Stress magnitude and its temporal variation at Mt. Asama Volcano, Japan, from seismic anisotropy and GPS

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ABSTRACT

The Earth’s stress regime is fundamental to its physical processes, yet few methods can determine absolute stress, and measurements of temporal variations in stress are controversial. The Global Positioning System (GPS) is used to measure stress-related deformation from magma movement, and stress causes aligned cracks whose orientations may be measured by seismic anisotropy. Here we show that changes in earthquake activity and ground surface deformation preceding and accompanying the eruption of Mt. Asama, Japan in 2004 correlate well with changes in the orientation and strength of seismic anisotropy. This correlation confirms the validity of anisotropy as a stress monitoring tool. It can be used to determine the crack aspect ratio \( (2.6 \times 10^{-5}) \) and provides a new method of constraining the difference between the magnitudes of the two horizontal stresses, \( \Delta S_H \), which is only about 10% of the vertical stress at Asama.

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1. Introduction

Stress drives all tectonic processes, but at present the only way to measure the absolute stress amplitude in the Earth involves drilling boreholes, which are expensive, intrusive and technologically difficult (e.g., Townend, 2006). Other techniques can provide some of the stress elements; geodetic methods such as GPS and InSAR determine strain changes, which are interpreted as changes in stress (Dzurisin, 2003). Earthquake records can be used to determine the change in stress after earthquakes (stress drops). Absolute stress tensor orientations and a single ratio involving stress amplitudes can be measured at the earthquake source by inversions of fault orientation data (focal mechanisms) (e.g., Arnold et al., 2005). Stress orientations can also be measured along the path between the earthquake and recorder by measuring seismic anisotropy, which is caused in the upper crust by aligned cracks that open and close in response to changing stress conditions (e.g., Nur and Simmons, 1969; Crampin, 1994; Boness and Zoback, 2006).

Monitoring of stress changes to predict earthquakes and volcanic eruptions has long been a goal of seismologists. However, claims of predictive capability are often disputed, and quantitative relations between observed stress indicators are lacking. Here we show that average seismic anisotropy delay time variations correlate well with strain changes measured from GPS, and we combine the two techniques to determine crack aspect ratios and to show that the difference between the magnitude of the two horizontal stresses, \( \Delta S_H \), is small (on the order of 10% of the vertical stress).

1.1. Anisotropy

Anisotropy causes two perpendicular components of shear waves to travel with different speeds, an effect called shear wave splitting or birefringence; the first arriving wave has a polarisation (\( \phi \)) parallel to the fast orientation of the anisotropic material and the delay time (\( d_t \)) between the two waves depends upon the integrated effect of anisotropy along the travel path. Aligned cracks cause hexagonal anisotropy with a slow axis of symmetry that is perpendicular to the crack plane orientations, so the fast orientation is parallel to the cracks (e.g., Nur and Simmons, 1969). Thus, \( \phi \) yields the average crack orientation, inferred to be parallel to the maximum horizontal stress \( (S_{Hmax}) \) and \( d_t \) is proportional to the path length and to crack density (e.g., Hudson, 1981). Studies claiming variations in shear wave splitting delay times as earthquake precursors are controversial because of concerns that spatial changes in earthquake sources were being misinterpreted as temporal changes in path properties (e.g., Peacock et al., 1988; Aster et al., 1990; Liu et al., 2004). However, changes in delay times in the damage zones after large earthquakes are clearly significant (e.g., Saiga et al., 2003).
The larger stress changes associated with deformation before volcanic eruptions provide an easier target for precursory studies, and shear wave splitting has recently been proposed as a method to monitor stress on volcanoes (Miller and Savage, 2001; Gerst and Savage, 2004; Bianco et al., 2006; Bianco and Zaccarelli, 2009). In New Zealand and Italy, changes in $\phi$ on the order of 90° were observed near volcanoes, associated with eruptive activity. The most likely causes were dyke intrusions that reversed local stress orientations (Gerst and Savage, 2004).

2. Data and setting

Mt. Asama is an active andesitic volcano located in central Honshu, Japan. An eruption that began 1 Sept. 2004 was preceded by a dyke intrusion, whose characteristics have been inferred from GPS measurements and from locations of volcanic earthquakes (Takeo et al., 2006).

We measured shear wave splitting on seismograms from all the available three-component seismometers monitored by the Asama Volcano Observatory during the period Sept. 2002 through May 2008 (Fig. 1; Table S1). We examined two types of events: Regional earthquakes determined by the Japan Meteorological Agency to be outside the Asama Volcano Observatory network but within 300 km of the volcano, and local earthquakes determined by the observatory (Urabe and Tsukada, 1992) to be within the network. Ray paths from local earthquakes travel solely through the volcano, so that anisotropy in the mantle or lower crustal mineral alignment will not affect the measurements. However, earthquake activity related to dyke intrusion will likely migrate, causing the ray paths to change. Spatial

Fig. 1. Tectonic setting and rose diagrams. (a) Tectonic setting of study area. The red box is the region outlined in part b. (b) Rose diagrams (circular histograms) of the fast polarisations of the shear waves for the three best filters at each station for the “steep” group of events (see Section 3.2), on a base map of the digital topography. Small roses are for stations with less than 10 measurements. Table S1 gives the numbers of measurements and averages for the stations within this figure. The blue line is the surface projection of the dyke (Takagi et al., 2005). The orange line is the orientation of $S_{max}$ measured from earthquake focal mechanism inversions (Townend and Zoback, 2006). The white line is the $-27\pm29°$ average orientation of $\phi$ from SKS phases measured (Long and van der Hilst, 2005) on stations east of longitude 136° and south of latitude 40°. 0221 and 0268 are the GPS stations whose spacing is plotted in Fig. 5. The blue box outlines the region enlarged in parts c and d. (c) Rose diagrams for events occurring before the eruption. (d) Rose diagrams for events occurring after the eruption. Yellow roses are for stations that were operating before the eruption.
changes are thus difficult to disentangle from temporal changes. Regional earthquakes occur mostly in the subducted Philippine and Pacific plates and their paths are affected by mantle and lower crustal mineral anisotropy as well as by crustal stress. They are also affected by laterally varying properties, but earthquakes far removed from the volcano should not have systematic variations in location that are correlated with magma movement. Therefore, changes in measurements from regional events that correlate with magma movement can be interpreted as temporal rather than spatial variations.

For the local events, we examined all a-type (tectonic) earthquakes located via the standard processing procedure (Urase and Tsukada, 1992). This includes stations run by Asama Volcano Observatory, the Japan Meteorological Association and Hi-net networks.

For the regional earthquakes, we searched the Japan Meteorological Agency (JMA) catalogue for all events over magnitude 2.0 within the interval bounded by longitudes 140.5° and 136.5°, and latitudes 38.5° to 34.5°, and examined all events that had triggered the recording system within 30 s of the earthquake origin times.

3. Methods

3.1. Shear wave splitting calculations

We use the Teanby et al. (2004) shear wave splitting code with modifications to allow automatic quality classification. The Teanby code uses at its base the Silver and Chan (1991) (SC) shear wave splitting analysis technique. The SC analysis is carried out on multiple measurement windows and cluster analysis determines the best window. The cluster that has the minimum variance is chosen as the best cluster, and a final SC measurement is made based on the best window within the cluster. The only manual step is to pick the S wave arrival. The results are free from operator bias, and large numbers of waveforms can be processed so that patterns can emerge from noisy data.

Local events are routinely checked by hand before cataloguing and we used these catalogued S arrival times. For the regional events, we used the TauP code (Crotwell et al., 1999) to determine the P and S arrival times expected for the event location and the IASPEI 91 arrival time model (Kennett and Engdahl, 1991). The waveforms were examined by eye and S wave arrival times were determined by comparing the expected arrival times with the waveforms, paying careful attention to changes in frequency content over time in the waveforms, and the relative arrival times of S and P waves because both P and S waves usually arrived later than predicted.

We use computer scripts developed by Wessel (2008) to automatically process data using the Teanby et al. (2004) cluster analysis splitting codes. The main advance in the scripts is in using multiple filters (Table S2) to find the frequency bands with the best signal-to-noise ratios and choosing the length of the measurement windows based on the period of the waveform. The measurement window has a minimum length of the dominant period, and a maximum length of 2.5 times the dominant period.

We make a small modification to the Teanby code to grade the measurements to minimize effects of cycle skipping, in which the splitting program may mismatch waveforms by an integer number of half cycles. If it is mismatched by one half cycle, then the fast and slow waves may be interchanged, and the determined delay time differs by one half period (Matcham et al., 2000). If it is mismatched by two half cycles, then the delay time remains the same but the determined delay time differs by a whole period. All clusters with numbers above a minimum threshold (set as 5 in the analysis here) are compared to the chosen “best cluster.” The measurements presented here fit the following criteria: If the average fast polarisation of any cluster of more than 5 windows is more than 8/18 radians from that of the best cluster, or if the average delay time differs from the best cluster delay time by more than 1/8 of the maximum allowed splitting time, then the measurement is rejected as possibly affected by cycle skipping (Figs. S1 and S2). We ran the analysis for the three best filters for each event–station pair. Waveforms that give good results for several filters will therefore be weighted more heavily than waveforms that give good results for only a single filter (Gerst and Savage, 2004).

Further grading of measurements passing the cycle skipping test is based on the normal errors of the final best splitting measurement (errors in δ must be less than 25°), the signal-to-noise ratio (it must be greater than 3), and two additional criteria to determine if the measurement could be “null,” which occurs when there is no splitting. Such null measurements occur when either there is no anisotropy, or if the initial polarisation is parallel or perpendicular to the fast orientation. We use the criterion suggested previously (Savage et al., 1996; Peng and Ben-Zion, 2004): If δ is between 0° and 20° or between 70° and 90° of the measured incoming polarisation direction, it is considered a null measurement. Furthermore, measurements with δt larger than a maximum value (1.0 s for regional events and 0.4 s for local events) are likely to be nulls and are removed. The local events and regional events used the same codes and filters, except for the difference in maximum allowed δt.

3.2. Event groupings

For angles of incidence at the surface of greater than 35°, converted phases can interfere with shear wave splitting measurements (Nutti, 1961; Booth and Crampin, 1985). However, low velocities at the surface result in near-vertical arrivals for many events. To account for this, many people consider only events within this “shear wave window” of 45° based on straight-line ray paths (i.e., assuming homogeneous velocity; e.g., Peacock et al., 1988). However, using only such straight-line ray path restrictions can unnecessarily delete too many events from analysis. We divided our results into several sets, depending on the angle of incidence and earthquake depth.

For regional events, one group (“steep” for steep incidence angles) was made up of events with straight-line angles of incidence less than 45°, following previous conventions. However, these events actually had angles of incidence between 0° and 17° from the vertical, calculated using the 2.0 km/s shear wave velocity in the top layer of the standard velocity model used to calculate hypocentres. All these earthquakes were deeper than 40 km. Events with larger incidence angles were broken into two groups based on whether their depths were shallower (“shall”) or deeper (“deep”) than 40 km.

All the “steep” events were within the so-called “Band-2” incidence of within 15° from the crack plane, in which the delay times are sensitive to crack density (Peacock et al., 1988; Crampin, 1999) (Fig. 2). Paths within the shear wave window but outside “Band-2” are in “Band-1” (Crampin, 1999). Although previous papers do not usually remark upon these properties, in “Band-2,” delay times and polarisations are relatively insensitive to angle of incidence and in “Band-1” delay times and polarisations are a strong function of back azimuth and angle of incidence. The negative delay times in “Band-1” (Fig. 2) arise because the delay times are calculated as a difference between two absolute directions, one of which is fast for vertical incidence angles. As the angle between the propagation direction and the crack plane continues to increase, the delay times decrease until they reach zero (the two waves arrive at the same time) and then for further increases in angles, the fast and slow components switch polarity. The former fast direction becomes slow and vice-versa, even for paths within the shear wave window.

Events in the “shall” and “deep” groups were measured at the stations closest to the summit that operated both before and after the eruption, for times up through the end of 2006. These paths had angles of incidence of 14° to 21° from the vertical (Calculated with a surface velocity of 2.0 km/s), which makes them within “Band-2” if their back azimuth is within 15° of the crack plane strike. Otherwise,
To calculate average parameters, we use Gaussian statistics for the delay times, and the Von Mises criterion (Mardia and Jupp, 2000), a circular analogue to the normal distribution, for the fast polarisations. Error bars presented are the 68% confidence intervals, even for waves that arrive within the shear wave window. These are shown by negative dt and changes in ϕ, even for waves that arrive within the shear wave window. Examples of waveform quality can be viewed in the measurement examples (Figs. S1 and S2).

### 3.3. Statistical analysis

To calculate average parameters, we use Gaussian statistics for the delay times, and the Von Mises criterion (Mardia and Jupp, 2000), a circular analogue to the normal distribution, for the fast polarisations. Table S1 includes information on the statistics. Error bars presented are the 68% confidence intervals. However, many of the distributions are bimodal and thus are not well described by normal distributions, so we recommend caution in interpreting the averages and standard errors (Fig. 1).

### 3.4. Stress modelling

We use program Coulomb 3.1 (Lin and Stein, 2004; Toda et al., 2005) to calculate the maximum horizontal stress (Lund and Townend, 2007) at different depths and positions based on a model including a dyke (Takeo et al., 2006) and two point (Mogi) sources (Takagi et al., 2005). The dyke extends between 3 and 5.1 km depth and has 0.8 m of opening. The small and large Mogi sources are at 0.25 km and 3.1 km depth, and correspond to 6400 m$^3$ and 200,000 m$^3$ of volume expansion, respectively. Young’s modulus is 8 × 10$^5$. The regional stress is assumed to have its intermediate stress σ2 as vertical (strike-slip stress regime) and is given by the weight of the rock above it, yielding a gradient of 21 MPa/km. Other stress regimes were examined, but the maximum horizontal stress was independent of the type of stress regime and depended only on the difference between the two horizontal stresses. The orientation of σ1 is −64°, parallel to the dyke, and the magnitudes of σ1 and σ3 vary in each case. For the near-isotropic horizontal stress case, we used σ1 = 21.05 MPa/km and σ3 = 20.95 MPa/km. For the case with horizontal stress as 10% of vertical, we use σ1 = 22 MPa/km and σ3 = 20 MPa/km. For Byerlee friction, σ1 = 26.0 MPa/km and σ3 = 16.0 MPa/km. The calculation of the misfit, θ, uses circular statistics (Mardia and Jupp, 2000), and is given by $θ = \cos^{-1}\left(f^{-1} \sum_{i=1}^{N} f(\cos(α_i-α_m))/N\right)$. This equation where $α_i$, and $α_m$ are the calculated ($S_{H,max}$) and measured (θ) angles, respectively, for each event $i$, and $f(x)$ is either the absolute value (L1 norm) or the square (L2 norm). Misfits for all models considered are in Table S3.

### 4. Results

In all, there were 32 permanent seismic stations recording data during the study period, and 30 gave reliable measurements (Fig. 1, Table S1). The other two were too noisy or had frequent power failures and so did not record much data. Examples of waveform quality can be viewed in the measurement examples (Figs. S1 and S2).

#### 4.1. Local events

Two hundred and fifty-five high quality measurements were returned from 97 events at 17 stations (Table S1). The frequency band with the largest numbers of high quality measurements (106) had high pass of 3 Hz, but frequencies up to 10 Hz were measured (Table S2). The depths of the events yielding high qualities ranged from 300 m above sea level to 2.2 km below sea level (Takeo et al., 2006). The path lengths range from 0.6 to 5.2 km, with a median of 3.5 km. The angles of incidence of these local events ranged from 2° to 75° using the velocity of 2.0 km/s that is used to locate the earthquakes. However, actual velocities near the surface are likely to be still smaller, because the near surface has low velocities (P wave velocities of less than about 2.5 km/s at most locations (Aoki et al., 2009), which translates to an S velocity of 1.4 km/s if a typical $V_p/V_s$ ratio of near-surface volcanic material of 1.7 is used (Rowlands et al., 2005)). We include only the events within the restricted angle of incidence of 45° in Fig. 3 and in the comparison of φ with Coulomb stress (Table S3). We show the whole set in the rose diagrams of Fig. 4. Summaries from both sets are available in Table S1.

#### 4.2. Regional events and time variation

There were 3644 events in the JMA catalogue fitting the “steep” criterion (Section 3.2), of which 525 events triggered the seismic network. Using the three best filters, 1305 high quality measurements were obtained from 276 events at 27 stations. The depths ranged from 61 to 366 km, but all except eight events had depths greater than 100 km. Station summary statistics are available in Table S1.

There were 273 measurements in the “shall” group. In the “deep” group (Section 3.2) there were 1183 measurements from events between 40 and 156 km (Table S1). The majority of the filters that provided the best signal-to-noise ratio for the regional events, including “steep,” “deep” and “shall” were low frequency, with 3 Hz as the maximum frequency filter for over 80% of the measurements.

At the longest running station, AVO, measurements of δt from deep regional earthquakes yield systematic changes with time, which correlate with GPS baseline length change between two stations situated on opposite sides of the inferred dyke (Takeo et al., 2006)
Moving averages reveal a strong correlation between $dt$ changes and GPS baseline length changes. However, statistical comparisons of moving averages are difficult due to the correlated errors. The relation between GPS and $dt$, which were made at different times, was quantified by three different tests. For one test, we fit a 10-degree polynomial to the GPS baseline length measurements. Then we compared the actual $dt$ measurements (not averaged) with the expected value of baseline length change from the polynomial fit. This yielded a correlation coefficient of 0.2 and a $p$-value of 0.0011 (the $p$-value is the probability that two variables with Gaussian distributions would yield such a correlation by chance). In another test, we determined 25 time windows of approximately 10 points for each $dt$ measurement, varying the numbers of points slightly so that measurements from the same event would remain in the same bin. The time windows varied in length between 21 and 118 days with a mean of 58 days. Then we calculated the average $dt$ and GPS baseline length in the same time windows and compared them to each other. We obtained a correlation coefficient of 0.6 and a $p$-value of 0.003. Finally, we calculated the averages of both GPS and delay time based on non-overlapping time windows of 60 days. Each time period has a variable number of points. The relation between the two sets of averages calculated in this way gives a correlation coefficient of 0.45 and a $p$-value of 0.02. Therefore the two time series are unlikely to be related by chance.

The distribution of $\phi$ for stations near the volcano changed dramatically after the eruption, changing by nearly 90° at most of the common stations (Fig. 1). Earthquakes with similar paths to the stations yield differing fast orientations before and after the eruption (Fig. 6) so the variations are not likely to be caused by varying paths. Stations close to the volcano that returned good measurements on regional earthquakes both before and after the eruption have average $\phi$ that change with time, but they vary in detail for different categories of events (Figs. 1 and 4; Table S1).

Average $\phi$ at station AVO rotates from about 30° in mid 2002 to about 70° just before the eruption on 1 Sept 2004 (Fig. 7). The rotation appears to correlate with the occurrence of b-type (volcanic) earthquakes, and precedes the eruption. After the main eruption, $\phi$ returns to its original value over a six month period. Average $\phi$ changes again in 2006, when the number of N-type (another type of volcanic earthquake, called “tornillo” in some studies) events increases. Station KUR was operating for less time than AVO, but more time than most other stations. It shows a similar, but smaller, variation in $dt$ for events in the “shall” group (Fig. 8).

5. Discussion

5.1. Quality of shear wave arrivals

Stations on the highest topography tend to have the smallest number of measurements and yield the most scattered results (Figs. 1 and 4). The S waves on these stations were difficult to pick, probably due to scattering, attenuation and topographic effects. Many S waves had poor signal-to-noise ratio, and it was difficult to decide where the S wave arrived. But most stations had some events with good signal-
Circular histograms of $\phi$ from different subsets of the regional events, as in Fig. 1: Panels on the left side (a and b) are from prior to the eruption on 1 September 2004 and the panels on the right (c, d, e) are from after that time. The patterns vary from station to station and from subcategory to subcategory, which are defined in Section 3.2, but most stations have distributions that change with time. Too few local events occurred before the eruption to make meaningful conclusions (Table S1).
to-noise ratio. We attribute this to attenuation and scattering affecting some paths more than others. It suggests that an attenuation analysis would help to constrain the regional structure. A general observation is that the P and S waves usually arrived later than predicted, and S was proportionately slower than the P waves. This is probably due to the slower velocities in the mantle wedge and crust in the volcanic region (Zhao et al., 2009).

5.2. Comparison of shear wave splitting with calculated stress orientations

There were few local a-type earthquakes before the eruption (Fig. 7). To compare anisotropy with the stress field after the eruption, we use the local earthquakes as a whole to avoid contamination from mantle paths. The population as a whole does not fit a Gaussian distribution, with a formal average of $-65 \pm 104^\circ$ for paths with steep incidence angles (>45°) (Table S1; Fig. 3). Yet they roughly match the stress field expected (Toda et al., 2005) for the magma sources determined from GPS (Takagi et al., 2005; Takeo et al., 2006) in a regional stress field with the most favourable orientation for the dyke emplacement and with 10% ΔS_H (Fig. 1; Table S3; Fig. 3). Measurements on paths that travelled close to the dyke tend to be perpendicular to the dyke, while measurements for paths that travelled further away are more dyke-parallel.

The splitting must be acquired between the earthquake source and the station. This means the anisotropic region is above the earthquake depths of 3–5 km below the surface. We calculate the stress field at different depths and compare the maximum horizontal stress, $S_{Hmax}$, to the $\phi$ measurements (Fig. 3) using three different assumptions: (1) Splitting is mainly caused by the portion of the path closest to the station. For these, the stress field at the station location is used for the comparison. (2) Splitting is mainly caused by the portion of the path near the earthquake. For these, the earthquake location is used. (3) The entire path contributes. We use the position halfway between the earthquake and station. Above and below the dyke extent of 3–5 km, the regional stress field dominates. Thus earthquakes occurring above the dyke should be least affected by the stress field of the dyke. We present results for a depth of 3.5 km here, because it is the shallowest region with a strong effect of the dyke (Fig. 3). Table S3 gives the misfits for all the combinations tested. Despite the obvious oversimplifications of our model, we obtain better fits when we include both the regional stress and the stress field from the dyke. We achieve the best result when we compare the stress field at 3.5 km with the measurements at the earthquake locations, which suggests that we are measuring deep stresses. We suggest that an
inversion for three-dimensional anisotropic velocity structure to compare to the stress measurements would be fruitful.

For a dyke to reverse the nearby orientation of $S_{\text{Hmax}}$, the difference between the two regional horizontal stresses ($\Delta S_{\text{H}}$) must be small (Gerst and Savage, 2004). Comparing the orientations of the $\phi$ measured from local earthquakes with models of the stress (Toda et al., 2005) from the dyke intrusion, our preferred model has $\Delta S_{\text{H}}$ equal to 10% of the vertical stress, which yields $\Delta S_{\text{H}}$ of only 6 MPa at 3 km below the surface (Fig. 3). This is lower than the average differential stress at 3 km of about 50 MPa from a compilation of borehole stress data (Townend and Zoback, 2000), which gives $S_{\text{Hmax}}$ parallel to the dyke throughout our model. However, volcanic areas such as Asama with cones that have many eruptions (polygenetic volcanoes) may have unusually low $\Delta S_{\text{H}}$ (Takada, 1994). The stress field as measured from earthquake focal mechanism inversions changes orientation near Mt. Asama (Townend and Zoback, 2006), which suggests a locally near-isotropic stress, and earthquakes greater than $M=4.0$ in the region are rare. Future studies using stress inversion from focal mechanisms on these small earthquakes are recommended to test this inferred low stress.

5.3. Calculating percent anisotropy from local earthquakes

Delay times of local earthquakes average $0.11 \pm 0.02$ s (Table S1). Delay time $dt$ is related to anisotropy by the following formula (e.g., Savage, 1999): $dt = L(v_1 - v_2)/(v_2 v_1)$, where $v_1$ and $v_2$ are the fast and slow velocities and $L$ is the path length over which the splitting occurs. For an order of magnitude calculation, we use the median path length of 3.5 km for $L$, $v_2 = 2$ km/s and an average delay time of 0.1 s for the local events. The fractional velocity ratio $(v_1 - v_2)/v_1$ is 0.057, or 6% anisotropy. Crack density $\rho$ is defined (Hudson, 1981) as the number of cracks $M$ times the cube of the crack radius $a$ divided by the volume $\rho=M(a^3/V)$. It can be calculated from the delay time and the average shear wave velocity $V_s$ using Hudson’s (1981) equation $\rho = 7d_tV_s/(8L)$. This gives a crack density of $4.4 \times 10^{-2}$, close to average values for crack density calculations from shear wave splitting (Crampin, 1994).

5.4. Mantle vs. crustal anisotropy

Fast polarisations from regional earthquakes measured away from the volcano are N–S (average $-1 \pm 4^\circ$) (Fig. 1). They do not line up
well with the regional stress orientations (Townend and Zoback, 2006), but are on average parallel to the fast orientations determined from SKS measurements in the region (Long and van der Hilst, 2005) (−27 ± 29°). Therefore, they are probably affected strongly by the mantle anisotropy. However, within the immediate vicinity of the volcano, there is a general tendency for close to the modelled dyke to exhibit ϕ perpendicular to the dyke, while stations further away have more dyke-parallel ϕ (Fig. 1). We infer that the measurements made on the volcano have been affected both by the mantle and by the local stress field.

As we have previously found at Mt. Ruapehu volcano (Gerst and Savage, 2004), different ϕ are obtained for the “deep” and “shallow” group (Table S1; Fig. 4), which may be caused by the greater influence of mantle anisotropy on the deep group.

For regional earthquakes, dt values average 0.3 to 0.4 s (Table S1). If anisotropy is constant with depth then the path length for the anisotropic region is 12 to 16 km, which is an average value for the depth below which cracks are assumed to be closed (Crampin, 1994). If the cracks progressively close with depth, then the anisotropy is present throughout a larger region. Fast polarisations caused by olivine orientation in the mantle may have been reoriented by the crustal stress field. The waveforms split by the lower layer may resplit in an upper layer, or if ϕ rotates slowly enough, dt can remain large while ϕ rotates to the stress orientation at the surface (Rumpker and Silver, 1998; Saltzer et al., 2000). We think that the waves have been resplit in the crust, based on the average incoming polarisation ϕ determined from the shear wave splitting code (Silver and Chan, 1991). In an isotropic earth, shear wave polarisations depend upon the focal mechanisms, but if there are two or more layers of anisotropy, the waves will be reoriented as they pass through each layer (e.g., Silver and Savage, 1994). The splitting code returns the fast orientation of the last layer, ϕ, and also ϕ, which is the orientation of the wave just before it arrived at the last layer. This ϕ will be the fast orientation of the second to last anisotropic layer if more than one anisotropic layer exists. The events deeper than 40 km yield ϕ of −31 ± 7° for 2488 measurements. This is within the 95% confidence interval of the −27 ± 29° average of ϕ from 11 nearby SKS measurements, which are phases that travel through the entire mantle and which were interpreted as caused by olivine orientation in the mantle above the subducted slab (Long and van der Hilst, 2005). The average dt of 0.6 ± 0.2 s for SKS phases in the region is consistent with the smaller dt of 0.3 to 0.4 s for regional events, which travel through only part of the mantle. Thus we think that the waves from the deep events are first oriented by the mantle, and then resplit by the crust.

5.5. Difference between “steep” and “deep” groups

Delay times for events at station AVO within the “deep” group (with shallow incidence angles) average 0.42 ± 0.03 s (Table S1), somewhat higher than those in the “steep” group (0.32 ± 0.02 s), despite the theoretical relationship that vertical cracks should yield lower delay times in this region (Fig. 2). The two groups also have similar hypocentral distances (172 km for the “steep” group and 176 for the “deep” group), with the biggest difference being a shallower depth of 83 km for the “deep” group versus 154 km for the “steep” group. Therefore this difference cannot be simply explained with one layer models and may be caused by the nonlinear interaction of splitting in the mantle and crustal layers, which would require modelling beyond the scope of this paper (Silver and Savage, 1994).

The most measurements and the best correlation between GPS baseline length change and delay time came from station AVO for events in the “deep” group, which have shallower incidence angles than the “steep” group (Fig. 5). Station AVO is close to the volcano and is located at the observatory, so that it is the longest operating station and it is fixed immediately if it fails, and therefore it is not surprising that it has the most measurements. Likewise, because of a restricted area, there were fewer events in the “steep” group; in particular, they didn’t occur during the time period with the most rapid variation in delay time (Fig. 9). However, the “steep” events follow similar patterns in dt to the “deep” group between late 2003 and mid-2007, and in ϕ after the eruption (compare Fig. 9 with Figs. 5 and 7). The 10-point moving average for the “steep” events yields similar results to the 20-point moving average for the “deep” set, because the averages cover similar time periods.

The average fast polarisations for the “steep” set at AVO before the eruption appear to be rotating in the opposite direction to the average for the events with shallower incidence angles (compare Figs. 7 and 9). However, close examination of individual measurements reveal that the

Fig. 9. Splitting measurements from station AVO in the “steep” group. Single measurements are purple circles. Red circles are 10-point moving averages. Other symbols as in Figs. 5 and 7.
difference in rotation is an artefact due to having fewer events in the “steep” group, combined with rapid rotation of the stress field.

5.6. Effect of angle of incidence and “Band-1 and 2”

In cracked media, theoretical delay times vary for paths coming in at different angles to the crack plane, even for constant crack density (Fig. 2). Therefore, changes in crack plane orientation could cause changes in splitting delay times. As explained in Section 3.1 “Event Groupings,” “Band-2” is a region of propagation angles that arrive within 15° of the crack plane (Peacock et al., 1988). Within this band the delay times vary slightly but not greatly with propagation direction. Outside this region, “Band-1,” the delay times change rapidly with both angle of incidence and with propagation direction. There is a small region where the fast and slow waveforms change positions, so that the fast polarisation changes by 90°.

The paths of the “steep” measurements all have incidence angles within 17° of vertical, and are all within the region denoted as “Band-2.” So, for these paths, the rotation in the crack planes should lead to little change in delay time. Therefore the time variation in delay time for these events is unlikely to be caused by rotation of the crack planes (Fig. 2).

The “deep” and “shall” groups are coming in at lower incidence angles. Some are within “Band-1” and some are in “Band-2.” In this region, if the cracks rotate due to a rotating stress field, constant paths between earthquakes and receivers will pass through varying angles with respect to the crack planes. They will then migrate between

![Fig. 10](image)

Recalculation of time variation at station AVO for deep events with shallow incidence angles, separating results into those regions where the angle between the propagation direction and the crack planes (see Section 5.6) is greater (“Band-1”; a and b) or less (“Band-2”; c and d) than 15°. Variations are observed with both sets, but since there is a wider region with crack planes less than 15°, there are more measurements in that region and they exhibit the same relationship with the GPS as the whole group did in Fig. 5.
“Band-1” and “Band-2.” Therefore, we test the hypothesis that the changing delay times (Fig. 5) as well as changes of 90° in fast polarisation (Figs. 1, 4, 6 and 7) may both be caused by the rotation of the crack orientations due to the stress field changing direction. As discussed in Section 5.4, the shallow and deep events exhibit different behaviour (Fig. 4), so here we concentrate only on the deep events. We separate the set of events defined as “deep,” into measurements whose paths went through the two bands. We divide our time periods into intervals over which the fast polarisations of the “steep” events have relatively constant values, and we use the average $\phi$ from the “steep” events to determine a crack plane orientation. We use these crack planes to determine whether events are inside or outside the 15° “Band-2” region (Fig. 10). Variations in delay time and the correlation with GPS path length changes are maintained in both groups. Therefore, we conclude that the variation in delay time is not controlled by changing crack orientations.

5.7. Time variations and crack aspect ratios

Travel paths near the surface affect the splitting measurements more than distant sections of the path (Rümpker and Silver, 1998; Saltzer et al., 2000), so that time-varying effects are most likely caused by near-surface stress changes, which can occur in seconds to minutes (Zatsepin and Crampin, 1997), rather than changes in mineral orientation at depth, which should occur over periods of thousands of years (Ribe and Yu, 1991). If we take the simplest interpretation that the changes in $\delta t$ are all caused by the changing widths of thin coin-shaped cracks (Hudson, 1981) and that the GPS baseline changes are also caused by opening and closing of these cracks, then we can relate the two measurements to give the aspect ratio $\alpha$, defined as the ratio of crack width to radius, as follows (Hatchell and Bourne, 2005): $\alpha = (4\varepsilon / \rho n^2)$ where $\varepsilon$ is the strain change. If we assume the strain occurs over the same 3.5 km path length for the GPS baseline change as for the shear wave splitting change, then $\varepsilon = 0.02 m/3.5 km = 5.7 \times 10^{-06}$. This yields $\alpha = 2.6 \times 10^{-05}$. It is independent of path length if the same length is used to calculate $\varepsilon$ and $\rho$. The small crack aspect ratio is indeed consistent with the assumptions of the model that the cracks are thin.

5.8. Comparison of delay time variations and polarisation variations

Surprisingly, neither the GPS baseline changes nor the delay time changes at station AVO have as strong a relation to the main eruption or to the numbers of b-type and a-type events as does $\phi$ (Figs. 5 and 7). $\phi$ is a proxy for stress orientation, while GPS and delay time measurements depend on the crack density and aspect ratios, and so are less directly related to the stress. One explanation is that, as the stress orientation changes it closes cracks of one orientation, while opening those in another orientation. In this way the numbers of cracks could remain constant even as the orientation changed. Perhaps the changing stress orientation occurred as the magma conduit geometry changed, which allowed an easier path for the magma to reach the surface.

6. Conclusions

Geodesy and seismic anisotropy can be combined to derive new information about volcanic regions such as the horizontal differential stress and the crack aspect ratios. Correlations of $\phi$ and $\delta t$ with volcanic activity furthermore suggest that magmatic intrusions disrupt the local stress field. Monitoring of this stress field by measuring anisotropy variations through shear wave splitting and possibly other techniques such as surface wave noise analysis (Brenguier et al., 2008), could be used in areas where geodetic information is unavailable, or where the changes are taking place too deeply to be observed by surface geodetic stations.

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Appendix A. Supplementary data


References

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