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 $3 \pm 2$ 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 $22 \pm 15$ 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 $4 \times 10^{18}$ 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.


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 $\sim 1$ cm/yr, and could be explained for the first three years following the event by either aftserslip 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 Pelteer, 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
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, $r_i = ax + by + c$, at each epoch. The phase ramp in interferogram $I_{ij}$ is the difference between the phase ramps at epochs $i$ and $j$, i.e., $r_{ij} = r_i - r_j = (a_i - a)x + (b_i - b)y + (c_i - c)$. 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 error parallel to the fault, but there to
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 $\sigma_{ij}^2 e^{-r_{ij}}$ [Hanssen, 2001], where $\sigma_{ij}$ is the interferogram error, $r$ is the distance between pixels and $d_{ij}$ is a characteristic e-folding distance. $\sigma_{ij}$ depends on epoch errors $\sigma_i$ and $\sigma_j$ 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, $\sigma_i^2$, 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. This cost function is the sum across the subset of interferograms $f_{ij}$, 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).

$$r_{los} = \frac{\lambda}{2\pi} \left[ T^T \Sigma_T^{-1} T \right]^{-1} r^T \Sigma_T^{-1} P$$

where $T$ is a vector containing time spans of the interferograms in the tree, $[t_{12}, t_{23}, \ldots t_{ij}]^T$, $P$ is a vector containing the phase data, $[\phi_{12}, \phi_{23}, \ldots \phi_{ij}]^T$ for each interferogram in the tree, $\lambda$ is the wavelength of the SAR instrument, and $\Sigma_T$ is the covariance matrix which takes into account atmospheric noise at each epoch and correlation between interferograms. The elements of $\Sigma_T$, $c_{ij-kl}$, describe the covariance between interferograms $I_{ij}$ and $I_{kl}$, and are given by

$$c_{ij-kl} = \sigma_i \sigma_k \delta_{il} + \sigma_j \sigma_l \delta_{kj} - \sigma_k \sigma_j \delta_{ij} - \sigma_l \sigma_i \delta_{kl}$$

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

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 $\sigma_i$ replaced by $\sigma_p$. 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, $v_k = v_1 + v_2 + v_3$, where $v_k$ is the LOS velocity calculated in 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, \( a \) 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 \( s_p > 0.8 \) 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

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 \( \pm 0.5 \) 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.

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 \( 3 \pm 2 \) mm/yr with a locking depth of \( 22 \pm 15 \) km.

3.2. 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 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). This yields a slip rate of \( -0.1 \pm 0.2 \) mm/yr across the ‘Kunlun’ fault. Inverting for the locking depth results in solutions that always hit the bounds so we fix the locking...
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 of 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 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, Xianshuine 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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