograms, it is necessary to view the data in terms of displacement rate rather than total displacement. The simplest way to do this is to stack all the interferograms and divide by the cumulative duration of the interferograms which are coherent at each pixel, however this technique suffers from the same limitations as the stacking method described earlier. Alternatively, each interferogram, which measures the total displacement between the dates of the master and slave acquisitions, can be converted into an average displacement rate by dividing by the time span of the interferogram. Assuming displacement builds up at a constant rate, this should be the same in any interferogram for a given pixel. A mean rate for each pixel can be estimated by finding the least-squares fit to these rate values using whichever interferograms are coherent at that pixel.

Both methods assume that the values in the interferogram are proportional to the duration of the interferograms. This is true for some tectonic signals but not the orbital or atmospheric components. If adjacent pixels are coherent in a different number of interferograms, this rate estimation method will divide the adjacent values by different

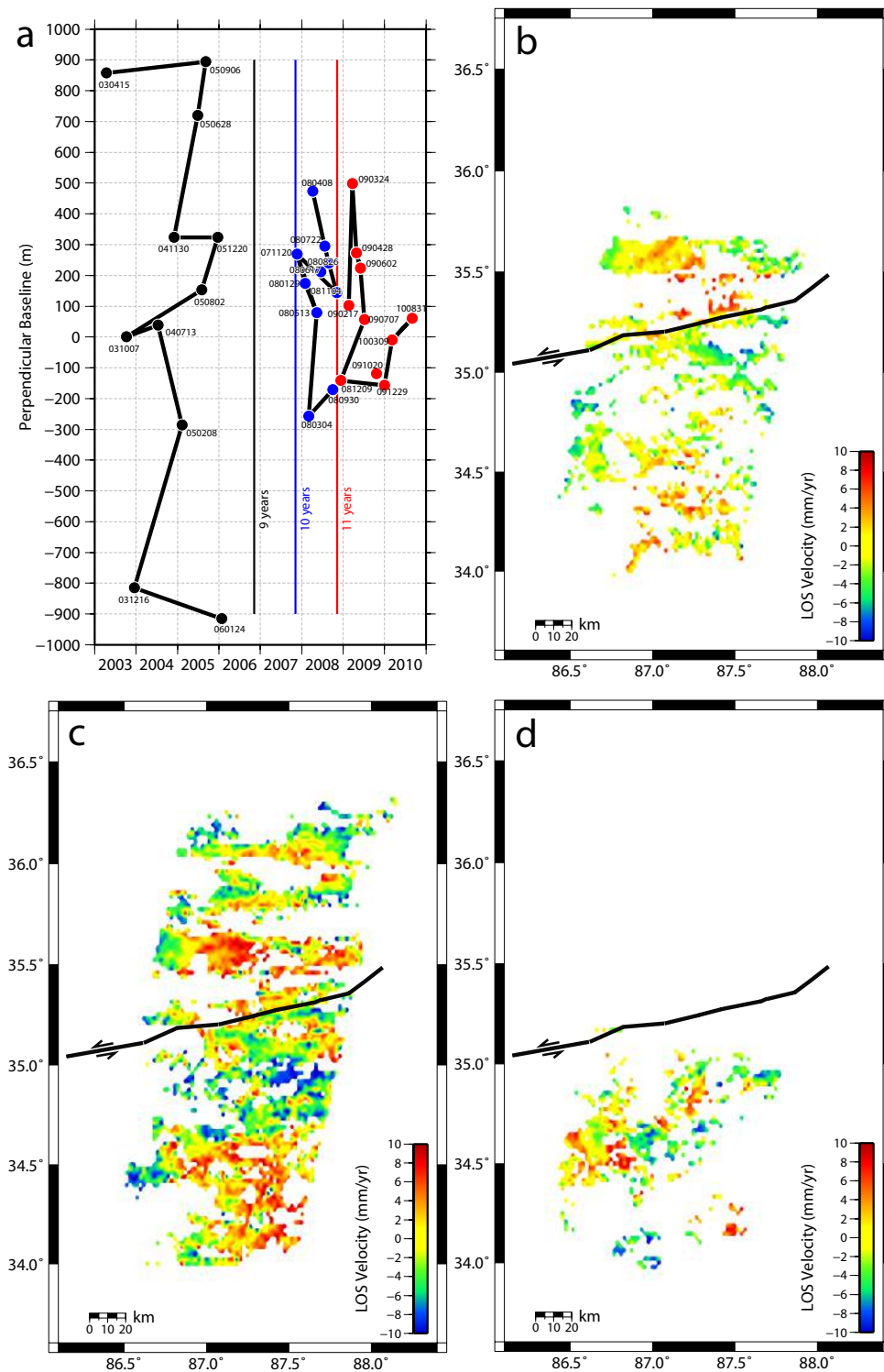


Figure 3.14: a) The baseline-time plot of the postseismic SAR acquisitions from track 305. The lines indicate the interferograms used in the each of the chains, with the colour of the solid circles indicating which stack each epoch belongs to. The solid vertical lines indicate the time after the earthquake. b) The chain showing velocity between 5.5 and 8 years after Manyi. The cumulative timespan of the chain is 2.3 years. c) The chain showing velocity between 10 and 11 years after Manyi. The cumulative timespan of the chain is 0.41 years. d) The chain showing velocity between 11 and 13 years after Manyi. The cumulative timespan of the chain is 1.49 years. The data coverage is so poor in this chain as it must go through date 091229 which has little data north of the fault

durations resulting in a jump. Therefore the orbital contributions should be removed before we calculate the rate at each pixel.

3.6.5 Network Corrections

Incoherence can limit the amount of usable data in the above stacking techniques. Network techniques developed by *Biggs et al.* (2007), allow information from all coherent pixels in all interferograms to be used and also uses an improved orbital correction routine. Using a network approach, it is possible to estimate an orbital error for a single acquisition, rather than solving the orbital error for each interferogram individually. Network approaches are generally only required for measuring small deformation signals, so they start with the assumption that the major contributor to any observed signal is noise, be it from atmospheric water vapour or an orbital baseline error.

The orbital ramp error is approximated as a planar ramp in the horizontal x and y direction. The slopes of orbital ramp (a_i, b_i) are solved at each epoch (i) in addition to a static offset (c_i) . Once the orbital co-efficients have been determined, the correction is then applied to the whole data set, an example of which is shown in Figure 3.15. The inversion solves the equations

$$(a_i - a_j)x_k + (b_i - b_j)y_k + (c_i - c_j) = \phi_{ij,k} \quad (3.6)$$

for all pairs of interferograms connecting the epochs i and j . Interferograms contain millions of pixels. Given the orbital errors and the signals we are studying are long wavelength, each data set is down-sampled to kilometre scale pixels to make the inversions more manageable.

Whilst the problem is over-determined as there are more equations (interferograms) than unknowns (epoch orbital parameters), the matrix is rank deficient as there is one fewer independent interferograms than the number of epoch acquisitions. Therefore a truncated singular value decomposition (SVD) is used to find the minimum norm solution. Once the coefficients have been found the appropriate tilts are removed from

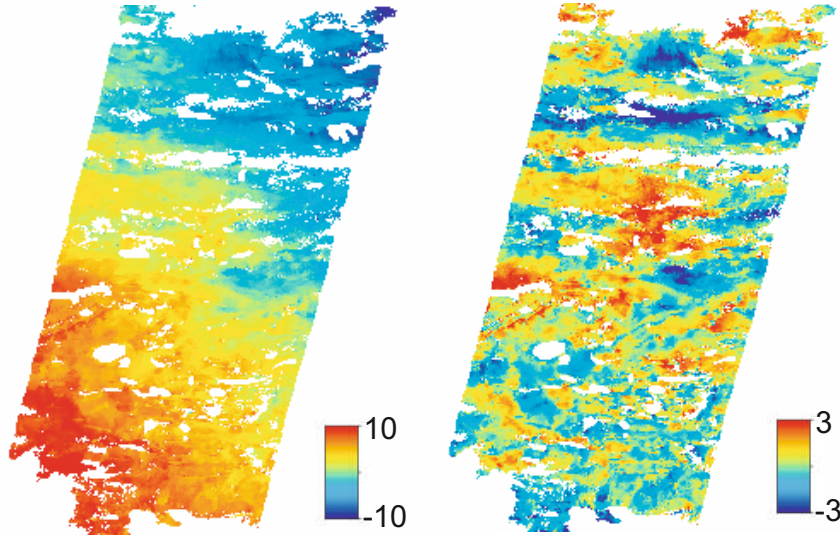


Figure 3.15: Left shows the unflattened unwrapped postseismic interferogram 041130-080408 from track 305. Right shows the same interferogram with calculated best-fitting linear plane removed. The tilt prevents the observation of any signal, that is only observable once it has been removed. The scale in both images is in radians and has different limits for both images

the interferograms, and the interferograms are now said to be flattened.

Spatial Correlations

The other major source of noise/errors in interferograms is from the atmosphere. A standard approach to modelling this noise is to fit a 1-D covariance function to the data to estimate the magnitude and spatial scale of the noise. We assume the noise covariance is radially symmetric and homogeneous throughout the interferogram (*Hanssen, 2001; Parsons et al., 2006*). Care must be taken when applying this assumption as it breaks down if there is significant topographically correlated water vapour or a weather front is passing through the area. This study only deals with data collected on top of the flat Tibetan Plateau, so the assumption should be valid for this area.

The 1-D covariance function is found from the radial average of 2-D auto-covariance function of the interferogram calculated from the cosine Fourier transform of the power spectrum (*Parsons et al., 2006*). A simple exponential function is used to characterise the covariance function,

$$c_{ij} = \sigma^2 e^{(-d_{ij}/\alpha)} \quad (3.7)$$

where c_{ij} is the covariance between two pixels i and j separated by a distance d_{ij} , and σ^2 is the variance and α the e-folding decay distance.

Temporal Correlations

To take into account the correlation between interferograms sharing common epochs and also the simple atmospheric noise, a variance-covariance matrix Σ_ϕ is formed. The element σ_{ij-kl} of the variance-covariance matrix Σ_ϕ corresponding to the interaction between two of interferograms ij and kl used to form the rate map, assuming that the atmospheric noise of each epoch is the same, is given by

$$\sigma_{ij-kl} = c_{ij-kl}\sigma^2, \quad \text{where } c_{ij-kl} = \begin{cases} 1 & \text{if } i = k \text{ and } j = l \\ 0.5 & \text{if } i = k \text{ or } j = l \\ -0.5 & \text{if } i = l \text{ or } j = k \\ 0 & \text{otherwise} \end{cases} \quad (3.8)$$

where σ^2 is the median atmospheric noise variance of the interferograms (as determined using Equation 3.7 on each interferogram). The correlation coefficient c_{ij-kl} indicates whether there is a common master/slave epoch, with independent interferograms having a value of zero. As the ratemap inversion is done on a pixel-by-pixel basis, the spatial correlation between pixels is not taken into account (exponential term in Equation 3.7). Any errors in the orbital corrections that trade off with deformation signals or topography-dependent atmospheric contributions are also not accounted for here.

3.7 Formation of a Deformation Map

A least-squares matrix inversion is performed on a pixel-by-pixel basis to determine the best-fitting line-of-sight rate of range change r_k for pixel k given the orbital flattened interferogram phase data Φ ,

$$\frac{4\pi}{\lambda}Tr_k = \Phi \quad (3.9)$$

where T is a vector of interferogram time-spans, $[t_{12}, t_{13}, t_{23} \dots t_{ij}]^T$ and Φ is the vector of individual interferogram phase, $[\phi_{12}, \phi_{13}, \phi_{23} \dots \phi_{ij}]^T$ and λ is the wavelength of the SAR antenna.

Starting with Equation 3.9 and incorporating the variance-covariance matrix Σ_ϕ , the rate determined at each pixel, r_k , and the associated error, Σ_r , are given by:

$$\frac{4\pi}{\lambda} T r_k = \Phi \quad (3.10)$$

$$\frac{4\pi}{\lambda} T^T \Sigma_\phi^{-1} T r_k = T^T \Sigma_\phi^{-1} \Phi \quad (3.11)$$

$$r_k = \frac{\lambda}{4\pi} \underbrace{[T^T \Sigma_\phi T]^{-1}}_{\Sigma_r} T^T \Sigma_\phi \Phi \quad (3.12)$$

As interferograms are conservative, the total displacement change between any two epochs is the same regardless of the chain of interferograms between them. This means that closed loops of interferograms contain redundancy. The sum of two interferograms, I_{ab} and I_{bc} is the same as interferogram I_{ac} where a , b , and c are the master/slave epochs. Interferogram I_{ac} is therefore redundant as it contains no new information. To decrease the calculation time, we only use independent, non redundant, interferograms in our inversions. For each pixel, a selection of independent interferograms is chosen by a Kruskal algorithm (*Kruskal*, 1956) which selects the least noisy interferograms (those with the lowest σ) in which the pixel is coherent. To improve the number of pixels in the velocity map, we use pixels that are coherent in four or more independent interferograms within the network, rather than every interferogram (as is required in previous stacking methods). The ratemaps of LOS velocity from the Manyi postseismic data calculated using network techniques are shown in Figure 3.16. The interferograms used in this technique are those without the second baseline estimation. The term Σ_r in Equation 3.10 provides an estimate of the error at each pixel. This error value is used to mask out any pixels with an error greater than the maximum rate observed in the ratemap. In this case the distribution of errors is not random, the pixels with the highest errors tending to be just to the south of the fault in a region which generally

shows positive LoS deformation. This means that previous static offset correction for each ratemap calculated during the flattening process must be incorrect, an issue which is addressed in the next section.

By comparing Figures 3.13, 3.14 and 3.16 it becomes apparent that using a network approach increases the number of coherent pixels in the velocity map. The ratemaps calculate an orbital correction that aims to flatten the ratemap. This correction over compensates as the area of deformation is not masked out during the flattening process so when comparing the data to any deformation models we must consider that there may be a residual linear ramp.

Ratemaps formed in this manner assume a constant velocity over the time period of the observations. If there is significant change in deformation rate during the interval of observations then a more complex time series approach is needed. To study the deformation immediately after the Manyi earthquake we use the data from *Ryder et al.* (2007), which is based on a time series approach and will be discussed in chapter 5. We also expand on the ideas presented here to incorporate variable errors at each epoch in the next chapter.

3.7.1 Topographic Correction

Throughout this ratemap formation we have assumed that there is no contribution of topographically-correlated water vapour-related atmospheric signals to the ratemaps. To test this is a reasonable assumption we plot the deformation ratemaps from Figure 3.16 against the DEM (Figure 3.17). For each of the ratemaps we find the best-fitting gradient (and intercept) using a linear least-squares inversion.

Figure 3.17 shows there is little to no topographic influence. There is a slight effect in the earliest ratemap (5.5-8 yrs) but the two later ones have an almost flat gradient. We apply a topographic correction by removing the linear trend found in Figure 3.17 from all three ratemaps.

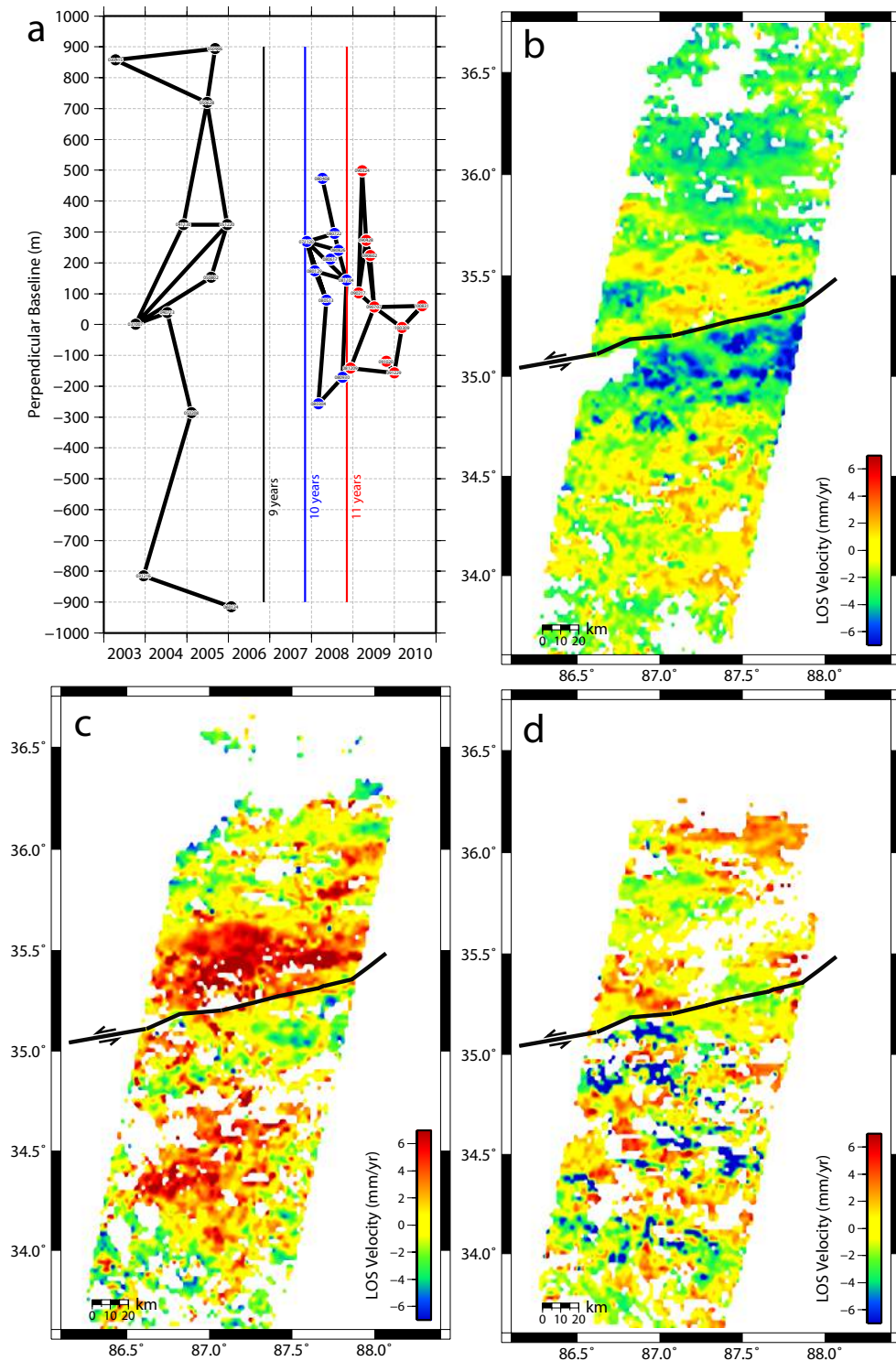


Figure 3.16: a) The baseline-time plot of the postseismic SAR acquisitions from track 305. The black lines indicate the interferograms used in the each of the networks. The dots indicate the epochs, with the colour indicating which network each epoch belongs to. The solid vertical lines indicate the time after the earthquake. b) The velocity map from between 5.5 and 8 years after Manyi. The mean velocity error over the whole ratemap is 3 mm/yr c) The velocity map between 10 and 11 years after Manyi. The mean velocity error over the whole ratemap is 4 mm/yr d) The velocity map from between 11 and 13 years after Manyi. The mean velocity error over the whole ratemap is 5 mm/yr.

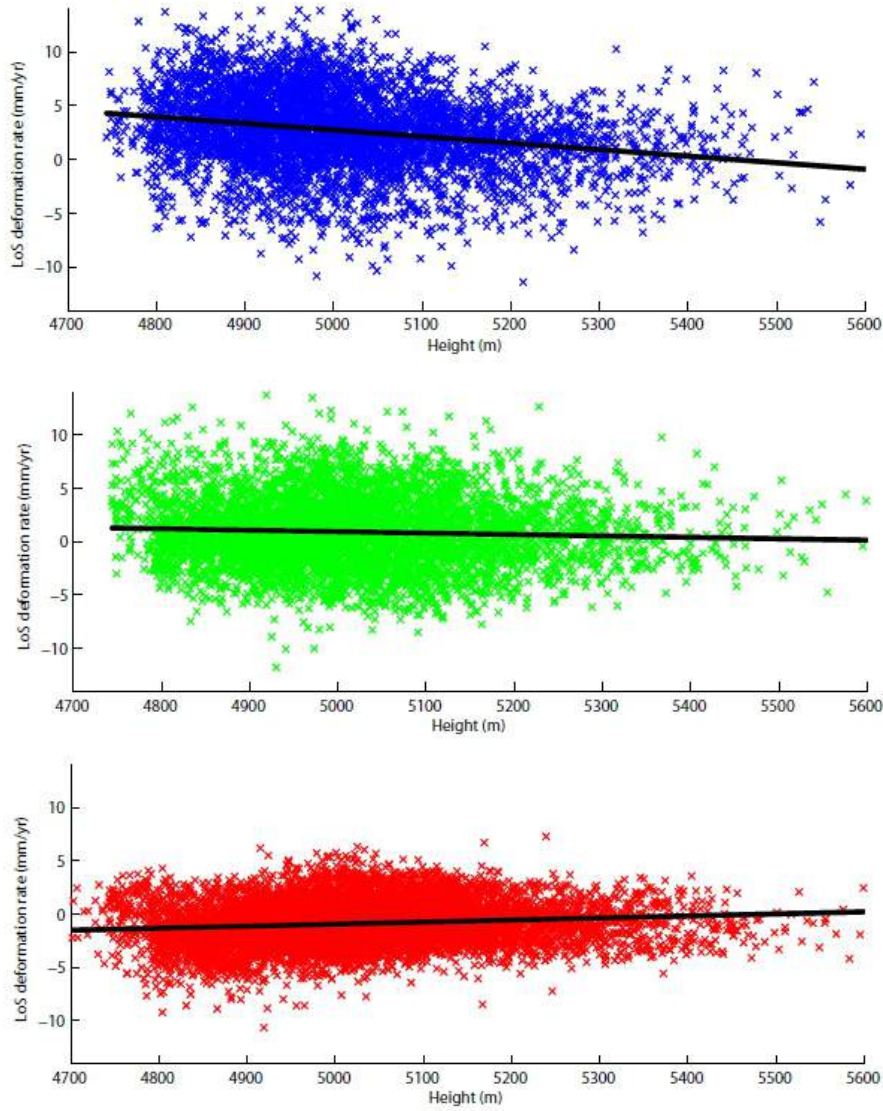


Figure 3.17: The points are colour coded according to the ratemap to which they belong. Blue points and lines correspond to the 5.5-8yr stack, green to the 10-11 and red to the 11-13 yr stack. The solid lines are the line of best-fit through the points. Gradient of the red green and blue lines are -0.0021 , -0.00030 and 0.00026 cm/m respectively. The error on the gradients are 0.00016 , 0.00026 and 0.00026 cm/m.

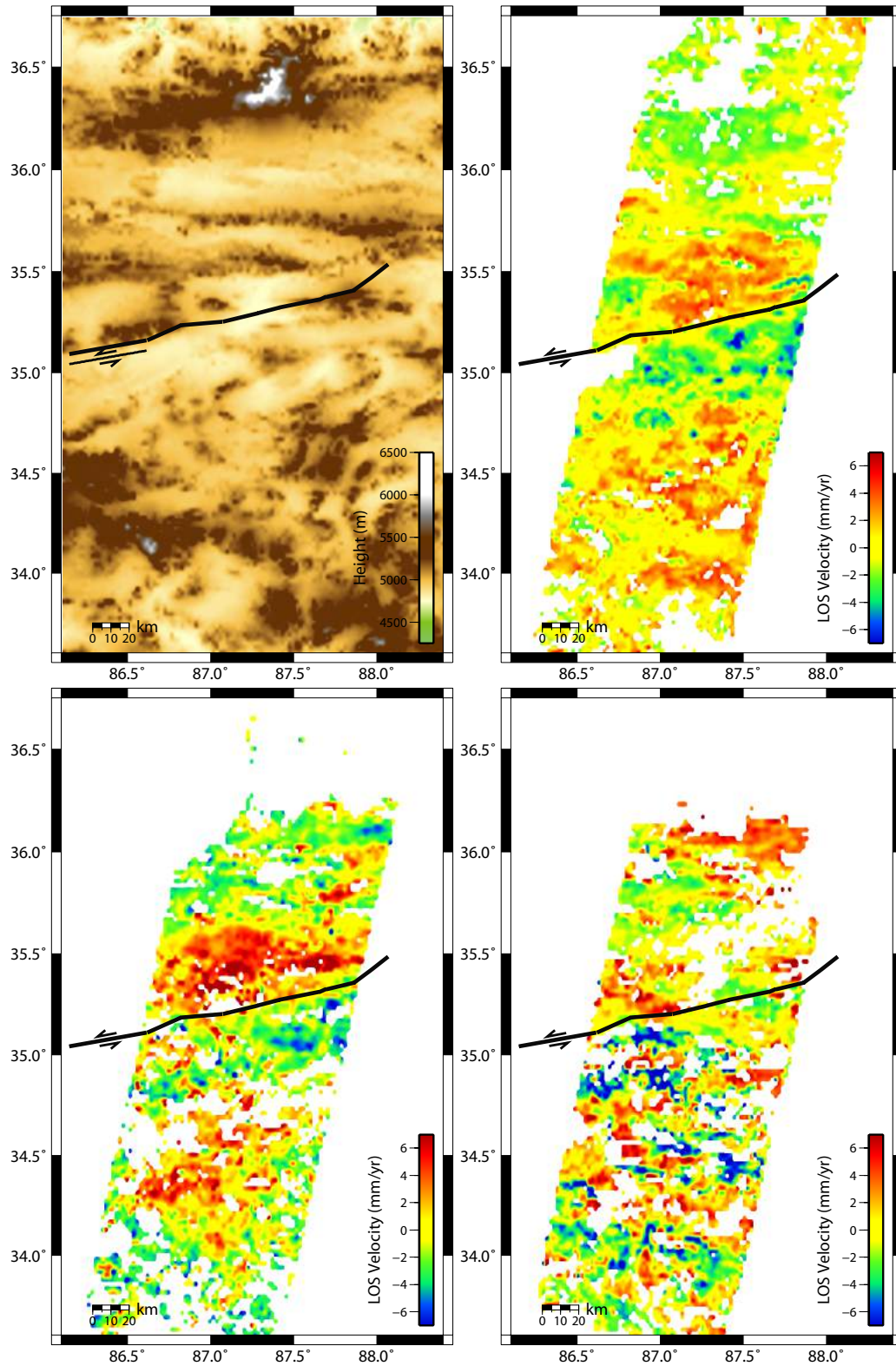


Figure 3.18: a) The SRTM DEM over the area of the interferograms. b) The topographically corrected velocity map from between 5.5 and 8 years after Manyi. The mean velocity error over the whole ratemap is 3 mm/yr c) The topographically corrected velocity map between 10 and 11 years after Manyi. The mean velocity error over the whole ratemap is 4 mm/yr d) The topographically corrected velocity map from between 11 and 13 years after Manyi. The mean velocity error over the whole ratemap is 5 mm/yr.

3.8 Are we observing postseismic motion?

The area around the Manyi fault is clearly still moving, even 12 years after the 1997 earthquake. Before we attribute the motion to postseismic deformation we must consider the possibility that the motion is not postseismic at all but the area has returned to a interseismic regime where stress is built up and will be released in the next 'Manyi' earthquake. In the next chapter we constrain the magnitude of interseismic strain accumulation across the Manyi Fault before the event.

Chapter 4

Interseismic Deformation prior to the 1997 Manyi Earthquake

4.1 Introduction

The previous chapter presented results that show the area around the Manyi is still, 10 years after the coseismic event, moving at rate of several millimetres/yr. A similar magnitude of displacement rate is observed in several InSAR studies (*Wright et al.*, 2004; *Taylor and Peltzer*, 2006; *Elliott et al.*, 2008) that measure the strain accumulation between earthquakes, known as the interseismic stage of the earthquake cycle. Such millimetre precision measurements of slip rates have only been feasible for the past few decades, meaning that it is still unclear where the postseismic stage of the earthquake cycle ends as it slowly transitions into the interseismic stage. In this chapter I aim to use and extend the techniques developed in the previous chapter to study the interseismic motion across the Manyi fault prior to the 1997 event, which will allow us to establish if the current observed deformation is due to some postseismic transient or whether the area has return to it's pre-earthquake state. This chapter is based on work contained in *Bell et al.* (2011).

4.2 Tectonic Background

The M_w 7.6, 1997 Manyi Earthquake in Northern Tibet (Figure 4.1) ruptured a 200 km long sinistral fault for which the coseismic slip distribution has been well established with InSAR and seismological methods (*Funning et al.*, 2007b; *Velasco et al.*, 2000). The initial postseismic motion deformed the surface at a rate of ~ 1 cm/yr and, for the first three years following the event, could be explained by either afterslip or viscoelastic relaxation (*Ryder et al.*, 2007).

To distinguish between these two possible mechanisms the postseismic time series must be extended; postseismic motion will decay over timescales (and length-scales) which are mechanism dependent (*Hearn et al.*, 2009; *Johnson et al.*, 2009; *Bruhat et al.*, 2011). The deformation rate from either mechanism will however, eventually decay to the point where the rate of deformation will become indistinguishable from the interseismic strain accumulation rate. Therefore, to establish the mechanism behind the current deformation in the area of the Manyi fault (*Bell et al.*, 2010), we must first constrain the long-term interseismic strain accumulation rate across the fault. Estimates of the pre-earthquake slip rate also allow us to constrain models of postseismic frictional afterslip (*Dieterich*, 1979; *Tse and Rice*, 1986; *Marone et al.*, 1991; *Perfettini and Avouac*, 2007).

Furthermore, obtaining accurate measures of interseismic strain accumulation can clarify the distribution of strain and faulting in this part of the Tibetan Plateau. The Kunlun fault clearly runs from 100°E to 90°E and possibly extends to 86°E at a latitude of 35.8°N (*Taylor and Yin* (2009) and Figure 4.1). However, whilst the 2001 Kokoxili (Kunlun) earthquake ruptured to almost 90°E , the rupture on the 1997 Manyi earthquake is left stepped southwards by 50 km. This region is devoid of any GPS sites (*Gan et al.*, 2007), hence InSAR measurements are essential for determining strain accumulation and hence, by inference, fault slip rates in this region, e.g. (*Elliott et al.*, 2008; *Taylor and Peltzer*, 2006). The distribution of strain and determination of fault slip rates is vital for discriminating between two end-member models of continental deformation in Tibet (Figure 1.2); 1) distributed deformation of a ductile lithosphere

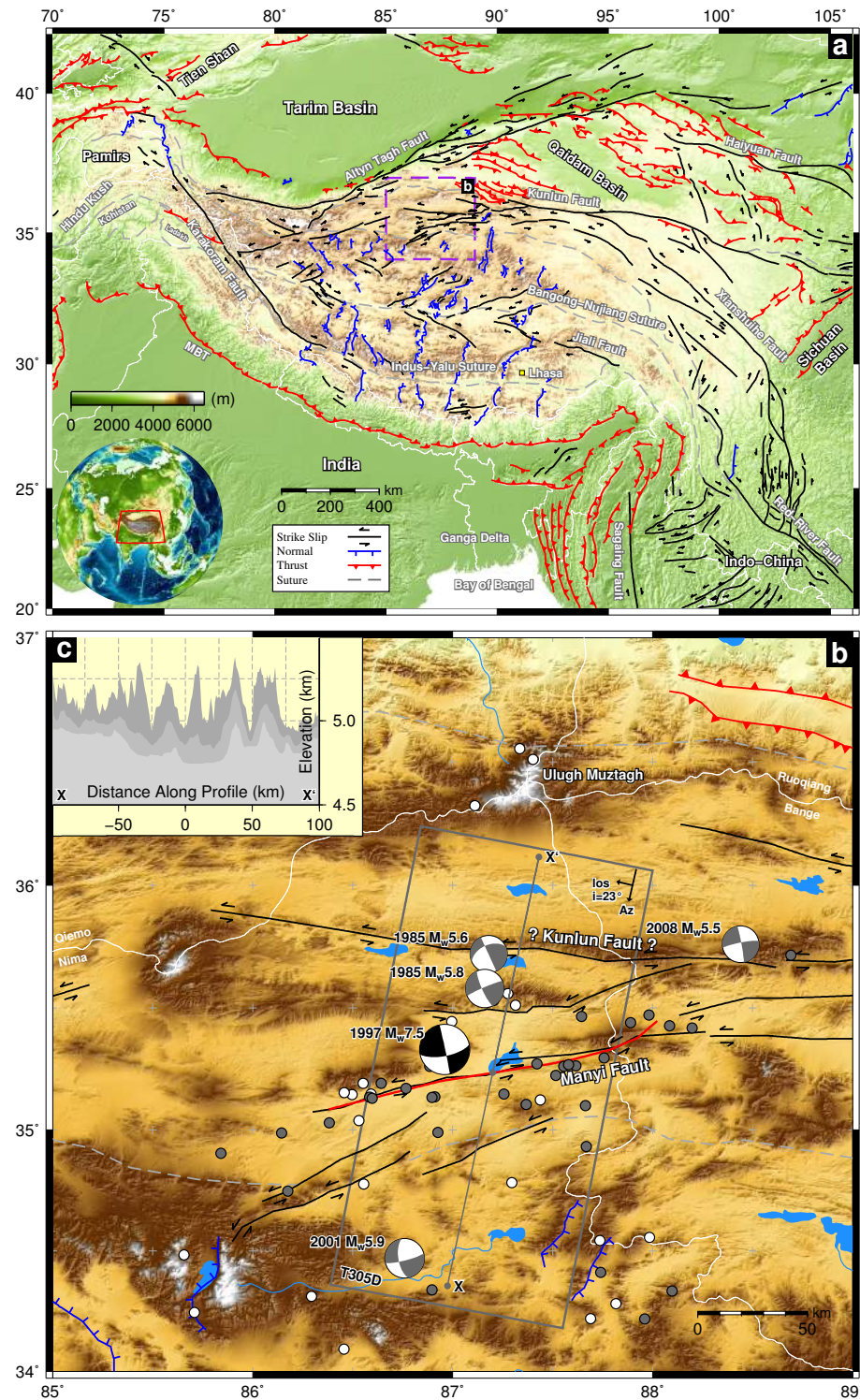


Figure 4.1: Fault and topographic map of the Tibetan Plateau and immediate surrounding region, with the study area delineated by a purple dashed box. Faults are from *Taylor and Yin* (2009). (b) Enlarged region of northern Tibet centered around the location of the Manyi earthquake (the red line indicates fault trace from *Funning et al.* (2007b)) and ERS coverage used (track 305 descending). The black arrows show the azimuth and look direction of the satellite. Focal mechanisms are earthquakes from the gCMT catalogue (1976-2010) with $M_w \geq 5.5$. The solid dots show the epicentre of events prior to (in white) and after (in grey) the 1997 event (*Engdahl et al.*, 1998) (c) Topographic profile along X-X' showing the limited relief of the plateau along the SAR track (50 km wide bins with min, mean and max profiles in grey-scale).

with brittle upper crustal failure on lots of faults e.g. *England and Molnar* (2005) or 2) quasi-rigid block rotation, where motion is concentrated on few faults along the block boundaries, e.g. *Meade* (2007).

4.3 Data Availability and Processing

The European Space Agency (ESA) launched one of the first InSAR capable satellites in 1992, ERS-1. That was followed by the launch of ERS-2 in 1995, which followed the ERS-1 satellite with a 1-day lag. These two satellites provided one of the earliest datasets across most of the Tibetan Plateau which been used for numerous interseismic studies in the area (*Wright et al.*, 2004; *Elliott et al.*, 2008). Unfortunately a gyroscope failure in 2000 compromised the pitch control for the satellites and made them largely unusable for InSAR. As the SAR antenna is an active device power constraints on-board the satellites restricted the majority of acquisitions to the daytime descending mode rather than the nighttime ascending mode.

We use ERS-1 and ERS-2 data from descending track 305 (Figure 4.1b and 4.2 that cover the majority (>80%) of the 1997 earthquake rupture. To prevent contamination of the deformation signal by the coseismic motion, we limit SAR acquisitions to those acquired before the 8th November 1997 earthquake. There are insufficient radar scenes on ascending tracks across the Manyi fault for interseismic studies.

The SAR data were processed using the JPL/Caltech software ROI.pac (*Rosen et al.*, 2004). An SRTM digital elevation model and Delft (ODR) orbits (*Scharroo and Visser*, 1998) were used to make topographic and orbital corrections respectively. The interferograms were filtered (*Goldstein and Werner*, 1998) and then unwrapped using a branch-cut algorithm (*Goldstein et al.*, 1988). The accuracy of the unwrapping was checked by testing phase closure loops of three interferograms and any errors that occurred during the unwrapping were either corrected manually or, if under 500 m in spatial extent, removed. We attempt to process every possible interferogram however the baseline between some dates was too large to produce an interferogram of sufficient

coherence.

4.3.1 Velocity Map Formation

The line-of-sight (LOS) velocity map in Figure 4.3 shows the annual LOS displacement change formed using a network of 20 interferograms (Figure 4.2) made from radar images at ten different epochs using the methods of *Biggs et al. (2007)* and *Wang et al. (2009a)*. As the deformation we are studying is of long wavelength (>10 km), we reduce the resolution of the interferograms to 800 m when calculating the LOS velocities to make the calculations more efficient.

Because our estimates of satellite positions are imperfect, an orbital baseline error remains, even after the effects of baseline separation have been corrected for during processing. This error manifests itself in the interferograms as a linear phase ramp. Interferogram I_{ij} , made from epochs i and j , has an orbital phase ramp: r_{ij} , that is dependent upon the position error at each epoch. The orbital errors can be thought of as contributing a phase ramp at each epoch,

$$r_i = a_i x + b_i y + c_i \quad (4.1)$$

The phase ramp in interferogram I_{ij} is then difference between the phase ramps at epochs i and j i.e

$$r_{ij} = r_i - r_j = (a_i - a_j) x + (b_i - b_j) y + (c_i - c_j). \quad (4.2)$$

We determine r_{ij} empirically by fitting a linear plane to I_{ij} and use these planes to determine the coefficients a, b, c at each epoch. These are then used to calculate and remove an orbital phase ramp from each interferogram. As the inversion is rank deficient, there being one fewer independent interferograms than epochs, the inversion uses singular value decomposition to find a minimum-norm solution. We expect the correction to remove correctly orbital phase errors parallel to the fault, but there to be a trade-off between the fault perpendicular ramp and the signal from tectonic

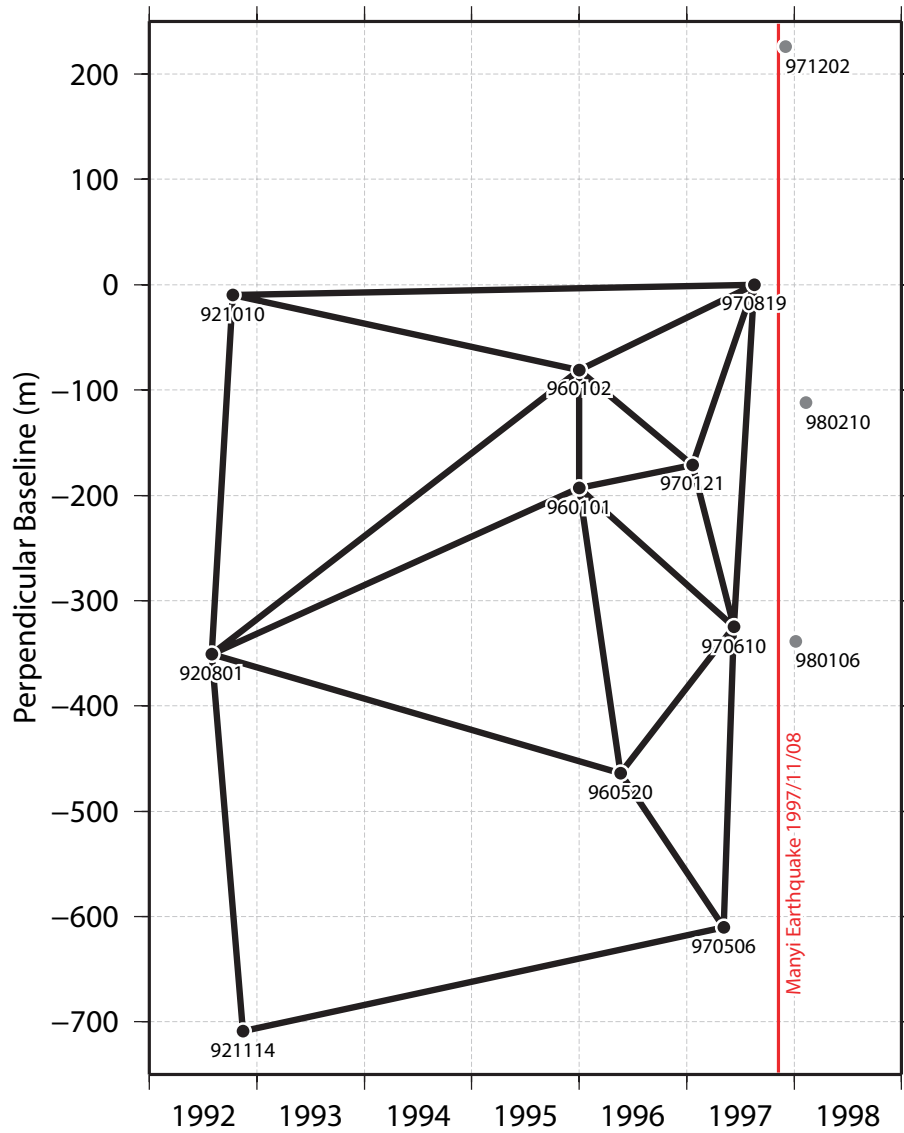


Figure 4.2: Baseline (relative to 19th August 1997 acquisition) versus time plot for all track 305 acquisitions used in estimating the interseismic strain accumulation across the Manyi fault. The solid black lines denote the interferograms used. The red line shows the time of the November 1997 Manyi earthquake.

strain accumulation. This would affect the slip rate estimate, so we assume the fault-perpendicular component of the correction contains some of the deformation signal, and we make a further correction when solving for the final slip rate.

We also determine a measure of the atmospheric noise for each interferogram. We assume the form of this spatially variable noise has a 1-D covariance function (*Hanssen*, 2001);

$$\sigma_{ij}^2 e^{-r/d_{ij}} \quad (4.3)$$

where σ_{ij} is the interferogram error, r is the distance between pixels and d_{ij} is a characteristic e-folding distance. σ_{ij} depends on epoch errors σ_i and σ_j such that $\sigma_{ij}^2 = \sigma_i^2 + \sigma_j^2$. By fixing d_{ij} to the mean value across all interferograms of 8 km, we can then estimate a variance at each epoch, σ_i^2 , using the same inversion technique used to find the orbital phase ramps.

As the 20 interferograms cover only ten epochs, there are redundant interferograms in the network. For each pixel an optimal tree, which comprises a subset of the connected independent interferograms, is calculated using a Kruskal algorithm that selects branches that minimise a cost function but still includes every epoch where a coherent pixel exists. The cost function is the sum across the subset of interferograms of f_{ij} , where f_{ij} is the fraction of incoherent pixels in interferogram I_{ij} . As there are few pixels which are coherent in all interferograms, we use pixels that are coherent in at least six of the interferograms of its optimal tree. We perform a pixel-by-pixel least-squares inversion to find the best-fitting LOS displacement rate r_{los} (Figure 4.3), from the appropriate tree of interferograms selected from Figure 4.2,

$$r_{los} = \frac{\lambda}{2\pi} [T^T \Sigma_T^{-1} T]^{-1} T^T \Sigma_T^{-1} P \quad (4.4)$$

where T is a vector containing time spans of the interferograms in the tree, $[t_{12}, t_{23} \dots t_{ij}]^T$, P is a vector containing the phase data, $[\phi_{12}, \phi_{23} \dots \phi_{ij}]^T$ for each interferogram in the tree, λ is the wavelength of the SAR instrument, and Σ_T is the covariance matrix which takes into account atmospheric noise at each epoch and temporal correlation between interferograms. We expand the temporal correlation used earlier (Equation 3.8) which assumed the noise/variance at epoch was the same, to incorporate a variable variance (σ_i^2) at each epoch. The elements of Σ_T , c_{ij-kl} , describe the covariance between interferograms I_{ij} and I_{kl} , and are given by

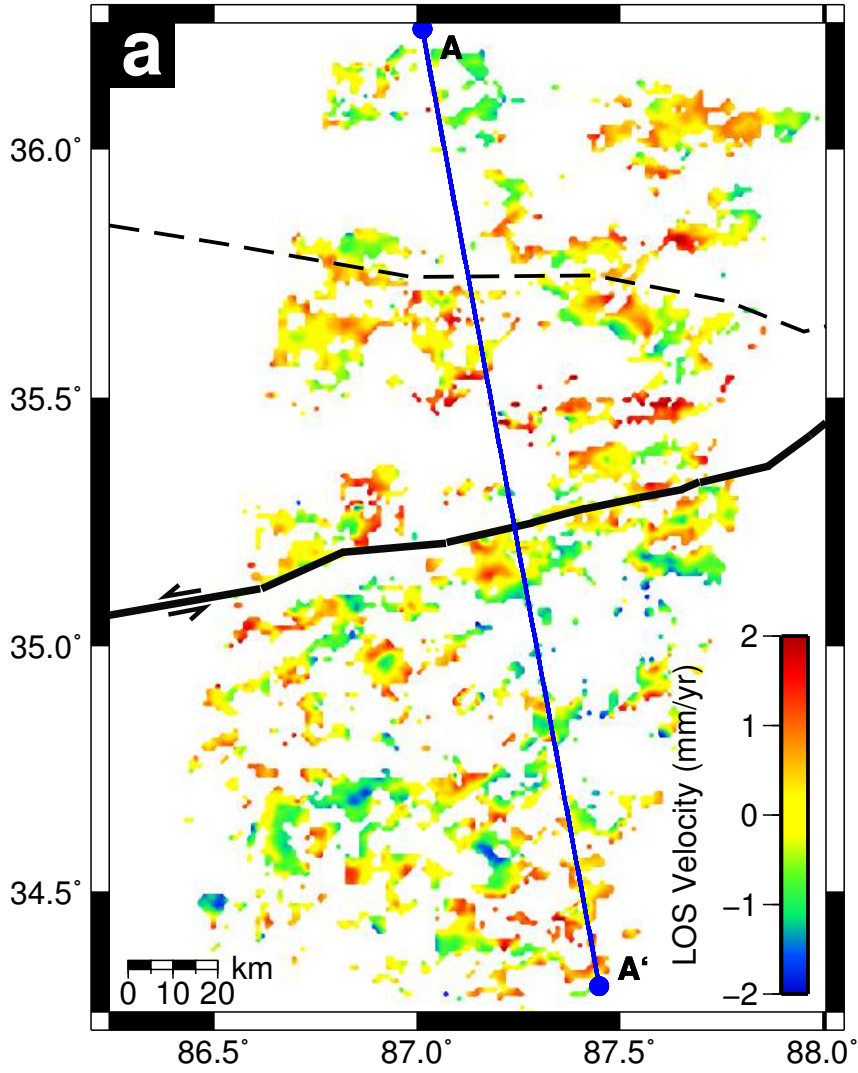


Figure 4.3: The LOS velocity map formed from the interferograms shown in 4.2. The thin solid blue line (A-A') denotes the location of the profile across the Manyi fault shown in Figure 4.7. The solid black line denotes the trace of the Manyi fault. The dashed black line denotes the trace of the possible extension of the Kunlun fault.

$$c_{ij-kl} = \sigma_i \sigma_k \delta_{ik} + \sigma_j \sigma_l \delta_{jl} - \sigma_k \sigma_j \delta_{kj} - \sigma_i \sigma_l \delta_{il} \quad (4.5)$$

where δ_{ij} is the Kronecker Delta. The rate error for each pixel in the inversion is given by vector σ_p from equation 4.6. A map of the error values for each pixel is shown in Figure 4.4.

$$\sigma_p = [T^T \Sigma_T^{-1} T]^{-1}. \quad (4.6)$$

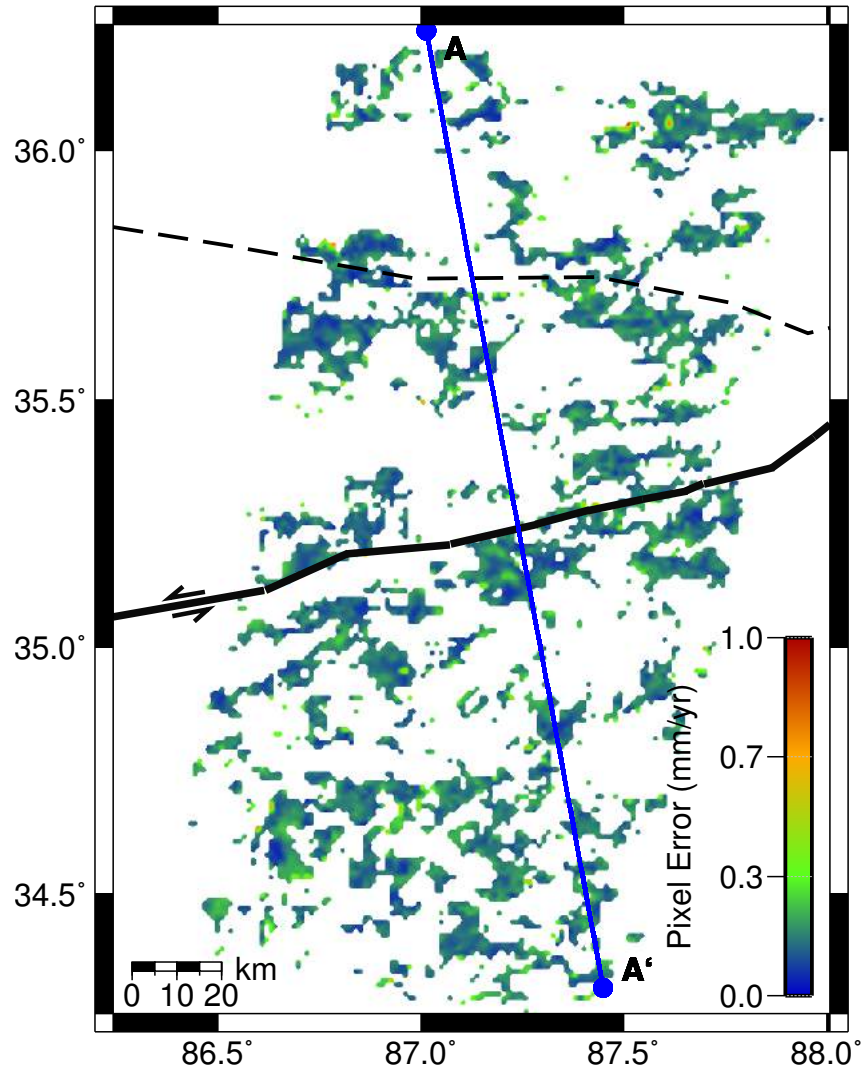


Figure 4.4: The error in the LOS velocity map (Figure 4.3) formed from the interferograms shown in Figure 4.2. The majority of pixels have a similar low error as they have the same tree of independent interferograms in their formation. Those with larger errors use noisier interferograms in their trees. The pixels with errors greater than 0.8 mm/yr have been culled in the figure. The thin solid blue line (A-A') denotes the location of the profile across the Manyi fault shown in Figure 4.7. The solid black line denotes the trace of the Manyi fault. The dashed black line denotes the trace of the possible extension of the Kunlun fault.

4.4 Modelling Interseismic Deformation

The interseismic phase is an integral part of the earthquake cycle. The locked, seismogenic layer behaves elastically as it is driven by far-field stress that is deforming the lithosphere below. Calculating the magnitude of interseismic strain accumulation across a fault can allow us to assess the seismic hazard it presents. Interseismic slip rate can be calculated using a variety of techniques.

Some methods look at the long term slip rates across the faults. Such techniques measure fault offsets in the field and use techniques such as cosmogenic dating (*Mériaux et al.*, 2004) or radioisotope dating (*Cowgill et al.*, 2003) to date the offset and thus calculate the slip rates. Others use geodetic techniques such as InSAR (*Wright et al.*, 2004; *Bell et al.*, 2011) and GPS (*Zhang et al.*, 2007) to map the current movement across the fault. Geodetic techniques provide measurements of the Earth's surface so it can be difficult to infer the style of deformation taking place at depth within the lithosphere. There are two types of interseismic strain accumulation model: 1) the shear zone model where deformation is distributed over a wide zone and is only localised at the surface (*Prescott and Nur*, 1981) or 2) the deep fault model (or back-slip) in which slip is confined to a discrete plane at depth (*Savage and Burford*, 1973).

4.4.1 Interseismic Models

Shear Zone Model

Bourne et al. (1998) use the shear zone model to explain the deformation below New Zealand. In these models the shear-strain rate at the surface due to the buried shear zone is equivalent to the sum of the distribution of screw dislocations (Figure 2.2) with infinitely small slip at depth. The velocity at depth however is filtered through an elastic lid of thickness d . This causes any features that are smaller than πd to be smoothed out and unresolvable. If we assume the locking depth is 15 km (*Funning et al.*, 2007b) then we cannot distinguish between one fault or a shear zone of width 50 km from surface measurements alone.

Savage and Burford Model

The Savage and Burford model used in this study (*Savage and Burford, 1973*) assumes a fault runs from the surface to infinite depth as a single discrete fault plane. As the frictional properties of rocks change with depth in the Earth (*Scholz, 2002*) we assume the top layer (to depth d) remains locked throughout the interseismic phase. Below depth d the fault creeps continuously and the surroundings behave elastically. The upper elastic locked lid responds to the deformation taking place below it. This is the situation described by Figure 2.4. The rate of surface deformation y at a perpendicular distance x from the fault locked to a depth d , moving at a slip rate of s is given by

$$y = \frac{s}{\pi} \tan^{-1} \frac{x}{d} \quad (4.7)$$

It should be noted that the above equation also assumes a constant rigidity (μ) both across the fault and with depth.

There are two unknowns in Equation 4.7: s and d . The equation scales linearly with s and the variation of the function with d is shown in Figure 4.5 which shows the near fault displacements (< 50 km) are very dependent upon the locking depth. We can also see how it is difficult to distinguish between orbital ramps, which would appear as straight lines, and the deformation signal. It is preferable to have data in the far field (> 100 km) where the deformation gradient is low and distinguishing between ramps and signal becomes easier.

4.4.2 Modelling the Manyi Fault

Using Equation 4.7 we can calculate the displacement pattern we would expect to see in a track 305 interferogram over the Manyi area (Figure 4.6). As we know the viewing geometry of the SAR satellite over track 305, we can project the motions into the satellite's line of sight. We assume that there is no component of vertical deformation across the fault as there is no significant topographic change across the fault. We take the fault geometry from the single fault model of *Funning et al. (2007b)* and extend

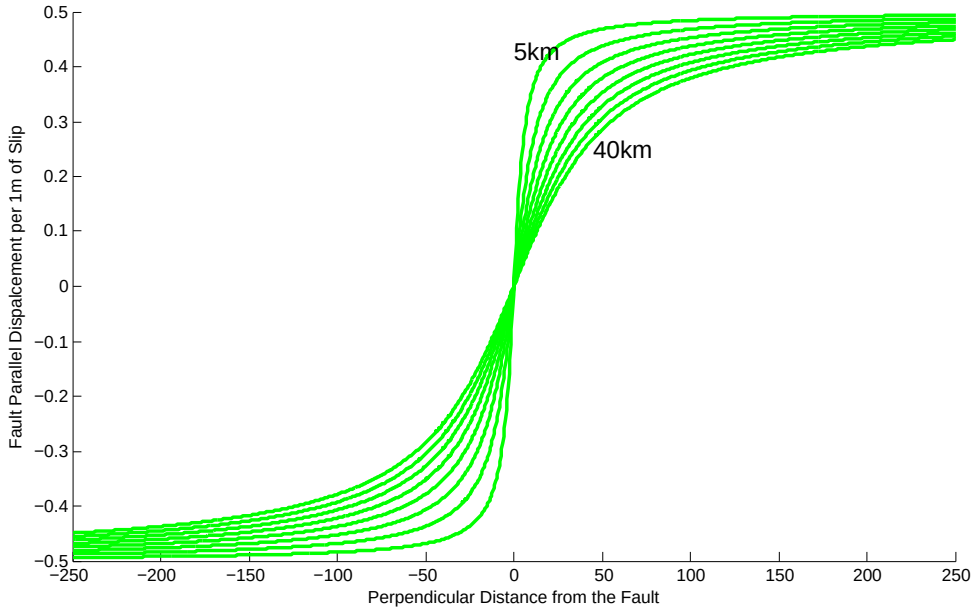


Figure 4.5: The change in length-scale of the deformation due an infinitely long vertical strike slip fault with a variable locking depth (from 5 km to 40 km). We can see how deep locking depths become increasingly more difficult to differentiate and how the interseismic signal often trade-offs off with linear tilts, especially when only the data within 100 km of the fault is available

it to 10,000 km long (to approximate an infinitely long fault) and allow slip of 1 m to occur on the buried fault. This forward model is calculated as a screw dislocation in an elastic half-space using the Okada formulation (*Okada, 1985*).

4.4.3 Inverting for a Slip Rate

To invert for the slip rate we must compare our predicted models to the data. We have removed any fault parallel orbital phase errors during the formation of the velocity map and we collapse the map onto a profile (Figure 4.7) perpendicular to the fault trace (A-A' in Figure 4.3). We subdivide the LOS velocity map (Figure 4.3) into 8 km width bins parallel to the fault (the covariance length used earlier) and calculate the weighted mean v and error σ_v of the rates in each bin. We also mask the model to remove the areas that are incoherent in the ratemap. We collapse the masked model onto the same profile. The mean of the model is not weighted as we assume there is no correlation between pixels in the models. Although the LOS velocity map is formed on

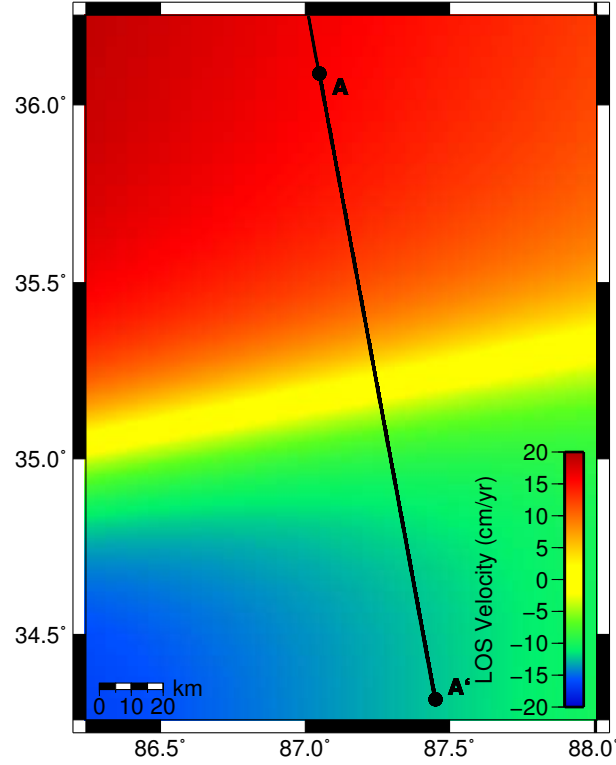


Figure 4.6: A forward model of the LOS displacement observed by an interferogram on track 305 of 1m/yr of slip across the Manyi fault. The interaction of the geometry of the fault (and thus motion across it) and the look direction of the satellite result in only 35% of the deformation being 'seen' by the satellite.

a pixel-by-pixel basis, we assume that the final values in the LOS velocity map will be spatially correlated with a covariance matrix of the same form used to estimate each epoch's atmospheric error, with σ_i replaced by σ_{p_i} ; the error on pixel i as shown in the error-map (Figure 4.4). Equation 4.3 then becomes

$$c_{ij}(\sigma_{p_i}^2 + \sigma_{p_j}^2)e^{-r_{p_i-p_j}/d} \quad (4.8)$$

We do this to ensure that pixels with large formal errors do not bias the average. We also remove any pixel where $\sigma_{p_i} > 0.8$ mm/yr as when there is a large difference between σ_{p_i} and σ_{p_j} the off-diagonal elements of the covariance matrix can become larger than the corresponding value on the diagonal. This can then cause numerical instabilities when finding the inverse.

We use the profile of the LOS velocity map along with the profiles of the interseismic

strain models to perform a weighted least-squares inversion of the slip rate assuming the LOS velocity is a result of three different components,

$$v_k = sg_k + ax_k + c \quad (4.9)$$

or in matrix form

$$b = \mathbf{A}x \quad (4.10)$$

$$\begin{pmatrix} v_1 \\ v_2 \\ \vdots \\ v_k \end{pmatrix} = \begin{pmatrix} g_1 & x_1 & 1 \\ g_2 & x_2 & 1 \\ \vdots & \vdots & \vdots \\ g_k & x_k & 1 \end{pmatrix} \begin{pmatrix} s \\ a \\ c \end{pmatrix} \quad (4.11)$$

where v_k is the mean LOS velocity calculated from the LOS velocity map profile at the point k located at a distance x_k along the profile, g_k is the Green's Function (*Savage and Burford, 1973*) that describes the surface displacement at point k in the direction of the satellite's line-of-sight (Figure 4.6), a is the magnitude of a fault-perpendicular orbital phase ramp and c is a static offset.

The above equation is not weighted according to the error (σ_{v_k}) of v_k . To weight the inversion we introduce the diagonal k by k variance matrix $\mathbf{\Sigma}_v$. The diagonal has the values $\sigma_{v_1}^2, \sigma_{v_2}^2, \dots, \sigma_{v_k}^2$ corresponding to the variance of pixels. Equation 4.9 becomes

$$x = \underbrace{[A^T \mathbf{\Sigma}_v^{-1} A]^{-1}}_{\Sigma_s} A^T \mathbf{\Sigma}_v^{-1} b \quad (4.12)$$

We make no attempt to remove topographically correlated atmospheric errors (*Elliott et al., 2008*) as there is limited topographic variation across the profile (Figure 4.1b).

4.4.4 Estimating the Magnitude of Errors

The least-squares inversion for slip rate provides formal error estimates Σ_s for the slip rate and the orbital ramp parameters. These are not realistic due to the non-linear relationship between slip and depth and also they do not account for the influence of the major source of error in InSAR studies: water vapour. To find more realistic values and uncertainties we use Monte Carlo simulations of atmospheric delay errors (*Wright et al.*, 2003). We use the statistics from the 1-D covariance function of the LOS velocity map calculated previously to create 1000 sets of simulated noise which is added to our LOS velocity map. The best-fitting slip rate and locking depth for each perturbed data set is determined by a parameter search across slip rate/locking depth space.

The best-fitting solution, with a slip rate of 2.8 mm/yr and a locking depth of 22 km (Figure 4.7), was the average of the results from each of the 717 inversions that did not hit the boundaries of the parameter space used. The parameter space bounds were 5-40 km for locking depth and -1-9 mm/yr for slip rate.

A slip rate of 2.8 mm/yr corresponds to a far-field LOS velocity of $\sim \pm 0.5$ mm/yr due to the geometry of the look direction of the satellite (shown in Figure 4.1a) and the strike of the fault. Given that 283 perturbed solutions hit the boundaries of the locking depth parameter space (but never the slip rate) it is clear the locking depth is not well constrained. Figure 4.8 shows the best-fitting models for each of the perturbed datasets. We use principle component analysis to calculate the orientation and size of an ellipse around the mean which contains 67% of the 717 perturbed solutions. This confidence ellipse gives estimates of the error in slip rate and locking depth of 1.8 mm/yr and 15 km respectively. We therefore estimate that the Manyi fault, prior to the 1997 event, was accumulating interseismic strain at a rate of 3 ± 2 mm/yr with a locking depth of 22 ± 15 km.

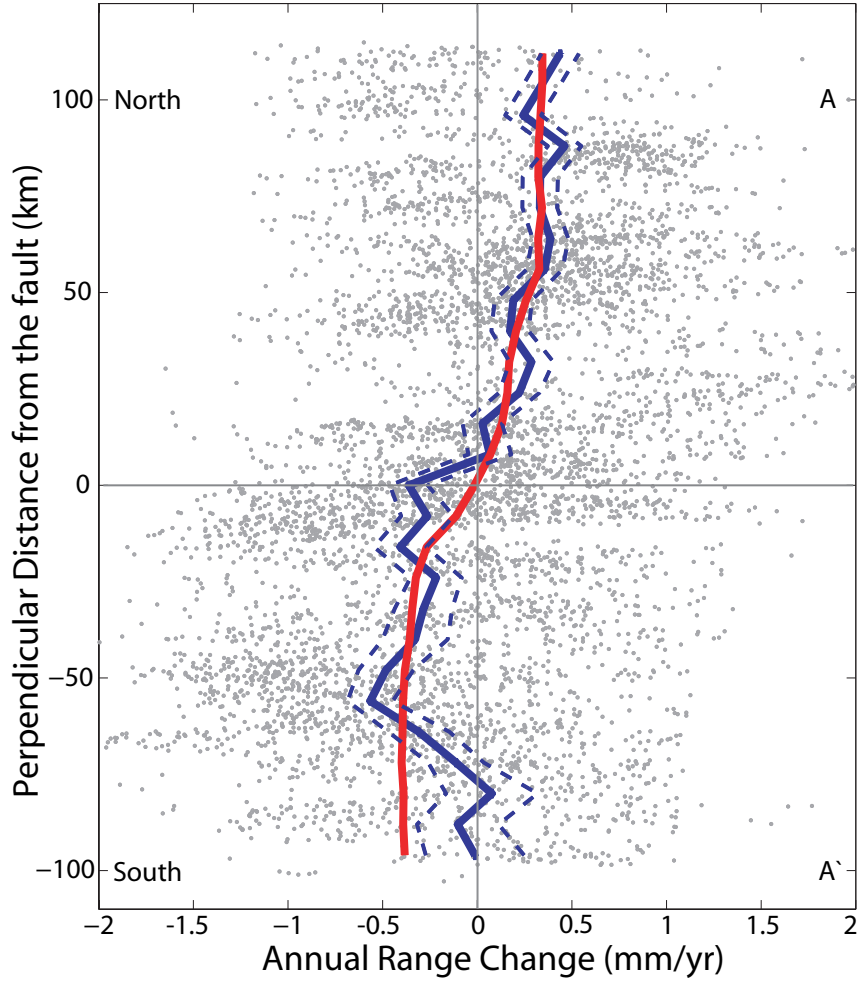


Figure 4.7: Profile across the Manyi fault of the LOS velocity map. The blue line is the weighted mean with dashed lines indicating the extent of the 2σ error. The black dots show the rate at each pixel once they have been corrected using the orbital ramp estimated in the inversion. The red line shows the best-fitting model (star in Figure 4.8)). Due to the geometry of the fault and the satellite look direction, 1 mm/yr of slip across the fault results in 0.35 mm/yr far-field LOS velocity.

4.4.5 Possible Extension of the Kunlun Fault

We have assumed that the only deformation observed in the LOS velocity map is due to motion across the Manyi fault, however the area covered by the map of LOS velocities also contains another mapped fault (*Taylor and Yin, 2009*) that could possibly be considered as the western extension of the Kunlun fault. Using our estimate of the slip rate across the Manyi fault we attempt to model any residual signal that could be due to slip on this fault. We remove the expected deformation from the our best-fitting

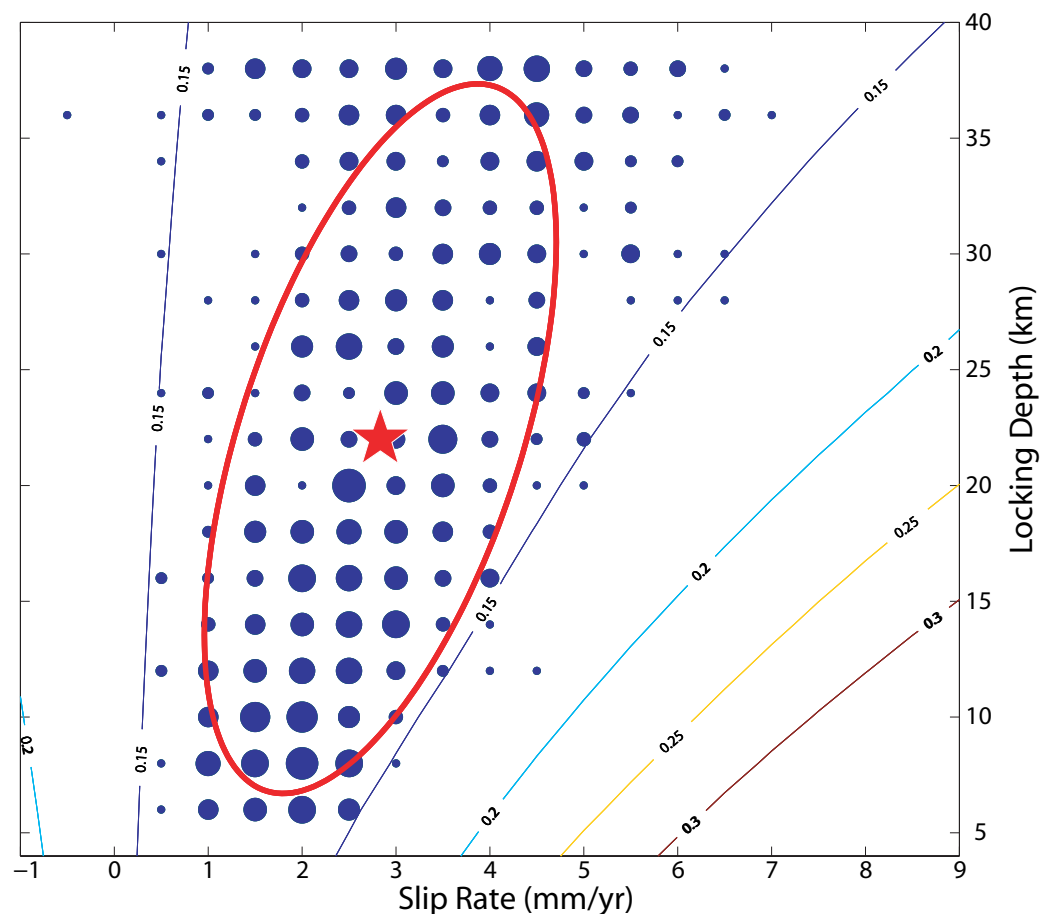


Figure 4.8: Slip rate-locking depth parameter used to determine the best-fitting solution to each perturbed data set for the Manyi fault. The blue dots show the grid points in parameter space that were used, with their area proportional to the number of solutions that fell on that point in the space, with the largest dot representing 17 solutions and the smallest 1. The contours show the RMS misfit (in mm) of the solutions at each point in the parameter space. The mean solution is shown by the red star. The red line indicates the location of the 67% confidence ellipse and shows the uncertainties in the average slip rate/locking depth.

model from the Manyi fault (Figure 4.7) and perform the same inversions as for the Manyi fault, only now with a profile taken perpendicular to the second fault. When inverting for both slip rate and locking depth, the solutions are unstable as over 900 of the 1000 solutions hit the bounds of locking depth. This is due to a lack of data to the north of the fault combined with the small magnitude of line-of-sight deformation. When we only invert for slip rate across this fault and fix the locking depth to that found for the Manyi fault (22 km), finding a best-fitting rate of -0.1 ± 0.2 mm/yr. We obtain the same results regardless of whether we solve first for the Manyi fault or the

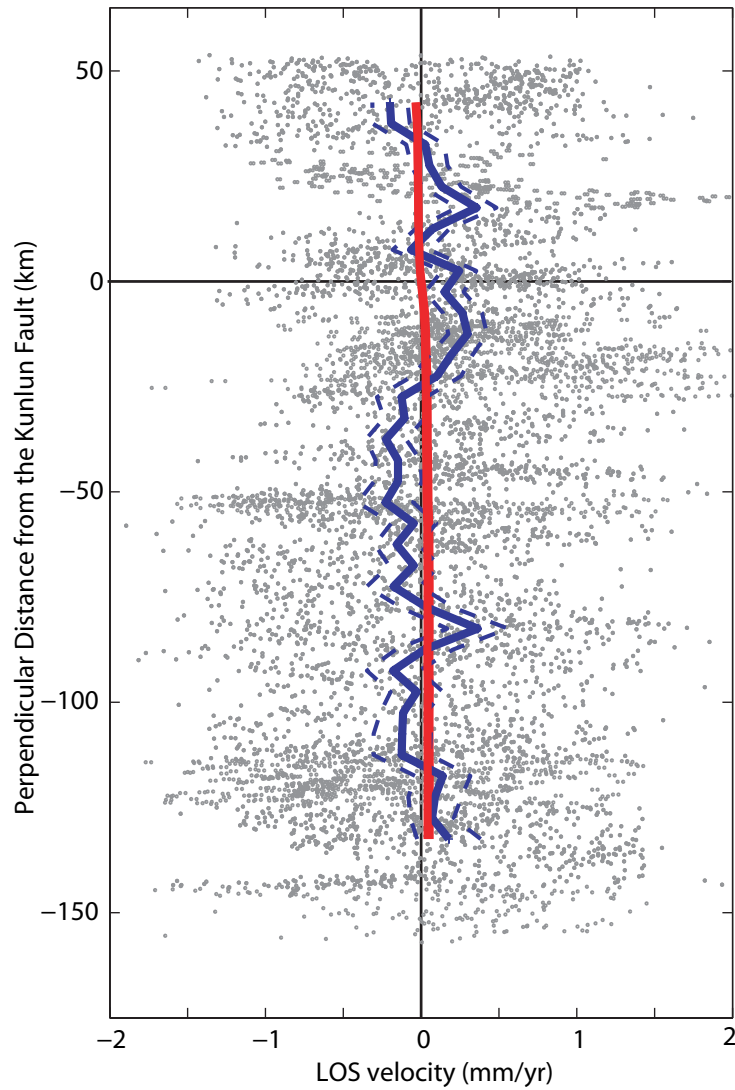


Figure 4.9: Profile across the Kunlun fault of the best-fitting solution to the data. The red line is model, blue is the weighted mean with the dashed blue lines indicated the location of the 2σ errors bounds. The black dots show the rate at each 800m pixel once they have been corrected using the orbital ramp calculated

'Kunlun' fault.

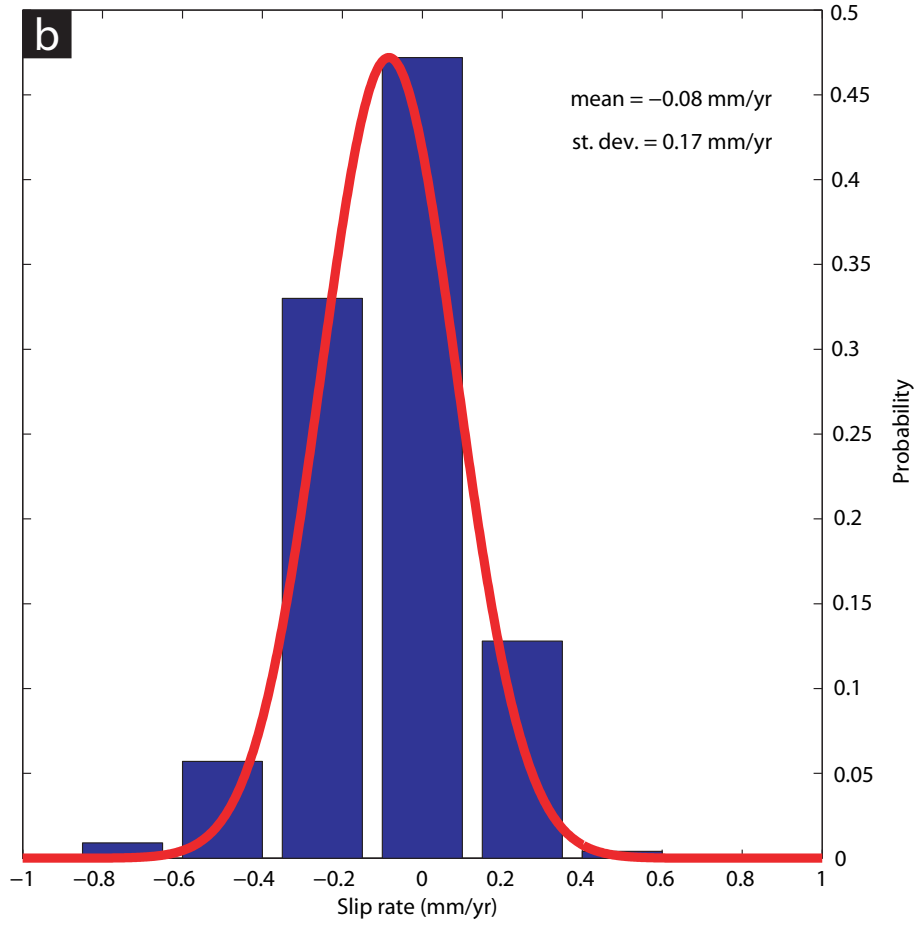


Figure 4.10: Histogram of the solutions to the 1000 perturbed dataset for slip rate on the extension of the Kunlun fault. The solutions are placed into 0.25 mm/yr bins and the blue bars indicate the probability of any one of the 1000 solutions falling into the bin. The red line shows a Gaussian distribution using the mean and standard deviation of the 1000 solutions.

4.5 Discussion

We estimate the rate of relative motion across the Manyi fault prior to the 1997 earthquake to be 3 mm/yr and near zero across the possible extension of the Kunlun fault at this longitude. The rate for the Manyi fault is about one third the rate of the plateau bounding Altyn-Tagh fault to the north (*Elliott et al.*, 2008) and main portion of the Kunlun fault to the east (*Zhang et al.*, 2004). Assuming the 1997 Manyi earthquake is representative of earthquakes on the Manyi fault, the average coseismic slip is 4 ± 1 m (*Funning et al.*, 2007b), and relative motion across the fault of 3 ± 2 mm/yr is constant over the whole interseismic cycle, our results indicate a recurrence time of 1300^{+3700}_{-700} yrs.

The slip estimate obtained here is one of the lowest slip rates determined by InSAR in the literature. Although the seismic hazard presented by this earthquake is small due to its remoteness, techniques such as those used here could be used to study the slip rates of faults in close proximity to dense urban areas that represent a significant seismic hazard.

Many Monte Carlo solutions for both the Manyi and 'Kunlun' faults hit the boundary limits of 0 and 40 km locking depth during the inversions. Our estimates of locking depth are therefore not well constrained. However, the preferred value of 22 km is in close agreement with the maximum depth of slip in the 1997 Manyi earthquake (20 km), as is to be expected if earthquakes occur in the seismically locked part of the crust. It is unsurprising the locking depth is poorly constrained as we do not have a large amount of data in the far-field due to data limitations. Far field data can be used to flatten the data before the inversions reducing the degree of freedom in the inversion.

4.5.1 Implications for Postseismic Deformation

Maps of the current LOS velocities around the Manyi Fault area on track 305 show LOS velocities of up to 8 mm/yr (*Bell et al.*, 2010). The LOS displacement rate expected if we assume the fault is still accumulating strain at the rate found in this study is only ~ 1 mm/yr, suggesting that, even 10 years after the 1997 event, the current deformation is likely to be predominantly postseismic. To model frictional afterslip, one of the major mechanisms for postseismic motion, an estimate of the interseismic rate prior to the event is useful. The interseismic rate (V_i) provides a lower bound to the pre-earthquake velocity and a constraint on the upper bounds of the pre-earthquake velocity (*Perfettini and Avouac*, 2007; *Barbot et al.*, 2009) which can help refine the range of viable afterslip models. We discuss this further in the next chapter which models the afterslip expected due to the Manyi Earthquake.

4.5.2 Deformation Models of Tibet

Reliable rates for the Altyn Tagh, Kunlun, Xianshuihe and Jiali Faults are typically 10 mm/yr (*Searle et al.*, 2011). A marked decrease in the slip rate is observed towards the ends of some of these more comprehensively measured strike slip faults, e.g the eastern Altyn Tagh fault (*Zhang et al.*, 2007) and the eastern Kunlun (*Kirby et al.*, 2007), although care must be taken when comparing slip rate estimates obtained from different techniques as they can have different timescales. Our observations support either a decay in slip rate also towards the western end of the Kunlun fault, or slip that has been distributed across a number of surrounding faults. In either case, our observations do not support a constant slip boundary model required by quasi-rigid block or micro-plate rotation (*Thatcher*, 2007; *Meade*, 2007). Block models require high slip rates concentrated on few faults. One of the major faults used in block modelling goes through the area covered in this study (Figure 4.11). Although the slip need not be on the Manyi Fault that is studied in detail here, it should be on one of the other faults in the study area. Such a slip rate would produce an obvious signal in the data which we do not observe.

4.5.3 Model Limitations

Throughout this study we have assumed that the medium accommodating the inter-seismic deformation is elastic. However *Ryder et al.* (2007) have modelled the rapid postseismic deformation in terms of viscoelastic relaxation. Simple interseismic strain models with an elastic layer over a viscoelastic half-space (*Johnson and Fukuda*, 2010), where relaxation times are much smaller than the earthquake repeat time, show a quasi linear variation of displacement across the fault, where we see deformation that is localised. This would suggest that the rapid rates of motion immediately after the Manyi earthquake are better explained by afterslip at depth (*Ryder et al.*, 2007). More complex viscoelastic models with larger viscosities in the lower crust and upper mantle (*Johnson et al.*, 2007), retain localised deformation throughout the earthquake cycle.

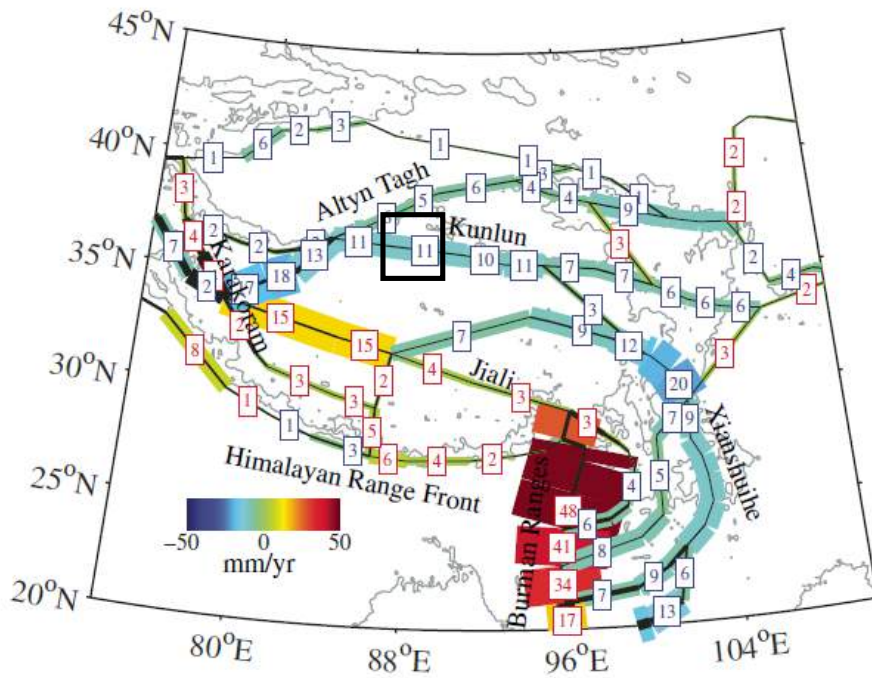


Figure 4.11: Adapted from *Meade (2007)*. The solid black box indicates the approximate area of this study. The coloured numbers show the predicted slip rates across the faults that are also marked in colour, the colour indicated the slip rate. An 11mm/yr slip rate is clearly seen to run straight through the area of study, something that would show up clearly in our data study

They show that using elastic models to describe viscoelastic behaviour could underestimate the long term slip rate across the fault by up to a factor of two. In this case the slip rate across the Manyi fault could be as high as 6mm/yr and have a repeat time half that estimated here. We discuss one such model in the discussion chapter of this thesis.

Chapter 5

Afterslip Modelling of the Manyi Earthquake

Afterslip was one of the preferred mechanisms to explain the postseismic deformation of the Manyi earthquake by *Ryder et al.* (2007). The models used in *Ryder et al.* (2007) were kinematic as the slip distribution was determined from the data itself. As the models are derived from the data there are few geophysical constraints to ensure the slip distribution is physically plausible. To derive a physically plausible afterslip model rate and state friction laws must be observed. In this chapter we model the afterslip that occurred after the 1997 Manyi earthquake using a series of rate and state models. We use the hot friction approximation of the rate and state equations that relate the stress field with aseismic slip to determine the patterns of slip expected for a variety of frictional parameters on the fault. We compare these models to the ERS data from *Ryder et al.* (2007) and the ENVISAT data from Chapter 3 to determine whether the afterslip models can explain the deformation patterns observed after the Manyi event.

5.1 Introduction

Afterslip was first documented in California, after the 1966 Parkfield earthquake (*Smith and Wyss*, 1968). It was then observed in the aftermath of other large events such

as the 1968 Borrego Mountains Earthquake (*Burford, 1972*), the 1979 Imperial Valley Earthquake (*Harsh, 1982*), and the 1987 Superstition Hills Earthquake (*Bilham, 1989*). These early studies showed that afterslip is quite distinct from seismic slip (*Scholz et al., 1969*) as the slip accumulates linearly with the log of the time and crucially, without any seismic signature. Afterslip is often thought to be a stable (aseismic) equivalent of aftershocks as both follow a similar logarithmic decay in energy release with time after the event (for aftershocks this trend was first described using the Omori Law (*Omori, 1894*)). As the behaviour is consistent across a number of different faults, afterslip represents a physical process fundamental to fault zones. Early attempts to understand the afterslip mechanism focused on shallow creep events (*Scholz et al., 1969; Crough and Burford, 1977; Wesson, 1988*) using rheologies that were based on plasticity. An anti-correlation was noted between the location of coseismic and postseismic slip, with the negligible postseismic slip in regions of high coseismic slip and vice versa (*Burford, 1972; Williams and Magistrale, 1989*). Although some of the models fit the data, they required unrealistic assumptions such as obeying different constitutive laws during the coseismic and postseismic phase. A much more fruitful avenue of investigation to explain afterslip lay within the realms of fault friction.

5.1.1 Stick-slip Behaviour

Rock sliding experiments by *Brace and Byerlee (1970)* observed stick-slip behaviour, with periods of no motion between the blocks followed by episodes of slip, and they used it to describe the behaviour of earthquakes. To describe frictional resistance with dynamic instabilities, followed by periods of no motion while the stress increases again, we can use an analogue model of a simple frictional spring-slider (Figure 5.1). In such a model the spring is taken to represent the stiffness of the loading machine in the laboratory apparatus or the elastic properties of the medium in which the fault is embedded.

To demonstrate how regular stick-slip behaviour can occur, we consider a case where the frictional force on the block F_r , is shown by the blue line in Figure 5.2. At point

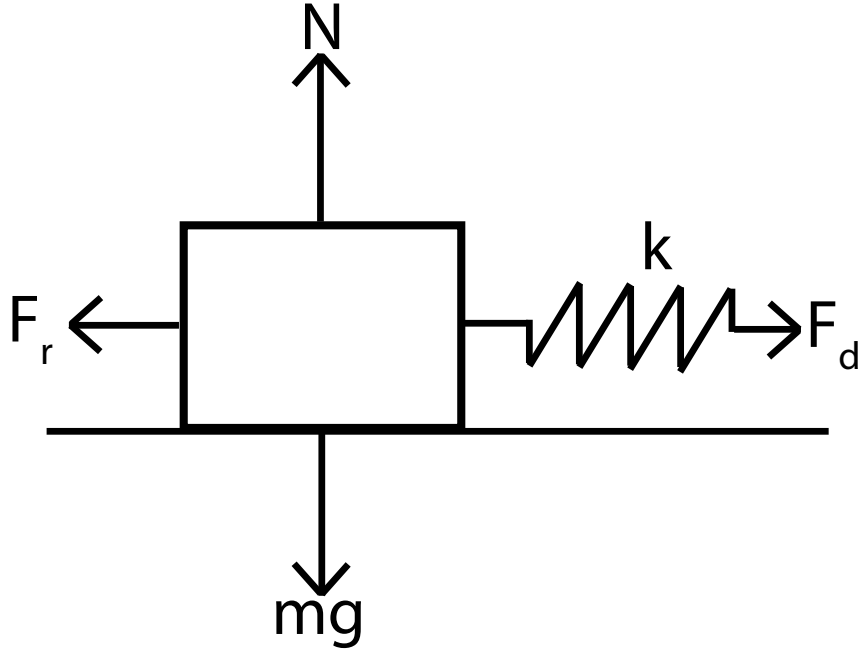


Figure 5.1: Frictional Spring Slider model used to model the stick-slip behaviour of faults. k represents the stiffness of the system F_d the loading force, N the normal stress, mg is the weight of the block and F_r is the resistive force. The behaviour of this model is described in the text.

As the force of the spring unloads, following the red line of slope k as F_r decreases faster than F_d). This produces a net force and hence acceleration of the slider. Once point B is reached F becomes the larger force, resulting in the deceleration of the slider until it stops at point C. If we assume no energy is lost during the instability, the pink areas between A-B and B-C are the same. We can therefore state that the condition required for an instability depends on the elastic properties of the medium and the friction properties of the contact and can be written as

$$\frac{\partial F_r}{\partial x} > k \quad (5.1)$$

The frictional behaviour shown in Figure 5.2 where frictional strength decreases with slip is known as slip-weakening behaviour. Slip weakening behaviour is observed when the static friction (μ_s) coefficient is greater than the dynamic friction coefficient (μ_d). For regular stick slip behaviour there must be a mechanism by which the

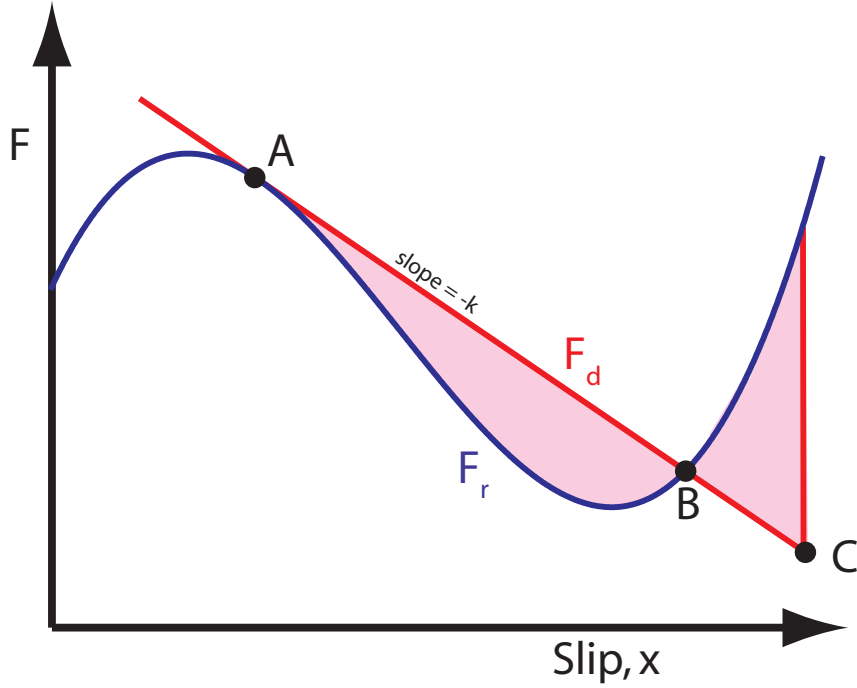


Figure 5.2: Force-displacement diagram of a theoretical force curve that is applied to block-slider system. In this case the frictional force falls more rapidly with slip than the spring system allows causing rapid accelerations.

dynamic friction eventually returns to the static value

Rabinowicz (1951) was the first to notice that when two surfaces are held together in contact, then μ_s increases with the logarithm of their time in contact t . He also found that there is a critical distance, D_c that must be exceeded when moving two surfaces against each other in order to change from the static to dynamic friction. These observations allow us to restate Equation 5.1 as, where σ_n is the normal stress.

$$\frac{(\mu_s - \mu_d) \sigma_n}{D_c} > k \quad (5.2)$$

This behaviour is observed in systems where μ_d decreases with the logarithm of the sliding velocity V , i.e velocity weakening. Instabilities in systems that are velocity strengthening (dynamic friction at steady state is higher than static friction) are quickly damped back to stable sliding.

To describe regular stick-slip behaviour of rocks in both velocity weakening and

strengthening zones *Dieterich* (1979, 1981); *Ruina* (1983) and *Rice* (1983) devised the slip-rate and state-variable frictional laws. The laws are empirical in nature and are based on laboratory experiments, so one must take care when extrapolating them from the lab scale (millimetres) to real faults (hundreds of kilometres). Despite these problems, they have been capable of reproducing a range of observed fault behaviours such as earthquake nucleation, earthquake rupturing, postseismic aseismic afterslip, depth of seismicity and the coupling at subduction zones.

Nowadays the popular rate and state laws are known as the Dieterich-Ruina law (Equation 5.3) which states that friction is a function of the sliding velocity V and a state variable θ which is time dependent (Equations 5.4 and 5.5).

$$\mu(V, \theta) = \mu_0 + a \ln \left(\frac{V}{V_0} \right) + b \ln \left(\frac{V_0 \theta}{D_c} \right) \quad (5.3)$$

where the state variable θ evolves according to either the aging law (*Dieterich*, 1981),

$$\dot{\theta} = 1 - \left(\frac{V \theta}{D_c} \right) \quad (5.4)$$

or the slip law (*Ruina*, 1983).

$$\dot{\theta} = -\frac{V \theta}{D_c} \ln \left(\frac{V \theta}{D_c} \right) \quad (5.5)$$

Although the two state evolution laws are fundamentally different in terms of their mechanical interpretation, they both describe the same laboratory observations well. Because the most appropriate evolution law is still uncertain (*Perfettini and Ampuero*, 2008), we will use the aging law as it is simpler.

In the case of static friction using the aging law, $\theta \approx t$, meaning θ can be approximated as the average age of the contacts. If, at a constant velocity V , the critical slip distance D_c is the distance through which a set of contacts is completely refreshed with a new set we can write

$$\theta = t = \frac{D_c}{V} \quad (5.6)$$

If we define μ_0 as the friction at some steady-state reference velocity V_0 we can write the friction at steady state as

$$\mu_{ss} = \mu_0 + (a - b) \ln \left(\frac{V}{V_0} \right) \quad (5.7)$$

Figure 5.3 (*Scholz, 2002*) illustrates the behaviour of an instantaneous increase in the load velocity followed by an instantaneous decrease (back to original velocity). The figure shows a sudden jump of magnitude a when the loading velocity increases. The friction then evolves to a new steady state value over the length-scale D_c which is b lower than the peak friction. A velocity reduction then causes a decrease in the friction, followed by a $1/e$ increase towards the steady state at low velocity. If the steady state friction when moving at a fast velocity is larger than the slow velocity (i.e. $a - b > 0$), the material exhibits velocity strengthening and it is energetically unfavourable to continue to rupture. It is impossible to nucleate earthquake ruptures in such regimes. If $a - b$ is negative, the friction is lower when moving at high velocity and instabilities can occur. Such conditions can cause earthquakes to nucleate and propagate.

5.1.2 Early Studies of Afterslip

Afterslip is defined as slip on a discrete fault or narrow fault zone which is observed directly on the surface or inferred via geodetic InSAR and GPS measurements of ground deformation.

The afterslip observed after four extensively studied strike-slip earthquakes is shown in Figure 5.4. The motion is measured using fiducial markers installed near the faults in the hours or days following the event. The magnitude of slip can vary greatly, with typically rapid rates initially, decaying logarithmically with time. The signature logarithmic decay is only observed for three events however, as there is negligible afterslip observed following the 1992 Landers Earthquake. The positions of the markers near the faults mean their relative positions could only be affected by motions near

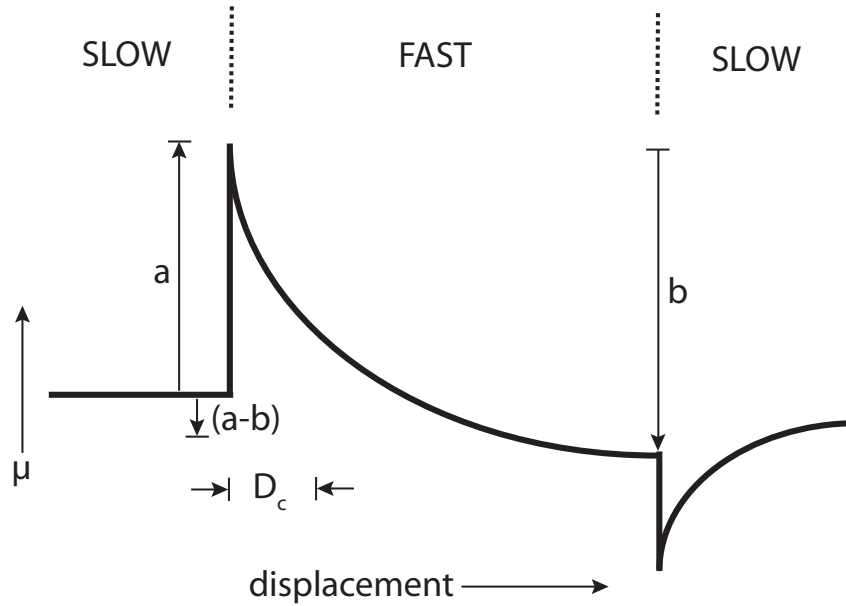


Figure 5.3: Friction versus displacement observed when a kick in loading rate is applied to a block. In this case $(a - b) < 0$ i.e velocity weakening. Based on *Scholz* (2002).

the surface. Shallow slip deficits were observed at all the events with the exception of Landers. The observed afterslip at the events in Figure 5.4 was a response to the shallow slip deficit. As Landers had no shallow slip deficit it had no afterslip.

Marone et al. (1991) introduce a scenario (Figure 5.5) that can be used to connect the observations of afterslip to the background theory of rate and state friction. The model has a wide zone of unconsolidated fault gouge down to depth h and lithified mature gouge below to a depth H , the base of the seismogenic layer. The figure shows the coseismic slip distribution from a typical buried strike slip fault. There is a slip deficit above depth h consistent with observations of many coseismic slip distributions (*Fialko et al.*, 2005). The variation of $a - b$ is also shown on the figure. Two transitions are indicated, one at h due to the change from unconsolidated gouge to consolidated and the other at H , due to the increase in temperature (*Scholz*, 2002).

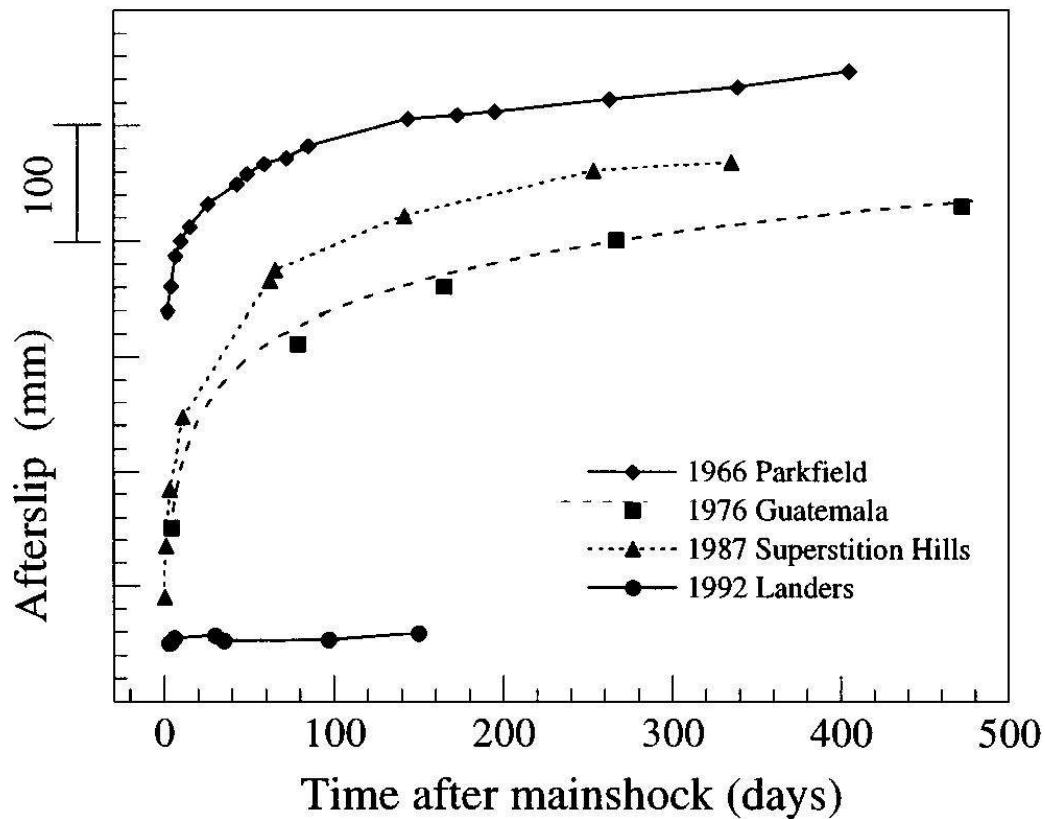


Figure 5.4: Observed motion following the 1996 Parkfield, 1976 Guatemala, 1987 Superstition Hills and 1992 Landers Earthquakes. Data show triangulation data of one point taken from *Smith and Wyss* (1968); *Bucknam et al.* (March, 1978); *Williams and Magistrale* (1989); *Sylvester* (1993) respectively. Of the four events, all but Landers show significant afterslip, often with the magnitude of afterslip approaching the magnitude of coseismic slip at the observation point. Landers notably does not exhibit any discernible postseismic motion. It should be noted that start of the four lines are offset vertically for visualisation purposes. Based on Figure 13, *Marone* (1998).

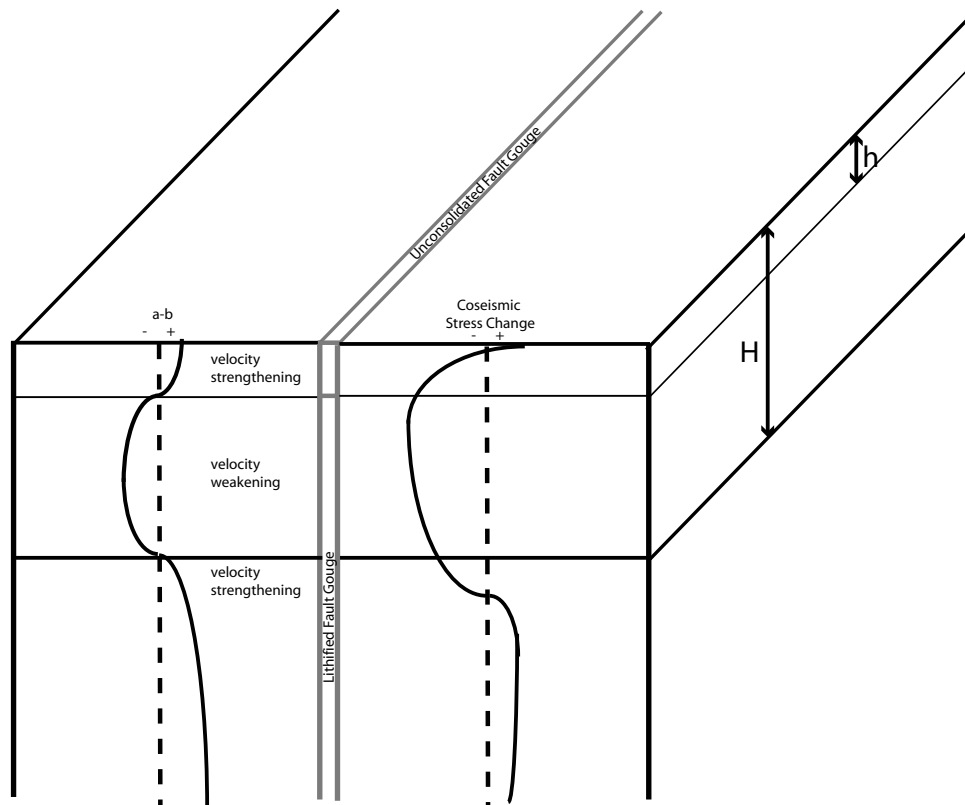


Figure 5.5: Model of a mature fault zone illustrating the relationship between the frictional and seismic properties. The curves show typical stress change associated with a Manyi-like earthquake and the variation in $a - b$ with depth. H denotes the thickness of the seismogenic layer and h denotes the thickness of the unconsolidated layer at the surface of a fault. Based on Marone *et al.* (1991).

5.2 Afterslip Modelling

5.2.1 Rate and State Friction Models

Earthquakes perturb the stress field in the region of a fault. This stress field then changes as postseismic slip (or flow) occurs. The rate-and-state friction laws can relate the shear stress change due to an event on a fault to the normal stress across it (Equation 5.8). This relationship has been frequently used to model seismic rupture propagation, slow slip events and frictionally controlled afterslip (*Scholz, 2002*). The equation can be parametrised such that the slip is either unstable (earthquakes) or stable (afterslip).

$$\tau(t) = \sigma_n \left(\mu_0 + a \ln \left(\frac{V(t)}{V_0} \right) + b \ln \left(\frac{\theta(t)V_0}{D_c} \right) \right) \quad (5.8)$$

where a , b , V_0 and μ_0 are frictional parameters from Equation 5.3 and τ and σ_n are the shear and normal stress respectively. θ is given by the aging law in Equation 5.4.

To model the frictional afterslip we use the approach of *Perfettini and Avouac* (2007) and assume that the afterslip is caused by brittle creep on a discrete fault plane in a pure velocity-strengthening regime. We also assume the sliding is stable (i.e the interface is not locked and allowed to heal). Therefore the following condition is true

$$\frac{d\theta(t)}{dt} = 0 \quad (5.9)$$

which means Equation 5.8 becomes

$$\tau(V_0) = \sigma_n \left(\mu_{ss} + (a - b) \ln \left(\frac{V_0}{V_{ss}} \right) \right) \quad (5.10)$$

This model is often referred to as the hot friction model, (*Linker and Rice, 1997*), and is used in other afterslip studies such as *Hearn et al.* (2009) and *Johnson* (2006). If we impose a static stress change at $t = 0$ that is produced by an earthquake, Equation 5.10 can be rewritten as the equation

$$V = V_{pre} e^{\left(\frac{\Delta\tau}{\sigma_n(a-b)}\right)} \quad (5.11)$$

where V_{pre} is the pre-earthquake velocity, $\Delta\tau$ is the shear stress change induced by the earthquake, and σ_n is the normal stress (*Scholz, 2002*).

The hot friction equation is unrealistic when describing motion as $t \rightarrow \infty$ and is discontinuous around $t = 0$. It was revised by *Lapusta et al. (2000)* and *Rice et al. (2001)* to

$$V = 2 V_{pre} \sinh \left(\frac{\Delta\tau}{(a-b)\sigma_n} \right) \quad (5.12)$$

In the above form of the equation V tends towards zero as $t \rightarrow \infty$, is continuous at $t = 0$ and reproduces the same behaviour for finite t . The revision is based on energetic arguments as we assume work done by afterslip is finite hence it must stop.

5.2.2 Viscous Shear Zone Models

The second type of afterslip model we consider is one where the fault zone at depth behaves as a finite width viscous shear zone. The creep velocity V in this case is

$$V = \frac{w}{\eta} \Delta\tau \quad (5.13)$$

where w is the horizontal width of the shear zone, $\Delta\tau$ the stress change and η is the viscosity of the shear zone.

5.3 Numerical Implementation of the Models

The inputs to the afterslip models are the fault geometry and coseismic slip distribution from *Funning et al. (2007b)*. We simplified the distributed slip model geometry by making the fault straight (using the fault centre location and strike of the single fault model) and vertically dipping. We do this primarily to improve efficiency. It is possible for afterslip to occur on lateral extensions of the coseismic fault plane but we

do not extend the model fault plane at its ends as we have no data in this area and also the slip at the edge of the fault planes is negligible. We do extend the fault plane with depth down to 40km but no further as any slip below 40 km manifests itself as a long wavelength ramp which would be removed during the orbital correction process, as shown in *Ryder et al. (2007)*. As the crust is velocity weakening above approximately 12 km depth (*Scholz, 2002*) the shallowest depth of the (afterslip) fault plane is set to 12 km. Although the top 4 km immediately below the surface is also believed to be velocity strengthening (Figure 5.5) our model does not allow afterslip to occur here for two reasons. Firstly, the deformation from very shallow fault patches will produce short wavelength deformation close to the fault. As the data processing requires significant smoothing, it is likely any such deformation will be adversely affected by the smoothing. Secondly, there is no significant slip deficit at the surface (Figure 3.6) so any deformation is likely to be relatively low magnitude.

The fault plane is discretised into 4x4 km patches (as in the coseismic slip model), resulting in N fault segments (in this case $N = 329$).

For each fault patch i we can state that

$$\tau(i, t) = \tau_{pre}(i) + \tau_0(i) + \tau_{aft}(i, t) \quad (5.14)$$

where $\tau_{pre}(i)$ is the initial, preseismic stress at plane i , which we assume to be zero, $\tau_0(i)$ is the stress change due to the earthquake, $\tau_{aft}(i, t)$ is the stress induced on plane i by afterslip on the whole (afterslipping) fault plane i.e from all fault patches.

$\tau_{aft}(i, t)$ is not only due to the slip that occurs on patch i but also all the surrounding patches of afterslip, and hence it can be rewritten as

$$\tau_{aft}(i, t) = \sum_{q=1}^N K(i, q) s(q, t) \quad (5.15)$$

where K is an elastic kernel the components of which represent the horizontal shear stress change on patch i due to a unit (horizontal) slip at patch q . $s(q, t)$ represents the afterslip at patch q and time t .

5.3.1 Initialising the Model

We assume that there is no significant preseismic stress on the extended fault as it was stably sliding during the interseismic period. Using this assumption, we combine Equations 5.12 and 5.14 at the time of the earthquake to state the afterslip velocity at any time t after it

$$V(i, t) = 2 V_{pre}(i) \sinh \left(\frac{\tau_0(i) + \tau_{aft}(i, t)}{(a - b) \sigma_n(i)} \right) \quad (5.16)$$

As we know the stress change from the earthquake at any point using the coseismic slip distribution and that $\tau_{aft}(i, 0) = 0$ at the time of the earthquake, we can calculate the initial postseismic afterslip velocities at any point. To work out the temporal evolution of the stress field we discretise the velocity into $\Delta s / \Delta t$ where Δs is the slip that occurs during (user defined) time Δt . For the first time step, Equation 5.16 can be written

$$\Delta s(i, \Delta t) = \Delta t V_{pre}(i) \sinh \left(\frac{\tau_0(i)}{(a - b) \sigma_n(i)} \right) \quad (5.17)$$

The slip ds is then substituted into Equation 5.15 to work out the perturbation to the stress field. This perturbed stress field is used to calculate the velocity for the next time step. This process is continued until the slip has been calculated for the specified time, in this case 1150 days. Other assumption that are necessary during the calculations of the slip which are:

1. No afterslip occurs (either pro-grade or retrograde) on patches where the stress change from the earthquake is negative. Areas where the stress change is negative are regions of high coseismic slip which we assume to be velocity weakening.
2. V is constant over the time interval dt for which the afterslip is calculated. As some models have a rapidly varying V immediately after the earthquake, we initially use a time step of 1 sec. Using this time step however would be prohibitive when attempting to calculate the slip for 1150 days. We therefore vary the time

- step from 1 sec (for one minute) to 1 minute (for one hour) to 1 hour (for one day) to 1 day (for 1150 days). This drastically reduces the run time of the modelling but keeps fine temporal resolution initially after the earthquake.
3. The change in normal stress due to afterslip ($\approx 10^7$ Pa) is ignored as it is insignificant in comparison with the normal stress from lithostatic pressure ($\approx 10^9$ Pa).
 4. Velocities are clipped to 0.3 m/s. In the low friction models there are patches where $\Delta\tau \gg (a - b)\sigma$. This produces unrealistically high velocity values and the models become unstable. At these velocities, dynamic effects become significant and produce seismic events. The threshold velocity (V_{dyn}) is given by $(2a\sigma c_s)/G$ where c_s is the shear wave velocity (*Rice et al.*, 2001) and G is the shear modulus. In the model with low friction V_{dyn} can be as low as 0.3 m/s, hence this provides a maximum threshold.
 5. We only use the horizontal component of shear parallel to the fault trace. We therefore neglect the vertical stress changes (and thus motions) which is justified as we do not include any gravitational effects in our modelling.

5.3.2 Afterslip Distribution to Ground Deformation

An afterslip distribution was calculated for each of the 1150 days after the Manyi event. Using the Okada formulation this slip distribution was converted to a three component (u_x, u_y, u_z) map of surface deformation at each of the time intervals in the *Ryder et al.* (2007) track 305 time series, i.e 25, 60, 90, 130, 165, 270, 305, 725 and 1145 days. Using the line-of-sight vector from track 305 we converted the deformation components into a deformation map in the line-of-sight of the satellite that is equivalent to the time series from *Ryder et al.* (2007). It should be noted that the original time series from *Ryder et al.* (2007) extrapolated back to immediately after the earthquake assuming the motion over the first 25 days was linear. Given the logarithmic behaviour exhibited by afterslip we do not assume we know anything about the first 25 days of motion,

and use the deformation at 25 days as the start of the time series when comparing the interferograms.

5.4 Modelling Results

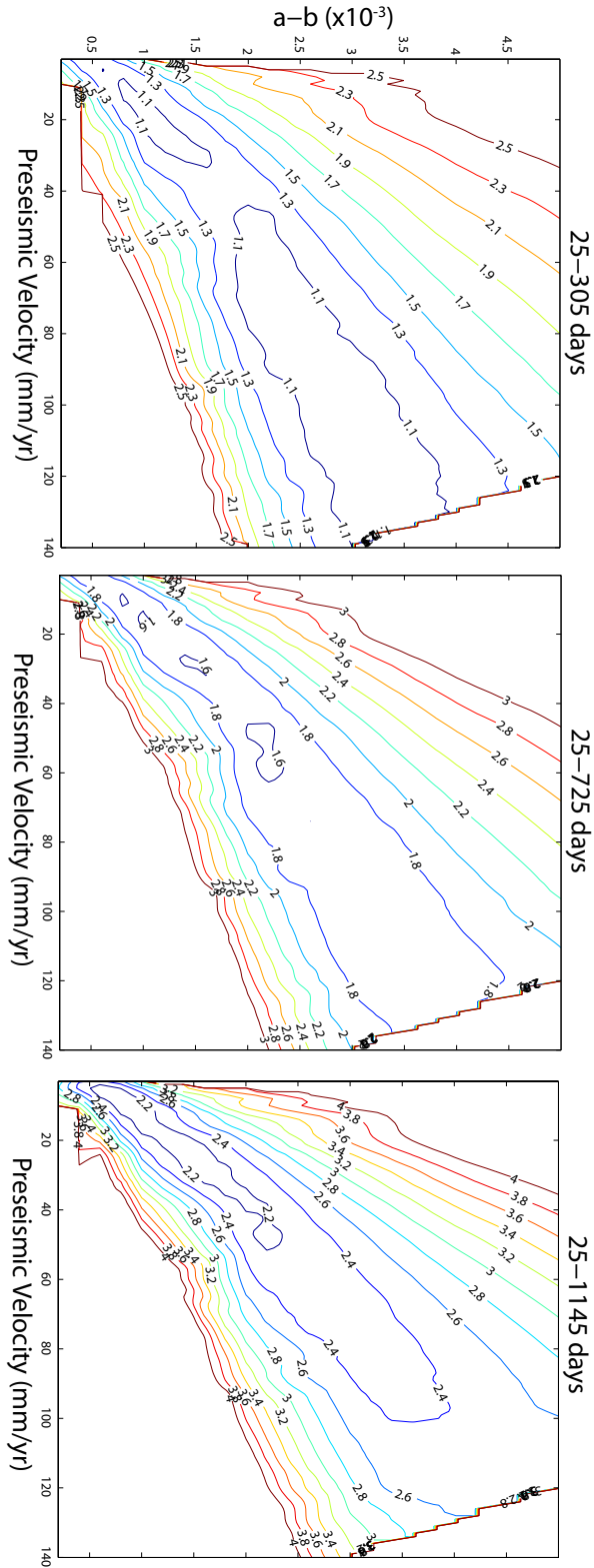
Each model is converted to an interferogram at each of the times points detailed above. The model is then subtracted from the time series and the residuals calculated. We must ensure however that any residual tilts are addressed. Tilts were removed from the initial dataset by *Ryder et al.* (2007) using data in the far-field. This can be problematic as the majority of models show some deformation on the edges of the data. We therefore remove a best-fitting linear ramp from residuals of each model at each time in the time series. This flattening uses only the data in the top and bottom 10km of the scene; this is sufficient to characterise any linear trends and does not contain any of the high (spatial) gradient deformation signals.

5.4.1 Frictional Afterslip Results

We run a series of models each with a different a and V_{pre} to find the best-fitting model to the data. The interseismic rate of 3 mm/yr determined in the previous chapter is used as a lower bound of V_{pre} . The upper bound of 140mm/yr is used as previous studies (*Perfettini and Avouac*, 2007) have found V_{pre} to be an order of magnitude higher than the interseismic rate. The frictional parameter, $(a - b)$, is varied between 2×10^{-4} and 5×10^{-3} based on ranges found in the literature (*Scholz*, 2002).

To find the best-fitting model (Figure 5.6) we compare the RMS residuals of the cumulative displacement at each time point summed over the first 725, 305 days and 1145 days for a series of models.

The RMS minimum does not correspond to a single best-fitting model; they show a series of best-fitting models in which $(a - b)$ trades off with V_{pre} . Table 5.4.1 shows a selection of models that lie within the RMS minima in Figure 5.6. These models are compared in the remaining figures of this chapter.



V_{pre} (mm/yr)	(a-b)
8	8×10^{-4}
20	1×10^{-3}
30	1.5×10^{-3}
50	2×10^{-3}
120	3×10^{-3}

Figure 5.7: The parametrisation of afterslip models that lie within the V_{pre} and $a - b$ trade-off. These models are compared in this chapter.

Profiles perpendicular to the strike and along the centre of the deforming region including all the data within a 20 km wide swath (Figure 5.9) are taken from each of the models displayed in Figure 5.8

For all models shown in Figure 5.9 there is little variation between the models within the error of the data (± 1 cm from *Ryder et al. (2007)*) with the exception of the 120mm/yr model when $t > 305$ days, which is expected from the later RMS wells (Figure 5.6). We therefore do not consider the 120mm/yr model any further.

Although the deformation patterns are similar, the early slip distributions of the models are quite distinct between the low and high V_{pre} models (Figure 5.10). This is largely due to the greater amount of slip in the first 25 days which is not constrained by our data.

A comparison of the moment release from the various models and the kinematic afterslip model of *Ryder et al. (2007)* is shown in figure 5.11. We can see that moment release is significantly greater for the model of *Ryder et al. (2007)* than any of the models that lie within the RMS minima. This is a consequence of the fault plane on which afterslip is allowed to occur being smaller in spatial extents in the models shown here than those shown in the kinematic inversion of *Ryder et al. (2007)*. The initial rate of moment release however is significantly higher in all the afterslip models than in *Ryder et al. (2007)*. This is due to the assumption of linear velocity rates in the

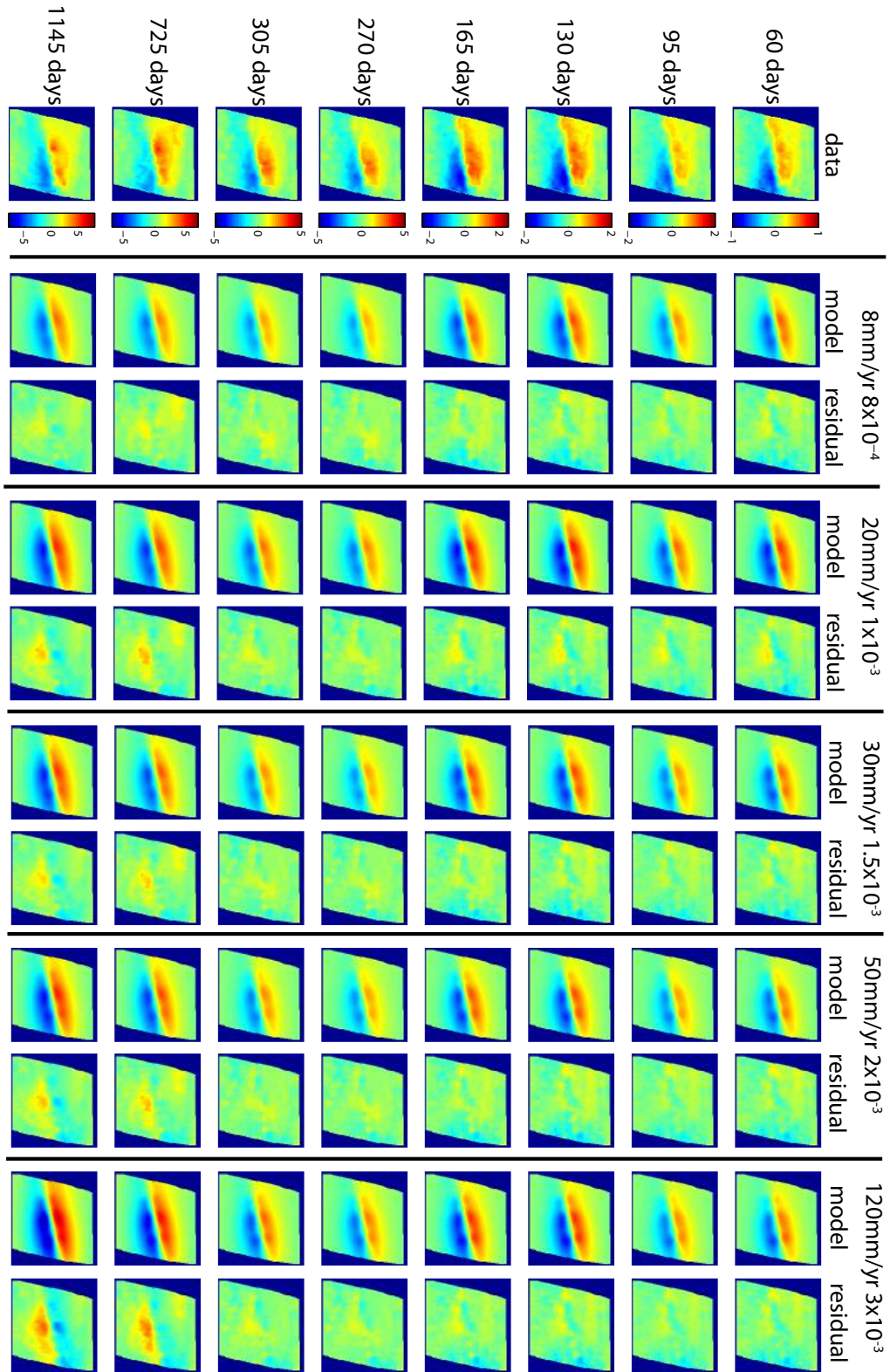


Figure 5.8: Line of sight deformation patterns and observed deformation at various time points from five rate and state friction models that lie within the trough in the RMS plot (Figure 5.6). Each panel is 150 x 150km in size.

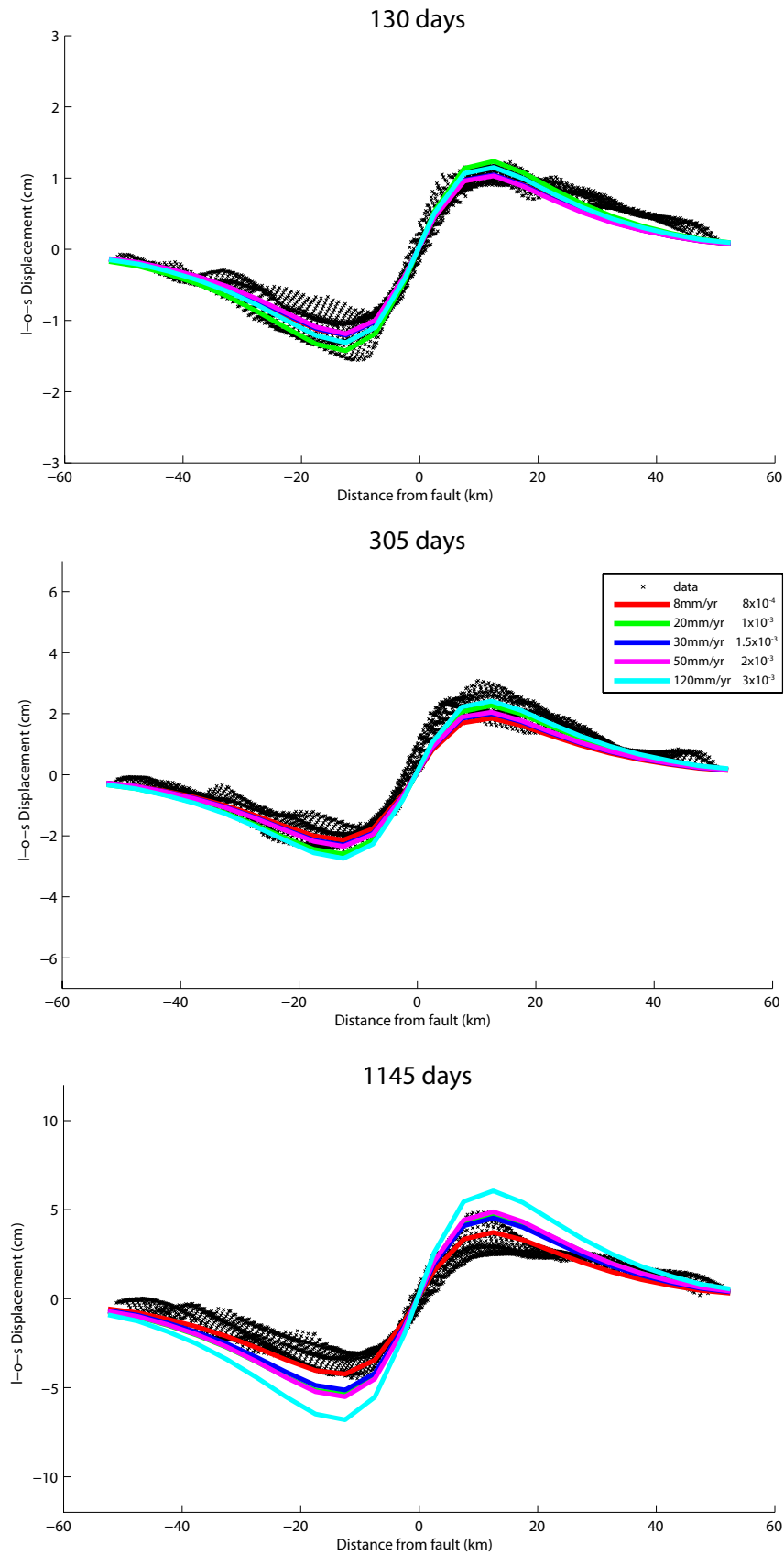


Figure 5.9: Line of sight deformation profiles at three time points (130, 305 and 1145 days) from five rate and state friction models (Figure 5.8) that lie with the trough in the RMS plot (Figure 5.6).

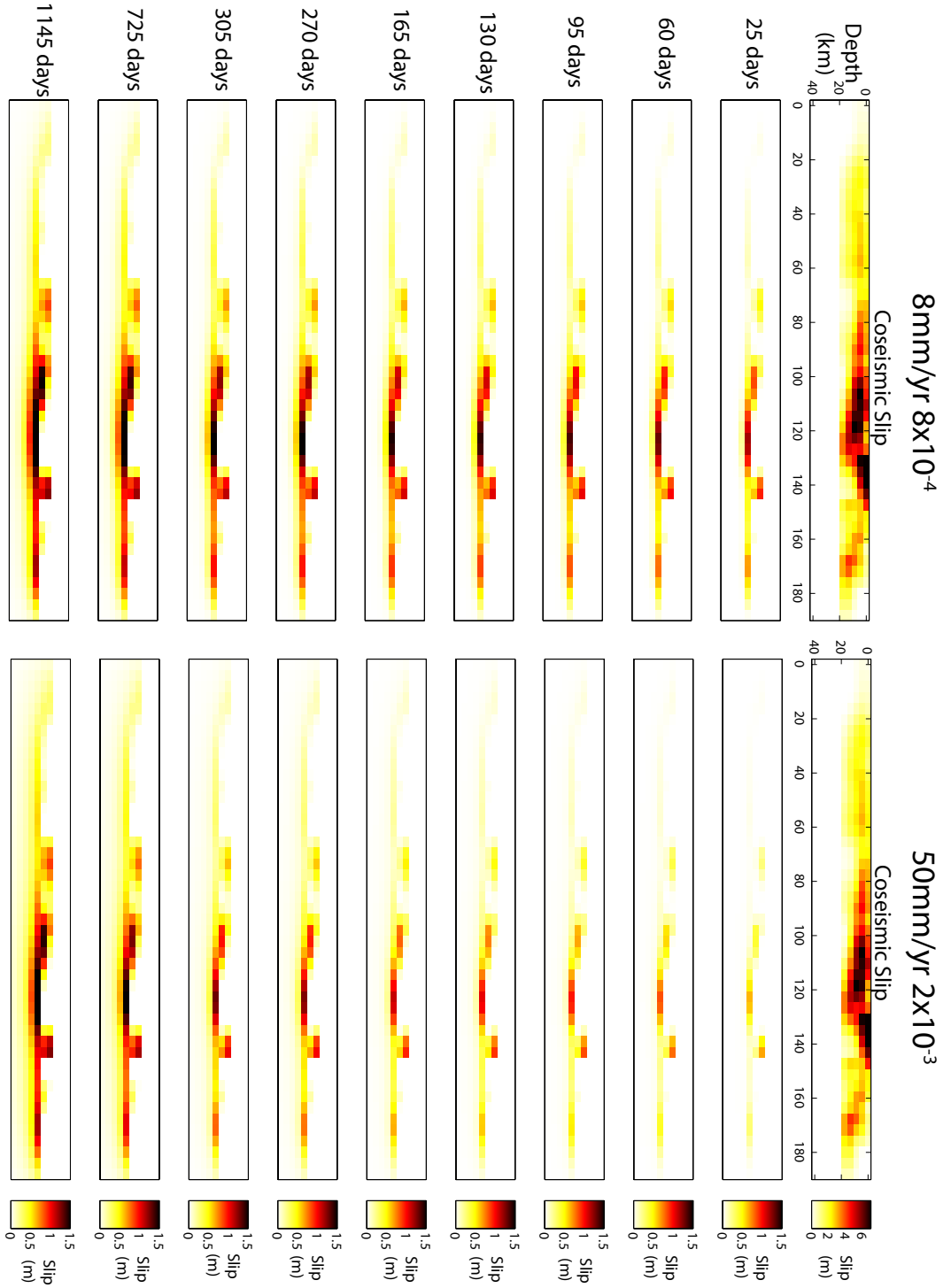


Figure 5.10: Modelled afterslip at various time points from the rate and state friction models with $v_{pre}=8\text{mm/yr}$ or 50mm/yr and $a-b$ of 8×10^{-4} or 2×10^{-3} respectively. The top panel denotes the coseismic slip distribution form *Running et al.* (2007b). There is a clear correlation between regions of high slip in the earthquake and regions of high afterslip. There is little afterslip in the deepest section of the fault model, largely due to (comparatively) small stress changes at those depths. It should also be noted that the scale is different between the coseismic and postseismic panels.

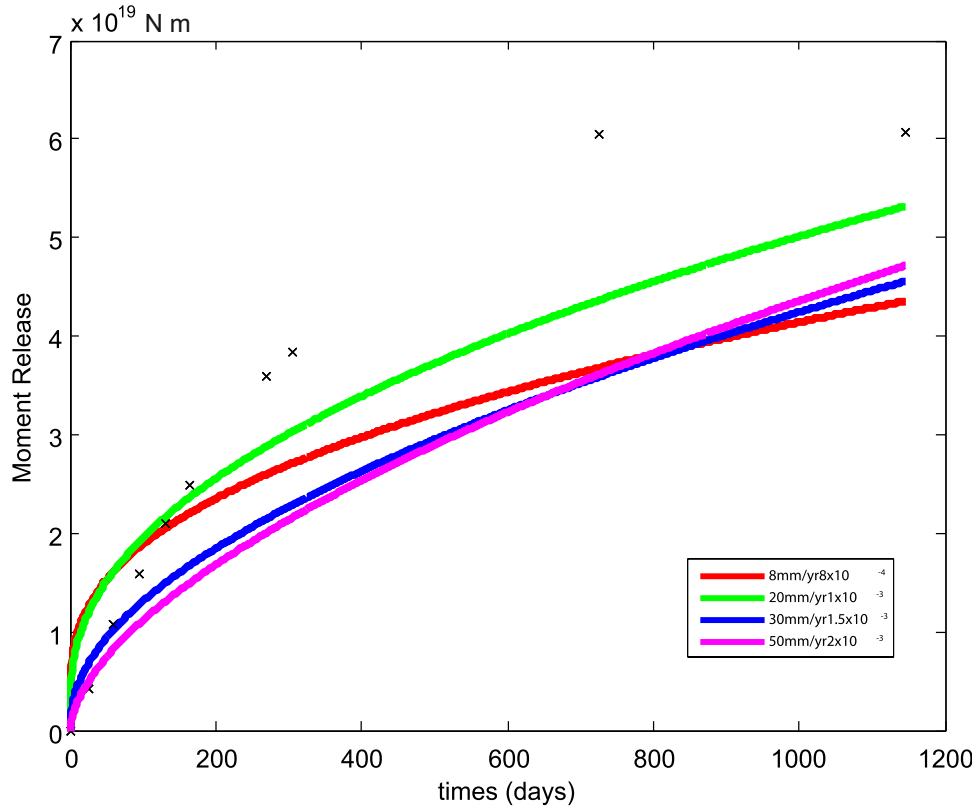


Figure 5.11: Seismic Moment release vs time for the two models. Black crosses denote the moment release from afterslip model in *Ryder et al. (2007)*. The moment was calculated from the cumulative slip across all the 4×4 km patches using a rigidity of 3×10^{10} Pa.

first, unconstrained, 25 days made by *Ryder et al. (2007)*.

Afterslip is a response to the increase in shear stress on the fault during the earthquake, and slip will continue to relieve the stress increase until it has been dissipated. We analyse the differences in how the shear stress is relieved on the fault plane as shown in Figure 5.12 to determine whether there is any difference in the stress state after 1145 days.

Both models in Figure 5.12 show similar distributions of stress after the 1145 days and importantly the stress has not been fully relieved on the plane. We extend the models to cover 13 years after the event and use the data from Chapter 3 to narrow the range of suitable models. Figure 5.13 shows the rate of moment release expected is different for each model during the 5.5-13 year period that is covered by the later ENVISAT data.

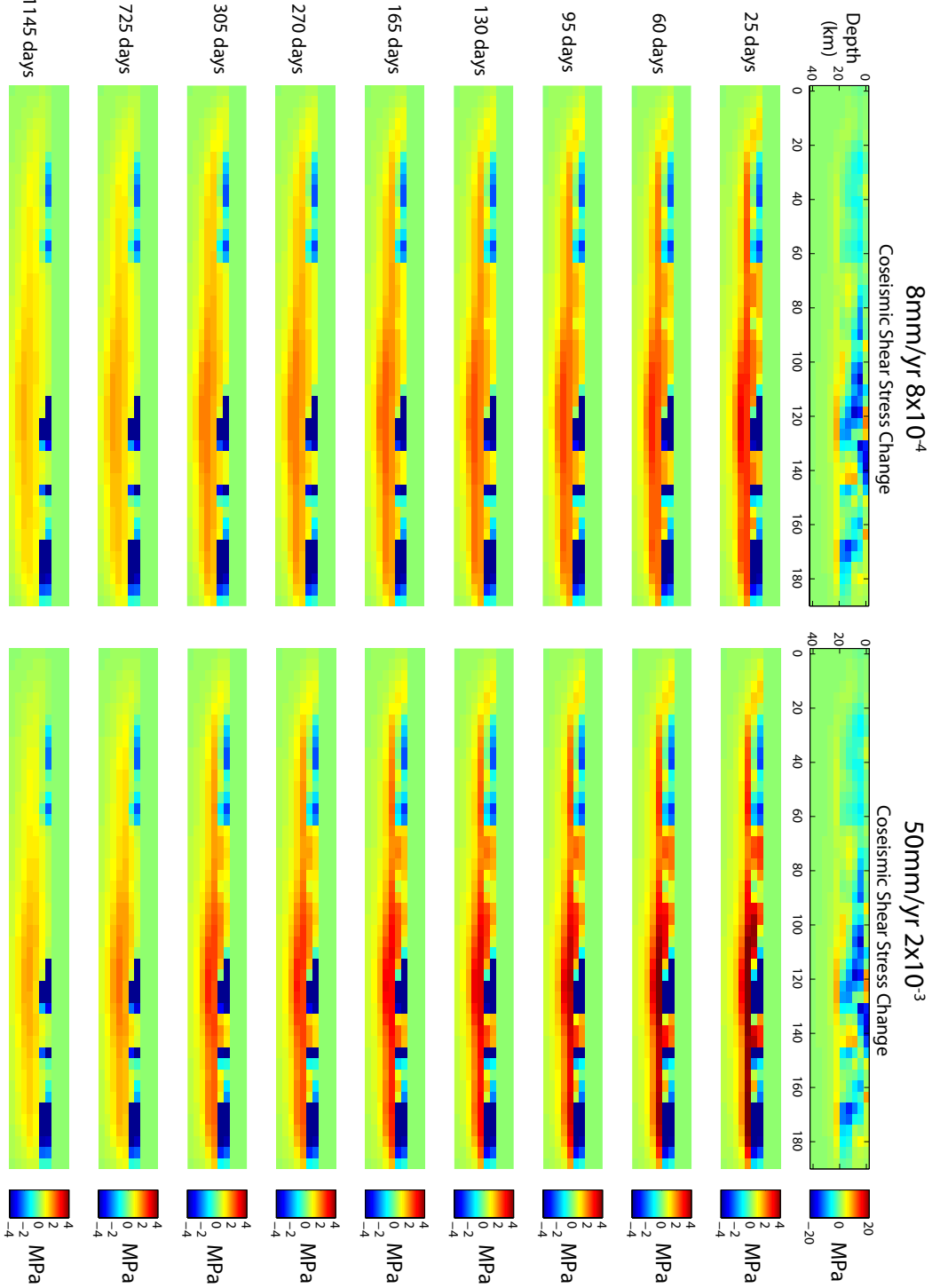


Figure 5.12: Shear stress at various time points from the rate and state friction models with $v_{pre}=8/50\text{mm/yr}$ and $a-b$ of $8\times 10^{-4}/2\times 10^{-3}$. The top panels show the coseismic stress change. Blue regions indicate a stress drop so no slip occurs on those patches hence the magnitude hardly changes over the course of the afterslip model.

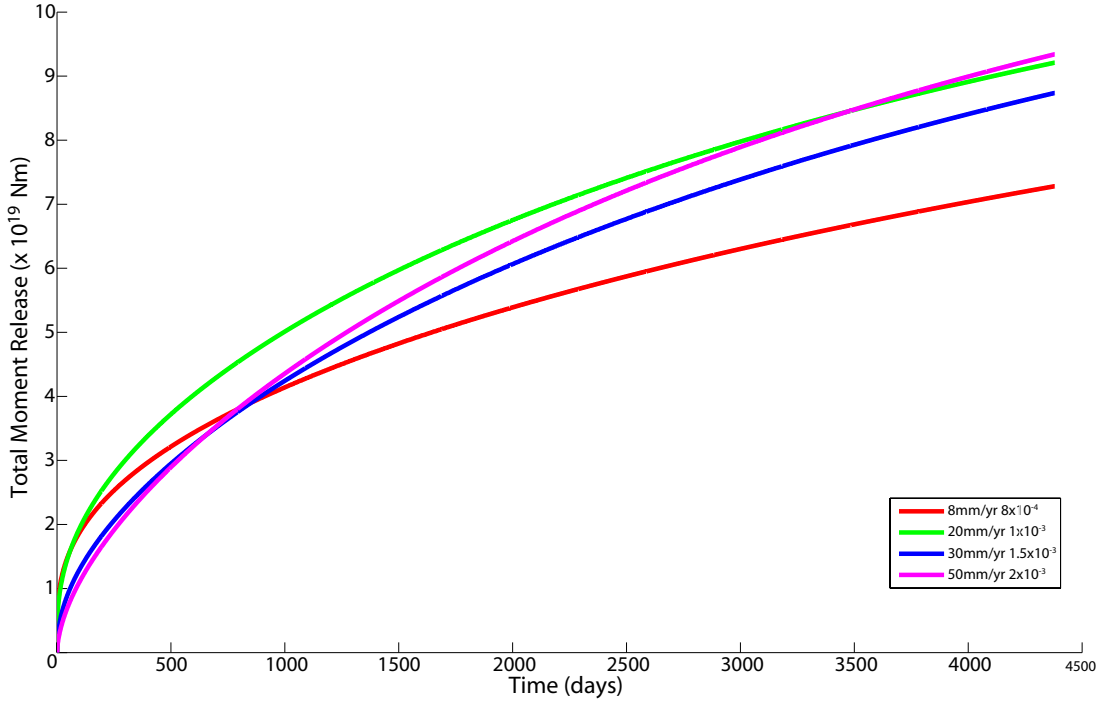


Figure 5.13: The total moment release across the whole plane for various rate and state models. The moment was calculated from the cumulative slip across all the 4 x 4 km patches using a rigidity of 3×10^{10} Pa.

5.4.2 Rate and State Modelling with the ENVISAT data

Our analysis of the ERS data shows there is a region of parameter space which produces models that fit the ERS data well. We calculate the deformation of these models in the ENVISAT observation period spanning 7.5 years starting 5.5 years after the Manyi event. We compare the deformation expected to the data processed in chapter 3 and produce RMS wells for each for the three l-o-s velocity maps (Figure 5.14). Profiles (using the same line as the profiles in Figure 5.9) are shown in Figure 5.15.

The shape of the RMS wells from both the ERS and ENVISAT data (Figure 5.6 and 5.14) suggest that the best-fitting model over the whole period of observation has a V_0 of about 10-20 mm/yr and an $(a - b)$ value around $1-2 \times 10^{-3}$.

Having concluded that a range of rate and state models can reproduce the observed deformation fields after the Manyi Earthquake, we now analyse the results of another afterslip model type, viscous shear zones before discussing the implications of the above

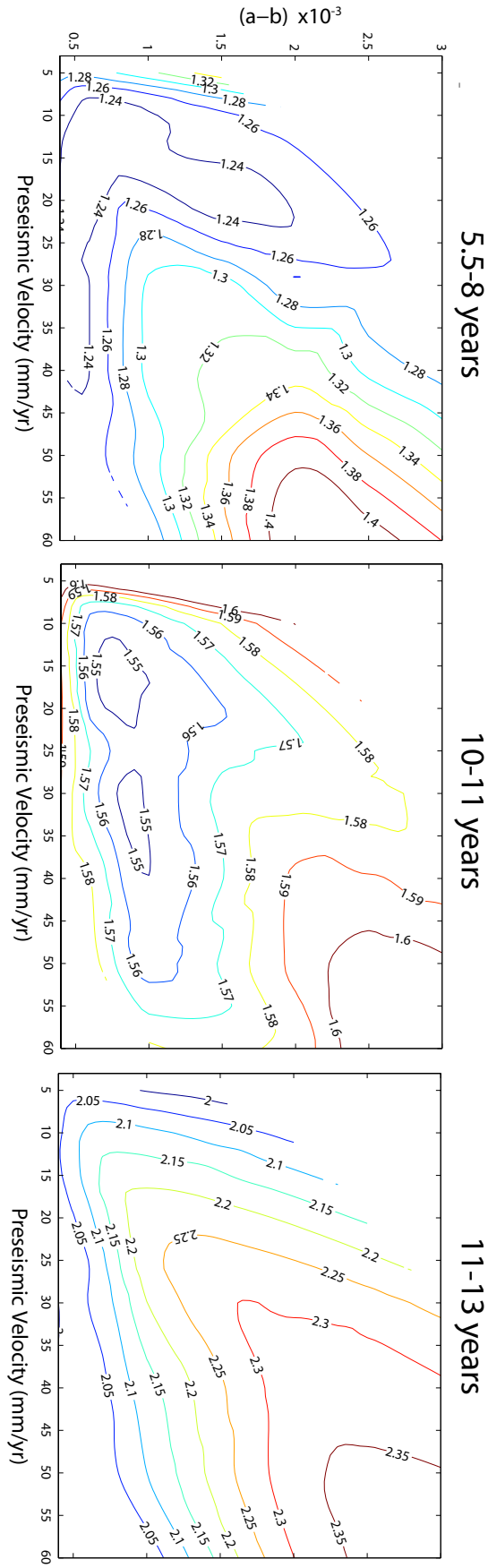


Figure 5.14: RMS misfit of parameter space for three observation periods (5.5–8 yrs, 10–11 yrs and 11–13 yrs) from rate and state friction models that lie with the trough in the RMS plot (Figure 5.6).

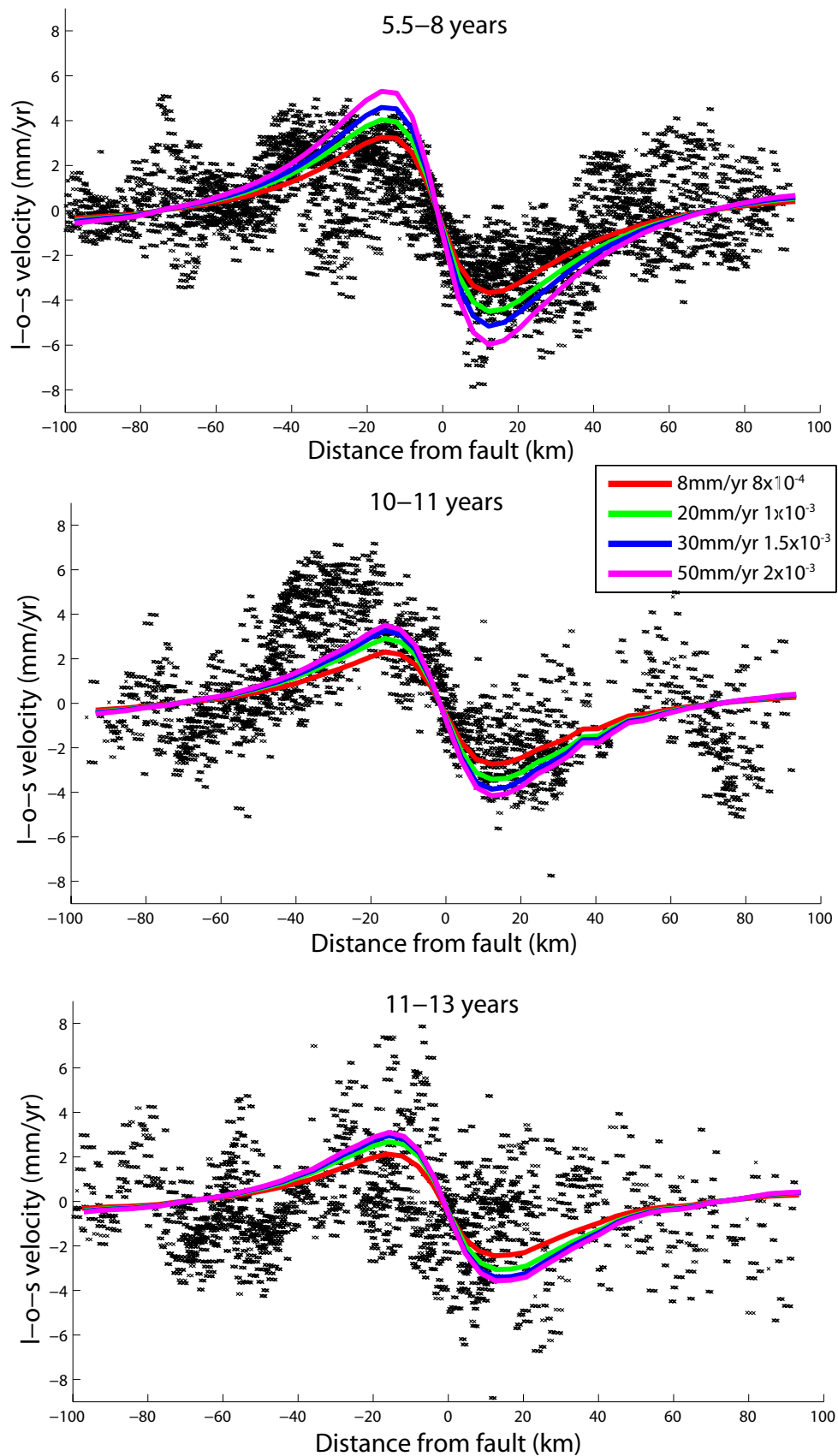


Figure 5.15: Line of sight velocity maps for three periods (5.5-8yrs, 10-11yrs and 11-13yrs) from four rate and state friction models that lie with the trough in the RMS plot (Figure 5.14).

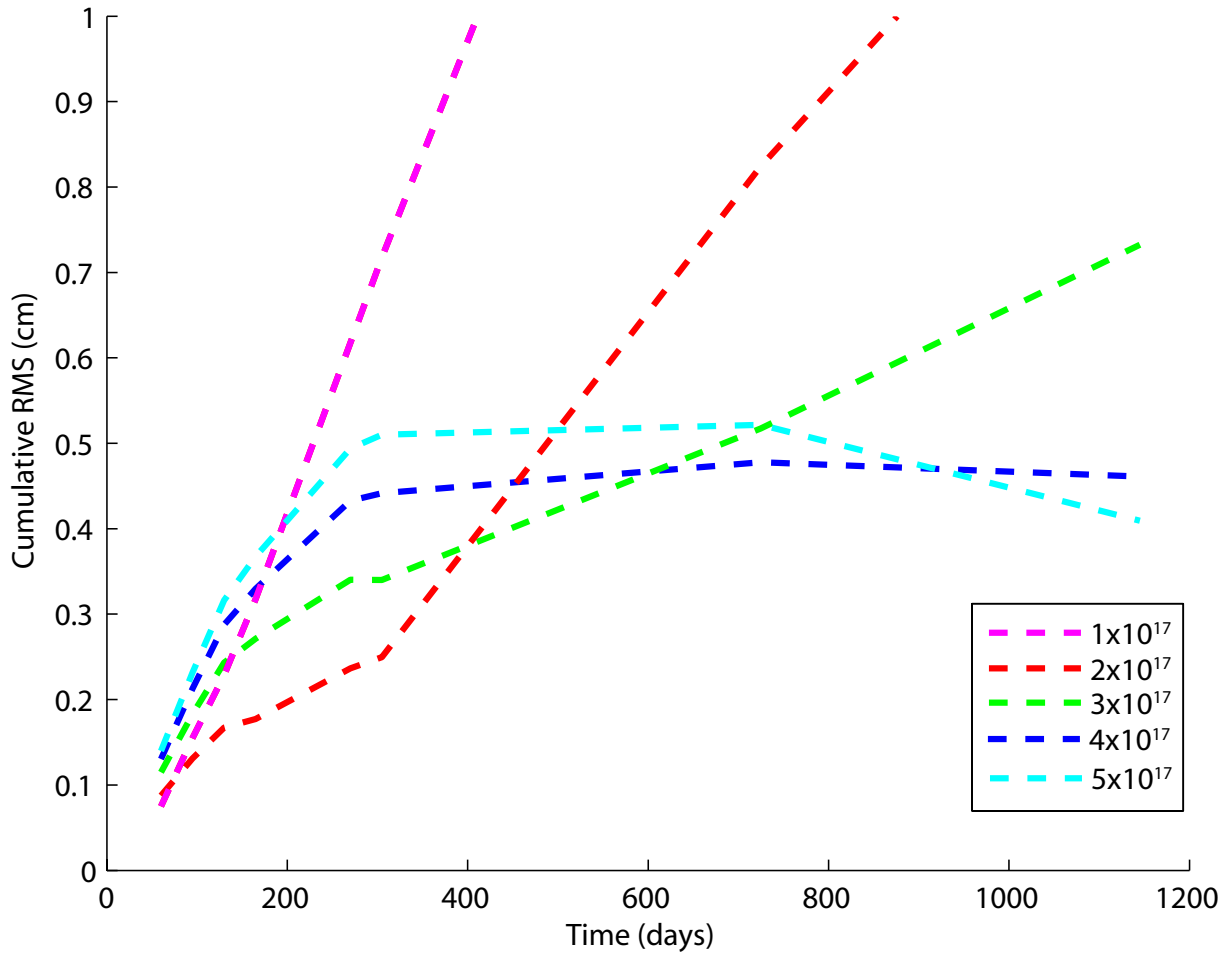


Figure 5.16: The RMS misfits of the cumulative l-o-s displacements from various shear zone models with a range of viscosities (Pa s).

results.

5.4.3 Viscous Shear Zone Results

For the purposes of this analysis we assume that the width of the viscous shear zone is 10 km which allows us to analyse the results in terms of viscosities alone. We run a series of models with a different η using the creep evolution relationship described in Equation 5.13. To find the best-fitting model we plot the cumulative RMS for a series of models with different viscosities (Figure 5.16) against time

The best-fitting viscosity model depends on time, as the lowest RMS solution corresponds to increasingly higher viscosity models as time goes on. The best-fitting initial

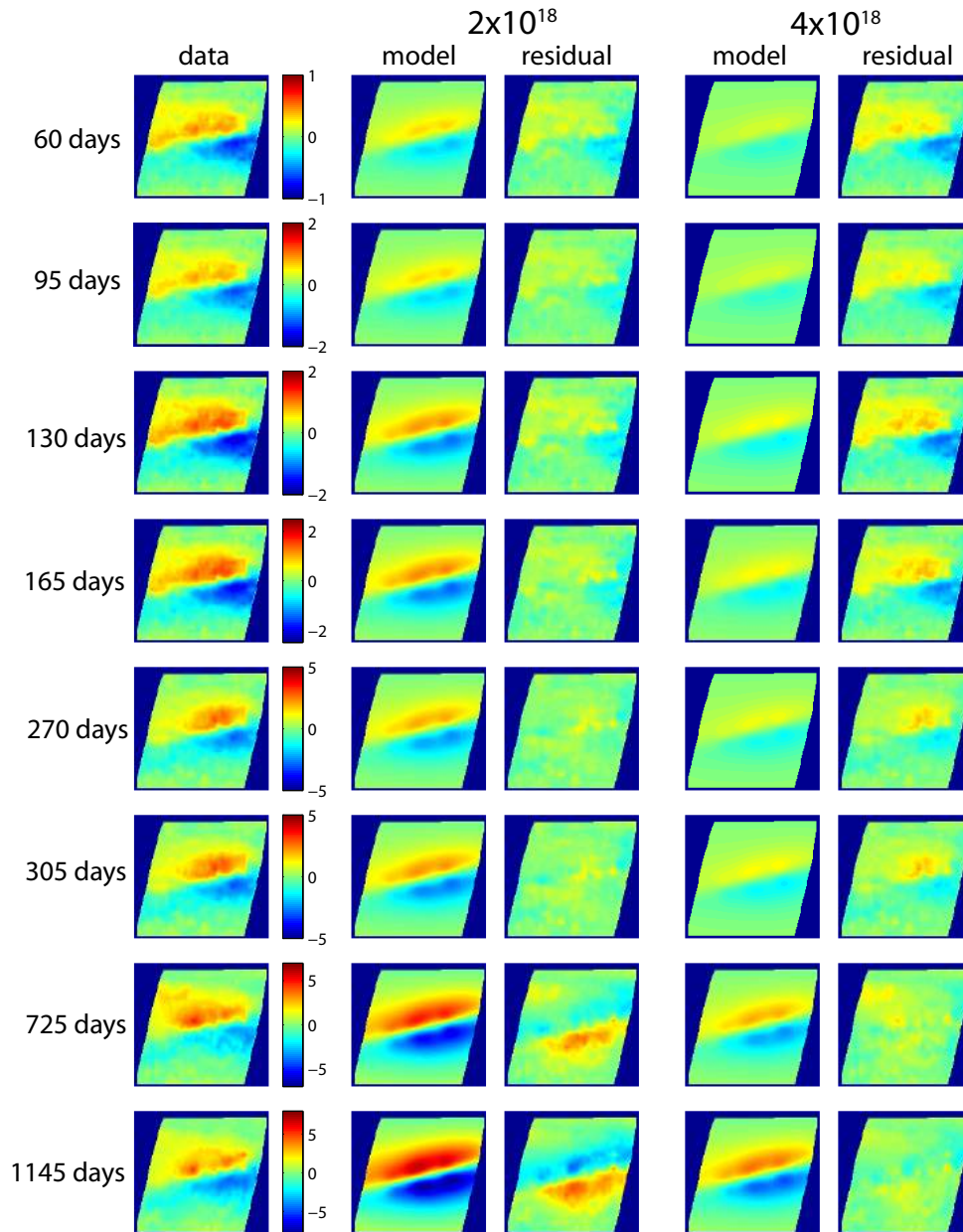


Figure 5.17: The data, model and residuals from a viscous shear zone model with a constant shear zone width of 10 km and viscosity of 2×10^{18} Pa s and also 4×10^{18} Pa s. Each panel is 150x150 km

model is the 2×10^{18} Pa s model and the best-fitting model at the end of the times series is the 4×10^{18} Pa s model (Figure 5.17).

As the data require an apparent increase in viscosity with time or decrease in width with depth, the constant width and viscosity shear zone model is not appropriate for describing the observed postseismic deformation. It does indicate however that another

class of model may be appropriate that invokes a power-law rheology. In a material with a power law rheology, the viscosity of the material is dependent upon the state of stress within the material: high stress results in low viscosity and vice versa. Such a model may be suitable here given that we observe an increase in apparent viscosity as the stress is relieved (i.e with time). Power law rheologies however are beyond the scope of this thesis.

5.5 Discussion

The modelling results presented show that the deformation pattern observed in the aftermath of the Manyi earthquake can be explained using a model based on the hot friction assumption of afterslip. We cannot define the exact parameters; just a range of suitable fitting ones. Kinematic inversion of afterslip based on inverting the observed SAR data reproduces the observed deformation pattern for the first 1145 days and can also match the full 13 years.

All the data used in this chapter (and thesis) are derived from InSAR. Collecting data over a particular area to study an earthquake is only possible when the satellite passes over the area in question. The the case of ENVISAT the interval between passis is 35 days. In this case the first postseismic SAR data to be acquired over track 305 occurred 25 days after the earthquake. This has implications for the modelling in this chapter. Firstly the coseismic model of *Funning et al.* (2007b) models the deformation of not only the earthquake but also the deformation that occurred in the 25 days (over track 305). Secondly the data in the time series does not start from 0 days after the earthquake but instead 25 days afterwards this means we have no data constraint for the first 25 days. The inclusion of the first 25 days of afterslip in the coseismic model would increase the coseismic stress change from the coseismic slip model. The predicted afterslip within the first 25 days is about $1\text{--}2 \times 10^{19}$ N m which is small in comparison with the total coseismic release of 2.2×10^{20} N m so we do not consider this further. The lack of any metric to measure the initial rate of afterslip is the primary reason for the

trade-off in model parameters as they are only significantly different in the immediate aftermath of the event for which we have no data.

5.5.1 Comparisons with previous afterslip results

We obtain a range of models with different $a-b$ values that provide a satisfactory match to the data. Laboratory-based friction experiments (*Ikari et al.*, 2007; *Ikari et al.*, 2011) show that one of the primary controls of friction on any fault is the mineralogy. Faults embedded in rocks rich in quartz and andesine feldspar have large coefficients of friction ($\mu > 0.6$), as do serpentine and kaolinite (≈ 0.65). Sheet silicate minerals however exhibit a range of frictional coefficients from 0.25 - 0.6. Materials with low coefficients of friction tend to have positive and (relatively) large $(a-b)$ values (10^{-2} - 10^{-3}). Materials with high coefficients of friction exhibit both velocity strengthening and weakening behaviour as $(a-b)$ has both negative and positive values ($\pm 10^{-3}$). As the Manyi fault is seismically active it must exhibit velocity weakening behaviour at least in part. This would suggest that the fault gouge around the fault is frictionally strong (*Ikari et al.*, 2011) which in turn suggests the Manyi Fault is reasonably mature as immature fault gouges tend not to show velocity weakening behaviour (*Moore and Saffer*, 2001). The low values of $(a-b)$ of all the best-fitting models in this chapter would support the suggestion that the Manyi Fault is reasonably mature.

Afterslip has been proposed as a mechanism to explain the postseismic deformation following a number of well studied earthquakes. Table 5.1 summarises the results found in these studies. The values of $(a-b)\sigma_n$ are reasonably similar considering the range of tectonic setting covered by the studies. The results of this study lie towards the higher end of the estimates found in literature. Laboratory experiments investigating $(a-b)$ values for wet granites, give results in the range of $(a-b) \approx 5 \times 10^{-3}$ - 5×10^{-2} (*Blanpied et al.*, 1995) which are an order of magnitude higher than those inferred in this study. This is also observed in the majority of models in Table 5.1 indicating that there may be a problem applying laboratory derived values to real, lithospheric scale faults.

Earthquake	Author(s)	$(a-b)\sigma$ (MPa)
1992 Landers	<i>Perfettini and Avouac</i> (2007)	0.48-0.53
1999 Izmit	<i>Hearn et al.</i> (2002); <i>Hearn et al.</i> (2009)	0.4
1999 Chi-Chi	<i>Perfettini and Avouac</i> (2004)	0.3-1.5
2001 Peru	<i>Perfettini et al.</i> (2005)	0.26
2002 Denali	<i>Johnson et al.</i> (2009)	0.5
2003 Tokachi-oki	<i>Fukuda et al.</i> (2009)	0.29-0.43
	<i>Miyazaki et al.</i> (2004)	0.2
2004 Parkfield	<i>Barbot et al.</i> (2009)	0.7
	<i>Johnson</i> (2006)	0.05-0.1
2005 Nias Simeulue	<i>Hsu et al.</i> (2006)	0.2-0.7
1997 Manyi	This study	0.6-1.5*

Table 5.1: Review of Afterslip Studies that use geodetic data. * indicates an average depth value (h) of 26 km was used to calculate σ_n where $\sigma_n = \rho gh$

5.5.2 Validity of the Model

In this study we have constructed a series of models that are capable of reproducing the observed deformation pattern over the Manyi region for 13 years after the event. Previous afterslip studies have tended, mainly due to a lack of data, to focus on the months and years immediately after the event. No other investigation detailed in Table 5.1 finds that afterslip can reproduce the deformation pattern for such a long time as we propose here. A series of assumptions were made when the models were calculated based on their validity for the initial few years only, so we must ensure that if afterslip is to be proposed as the mechanism behind the Manyi earthquake we should make sure the assumptions are valid for the duration of the modelling.

The rate and state friction models in this chapter are based on the hot friction assumption, which means that we do not include any healing effects in our models. *Perfettini and Ampuero* (2008) compare models that are rate dependent (RD) against rate and state dependent models and conclude that the behaviours are equivalent beyond the duration of the coseismic acceleration transients. This would suggest we are justified to neglect the state variable for the modelling performed here as we have

no information until 25 days. At this point both models are equivalent assuming the friction regime does not transfer back to a locked regime. Healing will only become significant and hence the change back to a locked value once $V \ll D_c/\theta$ (Equation 5.4). There is currently no consensus what value of D_c is appropriate for real faults. Laboratory experiments yield D_c values in the range of 10^{-6} - 10^{-4} m, but there are no agreements as to how the laboratory value should be scaled up to real faults (*Marone*, 1998). *Scholz* (1988) argues that D_c must be several orders of magnitude larger due to the topography of fault surface contacts whereas *Lapusta and Rice* (2003) argue that no scaling is required based on studies of the size of the smallest earthquakes. Current computational limits on the spatial resolution of models limit the range of D_c that can be tested numerically so models that do include the state variable (e.g *Fukuda et al.* (2009)) use D_c values of 10^{-3} - 10^{-1} m, meaning any studies of afterslip due to earthquakes cannot be used to derive the D_c value most appropriate for faults.

It is difficult to distinguish between afterslip and viscoelastic relaxation, especially at depth (*Perfettini and Avouac*, 2007). The next chapter aims to find the best-fitting viscoelastic model to describe the deformation pattern of the Manyi postseismic signal. I leave all discussion about whether afterslip or viscoelastic relaxation provides the most appropriate modelling until the final chapter.

Chapter 6

Viscoelastic Modelling of the Manyi Earthquake

6.1 Introduction

Any stress change within a viscoelastic medium will, over a long enough timescale, cause viscous flow to occur that alleviates the stress. The timescale over which the stress is relaxed depends on two things, the rigidity and viscosity of the medium (assuming a linear rheology). In Chapter 2 we discuss the equations that govern how both the stress and strain fields evolve as a function of time, viscosity and rigidity. In this chapter we create a series of different earth models that approximate the Manyi Fault area. We calculate each the deformation response of each model to the coseismic stress change induced by the 1997 Manyi Earthquake using the methodology described in Chapter 2.

We compare the deformation expected from each model to the two postseismic datasets: the ERS data covering 25-1145 days after the earthquake from *Ryder et al.* (2007) and the ENVISAT data described in Chapter 3 that covers 5.5-13 years after the event. We aim to find a model that corresponds to an RMS minima (in parameter space) when the model is compared with both datasets. We also want the RMS minimum to be stable in each of the datasets themselves. To obtain a measure of un-

certainty on the location of the RMS minima we perturb both datasets with correlated noise (as discussed in the previous chapters) and analyse how the minimum changes.

Viscoelastic modelling of the Manyi Earthquake was performed by *Ryder et al.* (2007). They analysed three types of models, simple Maxwell, channel models and Standard Linear Solid models. They found that a SLS model provided the best match to the data over the whole observation period. As discussed earlier in this thesis however SLS models are not realistic representations of the real Earth. In this chapter we analyse simple Maxwell, layered Maxwell and Burgers rheology models.

6.2 Viscoelastic Modelling

6.2.1 Parameterisation of the Viscoelastic Models

All the viscoelastic models in this chapter were calculated using the PSGRN/PSCMP code (*Wang et al.*, 2007) which was briefly described in Chapter 2. The code computes surface displacements at any user-defined time after an instantaneous stress change, in this case an earthquake. The user specifies the earth model and the fault (source) parameters used to calculate the coseismic stress change. In this case the fault parameters were based on the 11 fault model of *Funning et al.* (2007b) which assumes constant fault parameters on each of the 11 faults (Table 6.1). We did not use the distributed slip model (as used for afterslip modelling) as the computation time and disk usage increase linearly with the number of faults and using the distributed slip model proved prohibitive with the available computing resources.

We consider two Earth model geometries: a viscoelastic half-space overlain by an elastic layer which we refer to as a half-space model and layered model of a viscoelastic half-space overlain by a viscoelastic layer of width H which is overlain by an elastic layer (Figure 6.1) which we refer to as a layered model. Although it is possible to construct ever increasing numbers of layers we refrain from doing so as we introduce too many variables which are difficult to constrain and we are primarily interested in the bulk properties. In both instances the elastic layer is 15 km thick. This is sufficiently deep

Segment	Lat	Lon	Slip (m)	L (km)	D (km)	Strike	Dip	Rake
1	35.084	86.380	2.04	43.1	11.7	80	85	2.6
2	35.154	86.715	2.74	20.3	11.7	66	85	2.6
3	35.201	86.943	4.32	23.3	11.7	85	85	2.6
4	35.231	87.177	6.73	19.9	11.7	77	85	2.6
5	35.255	87.350	5.34	12.0	14.2	266	85	-9.3
6	35.270	87.482	4.91	12.5	14.2	258	85	-9.3
7	35.290	87.604	3.67	10.1	14.2	259	85	-9.3
8	35.306	87.680	3.16	4.3	14.2	247	85	-9.3
9	35.329	87.787	3.07	16.0	14.2	256	85	-9.3
10	35.378	87.925	0.18	11.8	14.2	233	85	-9.3
11	35.447	88.023	0.00	11.6	14.2	225	85	-9.3

Table 6.1: Earthquake fault parameters for the 11 fault model of *Funning et al.* (2007b) based on the inversion of SAR data. Segments are numbered from west to east and correspond to the faults in Figure 3.6. Lat and Lon are the latitude and longitude respectively of the centre and top of each fault patch. L denotes the lateral extent and D the maximum depth of the fault segment.

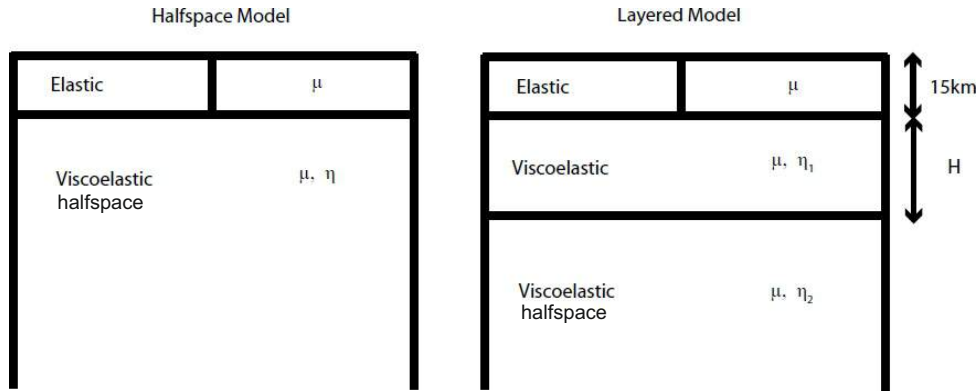


Figure 6.1: Cross section through the earth models used to model the viscoelastic deformation. Both models contain a 15km thick elastic layer at the surface, in which the (coseismic) fault dislocations are located. The model on the left denotes the half-space model, an elastic layer overlying a viscoelastic half-space with various user defined rheologies described in more detail in the text. The model on the right shows the layered model with an elastic layer overlying a viscoelastic layer (of thickness H) that overlies a viscoelastic half-space. The properties of the viscoelastic elements (η_1 and η_2) are varied as described in the text. All layers have the same rigidity μ .

that the coseismic model used in the viscoelastic modelling is fully contained within the elastic layer, a prerequisite of the PSCMP/PSGRN code. It is also consistent with the base of other coseismic models in the region (*Elliott et al.*, 2010).

Throughout this chapter we create a number of earth models in which the viscosity, η , is varied however in all earth models we use seismic velocities from *Steck et al.* (2009)

and an average crustal density of 2.7 g/cm^3 to define μ using the equation

$$v_s = \sqrt{\frac{\mu}{\rho}} \quad (6.1)$$

6.3 Results

For each of the different models we calculate the cumulative line-of-sight deformation at each time point in the timeseries from *Ryder et al.* (2007) (25, 60, 95, 130, 270, 305, 725 and 1145 days). We then calculate the residual deformation by subtracting the modelled deformation from the deformation data. We flatten each residual at each time point using the same algorithm used during the afterslip modelling in the previous chapter to remove any orbital tilt error. We then calculate the RMS misfit of the flattened timeseries data and model for each time point using the equation.

$$\sigma^2 = \frac{1}{n} \sum_{i=1}^n (d_i - m_i)^2 \quad (6.2)$$

To assess the goodness of fit of the models we sum σ for each of the time points in the observation period to calculate the cumulative RMS. This allows us to assess the size of misfit throughout the observation period and not just at the end.

6.3.1 Maxwell Half-space Models

We first model the simplest rheology, referred to as a Maxwell half-space rheology. The earth model used consists of an elastic layer overlying a viscoelastic Maxwell half-space (Figure 6.1). A Maxwell body relaxes over single timescale, referred to as the Maxwell time (t_M):

$$t_M = \frac{\eta}{\mu} \quad (6.3)$$

where η and μ are the viscosity the rigidity of the Maxwell body.

A Maxwell rheology was analysed by *Ryder et al.* (2007). The viscosity of the best-fitting Maxwell model was shown to increase as the observation period increased hence was disregarded. We repeat the modelling however as we are using different modelling algorithms, are treating the orbital tilt error corrections differently and also we have new data for which we would like to ascertain the best-fitting model. We vary the viscosity of the half-space to try and find the best-fitting Maxwellian viscosity for each point in the timeseries.

ERS data

When models are compared to the ERS data covering the first 1145 days of the post-seismic period from *Ryder et al.* (2007) (Figure 6.2) we can see that no single model matches the data well over the whole timescale of the observations. The viscosity of the best-fitting model appears to increase with increasing time (Figure 6.3) , increasing from 2×10^{18} Pa s after 25 days to $\approx 8 \times 10^{18}$ Pa s

ENVISAT data

Analysis of the first three years of postseismic data showed that models require an increase in the apparent viscosity over time to match the ERS data covering the first 3 years after the event. We performed similar analysis on the ENVISAT data from Chapter 3 to ascertain if the viscosity-time trend continued in the 5-13 years after the earthquake (Figure 6.4).

Figure 6.4 suggest that the best-fitting viscosity increases with time. The RMS curve for the 11-13 yr velocity map does not show a minima. This is a result of the lack of any coherent signal in the data and higher level of noise (discussed in the final chapter). To assess whether the increase in viscosity between the 5.5-8 year and 10-11 year is resolvable we perturbed the velocity maps with 100 sets of correlated noise (using the methods described in the previous chapters). The distribution of minima in the 100 datasets allow the error on the minimum misfit to be estimated (Table

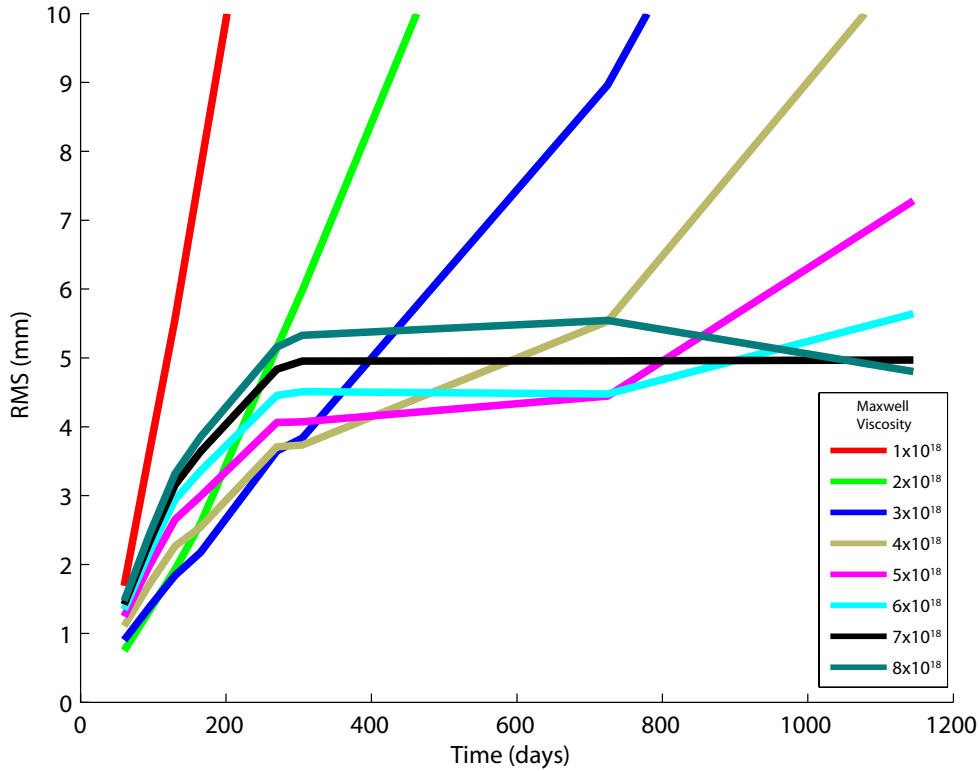


Figure 6.2: Cumulative RMS misfit curves for ERS data and models of a user-defined Maxwell rheology against time. The residuals have a best-fitting plane removed.

6.2). The errors show that the increase in viscosity is not resolvable over time for the ENVISAT data alone.

Velocity Map (years)	Mean Viscosity Minima ($\times 10^{19}$ Pa s)	Viscosity Error ($\times 10^{19}$ Pa s)
5.5-8	3.0	0.5
10-11	3.5	1
11-13	>10	-

Table 6.2: The mean and standard deviation are derived from the 100 RMS viscosity minima of the perturbed datasets. The minima for the 11-13 year searches all hit the bounds of the parameter search as the best-fitting solution for that data is no model (i.e infinite viscosity)

Summary

This results found in this section are similar to those found by *Ryder et al.* (2007), with an increase in the apparent viscosity as the time period of observations increases. Both studies found the increase to be robust by using perturbed datasets of noise. The

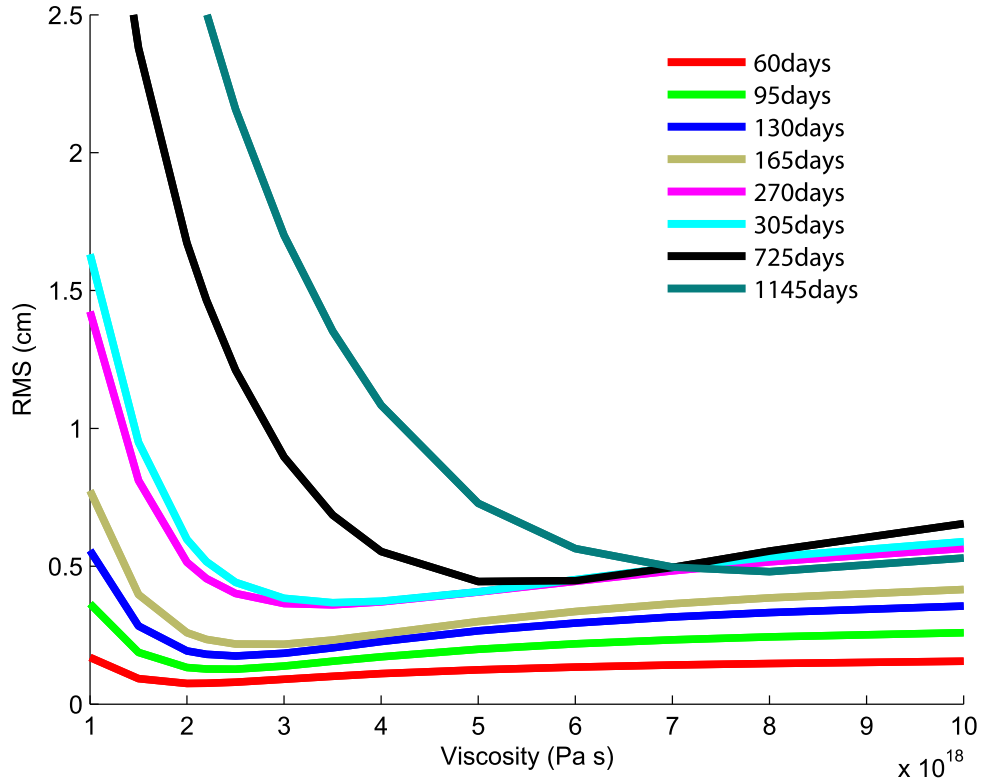


Figure 6.3: Cumulative RMS misfit curves as a function of Maxwell viscosity at constant time. Note the increase in RMS and viscosity increase of each minima with increasing time.

steepness of the curves in Figures 6.3 and 6.4 show that the lower limit of viscosities is significantly better constrained than the higher limit. It also shows that the RMS well becomes less well defined as time progresses.

The best-fitting viscosity for the ENVISAT data, which shows the deformation from 5.5-13 years after the event, is approximately an order of magnitude higher than the best-fitting viscosities for the ERS data initially after the earthquake. This suggests that the trend observed in the ERS data continues in the interval between the ERS and ENVISAT observation periods. There is also an indication that the viscosity increases during the ENVISAT observation period however the errors on viscosity minima for 5.5-8 and 10-11 year velocity maps are larger than the difference between them, therefore the increase in apparent viscosity is not robust using the data available in this study. Perturbed versions of the 11-13 year dataset do not reveal any indication of RMS minima confirming that the noise is too large. It is subsequently not used in any further analysis (although the results are shown)

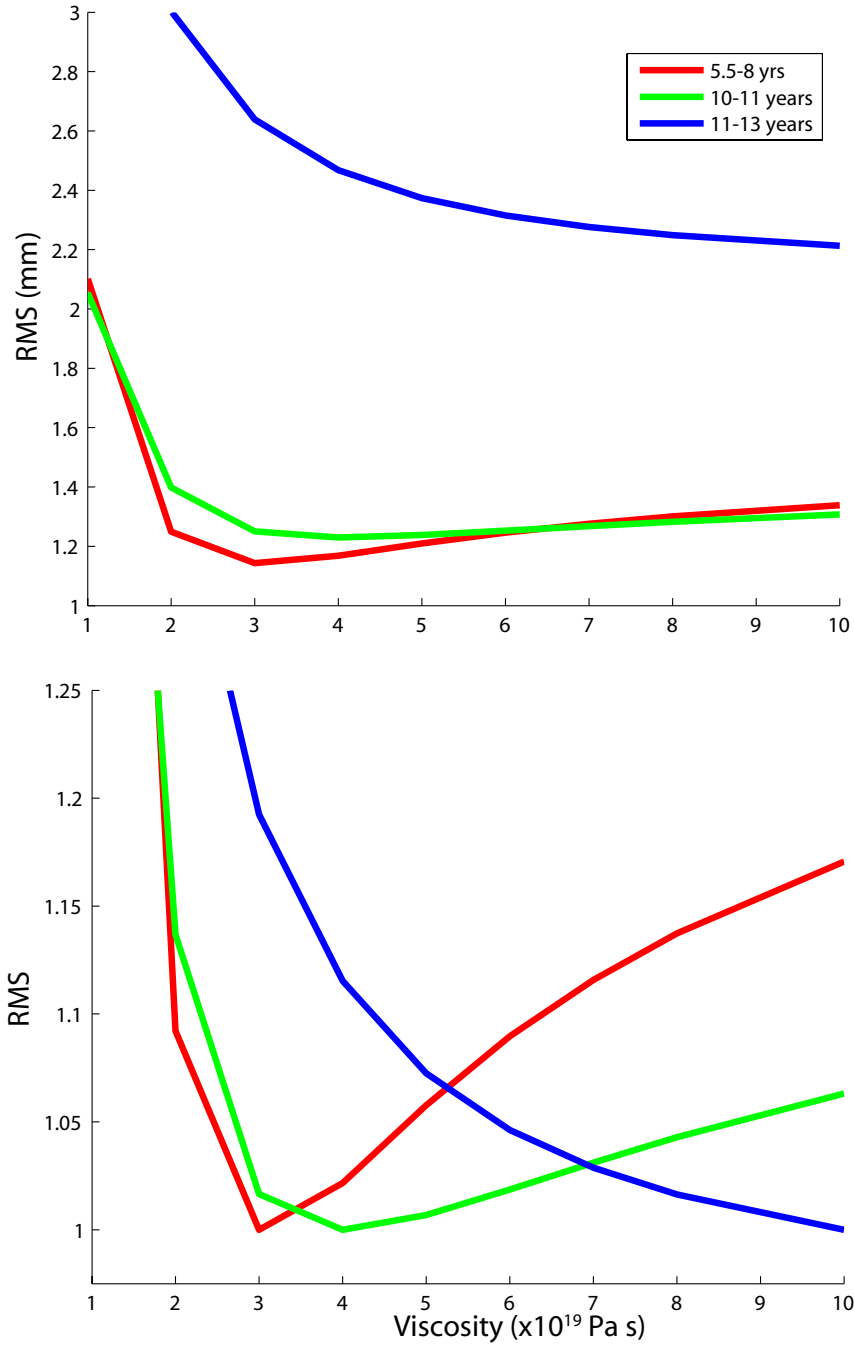


Figure 6.4: Top image shows the RMS misfit of each of the three velocity maps from chapter 3 and the bottom panel (b) is the same curves but with each ratemap RMS has been normalised to the minimum misfit.

We conclude, like *Ryder et al.* (2007), that a Maxwell half-space overlain by an elastic layer is not a realistic model to explain the postseismic deformation following the Manyi earthquake over the entire postseismic cycle.

6.3.2 Layered Maxwell Models

A single body characterised by a simple Maxwell rheology will always have a single viscosity and hence decay over one timescale. If we introduce a second Maxwell body with a different viscosity, the apparent viscosity inferred from surface deformation can be temporally varying. The simplest example of this is the layered earth model described in Figure 6.1.

A similar model of a Maxwell viscoelastic layer sandwiched between an elastic (surface) layer and an elastic half-space, was tested by *Ryder et al.* (2007) and found to match the data poorly. They found a model with a thin channel produced a pattern of deformation that is not consistent with the data and a thick channel was found to be indistinguishable from a half-space with the same viscosity.

This study does not analyse the viscous channel models of *Ryder et al.* (2007) but instead analyses models where the half-space can have a viscosity either higher or lower than that of the viscoelastic layer above it (Figure 6.1). We model the deformation using three different earth models with different viscoelastic layer thicknesses 15, 35 and 50 km, referred to as a thin, medium and thick layer.

ERS data

For each of our layered earth models we vary the viscosity of the the two Maxwell bodies and calculate the RMS misfit of the models to the ERS data (Figure 6.5). To determine error estimates on the bounds of the the minimum misfit models we perturb the initial dataset with correlated noise and find the location of the minimum misfit for each perturbed dataset. The mean location and error is then calculated from the range of minima found. To test whether the best-fitting solution is stable throughout the observation period we calculate the location of the minima (and associated error) after 305, 725 and 1145 days (Table 6.3).

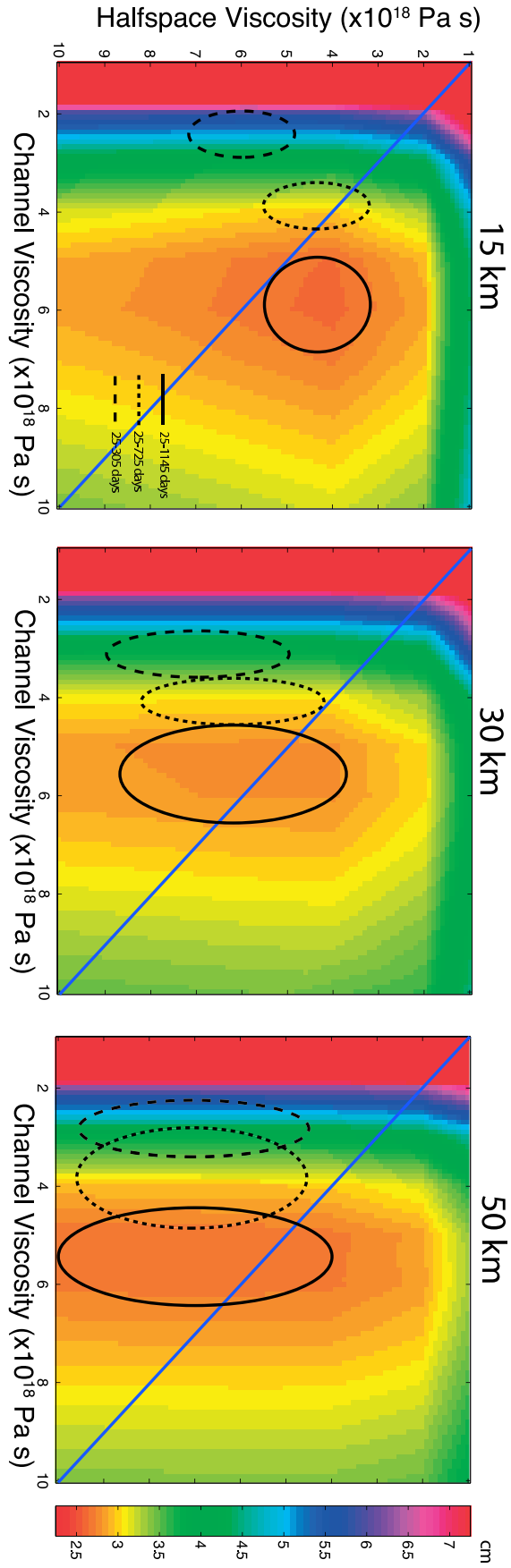


Figure 6.5: Best fitting parameter searches for the channel models. The coloured backdrop denotes the cumulative RMS misfit between the ERS data and models (described in the text) over the first 1145 days. The black ellipses denote the location and error of the best-fitting solutions after 305, 725 and 1145 days which are summarised in Table 6.3. The diagonal blue line lines there the channel models is equivalent to a Maxwell half-space model.

Channel Thickness (km)	Cumulative RMS period (days)	Channel Viscosity ($\times 10^{18}$ Pa s)	Half-space Viscosity ($\times 10^{18}$ Pa s)
15	20-305	2.5 ± 0.5	6.0 ± 1.0
	20-725	4.0 ± 0.5	4.5 ± 1.0
	20-1145	6.0 ± 1.0	4.5 ± 1.0
35	20-305	3.0 ± 0.5	7.0 ± 2.0
	20-725	4.0 ± 0.5	6.0 ± 2.0
	20-1145	5.5 ± 1.0	6.0 ± 2.5
50	20-305	2.5 ± 0.5	7.0 ± 2.5
	20-725	4.0 ± 0.5	7.0 ± 2.5
	20-1145	5.5 ± 1.0	7.0 ± 3.0

Table 6.3: The mean and standard deviation are derived from locations of the 100 RMS viscosity minima of the perturbed datasets.

Summary

The movement of the ellipses show that the best fitting model with time of all three channel models requires a change in viscosity of the viscoelastic layer over the initial postseismic observation period using ERS data. As this requires different models for different observation periods a single channel model cannot adequately describe the observed postseismic motion, therefore we do not compare the layered models to the ENVISAT data.

The difference (or lack thereof) between the mid and thick channel RMS shows the lack of resolution to deep deformation. Deformation in the deep halfspace results in a very long wavelength surface signal as it is filtered through the overlaying rock. The lack of resolution in the deep is a result of the orbital tilt correction removing any additional long wavelength signal from deep deformation.

The location of the RMS minima in all the plots in Figure 6.5 show that the channel viscosity is constrained a lot better than the half-space viscosity. It also shows that the best-fitting layered models differ little from a half-space model as the channel and half-space viscosity are very similar.

We have shown that one and two layer Maxwell bodies cannot provide an adequate description of the observed postseismic deformation following the Manyi event. It is possible to create models with increasing number of layers however we refrain from

doing so for two reasons. Firstly the number of variables and combinations of them becomes increasingly large and therefore more difficult to analyse in detail. The second run-time of the PSGRN/PSCMP increases dramatically as the number of layers increases. The combination of these two factors make modelling more than two layers unfeasible.

6.3.3 Burgers Rheology

The simplest model that can relax over multiple timescales is a half-space model with a Burgers rheology (*Bürgmann and Dresen, 2008*) in the viscoelastic half-space, the mechanical analogy to which is a Kelvin and Maxwell element joined together in series (Figure 2.5). The mechanical analogy to a Burgers body is similar to the SLS body, the only difference being the dashpot of the Maxwell element is not present in an SLS body. This extra dashpot means a Burgers body can relax completely like a Maxwell body. Like a SLS rheology Burgers rheology has multiple timescales of deformation, the feature of an SLS body which led *Ryder et al. (2007)* to adopt a preferred model with a SLS rheology. A Burgers rheology has been used to describe the motion across the Kunlun fault to the east of the Manyi area (*Ryder et al., 2011*). To describe the motion of a Burgers body, four variables need to be defined, the rigidity and viscosity in the Kelvin element and the rigidity and viscosity in the Maxwell element. We use the seismic velocities from *Steck et al. (2009)* to define the rigidity of both elements and the Kelvin and Maxwell (element) viscosities were varied.

ERS data

We perform a 2D parameter search of the best-fitting Kelvin and Maxwell viscosities to the ERS data covering the first 1145 days after the event (Figure 6.6). We also attempt to ascertain an estimate of the error in the minimum misfit using perturbed data in the same manner as described earlier (Table 6.4). Figure 6.6 shows that there is a stable minimum that is maintained throughout the observation period. There is also a second minimum which runs along the top of each plot. This minimum requires a low

Maxwell viscosity. Although it may match the data well we do not consider a Burgers model with a lower viscosity Maxwell element than Kelvin element. This is because the initial relaxation would occur across the Maxwell dashpot and not the Kelvin element. As this deformation is permanent the second viscosity would be largely irrelevant and the system would approximate a Maxwell rheology.

ENVISAT DATA

We compare the Burgers models to ENVISAT data to ascertain if the best-fitting model for the ERS data can explain the deformation observed in 5.5-13 years after the event (Figure 6.7). The RMS wells for the first two observation periods (5.5-8 and 10-11 years) show that the Kelvin and Maxwell element viscosities trade-off against each other however the RMS wells are very broad. Although a well exists for the 11-13 year model it is not stable and continuously hits the bounds so is not included in the analysis.

(Cumulative) RMS period	Kelvin Viscosity η_k ($\times 10^{18}$ Pa s)	Maxwell Viscosity η_m ($\times 10^{19}$ Pa s)
ERS data		
20-305 days	3.0 ± 1.0	6.0 ± 2.0
20-725 days	3.5 ± 1.0	6.5 ± 2.5
20-1145 days	4.0 ± 1.5	7.0 ± 2.5
ENVISAT data		
5.5-8 years	4.5 ± 1.5	7.5 ± 2.5
10-11 years	5.5 ± 2.0	7.0 ± 3.0
11-13 years	$3.0 \pm 2.0^*$	$7.0 \pm 3.0^*$

Table 6.4: The mean and standard deviation are derived from locations of the 100 RMS viscosity minima of the perturbed datasets. * denotes that the values are not reliable as the RMS well continues outside the parameter range.

Summary

The location and error of the minima are summarised in Table 6.4 and Figure 6.8. Again there is an indication of an increase in the apparent viscosity (of the Kelvin element) with time however it is not resolvable within the errors associated with the

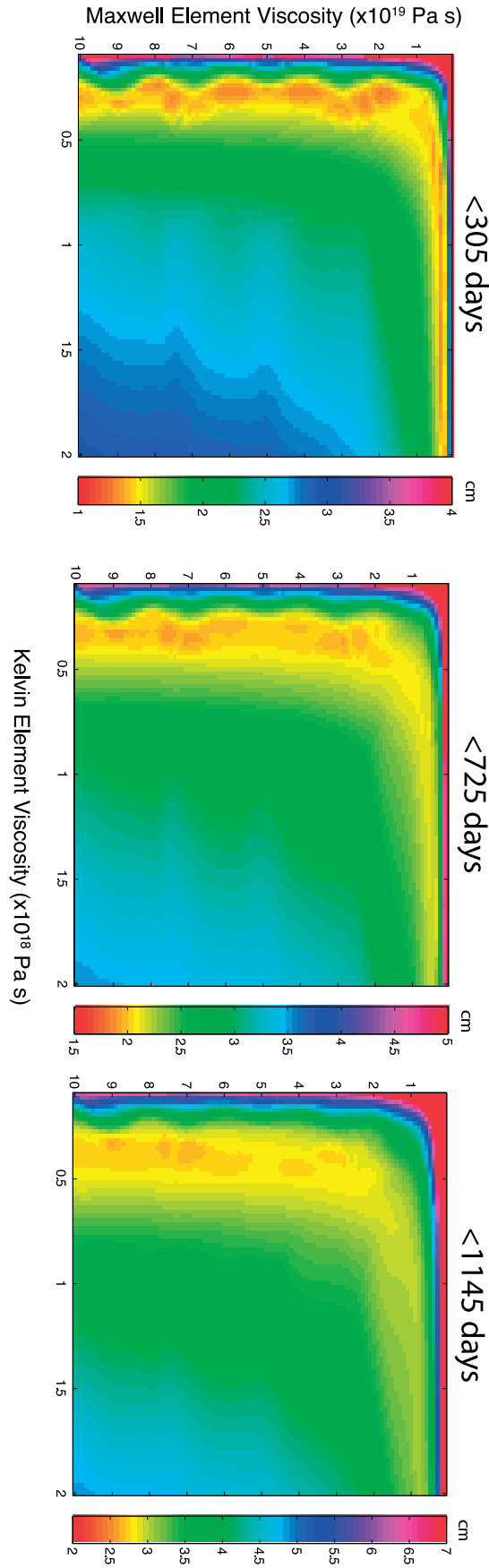


Figure 6.6: Best fitting parameter searches for the Burgers models and ERS data. The coloured backdrop denotes the cumulative RMS misfit between the data and models (as described in the text) over the ERS observation period in the title of each plot. We can see the location of the minimum RMS residual remains in approximately the same place throughout the observation period.

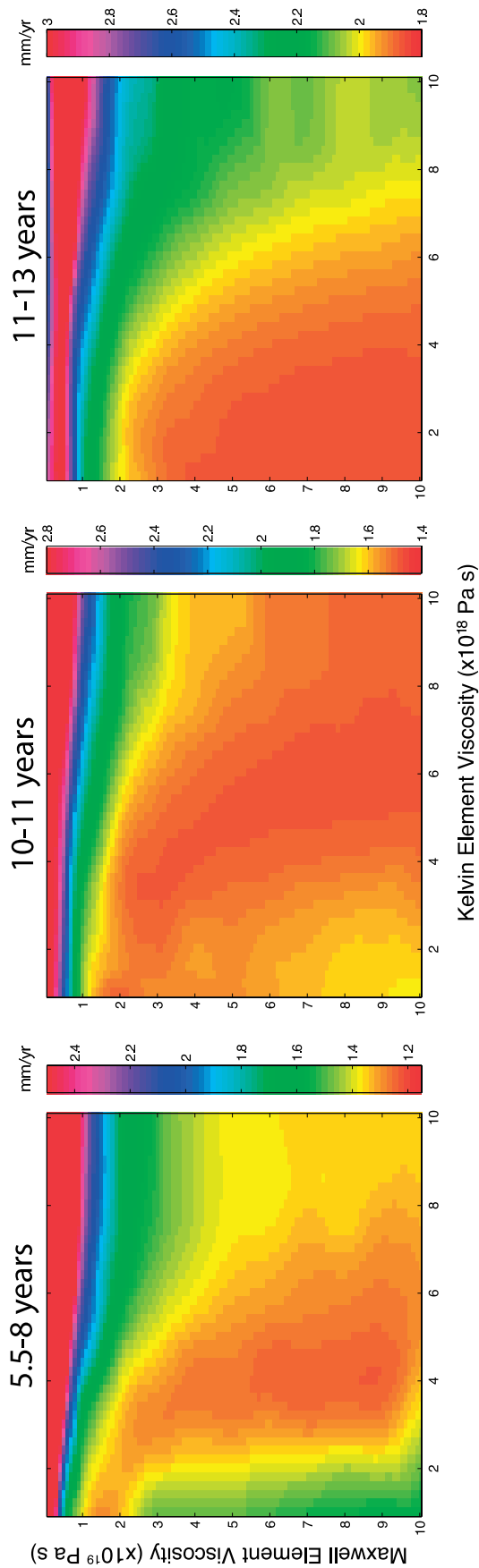


Figure 6.7: RMS parameter space of Burgers model with respect to 7.5 years of ENVISAT data. The minimum on the right panel (11-13 years) is bound by the edges of the plot. The location of the mean minimum misfit solution is given in Table 6.4 and Figure 6.8.

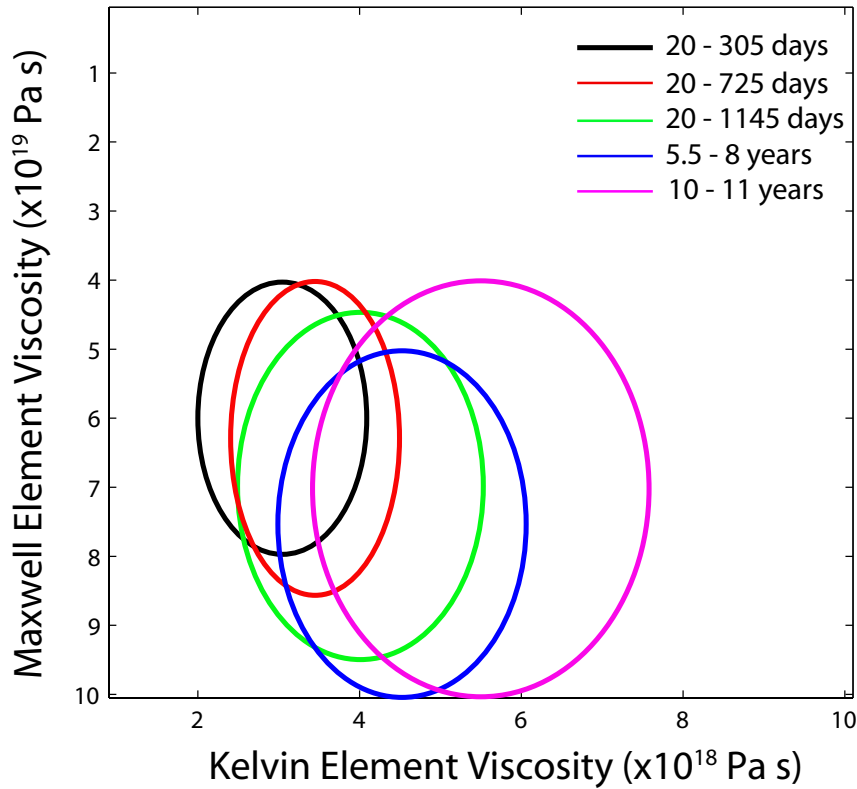


Figure 6.8: The location and error of the best-fitting minima solutions over five observation period that include both ERS and ENVISAT data. The values are derived from the range of minima in 100 perturbed datasets of the data.

data. The overlap of the error ellipses shows that there is a region of the parameter space that lies within the error bounds during all the observation periods analysed for the Burgers models.

6.4 Discussion

6.4.1 Viscoelastic Models

We modelled three types of earth-models during this chapter: Maxwell half-space, Layered Maxwell and Burgers half-space. No single model with a Maxwell half-space could match the data over the whole observation period for the ERS data. The best-fitting solution appeared to increase in apparent viscosity with time, a trend that appeared to continue with the ENVISAT data acquired 5.5-13 years after the earthquake. The EN-

VISAT suggest an increase in viscosity is preferred within the 5.5-11 year observation period however the increase is not resolvable.

The layered models show a similar trend to the Maxwell models especially when the layer is thin. As the layer thickness increases the difference in viscosity becomes less significant but is still resolvable. As time increases the ability to resolve an apparent increase in half-space viscosity with depth reduces. It is noted that the absolute misfit values of the channel models is comparable to Burgers model however the best model changes which is why it is disregarded. Another notable feature of these models is that, at early times, the best fitting models have a lower channel viscosity than halfspace viscosity. This is because of the shorter wavelength deformation pattern observed in the early stages. A stonger halfspace causes the deformation pattern to be more localised as is observed. This idea is dealt with further in Chapter 7 when discussing the *DeVries and Meade* (2013) channel model.

The Burgers modelling however does yield a model that, within error, can match the observed surface deformation over the entire study period (Figure 6.8). This model has a Kelvin element viscosity of $3.5\text{-}4.0 \times 10^{18}$ and a Maxwell element viscosity of $5.5\text{-}7.5 \times 10^{19}$ Pa s.

6.4.2 Implications of a Burgers Rheology

The best-fitting viscoelastic model is that of a Burgers body. We take the best solution as $\eta_m = 4.0 \times 10^{18}$ and $\eta_k = 6.0 \times 10^{19}$ Pa s. In the next chapter we compare the best-fitting Burgers Model to the best-fitting afterslip model, but we now consider the implications of a Burgers body model.

Burgers bodies are characterised by having initial elastic response followed by two deformation timescales, one that controls the initial deformation transient associated with the Kelvin viscosity element and the longer timescale deformation dictated by the Maxwell viscosity element. Burgers rheologies have been used by *Peltier et al.* (1981) to describe mantle rheology and the difference between upper and lower mantle viscosities (*Sabadini et al.*, 1987). Further studies of mantle rheology however have

shown that a Burgers model is only the first approximation as it cannot reproduce the observed frequency dependent Q of the mantle (*Ranalli, 1995*). To reproduce the weak frequency dependence we require several Kelvin elements. It is not clear however that Burgers Models could apply to the crust as would be the case observed here. There is little consensus as to whether the Burgers models are physically plausible or just good at mimicking other mechanisms that match the data (Barbot, pers comm).

Figure 6.9 shows the l-o-s displacement over time of one pixel in the deformation peak north of the fault (location 35.5°N , 87.0°E). It shows how the motion of one point changes over time. We show the deformation for three models, the preferred Burgers model of this study, the preferred SLS model of *Ryder et al. (2007)* and also a Maxwell half-space model with viscosity 4×10^{18} Pa s. We can see from the deformation curve how the SLS and Burgers models behave similarly in the first 1000 days which is the period covered by the ERS data. The two models diverge however after that, with the Burgers model continuing to deform at an appreciable rate whereas the SLS model slows down. This shows that with when considering only ERS data the Burgers and SLS are similar to each other and hence could both explain the dataset. The differences between them are only resolvable in the period of the ENVISAT dataset covering 5.5-13 years after the event where the deformation rates diverge dramatically.

6.4.3 Why the lack of resolution between models

The errors in Figure 6.8 show that the minima are not particularly well constrained. Figure 6.10 shows the difference between two models at the edge of the error envelope for the best-fitting model 10-11 years after the earthquake. We can see that the maximum difference between models $\leq 2\text{mm/yr}$ which is approximately the magnitude of the noise in the 10-11 year ratemap (1.7 mm/yr). In order to resolve between each model we require a lot smaller noise floor in the ENVISAT ratemaps.

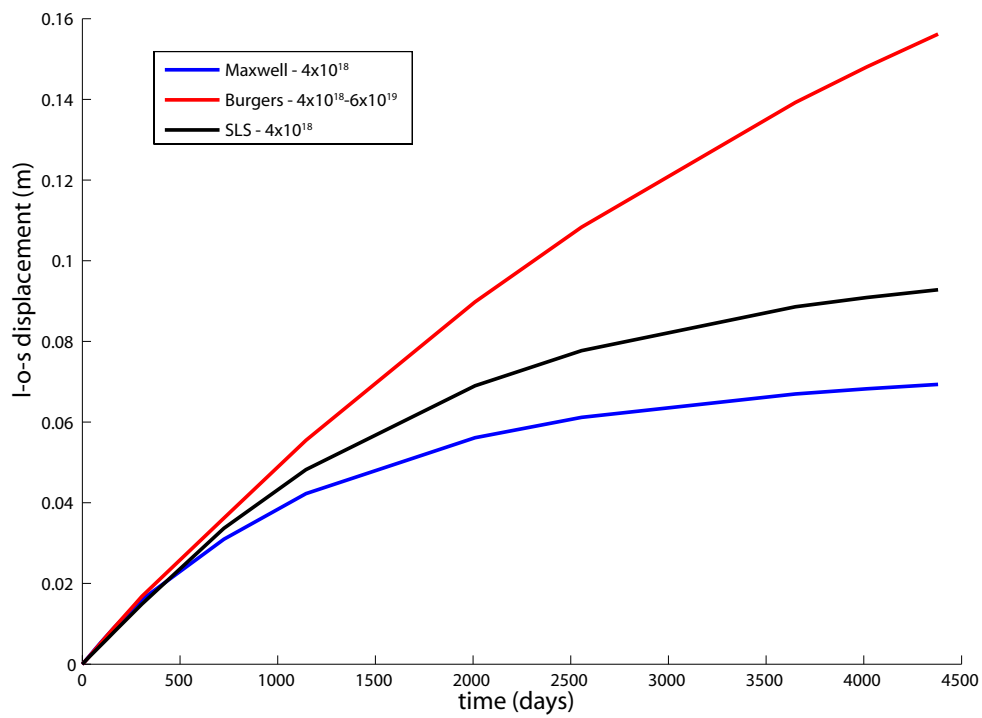


Figure 6.9: Line of sight displacement in the deformation peak north of the fault (35.5°N , 87.0°E). The models shown are the best-fitting models from this study, best-fitting from *Ryder et al.* (2007) and the best (but not well) fitting Maxwell model. It shows that the introduction of additional elements cause more motion longer after the event. It is this motion that initially made *Ryder et al.* (2007) prefer an SLS and also for this study to suggest a Burgers body.

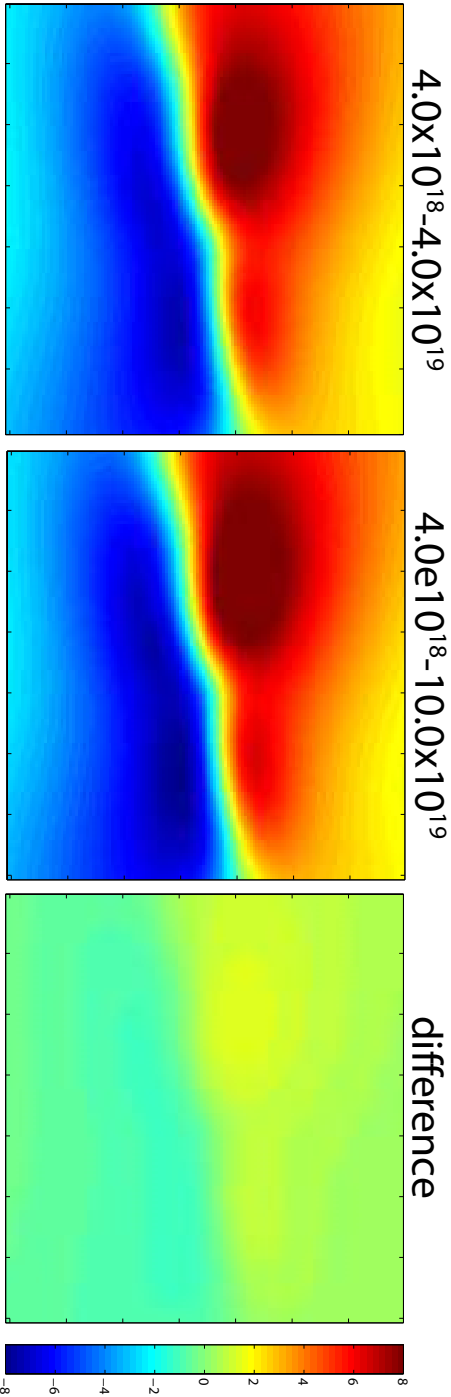


Figure 6.10: Difference in l-o-s deformation rate from 10^{-11} yr after the Manyi Earthquake. The Burgers models chosen lie in the edge of the error ellipse in Figure 6.8. The models do not have any linear ramp removed. They were calculated using PSGRN/PSCMP. All plots are on the same scale and are $150\text{km} \times 150\text{km}$. The difference between the peak l-o-s displacement is approximately 2mm/yr .

Chapter 7

Discussion

During this thesis a series of models of viscoelastic relaxation and afterslip were created to describe the postseismic motion for the Manyi Earthquake. Two postseismic datasets were used, one based on ERS SAR images acquired from 25 to 1145 days after the earthquake that is taken from *Ryder et al.* (2007). The other dataset uses ENVISAT SAR images from 5.5-13 years after the event. As the signal-to-noise ratio decrease with time we used a network approach to form three ratemaps which increase the ratio as described in Chapter 3. There are three major mechanisms of postseismic deformation: viscoelastic relaxation, frictional afterslip and poroelastic relaxation. We do not consider poroelastic relaxation as it was eliminated by *Ryder et al.* (2007) as implausible in this setting. We do however model afterslip and viscoelastic relaxation.

Previous studies of afterslip of the Izmit, Manyi and Gorny-Altai Earthquakes (*Hearn et al.*, 2002; *Ryder et al.*, 2007; *Barbot et al.*, 2008) use the ground motions to determine a distribution of afterslip. As these models are derived from the data, they always tend to fit the data well. However there is no geophysical mechanism behind the distribution and evolution of slip in these kinematic inversions. We attempt to produce afterslip models that are based on frictional rate-and-state laws, which relate afterslip to the coseismic stress change. We use a hot friction approximation which means the system is not affected by the state variable. This approximation assumes the postseismic rates are large enough that the frictional contacts do not stay in touch

for long enough to heal. This is likely to be true in the immediate aftermath of the earthquake however as the rates reduce this assumption becomes increasingly less valid. To incorporate healing effects we would have to include the state variable which is not known for real fault systems. We find a range of afterslip models which fit both datasets well.

We create a series of viscoelastic models of increasing complexity, starting with a simple Maxwell half-space below a surface elastic layer, followed by a viscoelastic layer sandwiched between an elastic surface layer and viscoelastic half-space. We then model a Burgers half-space beneath an elastic layer. We do not attempt to model a Burgers Layer as the number of free parameters would become unmanageable. We do not repeat the Standard Linear Solid (SLS) modelling of *Ryder et al.* (2007) because the Burgers model approximates an SLS as the Maxwell element viscosity tends to infinity. We find that the only viscoelastic model capable of reproducing the observed deformation field is a Burgers rheology model.

We compare the two best-fitting models for each deformation mechanism to ascertain if one model fits the data better than the other. We discuss why we do not attempt to combine the two mechanisms to form a potentially better fitting solution and also whether each mechanism is in fact mimicking the behaviour of another form of rheology which is not considered in this thesis, a power-law rheology.

Different postseismic studies of the same event (e.g Denali, Izmit, Parkfield) can produce different best-fitting models, even when sometimes using the same data. With this in mind we compare our results with three other published models of the Manyi postseismic deformation.

7.1 Afterslip vs Viscoelastic Relaxation

The previous two chapters modelled two distinct types of postseismic motion and determined that there is a model for each type of deformation mechanism which fits both the ERS (Figure 7.1) and ENVISAT data (Figure 7.2) well. We compare the

best-fitting model of a Burgers rheology with a Kelvin and Maxwell element viscosity of 4×10^{18} Pa s and 6×10^{19} Pa s respectively to an afterslip model with a preseismic slip rate of 10 mm/yr and $(a - b)$ of 1×10^{-3} . Although there are a series of well fitting afterslip models we choose the one with lowest $(a - b)$ as the evolution of the RMS well through time suggest that the lowest preseismic slip rate and $(a - b)$ are the most stable.

The RMS misfit of both models using both the ERS and ENVISAT datasets are shown in Figures 7.3 and 7.4 respectively. Perturbed datasets (as calculated in the previous chapters) are used to calculate the error of the misfit to determine if the RMS difference is resolvable.

Figures 7.3 and 7.4 show that the afterslip model fits slightly better than the Burgers model though for all time periods the error bars either overlap or are adjacent to each other. The only other studies of postseismic motion on the Tibetan Plateau focused on the Kokoxili Earthquake of 2001 is (*Ryder et al.*, 2011) and *Wen et al.* (2012). *Ryder et al.* (2011) used GPS and InSAR data and found a best-fitting Burgers rheology with a Kelvin and Maxwell element viscosity of 9×10^{17} Pa s and 1×10^{19} Pa s respectively which is similar to the results found in this investigation although both values are lower than those found here. *Wen et al.* (2012) use InSAR data and find an effective viscosity of 2×10^{19} Pa s. As they do not have InSAR data covering the early stages of the postseismic motion they limit their modelling to Maxwell bodies however the value they obtain is similar to the long-term viscosity from this study.

If we assume however that the motion is predominantly one mechanism there are implications for the rheology or fault properties around the Manyi Fault.

7.1.1 Afterslip Only

If the motion is solely due to postseismic afterslip using the models described earlier then there can be little to no viscoelastic relaxation. We know however that in the long term rocks behave viscously so viscoelastic motion should occur after large earthquakes as they cause significant stress changes below the seismogenic layer. If we do not observe

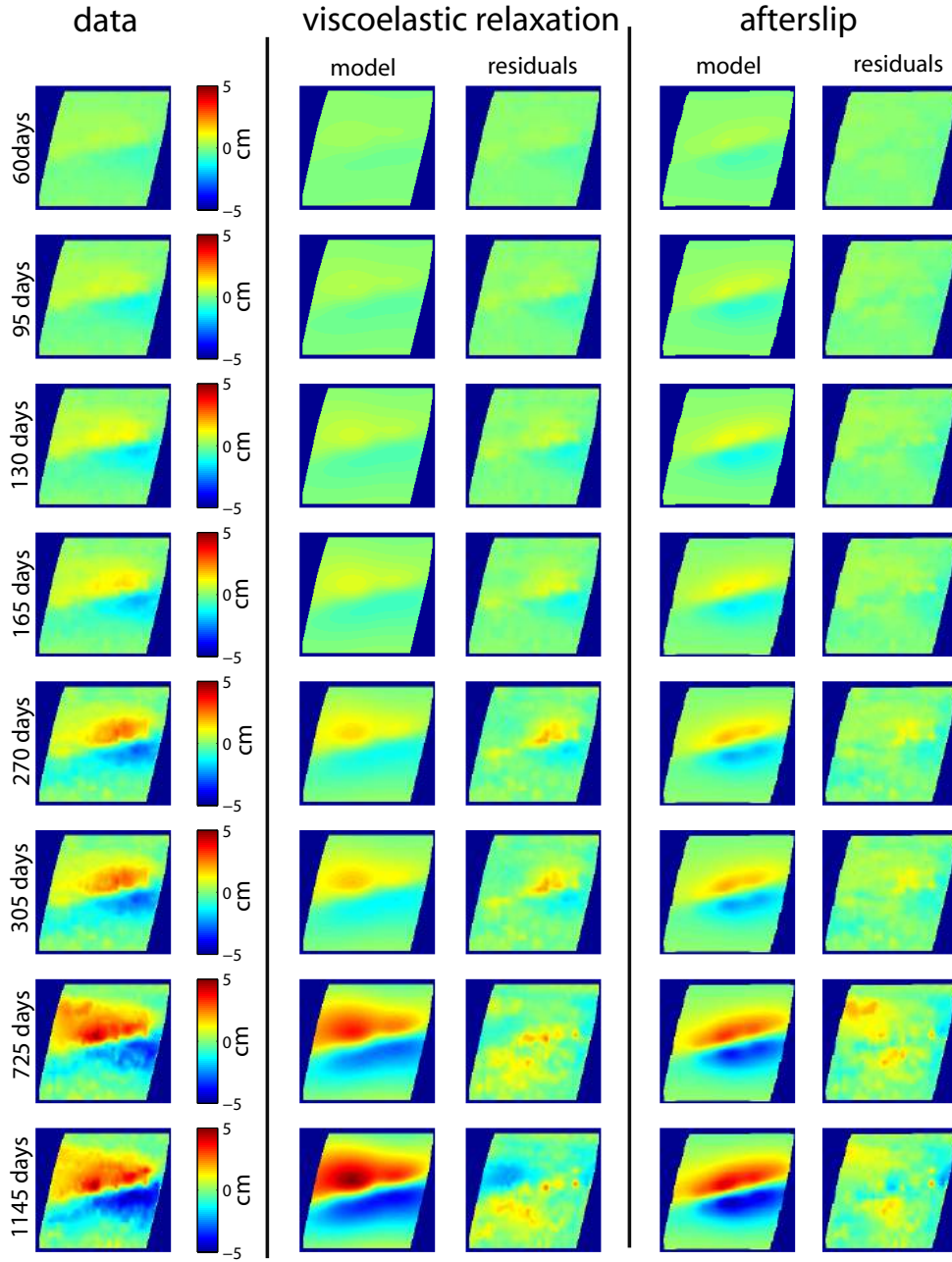


Figure 7.1: The cumulative displacement at various times after the postseismic. The three sets of images shown are the ERS timeseries from *Ryder et al. (2007)* the best-fitting viscoelastic model from this study and the best-fitting afterslip model as described in the text. All images are all measures of line-of-sight displacement in the look direction of track 305. Also shown are the residual deformation between the model and data.

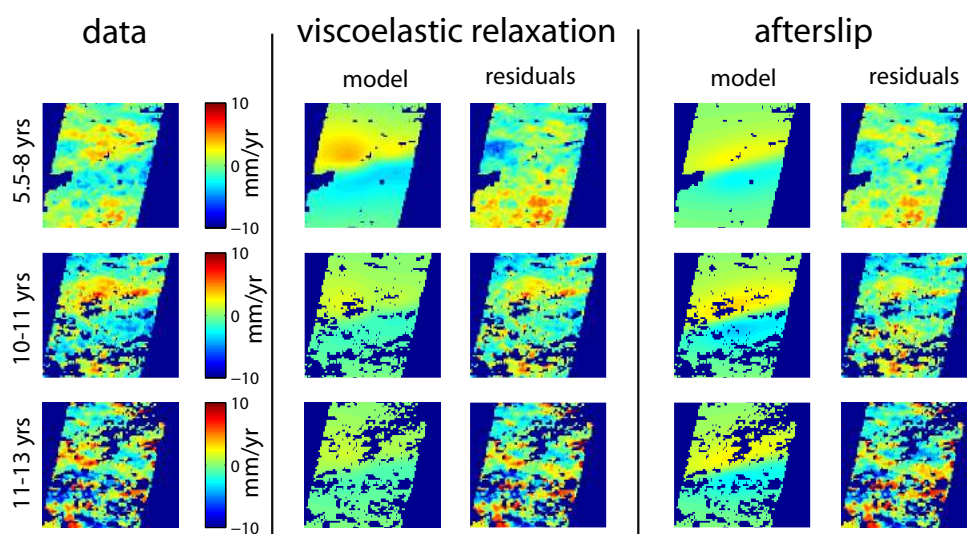


Figure 7.2: The displacement rates at various times after the earthquake. The three sets of images shown are the ENVISAT ratemaps from Chapter 3, the best-fitting viscoelastic model from this study and the best-fitting afterslip model as described in the text. All images show line-of-sight velocity in the look direction of track 305. Also shown are the residual between the model and data.

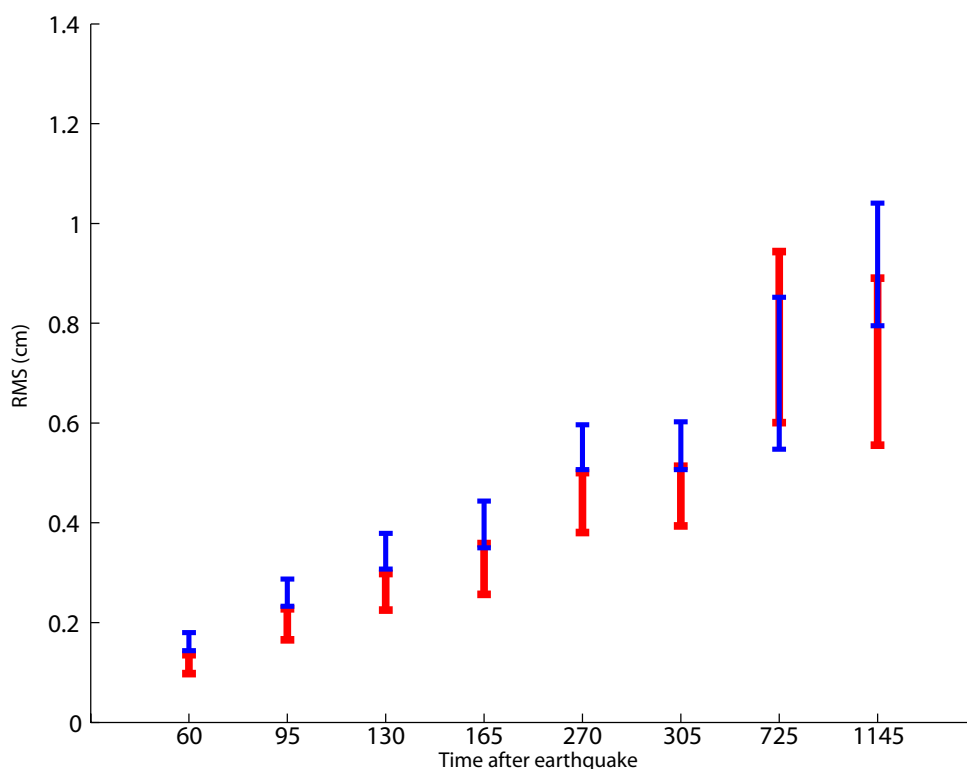


Figure 7.3: The RMS between the model and residual for the best-fitting afterslip (red) and viscoelastic model (blue) at various times during the ERS observation period. The error bars are estimated by calculating the distribution of misfit for a series of 50 perturbed datasets. It should be noted the x-axis is not a true scale in time but a scale of observation points.

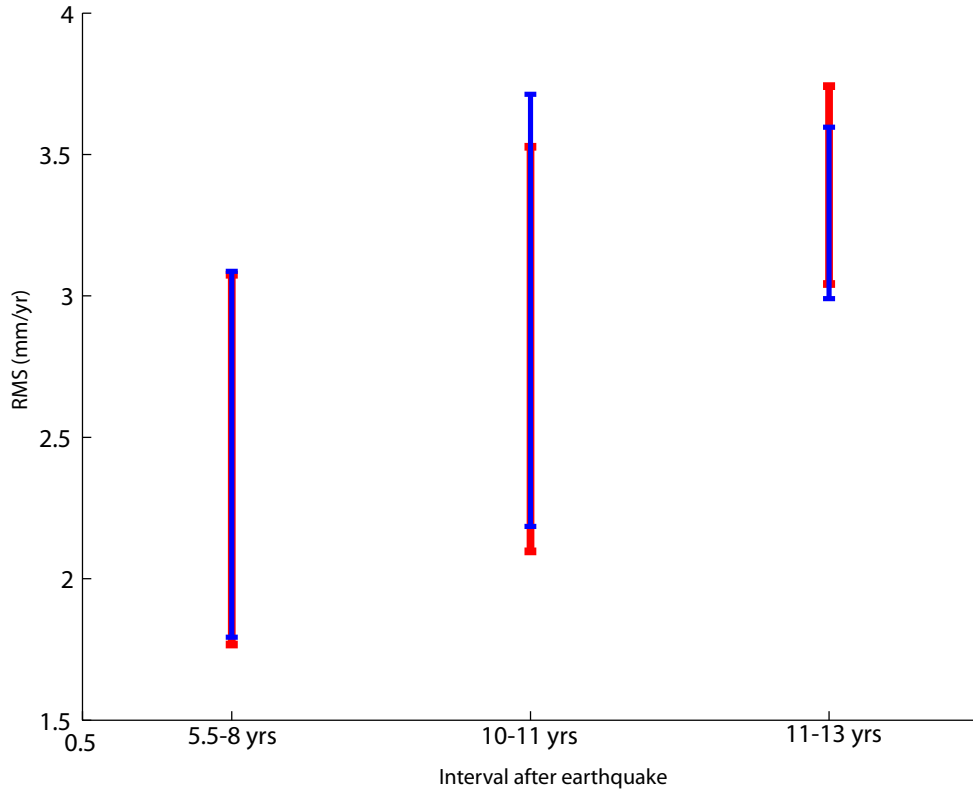


Figure 7.4: The RMS between the model and residual for the best-fitting afterslip (red) and viscoelastic model (blue) at various times during the ENVISAT observation period. The error bars are estimated by calculating the distribution of misfit for a series of 50 perturbed datasets.

any viscoelastic relaxation it means that the rates are slow and thus below the detection threshold of the data. As the rate of motion due to viscoelastic relaxation is smaller for more viscous rheologies the region below the seismogenic layer must be very viscous ($> 10^{20}$ Pa s).

If the signal were due to afterslip the Manyi Fault would show the longest duration of afterslip of any earthquake ever studied. It should be noted however that generally the observation period tends to be the limiting factor in most postseismic studies.

7.1.2 Viscoelastic Relaxation Only

If we assume the postseismic relaxation is solely viscoelastic relaxation of a Burgers body with a Kelvin and Maxwell element viscosity of 4×10^{18} Pa s and 6×10^{19} Pa s respectively then there is little to no afterslip. If this is the case than it would indicate

that either the friction ($a - b$) across the Manyi Fault is higher than those tested in this study or the critical distance for friction healing is much larger than the initial afterslip. Both of these may be the case in immature fault zones (*Marone, 1998*), however this is not expected to be the case across the Manyi Fault as it is part of the mature Kunlun fault system.

7.2 Multiple Mechanisms?

During this thesis both mechanisms have been treated independently. It is far more likely however that both are, at least in some way, responsible for the surface deformation that is observed. We do not attempt to form a combined model for two reasons. Firstly, if it is possible to match the data using one model alone, combining the deformation from two models is likely to produce a series of model combinations which match the data and would provide little useful geophysical information. Secondly combining two mechanism of stress change would require that each model respects the stress changes in the other. The methods used to form the two model types in this thesis cannot be combined so a new approach would be required.

7.3 Power-law Rheology?

The rheologies modelled in this thesis have all been Newtonian, i.e. the strain-rate is a linear function of stress. Laboratory experiments suggest that stress-dependent rheology would arise from the occurrence of dislocation creep as the primary mechanism of viscous deformation (*Carter and Tsenn, 1987*). We expect, for a power-law rheology, to observe an increase in apparent viscosity with time due to the decrease in stress, which is consistent with what we see in this study (Equation 7.1).

$$\dot{\epsilon} \propto \sigma^n \quad \eta_{eff} \propto \frac{1}{\sigma^{n-1}} \quad (7.1)$$

where n is the power law co-efficient.

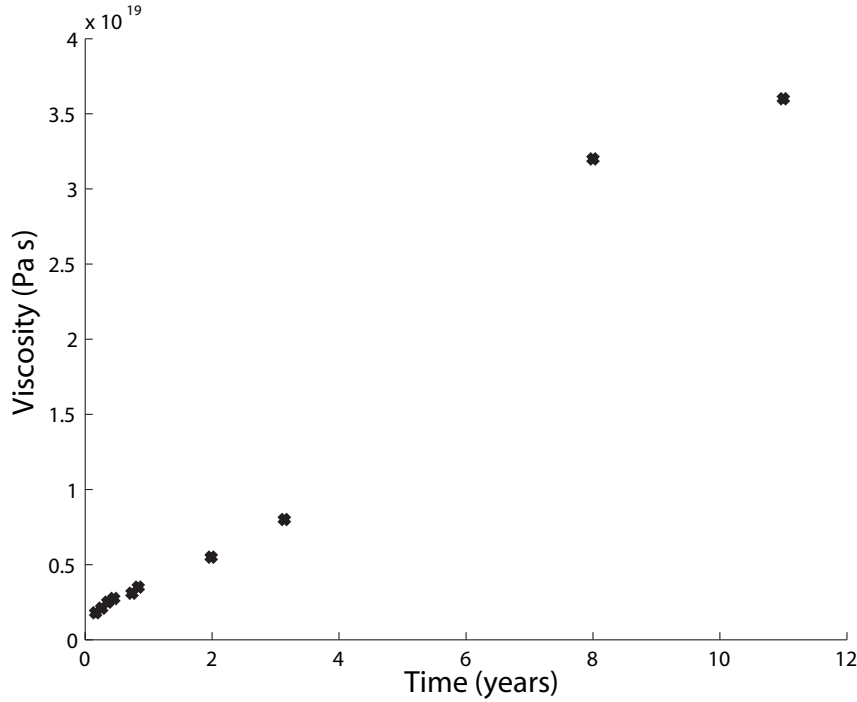


Figure 7.5: Best fitting apparent Maxwell viscosity over an 11 years time period. We fit the best fitting Maxwell viscosity at the time period of observation. We do not include the final ratemap from 11-13 yrs as it is so noisy.

The best-fitting Burgers model changes with time. As time increase the viscosities of both elements tends to increase, behaviour similar to that expected from a power-law rheology. However the errors make it impossible to eliminate or confirm a gradual increase in characteristic deformation timescale (Figure 6.8)

A range of afterslip models appear to fit the data. Whilst this could be due to uncertainties in the data it may suggest that the deformation is not due to afterslip but instead the afterslip models are mimicking a different phenomenon with similar behaviour. It is well documented that the exponential relationship between slip and stress in hot friction afterslip can mimic power-law behaviour (*Perfettini and Avouac, 2007*).

Figure 7.5 shows the best-fitting apparent Maxwell viscosity through time. It shows how there is a clear trend for increasing apparent viscosity, behaviour that is also expected from a power-law rheology.

7.4 Comparison with Other Postseismic Manyi Studies

There have been three previous studies focused on the postseismic deformation of the Manyi Earthquake. The first was the *Ryder et al.* (2007), the second by *Yamasaki and Houseman* (2012) and the lastly *DeVries and Meade* (2013). All studies use the same ERS data as in this study but all have come to slightly different conclusions about the best-fitting model for postseismic deformation after the Manyi event. We discuss the results of each study below and how their results compare to those in this study.

The first study by *Ryder et al.* (2007) showed that either afterslip or viscoelastic relaxation of a Standard Linear Solid with a viscosity of 4×10^{18} Pa s can describe the observed motion. This study shows that the deformation can be modelled using a Burgers rheology with Kelvin and Maxwell element viscosity of 4×10^{18} Pa s and 6×10^{19} Pa s respectively. The feature of both models that enable them to match the data is the variable characteristic timescale of deformation over the observation period. The only difference between the two models is the addition of a viscous element in series to the SLS model to form the Burgers model (Figure 2.5). The viscosity of the extra (Maxwell) element required in this study is an order of magnitude higher than the Kelvin viscosity and so only has a large effect over the decade timescale (ENVISAT data) and the behaviour of both the viscoelastic models is similar in the first few years after the earthquake. This explains why the SLS was sufficient to explain the data that was available to *Ryder et al.* (2007) as it only covered the first 3 years of postseismic signal.

The second study (*Yamasaki and Houseman*, 2012) uses a depth dependent viscosity model to describe the motion after 165 days. The first 165 days of motion are attributed to afterslip, a mechanism that is shown to be reasonable in both this study and *Ryder et al.* (2007). However, the afterslip models in this thesis show that the afterslip models may be valid beyond the first 165 days. The depth dependent model from *Yamasaki and Houseman* (2012) assumes that any deformation after 165 days is a result of viscoelastic

flow rather than afterslip. The afterslip models in this thesis all assume that the hot friction approximation is valid for the lifetime of the model. Eventually however, as the slip rate reduces, the hot friction approximation will break down, as the state variable can no longer be neglected and the frictional contacts will start to heal. The point at which this occurs can be determined for laboratory scale experiments but there is no current consensus as to when this should occur on real faults (*Marone, 1998*). The healing process will cause an abrupt halt of afterslip. The model of *Yamasaki and Houseman (2012)* would be valid if healing occurs at around 165 days. In the half-space models from this thesis we assume no depth dependence of viscosity. There is however a significant geothermal gradient with depth. As temperature is one of the major controls on viscosity so a depth dependence of viscosity is justified.

A third study *DeVries and Meade (2013)* combines pre-seismic data from *Bell et al. (2011)* and postseismic data from *Ryder et al. (2007)* to produce a model which explains both sets of observations. As this model attempts to explain two stages of the earthquake cycle simultaneously we discuss this study in more detail in the next section which looks at the earthquake cycle of the Manyi Fault.

7.5 The Earthquake Cycle of the Manyi Fault

It is desirable to have a single model capable of explaining both the preseismic and postseismic deformation rates. The studies in this thesis and previously treated the two stages of earthquake cycle independently. The study by *DeVries and Meade (2013)* uses what is known as a spun-up model to attempt to describe both stages using the same model. They study two Tibetan earthquakes, 1997 Manyi and the 2001 Kokoxili Earthquake. We will concentrate here on the Manyi event only.

DeVries and Meade (2013) use an earthquake cycle model which incorporates the effects of previous earthquakes across the Manyi Fault (which they assume to be a continuation of the Kunlun Fault). They assume the fault is infinitely-long with a long-term slip rate of 11.2mm/yr based on results from the measurement of the Kunlun fault

slip rate by *van der Woerd et al.* (1998); *Van der Woerd et al.* (2000); *Van Der Woerd et al.* (2002). Using this slip rate and the total slip in the Manyi Earthquake from *Funning et al.* (2007b) they assume the repeat time of earthquakes across the fault is 300 years. As the Kunlun Fault has accumulated many tens of kilometres of offset over the last ≈ 15 Ma (*Jolivet et al.*, 2003) it must have experienced a very large number of earthquakes. As the models assume a linear Maxwellian rheology the response at any given time is the sum of all the previous earthquakes (*Savage and Prescott*, 1978; *Hetland and Hager*, 2005). Eventually, as the number of earthquakes included increases, the interseismic velocity approaches a state which is cycle invariant. *DeVries and Meade* (2013) show that the cycle invariant velocities in the Fourier domain for a model with a mid-crustal viscoelastic layer can be stated as

$$v = \Delta u (C_1 e^{-\beta_1 t} + C_2 e^{-\beta_2 t}) \quad (7.2)$$

where Δu is the coseismic slip and C_1, C_2, β_1 and β_2 are functions of the mid-crustal viscoelastic layer thickness, depth of elastic layer, depth of coseismic rupture, Maxwell time, earthquake repeat time and wavenumber (see *DeVries and Meade* (2013), Appendix A).

Data from *Bell et al.* (2011) (Chapter 4) are used to describe the interseismic deformation rate for the five years immediately preceding the earthquake. They also use a statement from *Ryder et al.* (2007) to derive an average fault parallel deformation rate of ± 75 mm/yr in the first 725 days of postseismic motion at a distance 20 km from the fault. They then use these data to find the best-fitting model to each dataset independently. They ascertain that one model can be used to describe both datasets within the data errors; a thin viscous channel of width 10 km and viscosity 1×10^{18} Pa s, which lies above an elastic half-space and below a 20 km thick elastic layer which is assumed to have fully ruptured during the Manyi Earthquake. Figure 7.6 reproduces the results of the modelling in *DeVries and Meade* (2013). The model data are created using the MATLAB code provided with the paper, the interseismic data are taken from Chapter 4, however we also include a 20 km wide profile using the postseismic

timeseries data from *Ryder et al.* (2007) that describes the average rate of the first 725 days of postseismic motion. The postseismic data profile is taken perpendicular to the fault and is 20 km wide centred on the main deforming area.

Figure 7.6 shows that the data point used in the postseismic comparisons by *DeVries and Meade* (2013) does not correspond well with the timeseries data. This is because the point was derived from comments about one interferogram in *Ryder et al.* (2007) whereas the data used here is from the time series of *Ryder et al.* (2007) that has been smoothed both spatially and temporally during its formation. The data from *Bell et al.* (2011) that are used however contain an error that is not accounted for by *DeVries and Meade* (2013). As the deformation map is derived from InSAR there is a fault perpendicular tilt error/orbital correction that was simultaneously solved for when deriving the slip rate (as described in Chapter 4). We recalculate the best-fitting model for the pre and postseismic motion however we use the timeseries data from *Ryder et al.* (2007) for the postseismic data and also solve for a tilt error (Figure 7.7).

When tilt errors and more data are incorporated into the analysis the model proposed by *DeVries and Meade* (2013) is not suitable to explain the pre and postseismic motion near the Manyi Fault. Figure 7.7 also shows how the best-fitting apparent viscosity increases as the observation period increases. This observation is very similar to the results found earlier in this thesis when dealing with viscoelastic models. The model of *DeVries and Meade* (2013) assumed that the Manyi Fault is an extension of the Kunlun Fault and used this assumption to attribute a long-term 11 mm/yr slip rate across it. *Kirby et al.* (2007) show that the Kunlun Fault does not have a constant slip-rate at its eastern end where the slip is estimated to be around 3-4 mm/yr. If the same decrease applies to the western end of the Kunlun then their assumption is incorrect. Furthermore there are a number of possible western extensions of the Kunlun Fault (Figure 3.2) which could share the 11 mm/yr of slip across them meaning any one fault is moving comparatively slower. We repeat the modelling in Figure 7.7 however we use the long term slip of 3 mm/yr as determined in Chapter 4 and an earthquake

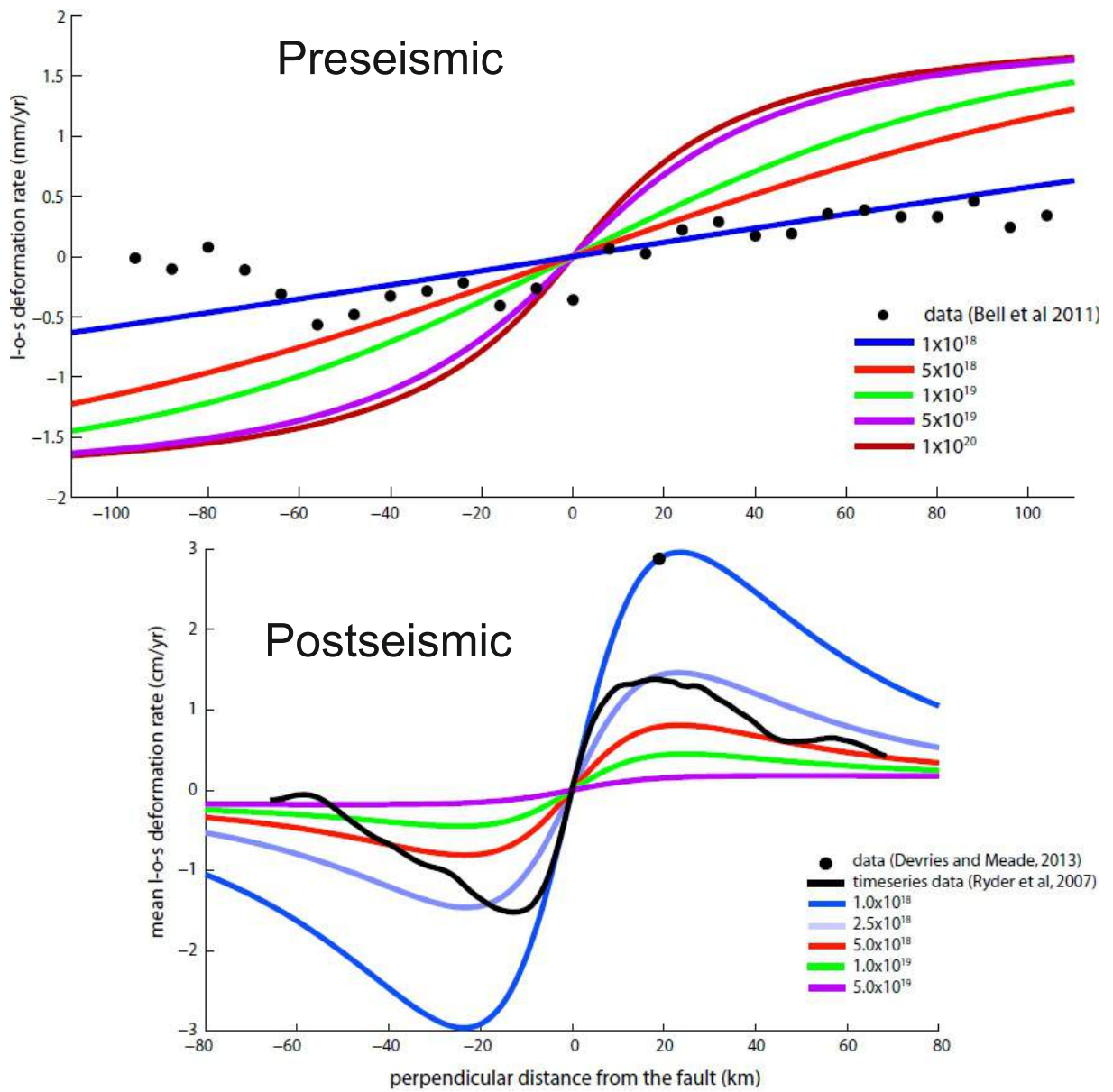


Figure 7.6: Preseismic (top) and postseismic (bottom) line-of-sight deformation profiles of various models and observations of the Manyi Earthquake. Models used to derive the deformation patterns have the geometry of the best-fitting model for the Manyi Earthquake in *DeVries and Meade (2013)* and viscosities ranging 1×10^{18} Pa s to 1×10^{20} Pa s. The interseismic data are taken from *Bell et al. (2011)*, the postseismic timeseries data are taken from *Ryder et al. (2007)* and the *DeVries and Meade (2013)* data point is taken from the supplementary materials of the paper. This figure is comparable to the upper left and upper middle panel of Figure 7 from *DeVries and Meade (2013)* although a factor of 0.35 (corresponding to the look vector of track 305) has been used to convert the models from horizontal motion to l-o-s motion.

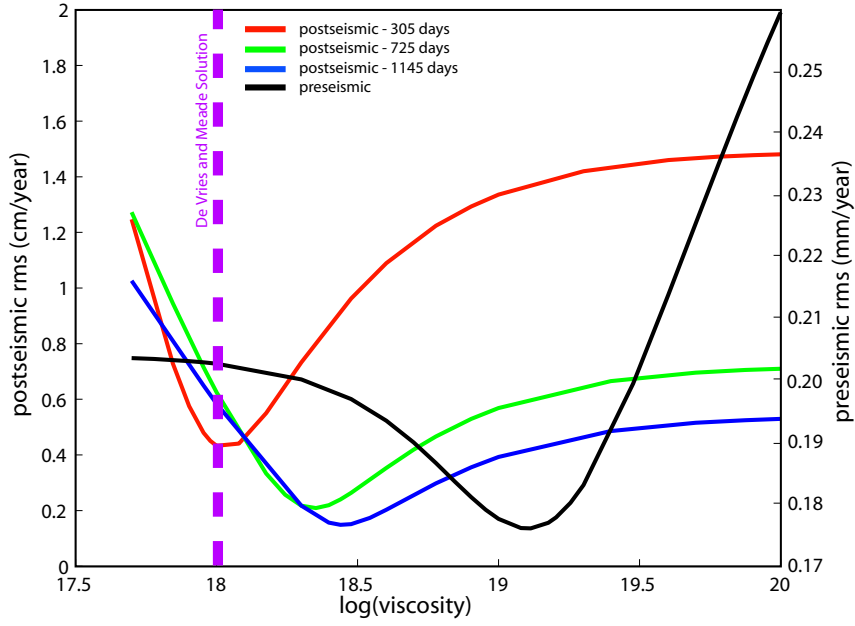


Figure 7.7: RMS misfit of the thin channel model geometries proposed by *DeVries and Meade* (2013). The best-fitting viscosity for both the post and preseismic data does not correspond to the same value as the best-fit found by *DeVries and Meade* (2013) (Purple line). We show the model misfit using three different observation periods using the ERS timeseries. The green line corresponds to the observation period in *DeVries and Meade* (2013). It can be seen how the best viscosity increases with the length of the observation period. Note all the postseismic deformation models are calculated from 25 days after the event.

repeat time of 1300 years. The best fitting solutions for this model setup (Figure 7.8) are similar in form to those in Figure 7.7 however the minima lie approximately an order of magnitude higher. Again there is no one solution which matches both the pre and postseismic motion well over the whole time period.

It is worth noting that when the interseismic strain accumulation rate of the model is reduced by an order of magnitude the best-fitting model viscosity increases by an order of magnitude. Figure 7.9 plots the best-fitting preseismic model viscosity for a series of models with different assumed slip rate. It shows that there is a linear relationship between the assumed slip rate and the best-fitting model viscosity. This makes the thin viscous channel model of *DeVries and Meade* (2013) highly dependent upon reliable a priori estimates of long-term slip rates something that we do not have across the Manyi Fault.

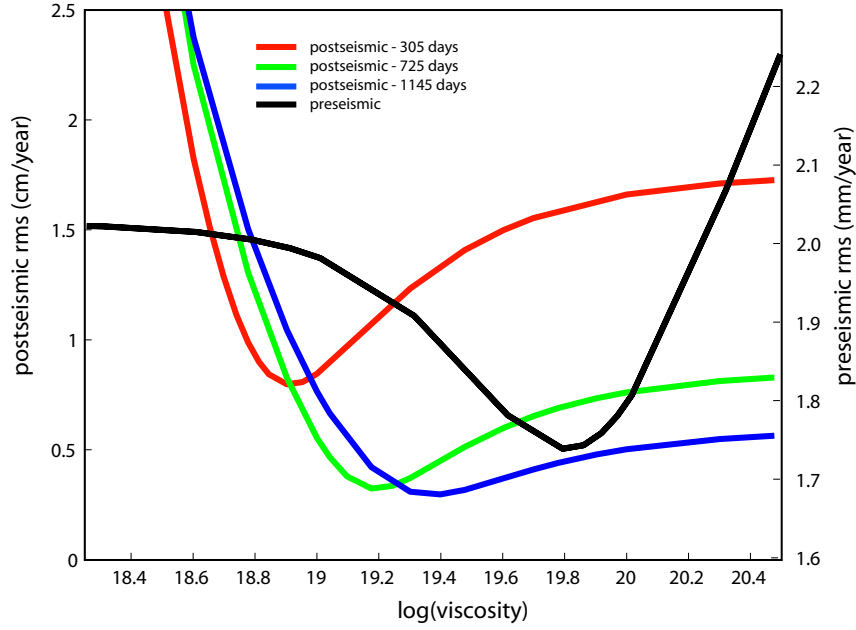


Figure 7.8: RMS misfit of the thin channel model geometry proposed by *DeVries and Meade* (2013) but with a slip rate of 3 mm/yr assumed rather than 11 mm/yr. The best-fitting viscosity for both the post and preismic data does not correspond to the same model. We show the model misfit using three different observation periods using the ERS timeseries. The green line corresponds to the observation period in *DeVries and Meade* (2013). It can be seen how the best viscosity increases with the length of the observation period. Note all the postseismic deformation models are calculated from 25 days after the event.

7.5.1 2-D assumptions

One of the reasons a thin channel model is preferred by *DeVries and Meade* (2013) is that it allows rapid near-field postseismic motions but still satisfies a constant far-field rate, a behaviour that is not possible using their equivalent half-space model. They also state that a channel model can produce higher spatial displacement gradients nearer the fault than a half-space model. The models used in *DeVries and Meade* (2013) however all assume an infinitely-long fault. We consider the steady state ($t \rightarrow \infty$) displacements due to a fault of finite length in the half-space model (Figure 7.10). We see that the length of the fault limits the far-field signal, effectively reproducing the far-field behaviour of the channel model. Earthquakes occur on finite length faults so the pattern shown here is more realistic.

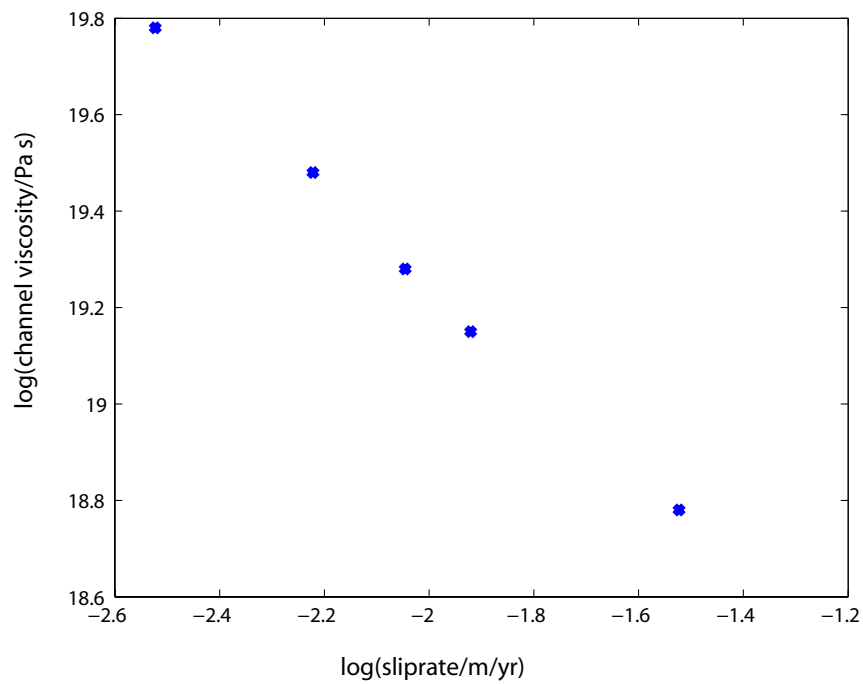


Figure 7.9: Model slip-rate against best-fitting preseismic model viscoisty. The slip rate is varied (between 3 mm/yr and 3 cm/yr) and the best-fitting channel viscosity is determined using the model in *DeVries and Meade* (2013) and the preseismic data from Chapter 4. Note the scales are both logarithmic

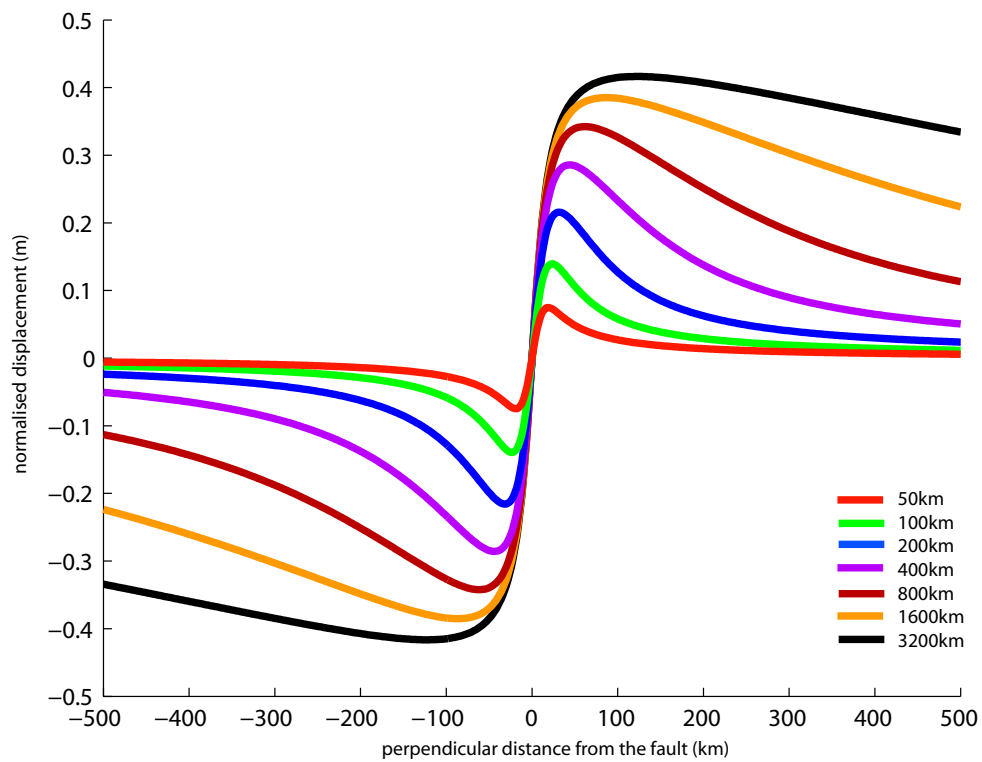


Figure 7.10: Postseismic deformation profiles perpendicular to a fault of varying length. The same rheology is used for all models (T_M of 1 year) and the same model geometry, a 15km elastic lid over a half-space. The models are calculated using PSGRN/PSCMP.

7.6 Conclusions

During this thesis we have demonstrated that:

- There is still significant postseismic motion of ± 8 mm/yr following the Manyi Earthquake from the period 2003-2011.
- The interseismic strain accumulation rate of the Manyi Fault is 3 ± 2 mm/yr with a locking depth of 22 ± 15 km.
- There is no evidence of significant strain on the fault to north of the Manyi Fault which may be an extension of the Kunlun Fault.
- Models with one characteristic timescale cannot reproduce the deformation pattern unless that characteristic timescale increases with time.
- The postseismic motion of the Manyi Earthquake can be explained by two mechanisms, afterslip of a fault with a range of frictional parameters and also viscoelastic relaxation of a Burgers body. The afterslip model is a slightly better fit to the data however the errors do not allow us to resolve the difference between the models.
- The results of this study are in agreement with both of the previous studies that focus on the Manyi postseismic motion only.
- A model that incorporates a series of historic earthquakes using a thin, low viscosity channel from 20-30 km depth cannot be used to explain the observed pre and postseismic motion of the Manyi Fault.

There are questions that remain as to the postseismic deformation mechanisms across the Manyi Fault despite the fact that we use the longest continuous (from the earthquake) observation period of all postseismic studies in literature.

7.7 Future Work

When this thesis was first devised the hope was that using more observations later after the Manyi earthquake would enable us to distinguish between the two mechanisms proposed by *Ryder et al.* (2007). This thesis has not been able to do this conclusively for a number of reasons. One of the reasons is that InSAR has an inherent error associated with the orbital baseline that exists in each interferogram. Although the orbital errors can be calculated using techniques described in the thesis they are still problematic when dealing with large scale (100 km) deformation such as postseismic motion. The far-field deformation from viscoelastic relaxation increases with time after the event. If this far-field effect is within the extents of the interferograms it is impossible to distinguish from any orbital ramp. In this thesis we removed any orbital ramp from the residuals but this adds more degrees of freedom in the inversions and thus makes models more difficult to differentiate. It would be preferable to not require an orbital correction of the timeseries. The majority of other studies of postseismic deformation have included GPS data (*Hearn et al.*, 2009; *Bruhat et al.*, 2011). These data can be used to calibrate the tilt errors within the InSAR and hence reduce the influence of orbital errors. Unfortunately placing GPS monitoring stations in the area of the Manyi fault is not realistic due to its remote location.

There is still significant noise in the ratemaps from the ENVISAT period. Unfortunately there are no further scenes that can be added to the ENVISAT series as the satellite is no longer operational. There is however another ESA satellite, Sentinel I, that will be launched in 2014 that will be able to determine the continuing motion in the Manyi area. Data from this satellite could be studied to ascertain the magnitude of any observing signal in the area and further refine the modelling.

As computational capacities increase modelling power-law rheologies become increasingly viable. There have also been numerous studies that propose laterally and depth variable rheologies. Using models of these rheologies and comparing with the data in this study could further constrain our understanding of the earthquake cycle.

A range of viable afterslip models with different V_0 and $(a - b)$ values are capable of

reproducing the observed deformation field. These models are very different however in the amount of slip that occurs during the the first 25 days after the earthquake. The slip in the first 25 days is contained within the coseismic interferograms. Removing the modelled slip from the coseismic model allows us to remodel the afterslip using revised slip distributions. This may allow us to constrain the range of viable afterslip models further as some may no longer match the data as well.

We have already discussed the limitations of modelling each deformation mechanism on its own. It is far more likely that the surface motion we observe is due to a combination of mechanism. A technique that combines both can solve for both model types simultaneously would allow us to produce integrated models which can distinguish between the contribution of each mechanism. We must exercise caution however as increasing the number of deformation mechanism that can be solved for simultaneously dramatically increases the number of free parameters in the inversion. It is possible that the best-fitting model is not be physically realistic as the problem is becomes increasingly less constrained. A possible solution to this would be to use Bayesian statistics to determine if the increases in free parameters can be justified with the available data

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Appendix A

Publication

The following pages reproduce the published paper below which constitutes the work from Chapter 4.

Bell, M., J.R. Elliott & B. Parsons (2011), Interseismic strain accumulation across the Manyi fault (Tibet) prior to the 1997 Mw 7.6 earthquake, *Geophysical Research Letters*, V. 38, L24303, doi:10.1029/2011GL049762.

Interseismic strain accumulation across the Manyi fault (Tibet) prior to the 1997 M_w 7.6 earthquake

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[1] The coseismic displacements caused by the M_w 7.6, 1997 Manyi strike-slip earthquake have been extensively studied. In order to assess whether the current deformation around the Manyi fault is due to one or more postseismic mechanisms and to constrain the rate-state models of after-slip, an estimate of the interseismic motion across the fault prior to the earthquake is needed. We use ESA ERS data to form 20 interferograms covering the five year period of 1992–1997, which are combined using a multi-interferogram method to calculate a map of line-of-sight velocities. Inverting this velocity map using a Monte Carlo method, we estimate relative motion across the fault of 3 ± 2 mm/yr prior to the 1997 earthquake, one third the rate of other major faults in the area such as the Kunlun and the Altyn Tagh faults. The locking depth is poorly resolved, but is estimated to be 22 ± 15 km. The localised pattern of deformation observed suggest that the viscosity of the lower crust and upper mantle in the Manyi area is greater than 4×10^{18} Pa s proposed by previous postseismic studies of the area. We find no evidence of significant deformation across possible westward extensions of the Kunlun Fault. These rates of interseismic deformation are much smaller than the rates still being observed today, 10 years after the event, indicating the current rates must be due to one of the postseismic deformation mechanisms. **Citation:** Bell, M. A., J. R. Elliott, and B. E. Parsons (2011), Interseismic strain accumulation across the Manyi fault (Tibet) prior to the 1997 M_w 7.6 earthquake, *Geophys. Res. Lett.*, 38, L24302, doi:10.1029/2011GL049762.

1. Introduction

[2] The M_w 7.6, 1997 Manyi Earthquake in Northern Tibet (Figure 1) ruptured a 200 km long sinistral fault for which the coseismic slip distribution has been well established with InSAR and seismological methods [Funning *et al.*, 2007; Velasco *et al.*, 2000]. The initial postseismic motion deformed the surface at a rate of ~ 1 cm/yr, and could be explained for the first three years following the event by either afterslip or viscoelastic relaxation [Ryder *et al.*, 2007].

[3] To distinguish between these two possible mechanisms the postseismic time series must be extended; postseismic motion will decay over timescales (and lengthscales) which are mechanism dependent [Hearn *et al.*, 2009; Johnson *et al.*, 2009; Bruhat *et al.*, 2011]. The deformation rate from either mechanism will however, eventually decay to the point where the rate of deformation will become indistinguishable from the interseismic strain accumulation rate. Therefore, to

establish the mechanism behind the current deformation in the area of the Manyi fault [Bell *et al.*, 2010], we must first constrain the long-term interseismic strain accumulation rate across the fault. Estimates of the pre-earthquake slip rate also allow us to constrain models of postseismic frictional after-slip [Dieterich, 1979; Tse and Rice, 1986; Marone *et al.*, 1991; Perfettini and Avouac, 2007].

[4] Furthermore, obtaining accurate measures of interseismic strain accumulation can clarify the distribution of strain and faulting in this part of the Tibetan Plateau. The Kunlun fault clearly runs from 100°E to 90°E and possibly extends to 86°E at a latitude of 35.8°N [Taylor and Yin, 2009] (Figure 1). However, whilst the 2001 Kokoxili (Kunlun) earthquake ruptured to almost 90°E , the rupture on the 1997 Manyi earthquake is left stepped southwards by 50 km. This region is devoid of any GPS sites [Gan *et al.*, 2007], hence InSAR measurements are essential for determining fault slip rates in this region [e.g., Elliott *et al.*, 2008; Taylor and Peltzer, 2006].

2. InSAR Observations

[5] To prevent contamination of the deformation signal by the coseismic motion, we limit SAR acquisitions to those acquired before the 8th November 1997 earthquake. We use ERS-1 and ERS-2 data from descending track 305 (Figures 1b and 2a) which covers the majority ($>80\%$) of the 1997 earthquake rupture. There are insufficient radar scenes on ascending tracks across the Manyi fault for interseismic studies.

[6] The SAR data were processed using the ROI_pac software [Rosen *et al.*, 2004]. An SRTM digital elevation model and Delft (ODR) orbits [Scharroo and Visser, 1998] were used to make topographic and orbital corrections respectively. The interferograms were filtered [Goldstein and Werner, 1998] and then unwrapped using a branch-cut algorithm [Goldstein *et al.*, 1988]. The accuracy of the unwrapping was checked by testing phase closure loops of three interferograms and any errors that occurred during the unwrapping were either corrected manually or, if under 500 m in spatial extent, removed.

[7] The line-of-sight (LOS) velocity map in Figure 2b shows the annual LOS displacement change formed using a network of 20 interferograms made from radar images at ten different epochs using the methods of Biggs *et al.* [2007] and Wang *et al.* [2009]. As the deformation we are studying is of long wavelength (>10 km), we reduce the resolution of the interferograms to 800 m when calculating the LOS velocities to make the calculations more efficient.

[8] Because our estimates of satellite positions are imperfect, an orbital baseline error remains, even after the effects of baseline separation have been corrected for. This error

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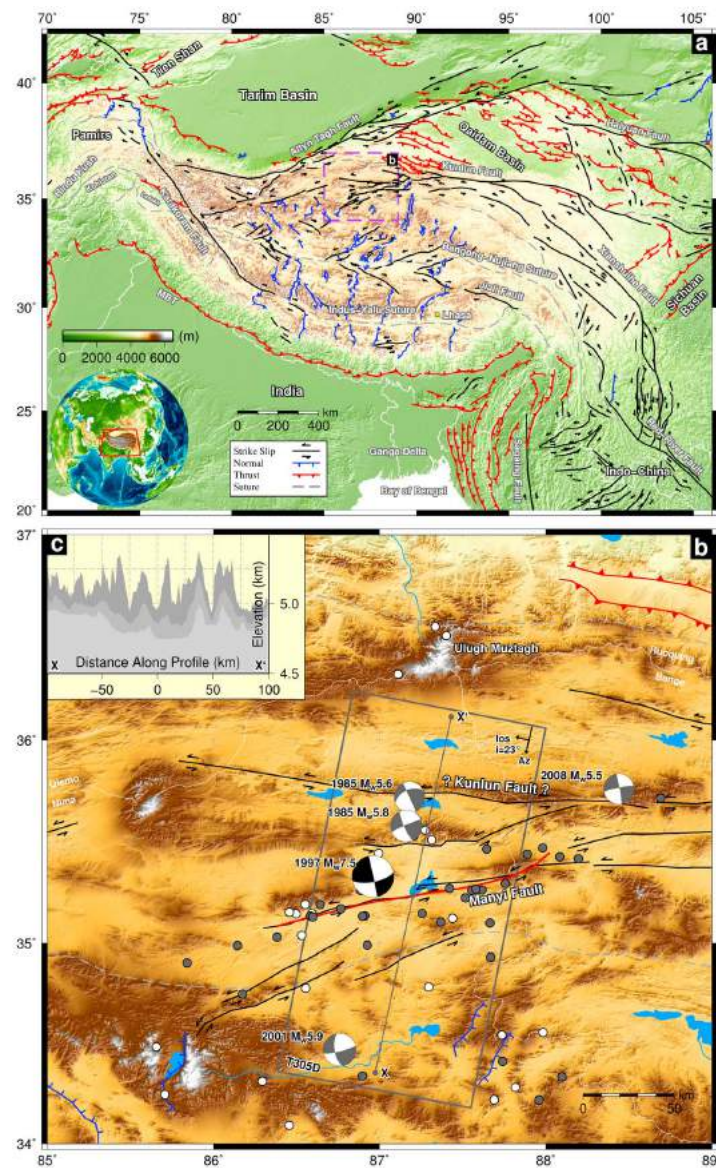


Figure 1. (a) Fault and topographic map of the Tibetan Plateau and immediate surrounding region, with the study area delineated by a purple dashed box. Faults are from [Taylor and Yin, 2009]. (b) Enlarged region of northern Tibet centered around the location of the Manyi earthquake (the red line indicates fault trace from Funning *et al.* [2007]) and ERS coverage used (track 305 descending). The black arrows show the azimuth and look direction of the satellite. Focal mechanisms are earthquakes from the gCMT catalogue (1976–2010) with $M_w \geq 5.5$. The solid dots show the epicentre of events prior to (in white) and after (in grey) the 1997 event. (c) Topographic profile along X-X' showing the limited relief of the plateau along the SAR track (50 km wide bins with min, mean and max profiles in greyscale).

manifests itself in the interferograms as a linear phase ramp. Interferogram I_{ij} , made from epochs i and j , has an orbital phase ramp: r_{ij} , that is dependent upon the position error at each epoch. The orbital errors can be thought of as contributing a phase ramp, $r_i = a_i x + b_i y + c_i$ at each epoch. The phase ramp in interferogram I_{ij} is the difference between the phase ramps at epochs i and j , i.e., $r_{ij} = r_i - r_j = (a_i - a_j)x + (b_i - b_j)y + (c_i - c_j)$. We determine r_{ij} empiri-

cally by fitting a linear plane to I_{ij} and use these planes to determine the coefficients a , b , c at each epoch. These are then used to calculate and remove an orbital phase ramp from each interferogram. As the inversion is rank deficient, there being one fewer independent interferograms than epochs, the inversion uses singular value decomposition to find a minimum-norm solution. We expect the correction to remove correctly orbital phase error parallel to the fault, but there to

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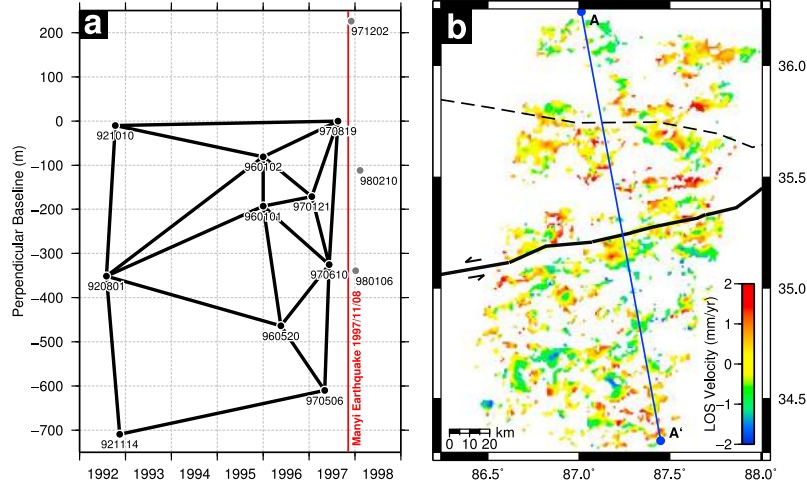


Figure 2. (a) Baseline (relative to 19th August 1997 acquisition) versus time plot for all track 305 acquisitions used in estimating the interseismic strain accumulation across the Manyi fault. The solid black lines denote the interferograms used. The red line shows the time of the November 1997 Manyi earthquake. (b) The LOS velocity map formed from the interferograms shown in Figure 2a. The thin solid blue line (A-A') denotes the location of the profile across the Manyi fault shown in Figure 3. The solid black line denotes the trace of the Manyi fault. The dashed black line denotes the possible extension of the Kunlun fault.

be a tradeoff between the fault perpendicular ramp and the signal from tectonic strain accumulation. This would affect the slip rate estimate, so we assume the fault perpendicular correction contains some of the deformation signal, and we make a further correction when solving for the final slip rate.

[9] We also determine a measure of the atmospheric noise for each interferogram. We assume the form of this spatially variable noise has a 1-D covariance function $\sigma_{ij}^2 e^{-r/d_{ij}}$ [Hanssen, 2001], where σ_{ij} is the interferogram error, r is the distance between pixels and d_{ij} is a characteristic e-folding distance. σ_{ij} depends on epoch errors σ_i and σ_j such that $\sigma_{ij}^2 = \sigma_i^2 + \sigma_j^2$. By fixing d_{ij} to the mean value across all interferograms of 8 km, we can estimate a variance at each epoch, σ_i^2 , using the same inversion technique used to find the orbital phase ramps.

[10] As the 20 interferograms cover only ten epochs, there are redundant interferograms in the network. For each pixel an optimal tree, which comprises of a subset of the connected independent interferograms, is calculated using a Kruskal algorithm that selects branches that minimise a cost function but still includes every epoch where a coherent pixel exists. The cost function is the sum across the subset of interferograms of f_{ij} , where f_{ij} is the fraction of incoherent pixels in interferogram I_{ij} . As there are few pixels which are coherent in all interferograms, we use pixels that are coherent in at least six of the interferograms of its optimal tree. We perform a pixel-by-pixel least-squares inversion to find the best fitting LOS displacement rate r_{los} , from the appropriate tree of interferograms (Figure 2),

$$r_{los} = \frac{\lambda}{2\pi} [T^T \Sigma_T^{-1} T]^{-1} T^T \Sigma_T^{-1} P \quad (1)$$

where T is a vector containing time spans of the interferograms in the tree, $[t_{12}, t_{23}, \dots, t_{ij}]^T$, P is a vector containing

the phase data, $[\phi_{12}, \phi_{23}, \dots, \phi_{ij}]^T$ for each interferogram in the tree, λ is the wavelength of the SAR instrument, and Σ_T is the covariance matrix which takes into account atmospheric noise at each epoch and correlation between interferograms. The elements of Σ_T , c_{ij-kl} , describe the covariance between interferograms I_{ij} and I_{kl} , and are given by

$$c_{ij-kl} = \sigma_i \sigma_k \delta_{ik} + \sigma_j \sigma_l \delta_{jl} - \sigma_k \sigma_j \delta_{kj} - \sigma_i \sigma_l \delta_{il} \quad (2)$$

where δ_{ij} is the Kroeneker Delta. The rate error for each pixel in the inversion is given by $\sigma_p = [T^T \Sigma_T^{-1} T]$.

3. Modelling Interseismic Deformation

[11] To model the interseismic strain accumulation, we assume uniform slip occurs on a buried infinitely long and deep strike-slip fault beneath a locked elastic lid of depth d [Savage and Burford, 1973]. The fault geometry is taken from the single fault coseismic model of Funning *et al.* [2007].

[12] As we have removed any fault parallel orbital phase errors, we collapse the LOS velocity map onto a profile perpendicular to the fault trace (Figure 2b). We subdivide the LOS velocity map into 8 km width bins parallel to the fault and calculate the weighted mean and standard error of the rates in each bin Figure 3a. Although the LOS velocity map is formed on a pixel-by-pixel basis, we assume that the final values in the LOS velocity map will be spatially correlated with a covariance matrix of the same form used to estimate each epoch's atmospheric error, with σ_i replaced by σ_p . We do this to ensure that pixels with larger errors are appropriately weighted. We use the profile of the LOS velocity map and models of interseismic strain to perform a least-squares inversion of the slip rate assuming the LOS velocity is a result of three different components, $v_k = sg_k + ax_k + c$, where v_k is the LOS velocity calculated in the LOS velocity

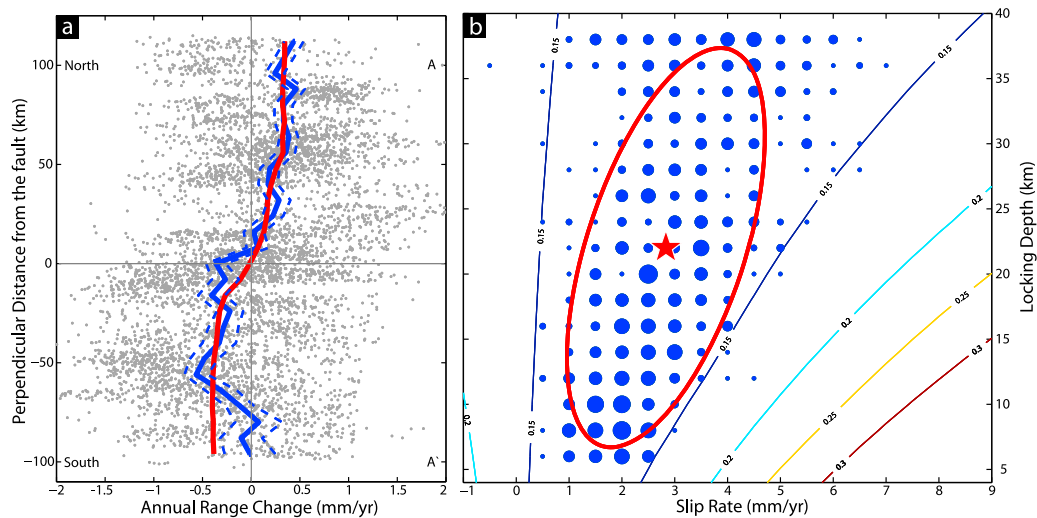


Figure 3. (a) Profile across the Manyi fault of the LOS velocity map. The blue line is the weighted mean with dashed lines indicating the extent of the 2σ error. The grey dots show the rate at each pixel once they have been corrected using the orbital ramp estimated in the inversion. The red line shows the best fitting model (star in Figure 3b). Due to the geometry of the fault and the satellite look direction, 1 mm/yr of slip across the fault results in 0.35 mm/yr far-field LOS velocity. (b) Slip rate-locking depth parameter space used to determine the best fitting solution to each perturbed data set for the Manyi fault. The blue dots show the grid points in parameter space that were used, with their area proportional to the number of solutions that fell on that point in the space, with the largest dot representing 17 solutions and the smallest 1. The contours show the RMS misfit (in mm) of the solutions at each point in the parameter space. The mean solution is shown by the red star. The red line indicates the location of the 67% confidence ellipse and shows the uncertainties in the average slip rate/locking depth.

map profile at the point k located at a distance x_k along the profile, g_k is the Green's Function [Savage and Burford, 1973] that describes the surface displacement at point k in the direction of the satellite's line-of-sight, a is the magnitude of a fault-perpendicular orbital phase ramp and c is a static offset. To avoid numerical instabilities during the inversion, we remove any pixel where $\sigma_p > 0.8$ mm/yr. We make no attempt to remove topographically correlated atmospheric errors [Elliott et al., 2008] as there is limited topographic variation across the profile (Figure 1b).

3.1. Estimating the Magnitude of Errors

[13] The least-squares inversion for slip rate provides a formal error estimate for the slip rate and the orbital ramp parameters. To find more realistic values and uncertainties we use Monte Carlo simulations of atmospheric delay errors [Wright et al., 2003]. We use the statistics from the 1-D covariance function of the LOS velocity map calculated previously to create 1000 sets of simulated noise which is added to our LOS velocity map. The best fitting slip rate and locking depth for each perturbed data set is determined by a parameter search across slip rate/locking depth space. Our best fitting solution, with a slip rate of 2.8 mm/yr and a locking depth of 22 km (Figure 3a), was the average of the results from each of the 717 inversions that did not hit the boundaries of the parameter space used. A slip rate of 2.8 mm/yr corresponds to a far-field LOS velocity of $\sim \pm 0.5$ mm/yr due to the geometry of the look direction of the satellite (shown in Figure 1a) and the strike of the fault. Given that 283 perturbed solutions hit the boundaries of the locking depth parameter

space (but never the slip rate) it is clear the locking depth is not well constrained.

[14] Figure 3b shows the best fitting models for each of the perturbed datasets. We use principle component analysis to calculate the orientation and size of an ellipse around the mean which contains 67% of the 717 perturbed solutions. This confidence ellipse gives estimates of the error in slip rate and locking depth of 1.8 mm/yr and 15 km respectively. We therefore estimate that the Manyi fault, prior to the 1997 event, was accumulating interseismic strain at a rate of 3 ± 2 mm/yr with a locking depth of 22 ± 15 km.

3.2. Possible Extension of the Kunlun Fault

[15] We have assumed that the only deformation observed in the LOS velocity map is due to motion across the Manyi fault, however the area covered by the map of LOS velocities also contains another mapped fault [Taylor and Yin, 2009] that could possibly be considered as the western extension of the Kunlun fault. Using our estimate of the slip rate across the Manyi fault we attempt to model any residual signal that could be due to slip on this fault. We remove the expected deformation from the Manyi fault and perform the same inversions as for the Manyi fault, only now with a profile taken perpendicular to the second fault (see auxiliary material).¹ This yields a slip rate of -0.1 ± 0.2 mm/yr across the 'Kunlun' fault. Inverting for the locking depth results in solutions that always hit the bounds so we fix the locking

¹Auxiliary materials are available in the HTML. doi:10.1029/2011GL049762.

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depth to 22 km. We obtain the same results regardless of whether we solve first for the Manyi fault or the ‘Kunlun’ fault.

4. Discussion

[16] We estimate the rate of relative motion across the Manyi fault prior to the 1997 earthquake to be 3 mm/yr and near zero across the possible extension of the Kunlun fault at this longitude. The rate for the Manyi fault is about one third the rate of the plateau bounding Altyn-Tagh fault to the north [Elliott *et al.*, 2008] and main portion of the Kunlun fault to the east [Zhang *et al.*, 2004]. Assuming the 1997 Manyi earthquake is representative of earthquakes on the Manyi fault, the average coseismic slip is 4 ± 1 m [Funning *et al.*, 2007], and relative motion across the fault of 3 ± 2 mm/yr is constant over the whole interseismic cycle, our results indicate a recurrence time of 1300^{+3700}_{-700} yrs.

[17] Many Monte Carlo solutions for both the Manyi and ‘Kunlun’ fault hit the boundary limits of 0 and 40 km locking depth during the inversions. Our estimates of locking depth are therefore not well constrained. However, the preferred value of 22 km is in close agreement with the maximum depth of slip in the 1997 Manyi earthquake (20 km), as is to be expected if earthquakes occur in the seismically locked part of the crust.

[18] Maps of the current LOS velocities around the Manyi Fault area on track 305 show LOS velocities of up to 8 mm/yr [Bell *et al.*, 2010]. The LOS displacement rate expected if we assume the fault is still accumulating strain at the rate found in this study is only ~ 1 mm/yr, suggesting that, even 10 years after the 1997 event, the current deformation is likely to be predominantly postseismic.

[19] Reliable rates for the Altyn Tagh, Kunlun, Xianshuihe and Jiali faults are typically 10 mm/yr [Searle *et al.*, 2011]. A marked decrease in the slip rate is observed towards the ends of some of these more comprehensively measured strike slip faults, e.g the eastern Altyn Tagh fault [Zhang *et al.*, 2007] and the eastern Kunlun [Kirby *et al.*, 2007]. Our observations support either a decay in slip rate also towards the western end of the Kunlun fault, or slip that has been distributed across a number of surrounding faults. In either case, our observations do not support a constant slip boundary model required by quasi-rigid block or microplate rotation [Thatcher, 2007; Meade, 2007] where slip is concentrated on few faults.

[20] Throughout this study we have assumed that the medium accommodating the interseismic deformation is elastic. However Ryder *et al.* [2007] have modelled the rapid postseismic deformation in terms of viscoelastic relaxation. Simple interseismic strain models with an elastic layer over a viscoelastic half-space [Johnson and Fukuda, 2010], where relaxation times are much smaller than the earthquake repeat time, show a quasi linear variation of displacement across the fault, where we see deformation that is localised. This would suggest that the rapid rates of motion immediately after the Manyi earthquake are better explained by afterslip at depth [Ryder *et al.*, 2007]. More complex viscoelastic models with larger viscosities in the lower crust and upper mantle [Johnson *et al.*, 2007], retain localised deformation throughout the earthquake cycle. They also show that using elastic models to describe viscoelastic behaviour could underestimate the long term slip rate across the fault by up to a

factor of two. In this case the slip rate across the Manyi fault could be as high as 6 mm/yr and have a repeat time half that estimated here.

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