Measuring S_{Hmax} with Stress-Induced Anisotropy in Nonlinear Anelastic Behavior

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Abstract

Mechanical stress acting in the Earth`s crust is a fundamental property that has a wide range of geophysical applications, from tectonic movements to energy production. The orientation of maximum horizontal compressive stress, S_{Hmax} can be estimated by inverting earthquake source mechanisms and directly from borehole-based measurements, but large regions of the continents have few or no observations. Available observations often represent a variety of length scales and depths, and can be difficult to reconcile. Here we present a new approach to determine S_{Hmax} by measuring stress induced anisotropy of nonlinear susceptibility. We observe that nonlinear susceptibility is azimuthally dependent in the Earth and maximum when parallel to S_{Hmax}, as predicted by laboratory experiments. Our measurements use empirical Green's functions that are applicable for different temporal and spatial scales. The method can quantify the orientation of S_{Hmax} in regions where no measurements exist today.

Introduction

Knowledge of the mechanical stress acting in the Earth's crust and lithosphere is important for a wide range of geophysical studies and applications[1–4] including plate tectonics[5,6], seismicity[7–11] and subsurface fluid behavior[9,12,13]. It is commonly represented as the orientation of the maximum horizontal compressive stress (S_{Hmax})[3,8,14–16]. Other information regarding the principle components is often not known, much less the full stress tensor. At regional to tectonic-plate scales, the orientation of S_{Hmax} is determined by plate boundary forces and tractions along the bottom of the lithosphere[17]. At local scales (<10 km), the orientation of S_{Hmax} may vary due to heterogeneities in density and elasticity, slip on faults[7,12], and pore pressure[12]. The orientation of S_{Hmax} is commonly estimated using borehole-based methods[18,19] and inverting earthquake focal mechanisms[20–24], and less commonly by measuring the orientation of stress sensitive geologic features[25]. Borehole-based methods are high-cost, point measurements with unknown applicability away from the borehole[26], and commonly applied in hydrocarbon producing regions[8]. Interpreting earthquake focal mechanisms is limited to seismically-active areas and requires an adequate monitoring network. Because of the limitations of these techniques, broad regions of continental interiors are poorly constrained where there are few or no measurements[15].

Rock samples in laboratory experiments typically exhibit anisotropic nonlinear elastic properties when a uniaxial stress is applied[27,28] (Figure 1). In the laboratory, the pressure derivative of the wave modulus (called nonlinear susceptibility, NS) is strongest when the angle between the uniaxial stress and the propagation of the probe wave is zero, and weakest when the angle is 90 degrees. The effect is greatest for compressional P-waves and the nonlinear elastic behavior is quantified by measuring this behavior. We investigate whether nonlinear elastic properties of the Earth are sensitive to the orientation of S_{Hmax} because of anisotropy in rock compressibility.

Rocks are heterogeneous materials with stress and strain dependent elastic properties, and finite, nonzero relaxation times (the slow dynamics)[29–31]. The relationship between stress, strain, and elasticity is complex in individual rock samples[32,33] with mechanical damage and weak grain contacts being primarily responsible for nonlinear elastic behavior[34]. Temperature, pressure, and the presence of fluids modulate the nonlinear behavior[34,35]. In the Earth, seismic velocities are commonly observed to be faster when rocks are compressed, usually interpreted as the closing of cracks[36,37], while they are typically slower after experiencing strong shaking, usually interpreted as the breaking or weakening of internal contacts[38,39]. After the disturbance, the material relaxes back to its original or a new metastable state, the process of slow dynamics. Thus, rocks are metastable in their elastic behavior and strongly influenced by relatively weak external forces perturbing the material structure[35,38,39].

We utilize this nonlinear elastic behavior in rocks and apply a new technique to passively monitor the orientation of stress in the lithosphere. Our approach to measure S_{Hmax} *in situ* relies on seismic velocity measurements that employ Empirical Green's Functions (EGF) derived from ambient noise recorded at multiple pairs of seismic stations[40]. Most studies in Earth that measure temporal changes in seismic velocities do so by differencing the phase in the coda part of the EGF[36,38,39]. The coda of the EGF follows the direct waves and is the result of scattered waves that travel through some volume between the two stations[41]. We can measure the velocity sensitivity to strain using a classic nonlinear acoustic approach known as the pump-probe method[33] where the material is strained with a low-frequency oscillation (pump) and the elasticity is monitored by measuring the travel time of a high frequency probe wave that is applied at different points in the pump cycle.

For this study, solid Earth tides are used as the low-frequency pump and EGFs are the high frequency probe. We perform this natural pump-probe experiment in two prototype studies located in north-central Oklahoma, U.S.A. and north-central New Mexico, U.S.A. (Figure 2). We selected north-central Oklahoma because of the ongoing induced seismicity, generated by decades of injected wastewater from oil and gas operations[42,43], that tends to occur on faults optimally orientated in the regional stress field[9]. North-central New Mexico was selected to test if we can resolve similar results in a geologic setting that straddles a continental rift and has a different stress field than Oklahoma[16,44] (Figure 2). In north-central Oklahoma, S_{Hmax} is oriented approximately N80E with some local variations[9], but in north-central New Mexico, S_{Hmax} is aligned nearly south-north along the Rio Grande Rift and rotates to a more east-west orientation in northeastern New Mexico[16]. The dominant faulting style in Oklahoma is strike-slip, though there is some normal faulting in the north in the vicinity of our study area while the faulting style is strongly normal faulting in northern New Mexico, associated with the Rio Grande Rift[8].

The Earth exhibits stress induced anisotropy of nonlinear susceptibility (NS) that is aligned with S_{Hmax} in two different geologic settings, matching observations from the laboratory when a uniaxial stress is applied to laboratory rock samples. Since our measurements use only ambient seismic noise, there are several advantages over existing methods for estimating the orientation of S_{Hmax}: (1) earthquake source properties are not used or required, (2) borehole measurements are not required, (3) sufficient seismic data exists in many regions of interest where traditional stress measurements are unavailable, and (4) the technique can be applied at a wide range of spatial and temporal scales.

Results

In both study areas, on average, the Earth is slower during extension than during compression by fractional velocities of 0.07% and 0.2% with uncertainties of 10% of the velocity change, for Oklahoma and New Mexico, respectively. This is consistent with the opening and closing of cracks, and the stiffening of internal contacts during compression. In Figure 3, we report NS as fractional velocity change, though NS is actually fractional velocity change per unit strain. Tidal strain is on the order 10^{-8} and it varies somewhat from cycle-to-cycle. Our results reflect the average peak-to-peak strain amplitude, which is discussed below.

Oklahoma

In Oklahoma, a fitted sine function to the results shows the maximum (negative magnitude) nonlinear NS occurs between 69°-86°, depending upon the selection of stations. Borehole measurements and a focal mechanism inversion estimate S_{Hmax} orientations between 71°-84° in the same region[9]. In this previous study, the reported S_{Hmax} azimuth is lower in the north than in the south (rotated counter-clockwise), and consistent with the values we observe from the maximum NS, see Figure 1 and Table 1 in Alt and Zoback (2017)[9]. Comparing our results to stress indicators used in the World Stress Map[25] (shown in Figure 2), our results are more consistent with borehole than earthquake measurements, which are inconstant with each other in some places. We note that considering only some of the northern stations produces ambiguous results. This ambiguity and a few positive values in Figure 3, indicating that velocities are faster during extension, may be the result of poroelastic effects and are discussed in more detail below.

New Mexico

In New Mexico, a fitted sine function to the results using all stations shows the maximum (negative magnitude) NS occurs at 178° (Figure 4). The maximum NS using only the 6 westernmost stations is 8° and using only the 6 easternmost stations is 163°, with the 3 central stations used in both subarrays. The regional stress indicators show a transition, moving west to east, from slightly southwest-northeast to south-north S_{Hmax} orientation within the Rio Grande Rift (Figure 2). Continuing east, a southeast-northwest S_{Hmax} orientation is expected, although no known stress indicators are available within this transition zone. Considering the 6 western stations, the NS predicts S_{Hmax} at 8°, which is consistent with stress indicators within the rift valley and mountains to the west. Using the 6 easternmost stations, NS predicts S_{Hmax} at 163°, which is rotated counterclockwise from the results of the western stations, and intermediate between the reported stress indicators within the Rio Grande Rift and the reported southeast-northwest stress indicators east of the study area. Our NS derived S_{Hmax} results for all 9 stations are in agreement with the average orientation of stress indicates within the footprint of the seismic array. Since no stress indicators exist for the eastern part of the study area, our results using the eastern stations suggest the observed clockwise rotation from east to west transitions into our study area. The results provide a clear example of constraining the stress field using passive seismic data in a region where no other estimates are available.

Discussion

The azimuthal dependence of NS closely tracks the orientation of S_{Hmax} in the Earth as shown in the laboratory[27,28]. In terms of elastic constants, what we measure is closely related to the 1-D nonlinear anelastic coefficient β which is the coefficient linearly related to strain in a Taylor expansion of Hooke's law (see, e.g., equations 7 and 8 in Johnson and Rasolofosaon[27] and Pantea et al.[45]). In single crystals and metals β is less than 10. In Earth materials it can be considerably larger, order 10^2-10^3 underscoring how very nonlinear elastic Earth materials can be. In the laboratory experiments, stress induced anisotropy is strongest for P-waves[28], which suggests our measurements describe the scattered compressional-wave energy either in the form of Rayleigh waves or P-body waves. Since our measurements are made with the vertical channels of a seismic station pair, we are measuring a specific component of the nonlinear coefficient β, which may be better represented as a tensor. Exploring the possible tensor properties of β is beyond the scope of this study but may be possible by developing a 6 component NS tensor using all combination of station channels. Such a measurement would be very useful in discerning the anisotropic mechanical damage variations in the upper crust.

The difference in strain between maximum extension and maximum compression for solid Earth tides is of the order 5×10^{-8}, which means the observed NS ($dv/d\varepsilon$, change in velocity over change in strain) is of the order 10^4-10^5. These results are similar in magnitude to those found by Takano et al.[46] near a volcano in Japan using EGF frequencies of 1-2 Hz, and an order of magnitude higher than those found by Hillers et al.[47] in California using frequencies of 2-8 Hz. In these two previous studies, the authors measured nonlinearity, but did not report azimuthal differences, nor the relationship between nonlinearity and stress-induced anisotropy.

In addition to the opening and closing of cracks in dry conditions, resulting in the softening and stiffening of internal contacts, there may be poroelastic effects. Under saturated conditions, pore pressure increases during applied compression and decreases during applied extension, with pore pressure having the opposite effect on the effective confining stress as the applied tidal stress. This effect is expected to be isotropic in most rocks[48]. If all station pairs in an array experience the same poroelastic conditions, the effect of stress-induced anisotropy is preserved, though curves shown in Figure 3 would shift upward (positive). In some cases we could even see positive values[47]. If different station pairs experience different poroelastic effects, this would complicate the estimation of S_{Hmax} since the observed azimuthal dependence of NS would no longer be due only to stress-induced anisotropy. In Oklahoma, we are able to get good estimates for S_{Hmax} despite likely contributions from heterogeneous poroelastic conditions[49] that are apparent when using a fewer number of station pairs. This may be why using only northern stations in Oklahoma produces ambiguous results, and that the sinusoidal fit is generally worse when using a fewer number of station pairs in Figures 3 and 4.

Measuring and modeling stress in the Earth's crust is challenging and it is important to match the length scale and depth to the desired application. Since stress heterogeneity likely exists at all scales[10,20,26], a measured stress or S_{Hmax} may not be representative of different length scales or depths. Our method has the potential to address some of these challenges. EGFs can be calculated using varying interstation distances to provide S_{Hmax} estimates at different horizontal length scales. Estimating S_{Hmax} at specific depths is more challenging but possible when using relative depths inferred through frequency content and coda time offset in the EGFs[38,41] but is a promising area of research. Perhaps most importantly, measuring the orientation of S_{Hmax} is not limited to locations with earthquakes or boreholes, and provides data driven constraints to regional estimates. Additional possibilities include calculating the time evolution of NS to obtain S_{Hmax}, which could reveal changes in relative amplitude and orientation. Temporal monitoring of subsurface fluid reservoirs or active fault zones may represent changes in pore pressures or fault zone properties during the loading cycle.

Long wavelength stress and deformation patterns in tectonic plates are generated from global scale mantle convection[50]. Smaller-scale patterns are related to the gravitational potential of a heterogeneous crust and lithosphere[51], or small-scale convection in the mantle[52]. The relative contribution of these two mechanisms is unknown. Ultimately, we do not know to what extent continental-scale stress models represent the actual stress field in regions with few or no measurements to constrain these estimates. Therefore, we cannot attempt to model or characterize mechanisms for these unknown heterogeneities. This method provides a dense and uniform metric of the orientation of S_{Hmax} across continental regions, which will improve stress models and our understanding of the underlying geodynamical processes.

We calculated EGFs as a function of tidal strain and azimuth in north-central Oklahoma and north-central New Mexico to constrain NS and derive the orientation of S_{Hmax}. Our results show in both study areas the seismic velocities are, on average, faster when the Earth is in compression relative to when the Earth is in extension. We observe stress induced anisotropy in nonlinear anelastic behavior, which is aligned with S_{Hmax} and provides a new technique to estimate the orientation of S_{Hmax} without focal mechanism inversions or borehole measurements. Large scale application of this method may resolve additional tensor properties of the nonlinear coefficient β, reveal how S_{Hmax} varies with horizontal length scales and depth, and how S_{Hmax} evolves temporally in regions such as fluid reservoirs and active fault zones.

Methods and Data

We use publicly available seismic data from the two study areas, north-central Oklahoma and north-central New Mexico. For Oklahoma, we obtained waveform data recorded by the Nanometrics Research Array (NX) from the Incorporated Research Institutions for Seismology Data Management Center (IRIS-DMC, www.iris.edu). The NX array consists of 30 broadband, 3-component instruments that recorded at 100 samples per second for about three years between mid-2013 and mid-2016 (red circles in Figure 2). For New Mexico, we obtained waveform data from 9 stations in Earthscope's Transportable Array (TA) from the IRIS-DMC. This subarray consists of 9 broadband, 3-component instruments that recorded at 40 samples per second for about two years between mid-2008 and mid-2010 (blue circles in Figure 2). We used only the vertical component for all seismic data.

Signal Processing

We organize the data into day-long segments and deconvolve the instrument response. When calculating EGFs from continuous broadband seismic data it is important to remove transient signals like earthquakes[53]. We remove earthquake signals from the data using the U.S. Geological Survey Comprehensive Catalog. We assign zeros with a taper for the waveform segments following three sets of earthquake criteria: (1) earthquakes with a minimum magnitude of 3.5 and maximum distance of 30 km from the array, between surface wave velocities of 2 and 5 km/s, (2) earthquakes with a minimum magnitude 5 and maximum distance of 2000 km from the array, between surface wave velocities of 2 and 7 km/s, and (3) earthquakes with a minimum magnitude of 6 at any distance, between surface wave velocities of 2 and 8 km/s. This resulted in the zeroing of 7.9% of waveforms for Oklahoma and 4.3% of waveforms for New Mexico. The disparity exists because there are more local earthquake in Oklahoma than in New Mexico.

Additionally, we clip all signals greater than 3 times RMS for each day-long segment to remove non-earthquake signals observed as emergent or impulsive noise.

Tidal Strain

The volumetric tidal strain is obtained using the software package SPOTL[54]. We use the volumetric strain component because we expect the nonlinear behavior to be localized on pre-existing faults, which may be at any orientation. We divided time into segments that fit in two strain magnitude bins, the top 25% and the bottom 25% where "top" refers to maximum extension and "bottom" refers to maximum compression.

Empirical Green's Functions

We cut and merge the day long preprocessed waveforms into segments appropriate for each stress bin and discard anything shorter than 30 minutes. We empirically determined that a station separation distance between 30 and 60 km produce the best EGFs for Oklahoma and selected station pairs accordingly. For the New Mexico stations we used all pairs. We calculated an EGF for each selected station pair, and segment for all bins using a phase cross correlation method[55] where we pre-whiten the spectrum before applying a phase cross correlation. There are approximately 780 segments for each station pair during the recording period, though the actual number for each pair varies based on data availability, data quality, and other factors.

Next, we describe the EGF stacking procedure (Figure 5). For each station pair we selected 14 day windows, and using the center of the window, selected all EGFs whose segment start time falls within +/- 7 days. Each EGF is scaled by the square root of the duration of the underlying time series and are all stacked. We calculate the Pearson correlation coefficient of each EGF with the stack. Any EGF that has a value of less than 0.5 is discarded and we produce a new stack with the remaining EGFs. We reevaluate the discarded EGFs and any that have a Pearson correlation coefficient greater 0.5 using the updated stack is re-included to create a new stack. The process is repeated until there are no discarded EGFs with a Pearson correlation coefficient greater than 0.5. Subsequent 14 day windows are calculated with a 7 day overlap. This stacking procedure is intended to include as many observations as possible while discarding outliers. The outcome is stacked EGFs for each station pair that represent 14 day windows with 7 day overlap for each of the two strain bins described above.

We sum the causal and acausal parts of the EGF and select the coda part as shown in Figure 5 to avoid direct wave arrivals. We determine the average phase difference and velocity change ($\Delta v/v$) between two stacked EGFs in a 30 second coda window for waves between 4 and 5 seconds period following the steps outlined in the wavelet method of Mao et al.[56]. We used a Morlet wavelet with $\omega_0 = 0.25$ Hz that corresponds to the periods we analyzed and allows us to recover the known phase shifts in simple synthetic examples. This method has the ability to measure phase shifts associated with changes in velocity as a function of frequency and coda offset time, but we are only interested and present the average velocity changes. Results measured by frequency and coda offset time likely contain depth information[38,41,56] but are beyond the scope of this study. Along with measuring phase shifts, we also calculate coherence between the two EGF stacks and discard any cases where the average coherence falls below 0.95. In our convention a negative ($\Delta v/v$) value means that the Earth is slower during extension than during compression, which we tested with synthetic examples.

The coda part of the EGF consists of scattered waves that are composed of surface waves and body waves. In general, the earlier part of the coda contains more scattered surface waves, while the later part contains more body waves, with the transition time governed by the scattering properties of the subsurface[41]. If the window of the measurements contains mostly surface waves, they are sensitive to the upper 2-3 km for Rayleigh waves between 4-5 second periods. If the window of the measurements contains mostly body waves, then the waves would be sensitive over a greater depth range, depending on the velocity of the scattered waves, and the scattering properties in the subsurface.

We group and stack the station pairs by azimuth to examine any directional dependence to the results. The azimuth for each pair is determined using the relationship between the more western station to the more eastern station so that values are always between 0 and 180, with 0 and 180 indicating south-north and 90 indicating west-east. For Oklahoma, we consider 9 azimuths in 20 degree steps. For each azimuthal interval, we average the dv/v values for all pairs whose azimuth is within +/- 20 degrees with wrapping. For example at an azimuth of 0 degree, we average paths between 0 and 20 plus between 160 and 180 degrees (which is equivalent to between -20 and 0 degree). For New Mexico we consider all station pairs individually because there are not enough station pairs to average in azimuthal bins. In addition to calculating the average $\Delta v/v$ at different azimuths, we fit a sine function, periodic on 2θ, to the results.

Uncertainties in our measurements are difficult to precisely estimate. When calculating EGFs, there is an intrinsic assumption that noise sources are equipartitioned, white, azimuthally uniform, and stationary. This is never true in the Earth, but we can take steps to reduce the influence of recordings that violate these assumptions. We assume that velocities in our study area vary measurably by no more than the amounts observed in other studies, less than +/- 1%[39,46,47]. Since we never directly compare data that has been recorded more than 14 days apart, seasonal variations are not expected to be important, including spectral content, azimuthal variations, and relative amounts of coherent and incoherent noise. By discarding EGFs that are not well-correlated and coda that are not highly coherent we avoid noise or signals that are not stationary. In both study areas, we stack over two years of data. The resulting uncertainty based on the variance in the accepted measurements suggests that uncertainties are at least an order of magnitude less than the magnitude of the measurements.

Data Availability

All data used in this study are available through the Incorporated Institutions for Seismology Data Management Center (www.iris.edu).

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Author Contributions

AAD and GHRB conceived the experiment, AAD performed all calculations, and contributed to the writing of this manuscript. PAJ, GHRB, and CWJ contributed to the interpretation of the results and the writing of this manuscript.

Competing Interests

The authors have no competing interests.

Figure 1.

Stress induced anisotropy in Nonlinear Susceptibility from Johnson and Rasolofosaon27. The vertical axis represents nonlinear susceptibility. The horizontal axis shows the angle between the orientation of the uniaxial stress and the direction of propagation of the probe wave. (AGU grants permission for individuals to use figures, tables, and short quotes from AGU journal and books for republication in academic works provided full attribution is included.)

Figure 2.

We measured the azimuthal dependence of nonlinear elastic behavior in north-central New Mexico and north-central Oklahoma. The blue circles are seismic stations from Earthscope's Transportable Array. The red circles are seismic stations from the Nanometrics Research Array. The short black lines indicate the direction of S_{Hmax} for stress indicators used in the 2016 World Stress Map, classes A, B, and C. The thick stress indicator lines are from borehole measurements and the thin stress indicator lines are from earthquake or geologic feature orientations. The rectangle in the top map indicates the position of the bottom map within North America.

Figure 3.

Shown are the seismic stations and azimuthal dependence of nonlinear susceptibility in Oklahoma. Red stations in the left panels are used to calculate fractional velocity changes shown on the right (see Fig. 2 for station locations). Vertical red bars represent uncertainties at the 99.5% confidence interval. The orange curve is the best fit sinusoidal function. The value listed is the azimuthal angle of maximum nonlinear susceptibility according to the fit sinusoid (negative fractional velocity changes).

Figure 4.

Shown is the azimuthal dependence of nonlinear susceptibility in New Mexico. The left, center, and right axes show the results using the 6 western stations, all stations, and the 6 eastern stations, respectively. The orange curve is the best fit sinusoidal function. The vertical red bars represent the fractional velocity change with uncertainties at the 99.5% confidence level for each station pair. The angle value reported on each axis indicates the azimuth with the maximum NS (negative fractional velocity change).

Figure 5.

Empirical Green's Functions, for Oklahoma (left), and New Mexico (right), ordered by inter-station distance. The black lines bracket the coda used for the velocity calculations.

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