IMPROVED UNDERSTANDING OF INTRAPLATE EARTHQUAKES IN THE SOUTHEASTERN USA WITH MATCHED FILTER DETECTION

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IMPROVED UNDERSTANDING OF INTRAPLATE EARTHQUAKES IN THE SOUTHEASTERN USA WITH MATCHED FILTER DETECTION

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SUMMARY

Most earthquakes occur along plate boundaries (also known as interplate earthquakes) and are caused by repeated accumulation and release of strain in tectonic plates moving past one another. However, the same driving forces causing interplate earthquakes does not account for intraplate earthquakes, which are located within the interiors of a tectonic plate. The relatively long recurrence intervals between large earthquakes, causal fault locations, and driving mechanisms of intraplate earthquakes present a challenge to understanding the seismic hazard in intraplate regions.

To better understand earthquake properties in intraplate settings, a more complete detection of earthquakes, precise locations, and magnitude estimations are critical. Traditional earthquake catalogs tend to miss smaller earthquakes buried in the background noise or coda waves of larger earthquakes, which results in an incomplete catalog. To overcome this, I use a matched filter method to detect microseismicity and build a more complete catalog. This technique applies cross correlation to detect previously uncatalogued events in continuous data by using the waveforms of known earthquakes as templates.

This thesis focuses on earthquake detection in the Southeastern United States, an intraplate region within the North American Plate hosting several seismic zones that are not well understood. In particular, I focus on the Piedmont Province, Eastern Tennessee Seismic Zone (ETSZ), and Middleton Place-Summerville Seismic Zone (MPSSZ).

In the Piedmont Province, I found that the 2014 Mw 4.1 Edgefield, South Carolina had an aftershock deficiency, suggesting that most of the strain was released during the mainshock. The mainshock also had a low stress drop which may account for the shallow calculated depth of this earthquake and the low number of aftershocks.
I examined the recent 2018 Mw 4.4 earthquake in the ETSZ, which had a similarly shallow depth, and very few aftershocks. I resolved the fault orientation on which the mainshock lies, which was previously unknown by finding focal mechanism of earthquakes near the mainshock and by performing rupture directivity analysis.

I detected and relocated microseismicity in the ETSZ using over 15 years of continuous data, yielding the most detailed complete catalog yet for this seismic zone, as well as magnitude estimations and more defined structure at depth. I found the greatest concentration along or to the east of the NY-AL Lineament, as defined by the magnetic anomaly, supporting the evidence that this feature’s origin is linked to seismicity in the ETSZ. I also examined seismicity around the Watts Bar Reservoir, near which the Mw 4.4 mainshock occurred, and found some evidence for Reservoir Induced Seismicity around this region. I also found limited evidence for hydrologically-driven seismicity due to seasonal rainfall in the shallow portion of the ETSZ, which contradicts some previous studies which hypothesize that most intraplate earthquakes are associated with the dynamics of hydrologic cycles.

In the MPSSZ, near Summerville, South Carolina, I detected new events during a temporary seismic deployment in 2011-2012 and relocated them. I found deep clusters and linear features which are not in line with the currently hypothesized extrapolation of the major fault plane in this region at depth. This has implications for the fault structures which are responsible for the 1886 M~7 Summerville, SC earthquake to be compared with future studies.
CHAPTER 1. Introduction

More than 90% of the total seismic energy is released along the boundary between tectonic plates, also known as interplate earthquakes. Compared to interplate seismicity, the causes of intraplate seismicity (i.e., earthquakes occurring within a tectonic plate) are not well understood. Tectonic loading rates are much higher along plate boundaries than in intraplate regions, resulting in longer earthquake recurrence in the latter. Intraplate earthquakes are more challenging to study not only because of their infrequent occurrence, but also because their recurrence intervals, causal fault locations and driving mechanisms are all poorly constrained.

Most of the largest known earthquakes in these regions are historical, with uncertain magnitudes and locations, as they occurred before the advent of modern seismological instrumentation. Seismological observation of present-day intraplate seismicity is sparse, and seismic stations are often separated by large distances of tens to hundreds of kilometers. Paleoseismic evidence suggests that we do not have a complete earthquake cycle record for many regions. This leads to large uncertainty about the seismic hazard in intraplate zones.

The Southeast United States hosts several intraplate seismic zones. The bedrock of the Eastern United States is colder, denser, and less fractured than that of the West Coast. This results in earthquake waves of a similar magnitude traveling farther with greater intensity on the East Coast than on the tectonically active West Coast. Hence, although earthquakes are less frequent than on the west coast and tend to have smaller magnitudes, the potential for damage is larger.

The focus of this thesis is on intraplate seismicity in three distinct regions of the Southeast United States. Many seismic zones are located here, including the Eastern Tennessee Seismic Zone (ETSZ), Middleton Place-Summerville Seismic Zone (MPSSZ), Bahamas Fracture
Seismic Zone (BFSZ), and at the periphery of this region, the Wabash Seismic Zone (WBZ), New Madrid Seismic Zone (NMSZ), and Central Virginia Seismic Zone (CVSZ). There is also the Piedmont Province that hosts another region of diffuse seismicity around the border of Georgia and South Carolina but is not named as a seismic zone yet. This thesis examines microseismicity in the ETSZ, MPSSZ, and Piedmont Province using a template matching method. This newly detected seismicity is further relocated with cross-correlation-derived differential times, and its magnitude is re-calibrated. An improved high-resolution catalog can help to better illuminate subsurface fault structures that host those microseismic events, moderate-size earthquakes, or historically large earthquakes, as well as spatial-temporal evolutions during its sequence.

This thesis focuses on utilizing continuous-time detection and relocation of seismic events to improve our current understanding of earthquakes in intraplate settings in the Southeastern United States. In Chapter 2, I first give an overview of the common methods used for many of the study regions in this thesis. Chapters 3-6 are case studies of different seismic zones in the Southeast. Chapter 3 focuses on the source properties of the 2014 Mw 4.1 Edgefield, South Carolina earthquake and on the aftershock productivity of this earthquake sequence. Chapter 4 focuses on the fault orientation and relocated seismicity associated with the 2018 Mw 4.4 Decatur, Tennessee earthquake, one of the largest events in the ETSZ in the past decade. Chapter 5 then examines over 15 years of continuous data in the ETSZ to better characterize this seismic zone, including seismicity distribution, general fault structure, relationship to the NY-AL Lineament, and potential for hydrologically driven seismicity in the shallow section of this seismic zone. Chapter 6 is about the detection of additional earthquakes associated with a recent 2011-2012 deployment in the MPSSZ, and on the comparison between my relocation of
seismicity to previous results in this seismic zone. In general, the chapters in this thesis are presented chronologically, in order of work completed throughout the last several years.
CHAPTER 2. Methods of Earthquake Detection, Relocation, and Magnitude Estimation

Earthquake detection, relocation, and magnitude estimation are critical components of seismological studies, providing the basis for subsequent analysis of spatio-temporal earthquake properties and subsurface structural imaging. In this chapter, I briefly describe each step in building an earthquake catalog.

An earthquake catalog is generally built by starting with an existing catalog during a certain time and in a certain region. The arrival times of these catalogued earthquakes are refined, if necessary, to create templates as input to matched filter detection. One such method is PSIRpicker, described in greater detail in the following sections. There are multiple methods of earthquake detection, matched filter detection and deep learning being the only algorithms applied in this thesis. However, a more holistic description of traditional detection methods is also presented for context.

The newly detected earthquakes have their phase picks refined by cross correlation and the differential times between earthquakes are computed in order to relocate the earthquakes in the new, more complete catalog. I describe two different methods of relocation, XCORLOC and hypoDD, in greater detail in the following sections. Lastly, the magnitudes of the newly detected earthquakes are estimated using relative amplitudes between the detections and the templates that detected them.

2.1 Earthquake Detection

An earthquake can be identified by the arrival of a P (compressional) phase and an S (transverse) phase, which are generated by a rapid shear slip along a fault plane (e.g., Shearer,
An even stronger indication of the presence of an earthquake is found by observing the moveout of P or S waves with time across several stations with increasing distances from a common source. Earthquake signal detection and seismic phase picking are challenging problems, especially for microearthquakes in the presence of high background noise.

For a local earthquake right beneath a seismic station, a P arrival has the clearest signal on vertical components, whereas S arrivals are the clearest on the two horizontal components of a station. The P phase arrives earlier than the S phase, due to the compressional waves traveling with higher velocity through the Earth’s rock than the transverse waves. Clear, impulsive changes in amplitude with a high signal to noise ratio (SNR) is an indication that a seismic phase has reached a seismic station at that time. Usually, the detection of seismic phases is followed by phase association, in which phases on different stations are attributed to a seismic event of common origin. There are several methods of detecting earthquakes and identifying the seismic phases that define them, described in greater detail below.

2.1.1 Traditional Methods

Traditionally, the detection of earthquakes and seismic phases on continuous waveforms was searched by manual inspection and picking by experienced experts. A large, impulsive amplitude increase indicates that an earthquake may have occurred around that time. However, the signal could also be due to anthropogenic activity (trains, mining blast, etc), instrument calibration or malfunction, so the presence of similar amplitude increases (at slightly different times) on nearby stations can remove some false detections.

Short-Time Average over Long-Time Average (STA/LTA) was one of the early algorithms used to detect potential earthquakes on continuous data (Allen, 1982). As the name
suggests, for a given time point in the continuous data, STA/LTA calculates the average values of the absolute amplitude in both a short window (STA) and a long window (LTA) preceding the point. The short-time window is sensitive to changes in amplitude around the point, while the long-time window provides a measure of the background noise level leading up to the time point. When the ratio of the STA divided by the LTA exceeds a threshold, an event is detected.

While STA/LTA has a quick run time, as the algorithm is quite simple, it also has some disadvantages. It needs an input threshold for the ratio, which requires careful consideration to decide on the desired false positive and false negative rates in the output. A threshold that is too low results in many false detections, while a threshold which is too high will miss many real seismic phases with relatively small amplitudes. Another problem is that STA/LTA is very sensitive to the background noise level, which would result in missing many of the smaller earthquakes, and those with other events in the long-term window leading up to the earthquake. Finally, STA/LTA only output the phase arrival time, but no additional information (i.e., phase type) is provided.

Another simple method of earthquake detection is that of envelope functions (Baer and Kradolfer, 1987; Earle and Shearer, 1994). This method uses the square of the envelope function, which is a smooth curve that outlines the extremes in amplitudes, and uses a method similar to that of STA/LTA to detect changes in this signal. This method is more sensitive to changes in amplitude, frequency, and phase. However, the arrival times of phases tends to be picked late (Sleeman and van Eck, 1999).

Other phase detection technique most involves computing higher order statistics such as skewness or kurtosis (Kuperkoch et al., 2010; Ross and Ben-Zion, 2014), and Auto Regressive-Akaike Information Criterion (AR-AIC) (Morita and Hamaguchi (1984); Sleeman and van Eck,
1999). This method entails modeling a time series as a multiple AR process, by using the Akaike Information Criterion (AIC). The AR-AIC method is typically used to automatically pick the P-phase arrival time around a small-time window around the given expected P-phase arrival time.

2.1.2 Matched Filter Detection

Matched Filter Detection is one of the best-performing methods to automatically detect earthquakes, given “template” waveforms of already identified (catalogued) earthquakes in a certain region. This method works by scanning through continuous waveforms for previously undetected earthquakes. The “similarity” between templates and a continuous waveform is measured by the normalized cross-correlation (CC) (Gibbons and Ringdal, 2006, Shelly et al., 2007; Peng and Zhao, 2009). In principle, the waveforms of an undetected earthquake should be identifiable by their similarity to the waveforms of a known event, provided these earthquakes occurred in similar locations with similar focal mechanisms (Schaff and Waldhauser, 2005).

For the study on the Mw 4.1 South Carolina sequence in Chapter 3 (Daniels et al., 2020), I use a GPU-based matched filter package developed by Meng et al., (2012). For the studies on the ETSZ and the MPSSZ, I implement the matched filter detection using the Python package EQcorrscan (Chamberlain et al., 2017). EQcorrscan can be found at https://eqcorrscan.readthedocs.io/en/latest/installation.html. Within this package, I apply the Fast Matched Filter (FMF) code as the function to calculate cross-correlations, which provides fast parallel time-domain correlations for CPU and GPU architectures (Beauce et al., 2017). As the dataset of over 15 years of continuous data in the ETSZ was large compared to what our machines could process, I take advantage of the Extreme Science and Engineering Discovery Environment (XSEDE) supercomputers with GPU resources using the Bridges system at the Pittsburgh Supercomputing Center (PSC) (Towns et al., 2014; Nystrom et al., 2015).
After applying the matched filter package (Meng et al., 2012) on the Mw 4.1 Edgefield earthquake, I used EQcorrscan for subsequent studies because it offers several advantages. EQcorrscan has built-in capabilities with ObsPy (Beyreuther et al., 2010), which has quickly become the modern means of earthquake analysis in Python. There are other packages that can be directly applied to the output of EQcorrscan, such as polarization analysis, arrival time refinement, and making the input files to hypoDD (Waldhauser and Ellsworth, 2000). For the 15-year study of the ETSZ, the former method (Meng et al., 2012) would have been infeasible because I would have had to download all the continuous data to our disks, which would have been too large. EQcorrscan automatically downloads waveforms temporarily as needed while it runs, then discards them from memory, making it possible to handle long-duration of template matching detection.

Regardless of which exact package is used, the workflow is as follows: prepare the continuous waveform (pre-processing, band-pass filtering, remove mean, filtering); process the templates (use catalog to cut template waveforms, compute the signal to noise ratio (SNR) of waveforms); run the matched filter detection; perform quality control and thresholding on the resulting detections. These steps are shown in the flowchart in Figure 2.1, in the context of the typical workflow for earthquake analysis.
For EQcorrscan, the SNR is calculated as the ratio of the maximum amplitude in the signal (template) window to the root-mean-square (RMS) amplitude in the whole window on the bandpass filtered waveforms. I choose a noise window starting 3.5 seconds before the arrival, which is different than the default 100-second window. I decided on this shorter window to be consistent with past studies in our group (e.g., Peng and Zhao, 2009; Meng et al., 2012, 2013), as the longer the noise window, the more likely another event is present.

On vertical channels, I cut the template window around P arrivals. On horizontal channels, I cut the template windows around S arrivals. This is because P arrivals have clearer
arrivals with greater amplitude on vertical channels, whereas S arrivals are clearer on horizontal channels. All three channels of templates are cross-correlated with the continuous data at each point in time, and the cross-correlation (CC) functions are stacked and normalized between −1 and 1. After obtaining the stacked station mean CC, the algorithm performs peak detection using a threshold above the noise level.

Depending on the study, there is some variation in the exact threshold used in matched filter detection. The threshold can depend on data quality, geologic properties of the local rock which can result in higher or lower average CC values, and the tolerance of particular studies for false detections.

The most commonly used threshold is an integer factor multiplied by the Median Absolute Deviation (MAD) (Shelly et al., 2007). The MAD is a measure of the variation of data, and is therefore useful in detecting anomalies and atypical patterns. For each time index, $i$, in the CC function, $x$, we can determine whether it is above the detection threshold, by first calculating the MAD of the data:

$$MAD = med_i(|x_i - med_i x_i|)$$  \hspace{1cm} (2.1)

where $med_i x_i$ is the median of all values $x_i$. This is done on a day-by-day basis for the continuous data. Then, we can determine whether a certain time point in the CC function, at time $i$, is a detection:

$$x_i = x_{\text{detection}} \text{ when } \frac{|x_i - med_i x_i|}{MAD} > \text{threshold}$$ \hspace{1cm} (2.2)

For the data in the ETSZ and in the MPSSZ, as was done before (Shelly et al., 2007; Peng and Zhao, 2009; Ross et al., 2019), I start analyzing the data by using a threshold of nine
times the MAD of the stacked CC sum to identify potential earthquake detections. For the study in the Piedmont Province, I analyzed data at a variety of thresholds: 9, 10, and 12 times the MAD. The probability of exceeding nine times the MAD for a normally distributed random variable is approximately $5.9e^{-10}$ (Shelly et al., 2007; Li and Zhan, 2018).

I impose a minimum of 5 seconds between detections, so that only the detection within a 5 second window with the highest value above the threshold will be saved as an output. Likewise, detections within a 5 second window that exceed the positive threshold are prioritized over detections that exceed the negative threshold. In the case that multiple templates detect the same event, the template that is associated with the detection is the one with the highest value above the threshold. The EQcorrscan algorithm returns the time of detection on the earliest channel of the template, while the matched filter technique developed by Meng et al. (2012) returns the estimated origin time of the newly detected event.

The arrival times of a detection need to be estimated, as only the time of a detection is returned by the algorithm. This detection time is set according to the earliest time in the template, which corresponds to the start time of the earliest trace in the template (earliest arrival time minus the 0.5 second pre-pick I cut the waveforms at), added to the shift required to move the template to the detection. I assign preliminary arrival times of the detection to be equal to the detection time added to the travel times of the template arrivals, as well as adding the 0.5 second pre-pick.
2.1.3 Deep Learning

Deep learning is a subfield of machine learning, formed with the goal of simulating the behavior of the human brain with the ability to “learn” from large datasets. Artificial or Convolutional Neural Networks (ANN, CNN) form the core of several recent deep-learning earthquake detection methods. Several competing recent deep-learning based earthquake tools are now commonly used to detect earthquake phases in continuous seismic data (Zhu et al., 2019; Kong et al., 2019; Mousavi et al., 2020).

Deep learning can provide robust earthquake detection and phase picking compared to traditional methods in seismology, with more sensitivity for smaller earthquakes and for events with lower SNR. A high-performing model can be attained using a large dataset of known earthquakes and arrival times (e.g., Mousavi et al., 2019). However, it’s not clear how these deep learning models react to data that differs to the datasets on which they were trained. Earthquakes from different datasets and study regions can exhibit different characteristics, so sometimes transfer learning is used to apply deep learning on a seismic region with smaller datasets, or with less history of seismicity (Jozinovi et al., 2022).

It is also challenging to adjust the input parameters on this class of algorithms, as it is seldom intuitive why some false detections are made, and why some earthquakes are not detected. Deep Learning is not yet used for real-time earthquake detection due to the novelty and uncertainty in reliability, but is becoming a more commonly used approach for earthquake detection in research mode (e.g., Tan et al., 2021).
2.2 New Procedures to Build High-Resolution Earthquake Catalogs

The input to matched filter detection consists of a catalog of known earthquakes, and arrival times of seismic phases at different stations. Since the waveforms are cut in a few-second window relative to the arrival times, it is necessary to have reasonable arrival times for each earthquake, within about 0.5 seconds or less of the true arrival times.

Likewise, the raw matched filter detections only have very rough, approximate arrival times. These arrival times are too imprecise to be used for relocation, or any method that requires a window relative to arrival times, such as magnitude estimation. There are several existing methods by which reliable, high-resolution seismic arrival times can be picked.

2.2.1 PSIRpicker

When an event has phase arrivals on either vertical or horizontal channels, but not both, it is likely that the data on the other component exists yet was not manually picked. It may be that the SNR was too low for the human analysts to pick the exact arrival times with confidence, or that they were looking at only one waveform at a time and did not have the context from waveforms at other stations that makes an arrival time (and an earthquake occurrence) more apparent. It is also possible that the arrival time was mistakenly not written into the catalog. In these cases, it is helpful to estimate the arrival times on missing channels, so that the signal of a seismic phase can be used for matched filter detection, even if the arrival time is not very clear or was initially not identified.

Typically, there are two ways to find phase arrival information given only an earthquake’s origin time, location, station locations, and waveforms. The first is to use a velocity model for that region to predict arrivals. The second is to detect sharp amplitude changes in the
signal, such as by the aforementioned STA/LTA method (Allen, 1982). However, the velocity model may have some inaccuracy and the STA/LTA method is sensitive to noise and other seismic phases.

PSIRpicer (Li and Peng, 2016) solves this problem without these disadvantages. Named for its “Predict-Search-Invert-Repeat” approach, PSIRpicer automatically picks the onset of P and S arrivals by first predicting initial picks using an input velocity model and using an SNR detector around the theoretical arrival times. A window length is needed to calculate the SNR function, and relates to a tradeoff between pick sharpness and stability. A longer window is typically used for the S arrival, because it tends to be noisier as there is coda from the P arrival.

Then the methods allow for a velocity model perturbation around a small time window using these initial picks. New arrivals are found using the inverted velocity model, repeating the process for a desired number of iterations. PSIRpicer is written in MATLAB and takes SAC data and a 1-dimensional velocity model as input.

**2.2.2 Refined Phase Picks Based on Cross Correlation**

In matched filter detection, the detected events have initial arrival times based solely on the template that detected the event. As a detection can occur several kilometers away from the template that detected it, some of the arrival time estimates can be incorrect (Zhang and Wen, 2015; Gao and Kao, 2020; Chamberlain et al., 2017). For this reason, I refine the estimated phase picks by shifting the template waveforms on top of the detection waveforms to a point where the waveforms are best aligned (Shelly et al., 2013).

The template waveforms are cross-correlated with the continuous data around the time of estimated detection, using a 2-second window around the estimated arrival time. A refined phase
arrival is defined as the time within this window of the highest normalized CC coefficient (Chamberlain et al., 2017). The arrival is removed from the detection if the CC value is less than 0. At a later time, if higher quality is desired, waveforms below a certain CC value are removed from the catalog of earthquake detections.

### 2.3 Earthquake Relocation

An earthquake’s hypocenter is characterized by three spatial coordinates: latitude, longitude, and depth, and one temporal coordinate, origin time. Earthquake location is a non-linear inverse problem wherein the hypocenters of earthquakes are estimated or refined. It is non-linear as the predicted travel times (between the earthquake and a station) are a function of distance and depth. Geiger (1912) introduced an iterative least-squares method for determining the location of an earthquake. The approach is to linearize the problem, by making small changes in the source coordinate point, and minimize the residuals between the predicted and observed arrival times as a least-squares problem.

Earthquake catalogs routinely contain location errors on the order of 1 km, or more, depending on station coverage and other factors (Bakun et al., 2011; Turquet et al., 2019). This uncertainty is larger than many fine-scale fault structures, which hinders the understanding of earthquake mechanics and spatio-temporal processes. By utilizing the timing differences between the arrivals of earthquakes recorded at stations in common, it is possible to determine more precise earthquake locations of a catalog. This process is known as relocation.

Earthquake relocation can result in precise locations of known seismicity, which can appear as clusters or linear features. When viewed on a map, relocations can delineate the
locations and orientations of earthquake-hosting faults which is vital knowledge for earthquake hazard mitigation.

I apply two separate relocation packages, XCORLOC (Lin, 2018) and hypoDD (Waldhauser, 2001), to relocate seismicity in the ETSZ and MPSSZ. This allows us to compare the relocation results and check for robustness of the solution within each study region. If a feature or pattern is present using both relocation methods, it is strong evidence of the feature truly existing.

Both relocation packages aim to improve relative earthquake location accuracy using both or either a catalog of phase arrivals and differential-time cross-correlation (CC) times between phases of different events (Waldhauser and Ellsworth, 2000). Earthquakes that have similar hypocenters typically have waveforms with high CC, so that accurate differential travel-time measurements for P and S waves can be obtained for event pairs at different stations. The differential travel times of P waves are measured on vertical components, while the differential travel times of S waves are measured on horizontal components.

For both methods, I compute CC based differential times between events. I use a 1.5-second-long waveform following each pick, cutting 0.2 seconds before the arrival time, and allow a pick to vary by up to 0.5 seconds when computing the CC between the waveforms of two events. I allow for sub-sample accuracy by applying a quadratic fit to the cross correlation. This allows for more accurate timing, and therefore locations, during relocation.

A minimum of 3 picks with a minimum cross-correlation of 0.6 between waveforms of different events is required. In the ETSZ, there are not many earthquake pairs within 5 km of one another and for event pairs greater than 25 km, there are not many pairs that meet the CC criteria. Therefore, I only use event pairs within 10 km of one another.
Both XCORLOC and hypoDD require an input regional velocity model. The model used for each method is shown in Figure S1. hypoDD requires a layer model, while XCORLOC requires a gradient model, so the model of Cameron et al. (2017), which is created for the region around the nearby Watts Bar Nuclear Power Plant, is adopted.

2.3.1 XCORLOC

A recent package developed to compute earthquake relocations is XCORLOC (Lin, 2018). XCORLOC can take both catalogued phase arrivals and cross-correlation based differential times as input to compute relocations. The initial estimates of the detections are gradually changed through an iterative inversion with a grid-search procedure using the hybrid L1-L2 norm as the misfit function (Lin, 2018).

The bootstrap resampling method which is built-in to this package estimates the uncertainties in relative hypocenter locations and origin times. This helps to better estimate the time delays between events caused by differences in the relative locations and times.

In the tests documented in Lin (2018), the relocation method of XCORLOC is less sensitive to outliers than hypoDD and may therefore results in more robust relocation estimates.

2.3.2 HypoDD

HypoDD is based on a double difference algorithm that reduces error by allowing un-modeled velocity variations. Double-difference residuals for earthquake pairs at each station are minimized by applying weighted least squares with the conjugate gradient method LSQR. HypoDD iteratively adjusts the vector differences between nearby pairs to solve for the relocations (Waldhauser, 2001).
HypoDD clusters seismicity based on the input parameters and returns the number of clusters, along with the number of earthquakes in each cluster, on the output screen. Damping is used when LSQR is the option chosen for inversion. The output from hypoDD gives a condition number (CND) for each iteration. The damping (DAMP parameter) should be modified in the input parameters based on the output when running each cluster, so that the CND returned for each iteration is ideally between 40 and 80 (Waldhauser, 2001). A small CND indicates that space might be rotated, but not distorted. A high CND is also not good, as it indicates space is distorted by a larger amount.

Due to the manual modification of the DAMP parameter, each cluster must be relocated individually. The DAMP parameters will depend on the number earthquakes in a cluster, as it tends to be larger for clusters with more earthquakes. If the damping is too high, the relocated earthquake hypocenters may not move, so this step is important.

2.4 Magnitude Calculation

Previous studies often use peak amplitude ratios to determine the magnitudes of newly detected events (e.g., Peng and Zhao, 2009). However, peak amplitude ratios are generally noisy, so I apply a recent method by Shelly et al., 2016 which avoids this known issue.

2.4.1 Principal Component Fitting to Estimate Relative Amplitudes

I apply the method of Shelly et al. (2016) to calculate the magnitudes of the newly detected earthquakes, which uses principal-component fitting between a template and a newly detected event to estimate relative amplitudes. For most detections, the magnitude of the detection is within one to two orders of magnitude of the template that detects it.
The method works as follows: for a template-detection pair, I calculate the amplitude ratio ($\alpha_i$) between each pair of waveforms across each station-channel i. Specifically, $\alpha_i$ is the ratio of the elements $v(2)$ and $v(1)$, which are the eigenvectors corresponding to the largest eigenvalue of the covariance matrix of the template and detection waveforms. I impose a positivity constraint on the amplitude ratio. The median value of all values of $\alpha_i$ for a template-detection pair is used to calculate the magnitude of the newly detected event:

$$M_{\text{new}} = M_{\text{template}} + c \log_{10}(\text{median}(\alpha_i))$$ \hspace{1cm} 2.3

The value of $M_{\text{template}}$ is the catalogued duration or local magnitude of the template earthquake. Most of the catalogued earthquakes only have a duration or local magnitude recorded, so I use Equation 2.3 to calculate the magnitude of the newly detected earthquakes. The value of $c$ describes the amplitude-magnitude scaling. I use a standard value of $c=1$ as the scaling factor (Shelly et al., 2016). I ensure that calculated magnitudes are reliable by requiring waveforms of detections to have a minimum CC of a certain threshold with waveforms of the template that detected the event on a minimum of a certain number of waveforms.
CHAPTER 3. The 15 February 2014 Mw 4.1 South Carolina Earthquake Sequence: Aftershock Productivity, Hypocentral Depths, and Stress Drops

In this study, I examine one of the largest recent earthquakes that occurred in the Piedmont Province, the 2014 Mw 4.1 Edgefield, South Carolina earthquake. I used matched filter detection to detect previously uncatalogued seismicity in the 2 weeks around this earthquake and found that this mainshock had an aftershock deficiency, according to Bath’s law. The multi-window coda spectral ratio method is used to estimate the corner frequencies for the Mw 4.1 mainshock and its only catalogued Mw 3.0 aftershock, and the stress drop for each event is calculated assuming a circular crack model and Brune’s model parameters. The resulting stress drops for these two earthquakes are 3.75 and 4.44 MPa, respectively. Hypocentral depths of the Mw 4.1 mainshock and Mw 3.0 aftershock are estimated by examining the differential time between a depth phase called sPL and P-wave arrivals, as well as by modeling the depth phase of body waves at shorter periods. The best-fitting depths for both events are around 3–4 km. The results in this chapter have been published in a peer-reviewed journal, Seismological Research Letters (SRL), as a research article (Daniels et al., 2020).

3.1 Introduction

The 15 February 2014 Mw 4.1 South Carolina Earthquake occurred in the Piedmont Province of the Southeast USA. Nicknamed the “Valentine’s Day Earthquake”, it was widely felt 100 miles away (U.S. Geological Survey, 2017; USGS “Did you feel it?” Intensity Report at https://earthquake.usgs.gov/earthquakes/eventpage/se610610/dyfi/intensity) and is one of the largest magnitudes to have occurred in the Southern Piedmont Province in decades.
Seismicity in the Piedmont region of Georgia and South Carolina is characterized by shallow depths of less than 5 km, small clusters, and local magnitudes typically less than 4 (e.g., Long, 2009). The major epicentral zones are near Lake Sinclair in central Georgia, the J. Strom Thurmond Reservoir at the border between Georgia and South Carolina, and reservoir induced seismicity near the Lake Jocassee and Monticello Reservoir areas (Bollinger et al., 1991). Most earthquakes are single isolated events except for concentrations of epicenters in the vicinity of reservoirs (Long, 2009) or major rivers (Costain, 2008).

On February 15, 2014 at 3:23:38 GMT, an earthquake occurred near Edgefield, South Carolina (Figure 3.1). The United States Geological Survey (USGS)’s National Earthquake Information Center (NEIC) catalog listed the moment magnitude (Mw) as 4.1. The epicenter (33.817°N, 82.092°W with a precision of 0.6 km, according to the USGS) was close to the state border of Georgia and South Carolina and was widely felt in both states. According to the USGS’s “Did you feel it?” webpage, vibrations were felt to a distance of approximately 300 km with a maximum intensity of V (Moderate) on the Modified Mercalli Intensity Scale.
Figure 3.1 Map of study region around the 2014 Mw 4.1 mainshock on 15 February 2014 at 3:23:38 GMT. Solid triangles indicate the 9 stations used to perform final detections for missing events from the ANSS catalog. Empty triangles indicate available Z9 stations not used in the analysis. Diamonds correspond to background events with magnitude greater than or equal to 3 since 1970 that occurred prior to the mainshock, whereas circles correspond to those which occurred following the mainshock. The focal mechanisms for the Mw 4.1 mainshock and Mw
3.0 aftershock are also plotted. The location of the 1974 ML 4.3 earthquake is also indicated. The Modoc Fault Zone is shown as the dashed line.

The epicenter is about 20 km east of the Thurmond Reservoir (known as the Clarks Hill Reservoir in Georgia), and is about 30 kilometers northwest of Augusta, Georgia. The Mw 4.1 mainshock was followed by a Mw 3.0 aftershock on February 16, at 20:23:35 GMT. However, no additional earthquakes were reported around the time of the mainshock. The USGS’s NEIC catalog listed the depth of the Mw 4.1 mainshock as 5.18 ± 0.8 km and the depth of the Mw 3.0 aftershock as 6.99 ± 1 km. Other reported solutions include a depth of 5 km by St. Louis University and the USGS’s moment tensor (MT) depth of 6 km for the mainshock. The USGS moment tensor solution can be found at https://earthquake.usgs.gov/earthquakes/eventpage/se610610#moment-tensor. This sequence was recorded by several stations in the region, the closest of which is 25.4 km from the epicenter. Permanent stations within 48 km include CO HODGE and CO HAW. There was also a temporary seismic network named SESAME (network code: Z9) (Fischer et al., 2010). The nearby Modoc Fault Zone is also shown in Figure 3.1.

Previous studies on earthquakes in the Piedmont Province of Georgia and South Carolina suggest that earthquakes near the Thurmond Reservoir occur on smooth, lubricated faults with low normal stress and multiple plane orientations, which may correspond to different joints in the rock (Johnston, 1980; Guinn, 1977; Marion and Long, 1980). The last earthquake with magnitude greater than 4 near this reservoir is the August 2, 1974 ML 4.3 earthquake, and the corresponding stress drop is 1.2 to 12 bars (or 0.12 to 1.2 MPa), depending on which method is used (Bridges, 1975). This event occurred about 37 km from the 2014 mainshock and was
located on the north side of the Thurmond Reservoir (Figure 3.1). Furthermore, the large b value in the Gutenberg-Richter relation for earthquakes in this region and other reservoirs suggests that many smaller magnitude earthquakes are triggered due to effects of reservoir impoundment (Gupta et al., 1972). The swarm behavior of these earthquakes can likely be explained through the triggering mechanism of decreased fracture or joint strength in the rock due to weathering or increased water pressure in a joint (Long, 2009). In particular, an asperity may fail by changes in hydrostatic pressure from a nearby reservoir, or lithostatic pressure and fault movement. When one asperity fails, fluids flow away from the failed joint and cause reduced strength in surrounding joints and fractures through increased fluid pressure (Long, 2019).

A plot of the background seismicity in the Thurmond Reservoir region (defined as events within 100 km of the epicenter of the Mw 4.1 mainshock) from the ANSS catalog is given in Figure 3.2. From 2010 to 2015 there are few events in general, with 2010 being an unusually quiet year and 2013 a comparatively seismically active year. The cluster of events in location and time from April 7 to April 27, 2013 may correspond to an earthquake swarm located 25-40 km west of mainshock. However, there are no catalogued events since 2010 within ten kilometers of the mainshock epicenter, as well as within several months following the mainshock, aside from the Mw 3.0 aftershock. The two years (2015 and 2016) following the Mw 4.1 event are relatively quiet compared to the background levels of seismicity. Among them, no events were reported in the region in 2016. It is unlikely that these detections are quarry blasts, as their times of day are mainly in the very early hours of the day or randomly throughout the day, with no peak near 8 am, 12 pm, or 5 pm, which are typical for quarry blasts in this region (T. Long, personal communication, 04/2019).
Figure 3.2 Background events from January 1, 2010 to August 1, 2017 in the region within approximately 100 km of the Mw 4.1 mainshock epicenter. The plot illustrates the magnitude and distance from the mainshock epicenter plotted against year. Solid points represent ANSS events used as templates, whereas unfilled points correspond to other recorded ANSS events.

Figure 3.1 marks nearby seismic stations in the study region, including those in the temporary SESAME experiment Z9 (Parker et al., 2013; Parker et al., 2016; Hopper et al., 2017).
A spectrogram from a nearby station Z9.D03 is shown in Figure 3.3. The spectrogram indicates that additional earthquakes temporally associated with the main event are not listed in the USGS catalog, motivating us to detect them with the matched filter technique. In this study, continuous waveforms from nine stations within 100 km of the epicenter are used, including Y55, D03, D04, D05, D06, HODGE, Y54, D07, and D08. Stations farther in distance from the mainshock epicenter have relatively lower signal to noise ratios (SNRs), especially for small-magnitude events.

Figure 3.3 (a) Raw vertical-component seismogram recorded at station Z9.D03 showing the 2014 Mw 4.1 mainshock. The bold dashed line denotes the origin time of the mainshock. (b) Log10-
based 2-16 Hz band-pass-filtered envelope function showing possible foreshocks and aftershocks. Those marked by dashed lines with magnitudes marked are events detected by the template matching method. (c) The corresponding spectrogram.

In this study, I conduct a systematic investigation of the source properties (i.e., hypocentral depths, stress drops, etc) of the Mw 4.1 mainshock and Mw 3.0 aftershock. To determine whether there are missing events around the mainshock, I apply a matched filter technique (Peng and Zhao, 2009) to scan through continuous waveforms to detect additional events covering the time period of February 8 to 22, 2014. More accurate depths of these two events are estimated by examining the differential time between a particular depth phase called sPL (Chong et al., 2010) and P wave arrivals. The stress drop for the Mw 4.1 mainshock and its Mw 3.0 aftershock are determined by the multi-window coda spectral ratio (MWCSR) method (Wu and Chapman, 2017). An updated seismic moment of the aftershock is calculated by computing the moment difference relative to the mainshock.

3.2 Matched Filter Detection

I download seismic data from the IRIS Data Management Center. The station distribution of the Z9 seismic network can be found at https://ds.iris.edu/gmap/#network=Z9&starttime=2010-01-01&endtime=2014-12-31&planet=earth. A 2-16 Hz band-pass filter is applied to the continuous three-component broadband seismogram data between 8 and 22 February 2014, as this filter provides the clearest signals for local earthquakes recorded at most stations. I also resample the continuous data to 40
sample/s to speed up the matched filter computation, as well as keeping enough high frequencies to distinguish events at short distances.

The two template events used first are the Mw 4.1 mainshock and Mw 3.0 aftershock listed in the ANSS catalog. Template waveforms are cut from the continuous waveforms starting from 30 seconds before to two minutes after the event. P and S-wave arrival times are manually picked by visual inspection for each waveform.

The matched filter method only uses template waveforms that have 12 or more channels with SNR greater than 5. To compute the SNR, the signal window spans 6s (1s before to 5s after) around the P and S arrivals. The noise window contains the 6s of signal ending 1s before the P wave arrival. The sliding-window cross-correlation (CC) function is computed by shifting every 0.025 s (i.e., one sample per step). The time window for computing the CC function is set to be 1 second before to 5 seconds after the P arrival time, and 1 second before to 5 seconds after the S arrival time, which should provide long enough waveform segments to capture the majority of the targeted P and S wave. For vertical channels, the P-wave time window is used, whereas for the two horizontal channels the S-wave time window is used.

Individual CC functions of all channels for each day are then stacked directly, resulting in one mean trace for each continuous day. The initial threshold value for a positive detection is equal to the median plus nine times the median absolute deviation (MAD) of the mean CC trace (Shelly et al., 2007).

To avoid over-counting events, duplicate detections are removed corresponding to multiple detections of the same target event with different template events (e.g., Meng et al., 2013). If there are multiple detections within two seconds of one another, only the detection with the highest CC value is kept. To obtain an approximation of the magnitude of the detected
events, the median peak amplitude ratio between each detected event and its best-matched template across all channels is calculated (Peng and Zhao, 2009).

Figure 3.4 shows an example of a newly detected magnitude 0.41 aftershock based on the matched filter method. The template event is the Mw 3.0 aftershock. The mean CC value for this detection is 0.427, much higher than the detection threshold of approximately 0.2. By visual inspection, there does indeed appear to be an event at the detected time. Stations HAW through D12 do not show the event as clearly as stations Y55A through D08, so it is more difficult to manually pick the arrival of the event for stations after D08.
Figure 3.4 Example of a detected M0.41 aftershock using the M3.0 aftershock as a template. (a) Mean cross-correlation (CC) values around the time of the detected events. The dashed line marks the detection threshold of 9 times the median absolute deviation. (b) A comparison between the 2-16 Hz continuous waveforms and the template waveforms. The continuous waveform is cut near the origin time of the detected event, which is represented by a vertical
dashed line. Station names are given in the first column on the right hand side of the waveforms. Cross-correlation values for each station are given in the second column, and approximate distances in kilometers from the detected event are given in the third column. The horizontal dashed line separates the stations used for stacking (above) from the stations not used (below).

The arrow in Figure 3.4 indicates the horizontal line that separates the 9 used stations (above the line), from those unused (below the line). For the nine stations included in this study, the event can be seen clearly and the CC values are high. For stations not included in this study, the CC values are generally low and the event is not picked up well at these stations with distances greater than 65 km from the mainshock.

As relatively fewer events are detected with the first two templates, the detections are used in turn as new templates. The same matched filter method is used with the new templates. In summary, there are 6 new events found using the first two templates (the Mw 4.1 mainshock and the Mw 3.0 aftershock), as well as 13 more found using the detections as templates, for a total of 19 detections (Table S3.1). Figure 3.5 gives the mean CC values of all detections by the templates at the top of each plot, without removing multiple detections within a two second window.
Figure 3.5 Mean cross-correlation (CC) values of all detections by the template marked on the top of each plot (20140215100556 Mw 4.1, 20140217103127, Mw3.0). From (a) to (g), the date
of each panel ranges between 12 February and 18 February 2014. The threshold of 9 times the median average deviation (MAD) added to the median cross-correlation waveform value is depicted by a dashed line on each plot. The vertical dashed line in panel (d) indicates the time of the mainshock on 15 February 2014.

Figure 3.6a shows that seven of the eight events before the mainshock are considered as positive ones only if the threshold is between 9 and 12 times the MAD. This suggests that most of the detections would not be considered as true events if a higher threshold were chosen. Hence, it is debatable whether their waveforms contain true events. The only M 0.68 foreshock with a relatively high mean CC value (0.311, or 19.78 times the MAD) occurred about 11 hours before the mainshock.

Using the 9 times MAD threshold, there are also new 11 aftershock detections. Four of these events occurred after the Mw 3.0 event (and hence could be considered as its secondary aftershocks), and most of the aftershocks occur on the same day as the mainshock. As shown in Figure 3.6a, several aftershocks are persistent even with the highest threshold of 15 times the MAD, suggesting that they are likely true events. Two of the aftershocks of the Mw 3.0 event have among the highest cross-correlation coefficients among all detections despite their low magnitudes.

Figure 3.6b plots the magnitude-frequency distributions for all detected events. If we exclude the Mw 4.1 mainshock and compute the Gutenberg-Richter (G-R) statistics for all detected earthquakes in the sequence around the epicentral region, the resulting a-value is 1.2 (based on the maximum–likelihood estimate). The a-value is the log-based-10 of the expected number of events with magnitude > 0 in a given time interval and area, and hence reflects the
productivity of the earthquake. Cumulatively, there are only 20 events with magnitudes larger than 0 other than the mainshock during the study period.

Figure 3.6 Distribution of two catalog events and newly detected events. (a) Magnitude of the events is plotted against time relative to the mainshock, in days. The gray scale of the data points corresponds to cross-correlation (CC) values of the events. The two template events are thus colored black, and the resulting detections have varying CC values. Detections are categorized
based on the threshold level for detection used. Circular data points represent events that are
detected at the highest threshold, 15 times the MAD added to the median. Triangular data points,
of which there are none, represent events detected at a threshold between the range of 12 to 15
times MAD plus the median. Square data points represent events detected at the lowest threshold
in the range of 9 to 12 times MAD plus the median. (b) Cumulative number of events with a
magnitude above a certain level, for all events detected excluding the mainshock. The dotted line
represents the best-fitting Gutenberg-Richter relationship between expected earthquake
magnitude and cumulative number of events. The star represents the magnitude 4.1 mainshock in
comparison to other events. The line fit represents an “a” value of , and a “b” value of 0.47.

3.3 Stress Drop Estimates

Next we apply the multi-window coda spectral ratio (MWCSR) method (Wu and Chapman,
2017) to obtain corner frequency and stress drop estimates for the Mw 4.1 2014 mainshock and its
Mw 3.0 aftershock. This method takes advantages of the multi-window spectral ratio method
(Imanishi and Ellsworth, 2006) and the averaging property of the S-wave coda (Mayeda et al.,
2007; Somei et al., 2014; Frankel, 2015; Wu et al., 2016), and has shown to produce more stable
spectral ratios compared to conventional empirical Green's function (EGF) or spectral ratio
methods (Wu and Chapman, 2017; Wu et al., 2018). The Mw 3.0 aftershock is considered as the
EGF of the mainshock and the procedures of Wu and Chapman (2017) are followed to process the
data.

The chosen coda window begins at twice the S-wave travel time and the entire coda
window length is set to 30 s. The entire coda window is then divided into five successive 10s long
sub-windows overlapped by half the sub-window duration (Figure 3.7a). Qualified traces from all
stations within 250 km are selected by examining the common decay characteristics of coda envelopes in the entire analyzed coda window over a broad frequency range (1-2, 2-4, 4-8, and 8-16 Hz) for this mainshock-aftershock pair (Figure 3.7b). The mean and standard deviation of the normalized envelope amplitude ratio (smoothed using a 10-s moving window) are required to be within 1.0±0.1 and less than 0.05, respectively, and only traces that meet the criteria are accepted. This selection scheme ensures that the coda waves decay in a nearly identical manner for the event pair, indicating very similar propagation and site effects. Therefore, taking the spectral ratio of qualified traces would remove the common propagation and site effects effectively, leaving an estimate of the source spectral ratio which can then be modeled with an assumed source spectral model (e.g., Brune, 1970, 1971; Boatwright, 1980).
Figure 3.7 The mainshock and the aftershock recorded at station Y54A at an epicentral distance of 55 km. (a) The transverse-component velocity of the two events. The noise, direct S wave and coda windows are denoted by shaded rectangles from left to right. The short horizontal bars mark the coda subwindows. (b) Comparison of coda decay rates for the two events. Left: narrow-band original (black and gray) and smoothed envelopes of the seismograms shown in (a). Right: the normalized coda amplitude ratio across the entire coda window.
The spectra of the pre-P noise window, the direct S-wave window, and the coda sub-windows are calculated using a multitaper approach (Prieto et al., 2009). First, the coda spectral ratios from the five sub-windows are stacked, and then the spectral ratios for all qualified traces are stacked. Note that the spectral ratios are resampled equally in the logarithmic domain to ensure equal weights at low and high frequencies, and the SNR threshold is set at 3. For comparison, the spectral ratios for the direct S-wave window are also calculated (Figure 3.8a, left panel).

Figure 3.8 The spectral ratios and corner frequency estimates. (a) individual (dot) and stacked (solid line) spectral ratios for direct S-wave windows (left) and multiple coda windows (right) measurement at all stations within 250 km; (b) The best-fitting Brune model (solid line). The dashed lines show the best estimates of model parameters (\(f_{c1}, f_{c2}, M_{ratio}\)). The shaded areas represent the 95% confidence intervals of the estimates; (c) The posterior distributions of the
model parameters. The solid lines mark the best estimates, and the vertical dashed lines outline the 95% confidence intervals. The diagonal panels show the posterior probability density functions of Mratio (top), fc2 (middle) and fc1 (bottom). The off-diagonal panels show the two-dimensional projections of the posterior samples (gray dots) of the MCMC simulations. The contours represent 0.5, 1, 1.5 and 2 standard deviations.

It is obvious that the spectral ratios derived from the multi-window coda waves show less scatter across stations when compared to those from the direct S waves (Figure 3.8a). Subsequently, the stacked spectral ratios are modeled with the Brune source model (Brune, 1970, 1971),

\[
\frac{u_1(f)}{u_2(f)} = \frac{M_{01} \left(1 + \frac{f}{fc_2}\right)^2}{M_{02} \left(1 + \frac{f}{fc_1}\right)^2} = Mratio \frac{\left(1 + \frac{f}{fc_2}\right)^2}{\left(1 + \frac{f}{fc_1}\right)^2}
\]

4.1

where \( Mratio \) is the ratio of the seismic moment of the two events, fc1 and fc2 are the corner frequencies of the mainshock and the aftershock, respectively. The two corner frequencies and \( Mratio \) are found simultaneously using a Markov chain Monte Carlo (MCMC) algorithm (Salvatier et al., 2016). The best estimates and uncertainties (95% confidence intervals) are determined from the posterior probability distributions (Figure 3.8c). The obtained corner frequencies for the mainshock and the aftershock are 2.18 Hz (1.86-2.55 Hz, 95% confidence interval) and 9.19 Hz (8.01-10.55 Hz, 95% confidence interval).
The stress drop is then calculated from seismic moment $M_0$ and corner frequency $f_c$ assuming a circular crack model and Brune's model parameters (Eshelby, 1957; Brune, 1970, 1971),

$$\Delta \sigma = \frac{7}{16} \frac{M_0}{r^3} = \frac{7M_0}{16} \left( \frac{f_c}{\kappa \beta} \right)^3$$  \hspace{1cm} (4.2)

$$r = \frac{\kappa \beta}{f_c}$$ \hspace{1cm} (4.3)

where $r$ is the source radius, $\beta$ is the shear-wave velocity at the source, which is assumed as 3.5 km/s, and $\kappa$ is a constant corresponding to the specific source model (0.372 for the Brune model).

The seismic moment of the mainshock is taken from SLU Earthquake Center regional moment tensor solutions, while that of the aftershock is calibrated through the fitted $M_{ratio}$, as the moment tensor inversion tends to have large uncertainties for small earthquakes. The SLU solution for the Mw 4.1 mainshock and Mw 3.0 aftershock can be accessed from the following sources:


The updated Mw for the aftershock is $2.91 \pm 0.05$, slightly lower than that from the moment tensor solution (Mw 3.02). The stress drop of the mainshock is 3.75 MPa with a 95% confidence interval of 2.36-6.01 MPa, which is low in the context of eastern North America (ENA), where moderate-sized earthquakes generally have high stress drops with typical values of ~10 MPa (e.g., Shi et al., 1998; Viegas et al., 2010; Viegas et al., 2012; Boatwright and Seekins, 2011; Wu and Chapman, 2017, and references therein). For the aftershock, this approach yields a stress drop of 4.44 MPa (2.95-6.73 MPa) (6.54 MPa 4.34-9.90 MPa if Mw 3.02 is applied).
3.4. Constraints on Depth

The focal plane solution of the Mw 4.1 mainshock is resolved via inversion of full waveforms filtered between 0.02-0.1 Hz by the SLU earthquake center. The inversion reported a focal depth of 5 km (SLU solution) and 6 km (USGS-MT). To further constrain the depth, the depth phases of body waves up to 2 Hz are modeled by comparing the observed and synthetic seismograms. Three component raw seismograms of the mainshock recorded at station Y55 are converted to ground velocity waveforms, using the transfer function in SAC with four corner frequencies of 0.001-0.003 Hz and 10-20 Hz. Then the ground velocity waveforms are low pass-filtered with a corner frequency of 2 Hz (number of poles 2, and number of passes 2). The CUS crustal model (Herrmann, 2013) and a modified version (M2) (Figure 3.9a) are used in computing synthetic seismograms with the frequency-wavenumber (FK) code of Zhu and Rivera (2002), and the SLU focal plane solutions are adopted in the calculation.

Figure 3.9 (a) Vp and Vs models. CUS model (Herrmann, 2013) is indicated with gray thick lines, and a modified model (M2) with thin dark lines. (b) Ray path for free surface P wave (bottom), and sPL (top). sPL forms from near surface coupling effects, including free surface P wave, sPhP, sPhPPPhP and later multiples.
The only difference between the CUS model and M2 is in the shallowest 1 km, which contains a more gradual velocity increase in M2. Synthetics with the CUS and M2 models both match the observations well. However, the M2 model is preferred, as the synthetic waveform around 2 seconds after the P arrival better fits the observed seismogram of the mainshock at station TA.Y55 (Figure 3.10). The negative and positive swing around 2 seconds after the P arrival is closely aligned for both the observed waveform and synthetic M2 model. In comparison, the observed waveform and synthetic CUS model do not match as well at two seconds after the P arrival, probably due to the sharp discontinuity modeled at a depth of 1.0 km. Therefore, the M2 model is adopted.
Figure 3.10 Comparison between observed (black) and synthetic (gray) waveforms of radial component for station Y55. The focal depth is indicated below each trace. Waveforms are normalized in amplitude.

The phase near 2.0 sec is probably the sPL phase as proposed by Chong et al. (2010), which is related to the upgoing S wave and converted P waves in the surficial layer (Figure 3.9b). In the case of a homogeneous half space model, the upgoing S wave converts to a horizontally
propagating P wave when the S to P reflection approaches the critical angle, and becomes the free surface P wave (Aki and Richard, 2002). However, for the case of a lower velocity layer above a half space, the conversion of upgoing S wave leads to sPhP (similar to the SsPmP, Langston, 1996, as well as its multiples such as sPhPPhP). That is why Chong et al (2010) named the wave trains of free surface P wave as sPhP and later multiples as sPL, due to the complicated coupling nature. This phase arrives between the P and S wave (Figure 3.11a), and is usually stronger on the radial component than the vertical component because of the nearly flat ray path of the P segment (Figure 3.9b).

Figure 3.11 (a) Observed Radial, Tangential and Vertical component of waveforms at Y55 station. (b) comparison between observed (black) and synthetic (gray) radial component for different depths. Low pass filter is used with a corner frequency of 2 Hz, number of poles 2 and number of passes 2.
The differential time between sPL and P waves is sensitive to focal depth. In Figure 3.11, synthetics with focal depths of 3-5 km are compared with the observed radial component at station Y55. The timing difference between sPL and P is too large at a depth of 5 km, and too small at a depth of 3 km. Instead, for a depth of 4 km, the observed and synthetic waveforms have the best match. Moreover, the synthetic Rayleigh wave at a depth of 3 km seems to be stronger than in the observed waveform, and weaker at a depth of 5 km, again suggesting that a depth of 4 km is more probable. Using the Shen-Ritzwoller profile (Shen et al., 2016), the observed and synthetic waveforms have the best match at a depth of 3 to 4 km. The focal depth of 5 to 6 km from SLU and USGSMT likely resulted from using longer period waves or stations at longer distances, where lateral variation of crustal structure may bias estimation of focal depth. For the Mw 3.0 aftershock (also low pass-filtered with corner frequency of 2 Hz), the sPL arrival is also observable at station Y55, but less obvious due to lower SNR. The differential time between the sPL and P arrivals for the aftershock is close to that of the mainshock, suggesting a similar focal depth.

3.5 Discussion

I detected foreshocks and aftershocks around the 15 February 2014 Mw 4.1 earthquake near Edgefield, SC. Starting with only two catalogued events (the Mw 4.1 mainshock and the Mw 3.0 aftershock) and iterating once with detections as new templates, the matched filter technique yields 19 possible other events (Table S3.1). These include at least one previously undetected M 0.68 foreshock, as well as several aftershocks including the M 0.41 event (Figure 3.4). The remaining 6 events with a threshold of 15 times the MAD can be easily identified on seismograms.
Approximately 95% of the detections have magnitudes less than 1. There is only 1 of the 19 detections with a magnitude between 1 and 3, with a magnitude of 1.92 (Figure 3.6a). According to the Gutenberg-Richter and the Bath’s Laws (Shcherbakov, 2004), the Mw 4.1 event is supposed to be followed by one magnitude larger than 3, and about 100 events with magnitude larger than 1 (with the b value of 1). Since the matched filter method detected many events with magnitudes smaller than 1, it is reasonable to assume that if there were additional events with magnitudes greater than 1, they should have been detected. It thus seems that there are indeed very few events with magnitudes between 1 and 3 in this sequence, which is unusual for an event of this size at plate boundary regions. However, this is not the only case when a Mw 4.0 type earthquake in the Eastern United States has been followed by very few aftershocks. The 2012 Mw 4.0 Waterboro, Maine earthquake is another example. Despite a dense temporary array being deployed in the days following the mainshock, only one aftershock was detected (Quiros et al., 2015). The 2017 Mw 4.2 Delaware earthquake is another case where there is a lack of larger magnitude aftershocks, with the only aftershocks having magnitudes less than Mw 2.0 (Kim et al., 2018). Similarly, the 2003 Mw 4.3 double earthquake in the central Virginia seismic zone had no aftershocks detected, despite a month of aftershock monitoring by a temporary local network (Kim and Chapman, 2005).

The relative lack of background seismicity occurring near the mainshock and the relatively low number of aftershocks (and the associated low a-value in the G-R fit) suggest that in general earthquakes are difficult to nucleate in these regions, even during the aftershock time period. However, the nearby 1974 ML 4.3 Clarks Hill Reservoir earthquake had prolific aftershocks, with several magnitude greater than 3, including the October 8, 1974 ML 3 and
December 3, 1974 ML 3.6 earthquakes, as well as many aftershocks with magnitudes less than 2 (Talwani, 1976).

An alternative explanation is that most of the accumulated strain was released by the mainshock. Another interesting feature of this sequence is the low calculated stress drops for the mainshock and its Mw 3.0 aftershock. Stress drop values of 3.75 MPa for the Mw 4.1 earthquake and 4.44 MPa for the Mw 3.0 earthquake are less than expected for tectonic earthquakes in the Central and Eastern US, which are typically on the order of ~10 MPa (e.g., Shi et al., 1998; Viegas et al., 2010; Boatwright and Seekins, 2011; Wu and Chapman, 2017; Huang et al., 2017; Boyd et al., 2017). The lower observed stress drops may be in part due to the depth of the earthquakes. The mainshock and its Mw 3.0 aftershock likely occurred at a depth shallower than initially estimated. Therefore, the energy released in the mainshock may be lower than previously thought. There is an observed increase in minimum stress drop with depth for eastern U.S. earthquakes, with the tendency for shallow earthquakes to exhibit lower stress drops (Long, 2019). The shallower depths may also account for the apparent deficit of aftershocks and triggered events from this earthquake. The ML 4.3 1974 Clarks Hill Reservoir earthquake had a similarly shallow depth of 1.5 km, and a relatively low stress drop of 0.12 to 1.2 MPa (Bridges 1975).

Historical records suggest on average a Mw 4.5 earthquake every 50 years in the region of the Thurmond Reservoir, with a depth limit of 4 km due to hydrostatic loading (Long, 2009). The 2014 Mw 4.1 earthquake, situated in the region near the reservoir, supports these estimates. Its proximity to the Thurmond Reservoir, like several other earthquakes in the region, indicates there may be some causal effect from the Thurmond Reservoir. The 2014 Mw 4.1 earthquake may also have ruptured the nearby Modoc Fault (Chowns, 1976), but whether the earthquake is
related to this fault is unknown. However, further analysis of the regional stress field, geologic structures, reservoir induced stress changes, and long-term earthquake behavior are needed to more definitively determine whether this seismicity is related to the reservoir.

3.6 Conclusions

The Mw 4.1 earthquake on February 15, 2014 near Edgefield, South Carolina was found to have an aftershock deficiency, suggesting that most of the strain was released during the mainshock. The observed stress drops of 3.75 and 4.44 MPa for the Mw 4.1 mainshock and Mw 3.0 aftershock, respectively, are less than expected for earthquakes in the Central and Eastern US, which are typically on the order of ~10 MPa. The lower observed stress drops may be in part due to the depth of the earthquakes. The mainshock and its Mw 3.0 aftershock likely occurred at a depth shallower than initially estimated, at 3 to 4 km. Therefore, the energy released in the mainshock may be lower than previously thought. The shallow depths and low stress drops are in line with the tendency of shallow earthquakes in the Eastern United States to exhibit lower stress drops. The shallower depths may also account for the apparent deficit of aftershocks and triggered events from this earthquake.
CHAPTER 4. Fault Orientation and Relocated Seismicity Associated with the December 12, 2018 Mw 4.4 Decatur, Tennessee Earthquake Sequence

In this study, I examine the recent 2018 Mw 4.4 earthquake in the ETSZ, which is one of the largest known earthquakes in this seismic zone. I use matched filter detection in the 8 weeks around the mainshock to detect missing earthquakes during this time. Earthquakes are relocated using two separate relocation packages, with cross-correlation based differential times, suggesting an E/NE-trending fault plane. I construct focal mechanism solutions for earthquakes near the mainshock using HASHpy and refine the focal mechanism solution and depth of the Mw 4.4 mainshock. I compare my solution for the mainshock depth of 5 to 6 km with my solution for the depth to the basement rock of 4.4 km, which is found using the S to P converted phase. Lastly, I perform rupture directivity analysis of the mainshock using the inversion of amplitude spectra from P-wave seismograms to estimate the apparent duration at regional stations, which reveals an approximately E-W bilateral rupture.

The results in this chapter have been presented at several recent conferences, including the 2021 Eastern Section of the Seismological Society of America (SSA) meeting (Daniels et al., 2021a) and the 2021 Fall American Geophysical Union Annual Meeting (Daniels et al., 2021b). The manuscript has also been accepted to the Eastern Section column of Seismological Research Letters (SRL) in May 2022 (Daniels et al., 2022).

4.1 Introduction

The December 12, 2018 Mw 4.4 Decatur, TN is one of the largest magnitude earthquakes to have occurred in the Eastern Tennessee Seismic Zone (ETSZ) in modern times.
The ETSZ is the second most active seismic zone in the eastern United States, after the New Madrid Seismic Zone (NMSZ) in the Mississippi Valley area, yet its origin is still not well understood (Powell et al., 1994). The ETSZ is defined by a roughly 300 km long, 100 km wide region of diffuse seismicity (Figure 4.1) that trends approximately 45 degrees from northeastern Alabama, north Georgia, northeastern Tennessee and western part of North Carolina (Powell et al., 1994; Brandmayr and Vlahovic, 2016; Hatcher et al., 2012; Chapman et al., 1997).

Figure 4.1 (a) Map showing topography and geologic provinces of study region and earthquakes in the ComCat catalog between January 1st 2005 and July 31st 2020. Each earthquake is plotted by a black open circle. The rotated rectangle shows the approximately boundary for earthquakes in the ETSZ. Only catalogued earthquakes within this region are input for template creation. The red star shows the epicenter of the December 12th, 2018 Mw 4.4 Decatur, TN earthquake. Triangles show the locations of stations operational within the 8 week study period and vicinity of the ETSZ. White box around mainshock indicates study region of interest near mainshock, in Figures 4.6 and 4.7. (b) Blue dots show catalogued seismicity between 1970 and August 1, 2020.
above magnitude 2. Focal mechanisms are taken from the USGS ComCat catalog between 1970 and August 1, 2020, as well as from Chapman et al., 1997. The 2003 M4.6 and 2018 M4.4 earthquakes are depicted by the red focal mechanism solutions. The red star denotes the location of the M4.7 1973 earthquake. The arrows depict the orientation of the maximum regional compressive horizontal stress, at N50°E (Davison, 1988).

Most earthquake epicenters occur in eastern Tennessee, which is characterized by highly tilted and faulted rock layers. The Valley and Ridge Province exhibits the highest level of seismic activity, although a few earthquakes also occur in a portion of the Blue Ridge and Appalachian Plateau Provinces.

The Valley and Ridge Province is underlain by basement rock which dates to the formation of the supercontinent Rodinia and the Grenville orogeny. After the breakup of Rodinia, rifting did not affect the basement rocks in the ETSZ and as such, the ETSZ is not related to any known Iapetan rift structures (Thomas, 2006). A decollement was subsequently produced by closing of the Iapetus Ocean during the Appalachian orogeny, resulting in the decollement being underlain by the Grenville-age basement rock and overlain by Paleozoic sedimentary rock.

The maximum depth of the decollement in the ETSZ is about 5 km, and is thicker in the east than in the west (Cook et al., 1983). More recent studies suggest that the depth to the basement rock in the Valley and Ridge Province is 6 to 12 km, and less than 2 km in the Blue Ridge and Cumberland Plateau Provinces (Brandmayr and Vlahovic, 2016).
Most earthquakes occur within the Grenvillian basement rock, below the Appalachian decollement, possibly indicating that seismicity is unrelated to the surface geology or the decollement (Bollinger et al., 1991, Powell et al., 1994).

Despite the ETSZ being one of the most active seismic areas in the eastern United States, the seismic hazard is difficult to assess, as there are no obvious surface faults attributed to current seismicity and as there has not been a recorded historical earthquake of magnitude greater than 5. The largest recorded earthquakes in the ETSZ include the 1973 Mb 4.7 Maryville, Tennessee earthquake (Bollinger et al., 1976, 1991), and the 2003 Mw 4.6 Fort Payne, Alabama earthquake (Withers et al., 2004).

Like other intraplate regions, earthquakes in the ETSZ are not typically attributed to mapped surface faults. As most earthquakes in this seismic region occur at mid-crustal depths below the Paleozoic decollement, it is possible that large earthquakes occurred during the Quaternary yet did not result in visible surface faulting. Searches for faults on the surface have revealed liquefaction and ancient faults, but present-day seismicity is not associated with these faults (Hatcher et al., 2012; Cox et al., 2022).

Paleoseismicity suggests that at least two M>6.5 earthquakes occurred in the ETSZ during the late Quaternary (Hatcher et al., 2012). Likewise, there is some evidence of three M>6 earthquakes in the ETSZ in the last 200,000 years, with two during the last 25,000 years (Warrell et al., 2017). Although historical recordings of moderate-size events in this region are available only back to about 200 years, estimates based on the geologic and tectonic setting of the ETSZ suggest that a future M>6 earthquake could be possible in this region (Hatcher et al., 2012).

The more recent December 12th 2018 Mw 4.4 Decatur, Tennessee earthquake (Figure 4.1) is the largest earthquake to occur in the ETSZ since the 2003 Mw 4.6 Fort Payne mainshock.
The 2018 earthquake provides a unique opportunity to examine the spatiotemporal seismic behavior of the ETSZ and the driving mechanism of seismicity in this region.

4.2 Background

The location and orientation of seismogenic basement faults in the ETSZ is poorly constrained (Powell et al., 2014). Understanding the cause of seismicity in the ETSZ is challenging because the seismogenic faults may be small, and the low strain rate results in infrequent earthquakes.

The dominant type of faulting in the ETSZ is strike-slip motion on steeply dipping planes. Focal mechanisms indicate a series of northeast trending, en-echelon basement faults, intersected by several east-trending faults. The NE and E trends can be interpreted as the dominant strike directions of the seismogenic basement faults (Chapman et al., 1997).

This can be grouped into two classes: N-S striking nodal planes with right-lateral slip and E-W striking nodal planes with left-lateral slip (Bollinger et al., 1991, Chapman et al., 1997). A smaller population of focal mechanisms includes NE-SW planes with right-lateral slip and NW-SE planes with left-lateral slip (Chapman et al., 1997).

Earthquakes in the ETSZ occur at depths of approximately 5 to 26 km, typically from the base of the detached Paleozoic thrust sheet down to mid-crustal depths (Vlahovic et al. 1998a). The crustal thickness is about 47 to 51 km in the southeastern part of the Blue Ridge Province in North Carolina, and extends to depths of 46 to 55 km in the Blue Ridge Mountains (Hawman, 2008). In the Tennessee Valley and Ridge province, the thickness shallows to 45 to 48 km, but then thickens to 54 km to the northwest (Parker et al., 2013).
The USGS estimated depth of the 2018 mainshock was 7.9 km, placing this earthquake on the shallower range of typical earthquake depths in the ETSZ. The shallow depth, relatively large magnitude, and location to the west of most ETSZ seismicity makes this earthquake anomalous as compared to the greater study region. This study examines in greater detail the source properties and potential fault structure that hosts the Mw 4.4 earthquake sequence.

4.3 Seismic Data and Pre-Processing

To study seismicity in the ETSZ, I use waveforms of earthquakes listed in the USGS Comprehensive Earthquake catalog (ComCat) located in the ETSZ and neighboring regions, defined as a roughly 350 km long by 180 km wide zone (Figure 4.1). This area was chosen to encompass the ETSZ and nearby seismicity, as to not exclude any events related to the ETSZ. Seismic data products, including cataloged events, their arrival times, and polarity picks, are downloaded from the USGS’s ComCat catalog at https://earthquake.usgs.gov/data/comcat/. I use 1,086 earthquakes inside this zone from January 1st, 2005, to July 31st, 2020, spanning over 15 years of catalogued seismicity to create templates for a matched filter detection. A plot of the cumulative catalogued seismicity with time is shown in Figure 4.2.
Waveforms from this catalog of events are downloaded from the IRIS Data Management Center (DMC) at https://ds.iris.edu/ and resampled at 40 Hz. I remove the mean of each waveform and apply a 2-16 Hz bandpass filter. To create “templates”, or waveforms of each earthquake, P picks are cut on vertical channels and S picks are cut on horizontal channels.

Many events seem to be missing catalogued phase picks on either vertical or horizontal channels. To add more phase arrivals to the templates, I apply PSIRpicker (Li and Peng, 2016), which automatically picks the onset of P and S arrivals by first predicting initial picks using an input velocity model. Using the ET1D velocity model (Figure 4.3) of Vlahovic et al. (1998b), I allow a 30% velocity model perturbation around which PSIRpicker searches for the initial predicted arrival times and inverts the velocity model. The window length for P and S picks are 0.8 and 1.5 seconds, respectively. PSIRpicker then predicts the new arrivals using the inverted
velocity model. The resulting perturbed velocities changed by an average of 6% for P velocities and 3% for S velocity using this method.

I cut the waveforms 0.5 seconds before a P or S arrival time to 3 seconds after the arrival time. Each earthquake must be recorded on at least 3 stations with a minimum signal to noise ratio (SNR) of 5 on the waveforms filtered from 2 to 16 Hz, using a noise window of 3.5 seconds before phase picks. These criteria result in 967 templates, which are shown in Figure 4.2.

4.3.1 Velocity Models

The most commonly used velocity model for the entirety of the ETSZ in recent publications is the ET1D model (Vlahovic et al. (1998b), which is shown in Figure 4.3. There is also an alternate velocity model, which focusses on modeling the shallow velocities near the Watts Bar Nuclear Power Plant (Cameron et al., 2017) and is most applicable to studying the Mw 4.4 mainshock and surrounding seismicity. For both models, I use a Vp/Vs ratio of 1.732.
Figure 4.3 Velocity models used for relocation and focal mechanism and depth refinement. Plotted are the models of Cameron et al., 2017 (Watts Bar model) and Vlahovic et al., 1998 (ET1D). An adapted cake layer model of Cameron et al., 2017 is used as input for hypoDD, while XCORLOC uses the gradient velocity model.

4.4 Matched Filter Detection and Refinement

I use a matched filter detection to scan through continuous waveforms for previously undetected earthquakes. The ‘similarity’ between templates and a continuous waveform is measured by the normalized cross-correlation (CC) (Gibbons and Ringdal, 2006, Shelly et al., 2007; Peng and Zhao, 2009). I examine continuous waveforms 4 weeks before to 4 weeks after the mainshock, in order to search for additional foreshocks and aftershocks.

In this study, I implement the matched filter detection using the python package EQcorrscan (Chamberlain et al., 2017). Within this package, I apply the Fast Matched Filter (FMF) code as the function to calculate cross-correlations, which provides fast parallel time-domain correlations for CPU and GPU architectures (Beauce et al., 2017).

As was done before (Shelly et al., 2007; Peng and Zhao, 2009; Ross et al., 2019), I use a threshold of 9 times the median absolute deviation (MAD) of the stacked CC sum to identify potential earthquake detections. I impose a minimum of 5 seconds between detections, so that only the detection within a 5 second window with the highest value above the threshold will be saved as an output. Likewise, detections within a 5 second window that exceed the positive threshold are prioritized over detections that exceed the negative threshold. In the case that multiple templates detect the same event, the template that is associated with the detection is the
one with the highest value above the threshold. The algorithm returns the time of detection on the earliest channel of the template.

I then estimate arrival times for each detection. The difference between the template origin time and the earliest template pick allows us to estimate the phase arrival times of a detection. For each candidate detection, I refine the estimated phase picks. The template waveforms are cross-correlated with the continuous data around the time of estimated detection, using a 2-second window around the estimated arrival time. A refined phase arrival is defined as the time within this window of the highest normalized CC coefficient. The arrival is removed from the detection if the CC value is less than 0.

This results in 3,719 candidate detections within the 8-week period of this study, including some self-detected templates. However, several of these detections contain impulsive wavelet-like signals at station ET.GRBT that correlate well with some templates (Figure 4.4). These false detections are in part attributed to interference caused by the testing of a new radio communication at another station (personal communication, Mitch Withers, CERI, 11/12/2020), and are removed by the subsequent steps.
Figure 4.4 Example of a detection from the unrefined output of EQcorrscan matched filter detection. This detection had several continuous waveforms, shown in black, that do not look like true earthquake waveforms but have a high enough cc to the template, in red, that detected them to be considered a detection. As this detection was made on only one station, GRBT, it was not included in the final set of detected earthquakes.

To further refine the candidate detections, I apply additional criteria to rule out false positives. Each waveform on which a detection is made must have a minimum SNR of 1. Candidate detections must also be detected at a minimum of 3 stations, to minimize the probability that the detection contains anthropogenic activity or local noise.

The catalog of refined detections from EQcorrscan has 930 earthquakes during the 8 weeks around the mainshock (Table S4.1), including 28 templates. There appear to be a couple microearthquakes in the days leading up to the mainshock, as well as several aftershocks that are not listed in the original USGS catalog (Figure 4.5a). Figure 4.5b shows waveforms and
spectrograms at nearby station ET.BCRT around the Mw 4.4 mainshock. The matched filter technique detected several early aftershocks listed in the USGS catalog, including one new event right following the mainshock (Figure 4.5b). Next, I determine the magnitudes of the newly detected earthquakes and relocate them based on waveform cross-correlation.

Figure 4.5 (a) Magnitudes (Md) of templates (blue circles) and calculated magnitudes of detections (red circles) plotted as a function of time in the 4 weeks before to 4 weeks after the Mw 4.4 (Md 3.88) mainshock in the ETSZ. (b) Raw vertical-component seismogram recorded at station ET BCRT showing the 2018 Mw 4.4 mainshock. Time scale is 20 minutes before origin time to 100 minutes after. The bold dashed line denotes the origin time of the mainshock. (c)
Log10-based 2-16 Hz band-pass-filtered envelope function showing possible foreshocks and aftershocks. Those marked by dashed lines with magnitudes marked are events detected by the template matching method. Only detections with SNR>1 on at least 3 stations are included. (d) The corresponding spectrogram. Only detections with SNR>1 on at least 3 stations are included.

### 4.5 Magnitude Calculations

I apply the method of Shelly et al. (2016) to calculate the magnitudes of the newly detected earthquakes, which uses principle-component fitting between a template and a newly detected event to estimate relative amplitudes. Equation 2.1 is used to calculate the magnitude of the newly detected earthquakes. The USGS ComCat catalog lists the Mw 4.4 mainshock as a duration magnitude of Md 3.88. The value of $c$ describes the amplitude-magnitude scaling. I use a standard value of $c=1$ as the scaling factor (Shelly et al., 2016). I ensure that calculated magnitudes are reliable by requiring waveforms of detections to have a minimum CC of 0.5 with waveforms of the template that detected the event on a minimum of 3 waveforms.

### 4.6 Earthquake Relocation

The templates taken from the ComCat catalog are originally located using ANSS Hyp2000, the stand-alone version of the Y2K compliant Hypoinverse (Klein, 2002). Because the detections have similar waveforms to their template events, they likely occurred nearby. Hence, I assign the initial location estimate of a detection to be that of the template that detected it. In order to find more accurate locations of the detections, I relocate both the catalogued and detected earthquakes.
I apply two separate relocation packages, XCORLOC (Lin, 2018) and hypoDD (Waldhauser, 2001), to relocate the events. This allows us to compare the relocation results and check for robustness of the solution. Both relocation packages aim to improve relative earthquake location accuracy using a differential-time cross-correlation relocation method (Waldhauser and Ellsworth, 2000). For both methods, I first compute CC based differential times between events. I use a 1.5-second-long waveform following each pick, cutting 0.2 seconds before the arrival time, and allow a pick to vary by up to 0.5 seconds when computing the CC between the waveforms of two events. I allow for sub-sample accuracy by applying a quadratic fit to the cross correlation.

A minimum of 3 picks with a minimum cross-correlation of 0.6 between waveforms of different events is required. There are not many earthquake pairs within 5 km of one another and for event pairs greater than 25 km, there are not many pairs that meet the CC criteria. Therefore, I only use event pairs within 10 km of one another.

Both XCORLOC and hypoDD require an input regional velocity model. The model used for each method is shown in Figure 4.3. HypoDD requires a layer model, while XCORLOC requires a gradient model, so the model of Cameron et al. (2017), which is created for the region around the nearby Watts Bar Nuclear Power Plant, is adopted.

4.6.1 XCORLOC Relocation

For XCORLOC, the cross-correlation based differential times are input to compute relocations. The package can be found at https://sites.google.com/view/guoqing-lin/products/xcorloc and is maintained by Dr. Guoqing Lin. The initial estimates of the detections are gradually changed through an iterative inversion with a grid-search procedure.
using the hybrid L1-L2 norm as the misfit function (Lin, 2018). Table S4.2 lists the exact input parameters for XCORLOC. Relative location uncertainties are estimated using bootstrap resampling. For the 300 cross-correlation based relocated events, the median estimated horizontal uncertainty is 0.7 km, and the median estimated vertical uncertainty is 1.1 km.

4.6.2 HypoDD Relocation

HypoDD, a similar relocation package, can be found at https://www.ldeo.columbia.edu/~felixw/hypoDD.html. Approximately 5 clusters with at least 5 earthquakes each are relocated in the region of the ETSZ with appropriate damping parameters. The input parameters to hypoDD are provided in Table S4.3. The median horizontal uncertainty for the 230 relocated events using this method is 0.2 km and the median vertical uncertainty is 0.3 km.

4.6.3 Comparison

Locations of relocated seismicity using both methods are shown in Figure 4.6. The mainshock sequence and other nearby seismicity appears to have focal depths of approximately 4 to 5 km. Table S4.4 contains the times, locations, depths, and magnitudes calculated for the events in the 8-week study period for which magnitudes and XCORLOC relocations were achievable. There appears to be a linear feature trending ENE along which much seismicity near the mainshock lies. This appears to delineate the fault on which the mainshock occurred.
Figure 4.6 Templates (blue circles) and detections (orange circles) near the mainshock, relocated using XCORLOC (panels (a) and (b)) and hypoDD (panels (c) and (d)). Red star is the Mw 4.4 mainshock. Black square is the location of the Watts Bar Dam and black diamond is the location of the Watts Bar Nuclear Power Plant. Dashed lines represent the interpreted location of the fault on which the mainshock occurred. Note that XCORLOC and hypoDD relocate a slightly different sub-set of events.

4.7 Focal Mechanisms of Nearby Seismicity

Of the 11 events that lie on or close to the mainshock fault, only the mainshock has a catalogued focal mechanism. The USGS has two reported solutions, both of which are mainly
strike-slip at a depth of either 11.5 km or 5 km. The USGS moment tensor solution for the Mw 4.4 mainshock can be found at https://earthquake.usgs.gov/earthquakes/eventpage/se60247871/moment-tensor. There is also a solution from SLU’s regional moment tensor solution (Herrman, 2013), which is similarly strike-slip, at a depth of 4 km. The SLU solution can be found at http://www.eas.slu.edu/eqc/eqc_mt/MECH.NA/20181212091443/index.html. The beach balls of these solutions are shown in Figure 4.7b.

I wish to determine the type of fault motion for other nearby events, in order to determine the location and orientation of the mainshock fault. Using the 11 events clustered around the mainshock as input, I apply the method of Hardebeck and Shearer (2002) to calculate the focal mechanisms of these events, with the python module HASHpy (Williams, 2014). This method allows for errors in polarities and takeoff angles and performs multiple trials with variations on the source location and velocity model. The USGS catalog contains several polarities for the events, and I add several manual polarities on waveforms that I deem have strong impulsive first motions. These additional polarities are only made on waveforms with a minimum SNR of 2 when filtered between 2 and 16 Hz.

I require that events have at least 7 polarities, a maximum azimuthal gap of 135 degrees, a maximum takeoff angle gap of 90 degrees, and an estimated fraction of bad picks of 0.1. The resulting focal mechanism solutions are plotted in Figure 4.7a, at the relocated epicenters of the 4 events where the criteria were met. The first motion data for these events are plotted on beach balls in Figures 4.8 to 4.11. Table S4.5 lists the nodal plane solutions for these events.
Figure 4.7 (a) Focal mechanism solutions of mainshock and nearby earthquakes using the method of Hardebeck and Shearer (2002) plotted at the locations of relocated seismicity using XCORLOC. Red star denotes relocated position of mainshock. Black square is the location of the Watts Bar Dam and black diamond is the location of the Watts Bar Nuclear Power Plant. (b) Same, for hypoDD relocations. One additional earthquake with an associated focal mechanism is relocated. Focal mechanisms 1 through 4 are listed, in order, in the Supplementary Material. (c) Alternative focal mechanism solutions for the mainshock, (top) from SLU at a depth of 4 km, (middle) from preferred USGS at a depth of 11.5 km, and (bottom) from USGS at a depth of 2 km.
Figure 4.8 First motion data for earthquake with origin time 2018-06-10 11:51:57.85 using HASHpy.

Figure 4.9 First motion data for earthquake with origin time 2018-12-12 9:14:43.61 using HASHpy.
Figure 4.10 First motion data for earthquake with origin time 2018-12-12 9:27:07.94 using HASHpy.

Figure 4.11 First motion data for earthquake with origin time 2018-12-14 00:24:06.17 using HASHpy.
My calculated mainshock focal mechanism is very similar to the preliminary focal mechanisms by the USGS and SLU, with very similar strike, dip, and rake. Other regional catalogued focal mechanisms from Chapman et al. (1997) are shown in Figure 4.12, with an additional event near the mainshock having a similar focal mechanism to the mainshock.

Figure 4.12 Regional map view around mainshock, with my HASHpy solutions as blue beach balls, and catalogued beach balls from Chapman et al., 1997 as red beach balls. The event near the mainshock from Chapman et al., 1997 occurred on December 22, 1985 and is plotted at the catalogued location, as it is outside the time frame of the study period for relocation. Blue dots are relocated templates using XCORLOC, and orange dots are relocated new detections using XCORLOC.
The three other events near the mainshock have strike-slip focal mechanisms in a very similar orientation to the mainshock. The focal mechanisms are consistent with left lateral faulting along an East/North-East (E/NE) trending fault. This corroborates the evidence that the relocated seismicity delineates the fault along which the Mw 4.4 mainshock occurred.

4.8 Focal Mechanism and Depth Refinement of the Mainshock

In order to refine the depth and focal mechanism solution of the mainshock, I apply the cut-and-paste (CAP) method (Zhu and Helmberger, 1996), which minimizes the L2 norm of the difference between the observed and synthetic waveforms using a grid search to solve for the best moment magnitude, source depth, and focal mechanism. CAP is maintained by Dr. Lupei Zhu and can be found at http://www.eas.slu.edu/People/LZhu/home.html. Each waveform is partitioned into a Pnl and surface wave segment, which are allowed to shift independently to achieve the best correlation between observed and synthetic waveforms. This process may suppress some of the error attributed to the velocity model.

I download the waveforms of the 22 stations within 200 km of the mainshock epicenter, as well as from the nearby Watts Bar Nuclear Power Plant station (Graizer et al., 2020), two minutes before the mainshock origin time to three minutes after. The east and north components are rotated to the radial and transverse directions, respectively. The instrument response of each waveform is also removed, and I convert the ground velocity waveforms to units of cm/s in order to calculate a reliable moment magnitude.

I input the P arrival times and the CAP algorithm separates each waveform into Pnl and surface wave sections. The maximum length for the Pnl and surface wave windows are 35 s and 70 s, respectively, and I allow a maximum time shift of 1 s for the Pnl and 3 seconds for the
surface wave. The filter for Pnl waves is 0.05 to 0.3 Hz, and the filter for surface waves is 0.05 to 0.1 Hz. These frequency bands are chosen to maximize the SNR of the data and to avoid high-frequency noise observed at several stations. Data from several stations is contaminated by low frequency noise and is removed, resulting in 11 stations with clear arrivals as input.

To create the synthetic seismograms, the Green’s function at each station’s epicentral distance is calculated using the Haskell propagator matrix frequency-wavenumber (FK) method (Zhu and Rivera, 2002). I create synthetic waveforms for candidate source depths from 1 to 12 km, using the ET1D velocity model as input. To find the best focal depth, I perform the grid search in the depth range of 1 to 12 km to find the waveforms which yield the highest variance reduction. Figure 4.13 shows the best fit solution for depths in this range. The depth that returns the solution with the minimum RMS is the final source depth.
Figure 4.13 Resulting RMS of solutions for different candidate source depths using CAP. The focal mechanism solution at each depth is plotted as a beach ball, with the number above representing the calculated magnitude.

Figure 4.14 shows the waveform fit of the best solution. Synthetic seismograms are plotted on top of observed waveforms, with time shifts. The figure depicts a good match between synthetic and observed waveforms, with a 47% variance reduction. There are some stations that do not have fits as well as others, including the nearby Watts Bar station. The best solution corresponds to a moment magnitude Mw 4.52 at a depth of approximately 5.9 ± 0.6 km, with a strike, dip, and rake of 81°, 78°, and −3°, respectively, which is consistent with the SLU’s regional moment tensor solution for this event (Figure 4.7b).
I also try an alternate velocity model, which focuses on modeling the shallow velocities near the Watts Bar Nuclear Power Plant (Cameron et al., 2017). The results of this CAP fit are shown in Figure 4.15, and estimate the mainshock depth to be 5 ± 1 km. From the CAP solutions, the mainshock depth is therefore between 5 and 6 km.

Figure 4.14 Focal mechanism solution for the December 12, 2018 Mw 4.4 mainshock using CAP and ET1D as the input velocity model. Black waveforms (observed ground velocity) are superimposed with red waveforms (synthetic seismogram). Station names are at the left of each row, with epicentral distance in km/constant shift in seconds. The leftmost two columns denote the waveform fits of the vertical and radial components of the Pnl waves, and the next columns
show the fit of the vertical, radial, and tangential components of the surface waves. The numbers below each trace are the time shifts, in seconds, and the cc coefficients.

Figure 4.15 Focal mechanism solution for the December 12, 2018 Mw 4.4 mainshock using CAP with the velocity model of Cameron et al., 2017. Black waveforms (observed ground velocity) are superimposed with red waveforms (synthetic seismogram). Station names are at the left of each row, with epicentral distance in km/ constant shift in seconds. The leftmost two columns denote the waveform fits of the vertical and radial components of the Pnl waves, and the next columns show the fit of the vertical, radial, and tangential components of the surface waves. The numbers below each trace are the time shifts, in seconds, and the cc coefficients.
4.9 Mainshock Rupture Directivity

Rupture directivity offers insight into the source processes of an earthquake and can help identify the rupture plane, in order to discriminate it from the auxiliary plane. I follow the method of Cesca et al. (2011), which is based on the inversion of amplitude spectra from P-wave seismograms to estimate the apparent duration at each station.

All available stations with P arrivals of the mainshock are selected, using the north, east, and vertical orientations. I deconvolve the instrument response from the raw velocity data and convert to displacement. The inversion process is limited to P-wave time windows, so I cut the waveforms from 5 seconds before the P arrival to the S arrival and apply a 5% cosine taper to either end of the data to minimize the effect of noise and the S arrival. I use a bandpass filter of 0.05 to 1 Hz. The ET1D velocity model is used in computing synthetic seismograms with the frequency-wavenumber code of Zhu and Rivera (2002), which are also filtered and cut in the same manner as the observed waveforms. A different synthetic seismogram is computed for each candidate apparent source duration, using a triangular pulse as the source-time function.

The next step is to compute the amplitude spectrum of both the observed and synthetic waveforms, as the input to invert for apparent source duration. I perform a grid search for possible apparent source durations at each station, in the range of 0.2 to 5 seconds, with an increment of 0.2 seconds. In general, I observe a smooth single-minimum curve of misfit versus apparent duration. However, a few stations, especially those close to the mainshock, have a P-window too short to be reliably used, a poor fit between synthetic and observed waveforms, low-frequency noise in the range of interest, or a combination thereof, and hence they are not used. Figure 4.16 illustrates an example of the main steps as viewed for one station.
Figure 4.16 Example of the steps followed to invert for the apparent source duration at one station. Left: Filtered displacements (black lines) and synthetic seismograms (red lines) for the point source model corresponding to the lowest misfit. Center: Amplitude spectra comparison, with red lines corresponding to the spectra of the best fitting synthetic spectra. Right: Amplitude spectra misfit values for different source durations. Best solution is identified by a red star.

Inversion results are shown in Figure 4.17. Stations are colored by the apparent duration and show a directivity effect which is modeled by a sinusoid curve. The azimuthal distribution of apparent duration indicates that the mainshock involved a bilateral rupture, with stations to the East and West of the epicenter experiencing shorter apparent durations, in general, compared with stations to the North and South.
Figure 4.17 Inversion results for the directivity of the Mw 4.4 mainshock. (a) Colored plot of inverted apparent duration at stations located around the epicenter as triangles, with red to blue indicating increasingly longer apparent durations. Open circles represent stations that were not used in the inversion. Beachball is plotted at location of Mw 4.4 mainshock. (b) Apparent duration versus azimuth for each station (black dots) and the best fitting model (solid line).
4.10 Basement Depth Estimate near epicenter of Mainshock

As the mainshock depth is close to the estimated boundary between the Grenvillian basement rock and the overlying Paleozoic sediment, I wish to constrain the location of this event relative to the decollement. I compute a refined basement depth near the epicenter of the Mw 4.4 mainshock. One technique uses the travel-time difference of S, P, and Sp phases, along with an estimated angle of incidence, to infer the thickness of the overlying sediments (Nolte and Tsoflias, 2021).

An Sp (S-to-P) converted phase can be efficiently produced at a sediment-basement boundary with an abrupt velocity contrast. The Sp wave has an impulsive arrival and is distinct from the S arrival as it has the highest amplitude on the vertical component. Here I examine accelerograms from the nearest available station to the Mw 4.4 mainshock. This station is located at the nearby Watts Bar Nuclear Power Plant, which is approximately 5 km from the epicenter and had triggered accelerograms during the mainshock. Seismic data recorded at Watts Bar nuclear power plant are not open to public and can be requested by contacting trsmith1@tva.gov.

I use the accelerogram recordings from station 75A, at the unit reactor building on the floor slab, as this station has the clearest signals. The triggered waveforms of this station are plotted in Figure 4.18a. I calculate an S-Sp time of 0.51 seconds and an S-P time of 0.74 seconds. I use a built-in method in ObsPy (Beyreuther et al., 2020), which uses the orientation of the particle motion vector around the P arrival window, from 0.2 s before the arrival to 0.5 s after, to calculate the angle of incidence, which is estimated at 29 degrees.
Figure 4.18 (a) Three component (vertical, longitudinal, and tangential) accelerogram at Watts Bar Plant station 75A. Blue line marks the Sp arrival and red line marks the S arrival. (b) Diagram of a two-segment travel path for the earthquake. The two paths extend from the hypocenter to the conversion interface and from the conversion interface to the surface, which can be modeled as two triangles. The hypotenuse of the upper path is the ray path, \( f \), as the following diagram illustrates. Solving for the distance, \( d \), results in the approximated depth to the interface at the decollement.

The method assumes a two-segment travel path for the earthquake, as shown in Figure 4.18b. The two paths extend from the hypocenter to the conversion interface and from the conversion interface to the surface, which can be modeled as two triangles. The hypotenuse of the upper path is the ray path, \( f \), as the following diagram illustrates.

The S-Sp time arises from the travel time difference of the waves along segment \( f \). Likewise, the Sp-P time arises from the travel time difference of the waves along the ray path from the earthquake to the conversion interface. Once we solve for \( f \), the vertical side of the
upper triangle can be solved for the depth to the basement rock. I use the following equation from Nolte and Tosflias (2021) to estimate the depth to the basement rock:

\[ f = \frac{1}{\left( \frac{1}{v_{up}} - \frac{1}{v_{us}} \right)} t_{S-S_p} \]  

The average of the top 5 km layer velocities of Cameron et al. (2017) is used for the top layer above the decollement. The estimated basement depth is 4.4 km.

4.11 Discussion

4.11.1 Relative Lack of foreshocks and aftershocks

Only a small fraction of detections are relocated by either XCORLOC or hypoDD. XCORLOC and hypoDD relocate 13% and 10% of input detections which are not templates, respectively. This is in part due to several detections being made on only three stations, and therefore with few waveforms that have high enough correlation to nearby events, they did not pass the thresholds for relocation. A few of the detections could also be false, but waveforms indicate that a majority are likely small-magnitude earthquakes, or anthropogenic events such as mining blasts. However, these detections are not all necessarily located near the mainshock, as the relocated clusters show in Figure 4.19. Locations of earthquakes farther from the mainshock site are likely less accurate, as I chose a velocity model that is best suited for the area around the mainshock region. The detections that are relocated by XCORLOC within 25 km of the mainshock epicenter have duration magnitudes of 1.2 to 1.5, so it is unlikely that there are larger magnitude detections that are excluded near the mainshock, as these events should have been detected on more stations with clearer signals.
Figure 4.19 Templates (blue circles) and detections (orange circles) relocated using XCORLOC (panels (a) and (b)) and hypoDD (panels (c) and (d)). Red star is the Mw 4.4 mainshock. Panels (a) and (c) show original catalogued and detected earthquake locations, and panels (b) and (d) show relocated seismicity. The clusters of seismicity could be due to a combination of mining induced earthquakes and mislocation, a subject that will be further investigated in a subsequent study.
There does not appear to be a statistically significant increase in seismicity leading up to the mainshock. The months leading up to the mainshock saw a below average rate of seismicity within 25 km of the mainshock (Figure 4.20, Figure 4.21). Similarly, there is not a clear increase in seismicity in the days to hours before the mainshock.

Figure 4.20 Cumulative Number of earthquakes within 25 km of mainshock epicenter (blue line). Black dots denote individual catalogued earthquakes by their magnitude (Md), and red dots are detected events using matched filter detection. Black vertical line is at mainshock origin time.
Figure 4.21 Background (catalogued) earthquakes within 50 km of the mainshock between January 1st 2005 and July 31st 2020, plotted by (a) magnitude (Md) and (b) distance (km) from mainshock epicenter. Filled circle is mainshock.
4.11.2 Fault Orientation

Few, if any, earthquakes in the ETSZ are attributed to any named faults (Steltenpohl, 2010). It may be particularly difficult to determine where faults lie, as most of them do not break to the surface and are not continuous (Warrell et al., 2017; Hatcher et al., 2012; Bollinger et al., 1991). Like other earthquakes in the ETSZ, the Mw 4.4 mainshock has no associated mapped fault. The relocated seismicity of the Mw 4.4 mainshock using both relocation methods of XCORLOC and hypoDD reveals a E-NE linear trend of seismicity. The feature is consistent with a left-lateral motion on a E-NE striking fault, as evidenced by both the USGS’s and SLU’s focal mechanism solutions and several of my calculated focal mechanisms for nearby earthquakes. Most of the focal mechanism solutions indicate strike-slip motion, which is representative of the faulting style of the ETSZ.

Rupture directivity indicates rupture along an approximately East-West fault, which is consistent with my interpretation of a E-NE trending fault. HypoDD relocates the mainshock to the center of the inferred mainshock fault, while XCORLOC relocates it to the W-SW edge of the relocated seismicity. HypoDD’s relocation of the mainshock to the center of the fault supports bilateral rupture toward either direction of the fault found by the directivity analysis, but the relocation uncertainties preclude clearer relative locations.

Relocated aftershocks are mainly within 10 km of the mainshock epicenter and do not spread farther within the 8-week study period. There are several scattered events about 40 km away and another cluster further away, but these are not considered as cases of remote triggering by the mainshock. This suggests that this moderate-sized mainshock had a limited spatial effect and likely did not trigger seismicity elsewhere.
4.11.3 The Origin of the Mw 4.4 Mainshock

The December 12, 2018 Mw 4.4 mainshock is one of the largest earthquakes recorded in the ETSZ. Its location, depth, and spatiotemporal properties make this earthquake unlike most seismicity in the seismic zone. The sets of USGS, SLU, and my focal mechanism solutions for the mainshock all indicate strike-slip faulting on steeply dipping planes, which is consistent with the dominant type of faulting in the ETSZ. The mainshock and nearby events, which appear to lie along the same fault, mostly appear consistent with ENE/WSW left-lateral strike-slip motion on the ENE/WSW striking planes. This is somewhat consistent with the dominant type of faulting in the ETSZ, wherein most of the seismicity exhibits right-lateral slip on N-S striking planes and left-lateral slip on E-W striking planes, with some minor seismicity on NE-SW planes with right-lateral slip (Bollinger et al., 1991; Chapman et al., 1997).

The CAP solution for the mainshock depth of 5 to 6 km suggests that the mainshock was shallower than the estimate of the USGS. Figure 4.22 illustrates the depths of seismicity near the mainshock epicenter, compared to seismicity along a cross-section through the ETSZ. Although most seismicity in the ETSZ has depths greater than 10 km, the mainshock sequence and other catalogued seismicity on the mainshock or nearby faults has focal depths of approximately 4 to 5 km. The relocated seismicity also has larger depth towards the east (Figure 4.22d).
Figure 4.22 (a) Map of ETSZ. Black dots represent events relocated by XCORLOC that are within 20 km of the cross-section A-B, and gray dots represent those events which are farther away. Red star represents location of mainshock. Dashed line shows approximate location of the
NY-AL Lineament. (b) Plot of events within 20 km of the cross-section A-B, plotted as a function of distance along the cross-section and depth. Red triangles represent the subset of events which are within 5 km of the mainshock’s epicenter. Yellow star shows location of mainshock, at the CAP solution of 4 km depth. Black dashed line shows the typical 5 km depth of the decollement in the ETSZ, while the red dashed line shows the 4.2 km depth estimate for the depth to the basement. Vertical dashed line depicts approximate location of the NY-AL Lineament, as it transects the cross-section A-B. (c) Zoom-in around mainshock and nearby seismicity. The empty star represents the XCORLOC depth of the mainshock. The two filled stars represent the depths from CAP solutions using the two different velocity models. Red triangles represent the subset of events which are within 5 km of the mainshock’s epicenter. (d) Same as (c), but for an along-strike (E-NE/W-SW) cross section. (e) Same as (c), but for a fault-perpendicular (N-NW/S-SE) cross section.

My solution and the SLU solution suggests that the mainshock fault is shallow and lies very close to the decollement, which is generally considered to be at ~5 km depth in the ETSZ. My estimated depth to the decollement beneath the mainshock is approximately 4.4 km. Results are consistent with previous studies of thickness variations of sediment overlying the basement rock being thicker in east than in west (Cook and Vesudevan, 2006, Brandmayr and Vlahovic, 2016). This depth places the mainshock definitively within the basement rock and not in the overlying sedimentary rock. However, as most seismicity associated with the ETSZ occurs well beneath the decollement in the range of 5 to 26 km, this raises the possibility of a different driving force for this fault.
My results demonstrate that detailed analysis of the December 12, 2018 Mw 4.4 mainshock sequence can provide new insight into fault structure in the ETSZ, which is typically elusive in this seismic zone. However, the exact origins of the mainshock and this fault are not clear. As the Mw 4.4 mainshock occurred near the western edge of the ETSZ, there were few stations in operation to the west of the epicenter during the 8-week study period. I note that better azimuthal coverage of the stations, especially adding stations near the mainshock to the west, would significantly improve the calculation of focal mechanisms of this study and provide a clearer picture of rupture directivity of the mainshock.

Several recent studies have found possible paleoseismological evidence of M>6 earthquakes in the ETSZ during the late Quaternary (Hatcher et al., 2012; Warrell et al., 2017; Cox et al., 2022). In addition, Jobe et al. (2022) found possible surface expressions that are broadly consistent with the left-lateral E-W trending lineaments and previously mapped faults along the ETSZ. Figueiredo et al. (2022) also found the 2020 M5.1 Sparta, North Carolina earthquake ruptured along a pre-existing shallow fault structure to the surface, which was not identified previously. While these geological studies did not cover the region that hosted the 2018 M4.4 mainshock, I argue that further joint studies are needed to determine the causes of shallow seismicity in the broader seismic zone of the ETSZ and elsewhere in the CEUS.

4.12 Conclusions

I examined the recent 2018 Mw 4.4 earthquake in the ETSZ, which had a similarly shallow depth, and very few aftershocks, compared to the 2014 Mw 4.1 Edgefield, SC earthquake. I resolved the fault orientation on which the mainshock lies, which was previously unknown by finding focal mechanism of earthquakes near the mainshock and by performing
rupture directivity analysis. Focal mechanism solutions, relocated epicenters, and rupture
directivity all suggest that the mainshock ruptured on an approximately E-NE trending fault with
left-lateral motion. I also conclude that the Mw 4.4 mainshock occurred at a depth of 5 to 6 km,
which is below my estimated depth to the basement rock below the decollement of
approximately 4.4 km.
CHAPTER 5. Characterization of seismicity in the Eastern Tennessee Seismic Zone (ETSZ) using over 15 years of continuous detections

This study examines over 15 years of continuous data in the ETSZ, using matched filter detection to find microseismicity not included in the original earthquake catalog. I used about 30,000 GPU-hours on the XSEDE Bridges supercomputer, which took nearly a year and a half of run time. The new catalog of detections yields the most detailed complete catalog yet for this seismic zone, as well as magnitude estimations and more defined structure at depth. I also examined seismicity in the vicinity of the Watts Bar Reservoir, which is located about 5 km from the epicenter of the Mw 4.4 December 12, 2018 Decatur, Tn earthquake, to search for Reservoir Induced Seismicity (RIS). I also examined seismicity in the entire ETSZ to search for a correlation between shallow earthquakes and seasonal hydrologic changes.

The results in this chapter have been presented in a recent conference, the 2021 Fall American Geophysical Union Annual Meeting (Daniels et al., 2021b). This work is funded by a USGS (United States Geological Survey) Earthquake Hazards Program Grant and the final technical report was submitted to the USGS on July 3, 2022.

5.1 Introduction

As mentioned in Chapter 4, the Eastern Tennessee Seismic Zone (ETSZ) is the second most active seismic zone in the eastern United States, after the New Madrid Seismic Zone in the Mississippi Valley area, yet its origin is still not well understood (Powell et al., 1994). The ETSZ is defined by a roughly 300 km long, 100 km wide region of diffuse seismicity that trends approximately 45 degrees from northeastern Alabama, north Georgia, northeastern Tennessee.
and western part of North Carolina (Powell et al., 1994, Brandmayr and Vlahovic, 2016, Hatcher et al., 2012, Chapman et al., 1997), as Figure 5.1 depicts.

Most earthquakes epicenters occur in eastern Tennessee, which is characterized by highly tilted and folded rock layers. The Valley and Ridge Province exhibits the highest level of seismic activity, although many earthquakes also occur in the Blue Ridge Province. Most earthquakes occur within the Precambrian basement rock, below the Appalachian decollement. My recent study around the 2018 Mw 4.4 earthquake epicenter suggests a depth of approximately 4.4 km to the decollement in the western part of the ETSZ, described more in depth in Chapter 4.
Figure 5.1 Map showing topography of study region and earthquakes in the ComCat catalog between January 1st 2005 and July 31st 2020. Each earthquake is plotted by a black open circle. The rotated rectangle shows the approximately boundary for earthquakes in the ETSZ. Black lines denote New York–Alabama (NY-AL) and Clingman (CL) magnetic lineaments. White stars indicate epicenters of largest earthquakes recorded in the ETSZ; The 1973 Mb 4.7 Maryville, Tennessee earthquake, the 2003 Mw 4.6 Fort Payne, Alabama earthquake, and the 2018 Mw 4.4
Decatur, Tn earthquake. Pink triangles show stations on which catalogued earthquakes have arrival times.

Despite being one of the most active seismic areas in the eastern United States, the seismic hazard of the ETSZ is difficult to assess, as there are no obvious surface faults, and no recorded historical earthquakes of magnitude greater than 5. The December 12th 2018 Mw 4.4 Decatur, Tennessee earthquake (Figure 5.1) is the most recent moderate-size earthquake in the ETSZ. Additional information about this event can be found in Chapter 4.

In this study, I perform a systematic detection of the seismicity between January 1, 2005 and July 31, 2020 with a template matching technique, followed by relocations with waveform cross-correlation to reveal characteristics of seismicity at depth.

5.1.1 Focal Mechanisms and Faults in the ETSZ

The location and orientation of seismogenic basement faults in the ETSZ is poorly constrained (Powell et al., 1994). Hypocentral patterns are diffuse and nodal planes do not align with any known major basement structures. This suggests that earthquakes occur on many small reactivated faults rather than on large, continuous faults. This may account for the lack of large M > 5.5 earthquakes in the region (Bollinger et al., 1991).

The dominant type of faulting in the ETSZ is strike-slip motion on steeply dipping planes. This can be grouped into two classes: NNE-SSW striking nodal planes with right-lateral slip and E-W striking nodal planes with left-lateral slip (Bollinger et al., 1991, Chapman et al., 1997, Powell et al., 1994). Focal mechanisms indicate a series of northeast trending, en-echelon basement faults, intersected by several east-trending faults.
The modern-day compressive stress field has existed since the Cretaceous, about 70 million years ago. This resulted from continental resistance to plate motion and ridge-push along the East Coast (Zoback and Zoback, 1991). The orientation of the maximum regional compressive horizontal stress is N50°E (Davison, 1988), which agrees with (Zoback and Zoback, 1991) and the primarily strike-slip and normal orientation of focal mechanisms (Figure 4.1b). Present-day seismicity may occur on ancient faults being reactivated in the current stress field (Chapman et al., 1997, Wheeler, 1995).

5.1.2 The New York – Alabama (NY-AL) Lineament

However, the question remains as to the locations and extent of these faults. Several models to explain the source of seismicity in this region have been proposed.

About 80 to 90 percent of Tennessee-North Carolina earthquake epicenters lie between deeply situated linear structures or boundaries that have been detected due to their magnetic field (King and Zietz, 1978; Nelson and Zietz, 1983).

The most notable such feature is the New York-Alabama (NY-AL) Magnetic Lineament, which is not seismogenic itself, and is defined by the gradient in magnetic anomalies and associated with Bouger gravity lows (Powell et al., 1994, 2014, King and Zietz, 1978). Seismicity appears to be spatially correlated with this magnetic and gravity lineament, which spans 1300 km from New York to Alabama, as shown in Figure 5.1. This feature likely represents a major crustal boundary in the Grenville orogeny or a buried right-lateral strike-slip fault attributed to the Grenville orogeny (King and Zietz, 1978, Steltenpolhl et al., 2010). The majority of ETSZ seismicity occurs near a part of the NY-AL lineament, and especially to the east of it.
However, along most of this lineament lie seismically inactive regions. Since the New York-Alabama Magnetic Lineament has a much larger extent than the ETSZ and the regions of highest earthquake density are not along the NY-AL lineament, others have argued that this feature is not correlated with the ETSZ (Hatcher et al., 2012, Bollinger et al., 1991). The lineament bounds much, but not all of seismicity, as some epicenters occur to the west of the lineament. As a result, the exact relationship between this geophysical anomaly and seismicity in the ETSZ is unclear (Hatcher et al., 2012, Powell et al., 2014). Furthermore, the NY-AL lineament has an approximate strike of 30 degrees east of north, which is not consistent with the dominantly NE and E striking planes of seismicity in the ETSZ (Bollinger et al., 1991).

The above observations suggest that the NY-AL lineament could be a crustal boundary, which parallels the overthrust Appalachians. The NY-AL lineament could separate the North American cratonic basement crust of Grenville age from the more heterogeneous basement crust underlying the Valley and Ridge province and Blue Ridge province. One challenge to this view is that the presence of subducted Grenville age rocks to the east in the Valley and Ridge province and Blue Ridge province indicates that the NY-AL lineament is a structural feature within the Grenville basement rock, instead of a boundary between blocks of different compositions or ages (Bollinger et al., 1991). There is another magnetic anomaly in the region, the Clingman magnetic lineament (Powell et al., 1994), which may also be a structural discontinuity and possibly an ancient fault (Nelson and Zietz, 1983).

Most seismicity in the ETSZ occurs between the NY-AL lineament, to the west, and the Clingman lineament, to the east, in a region called the Ocoee block. This could suggest that the lineaments bound the seismogenic basement block, therefore influencing the locations of
regional seismicity (Johnston et al., 1985). Powell et al. (1994) suggests that the Ocoee block is relatively weak and that regional stress builds up in this block, resulting in the ETSZ.

The weak-block theory is also supported by the presence of a low-velocity zone in which the NY-AL lineament is situated and a higher velocity zone to the east (Powell et al., 2014). The ETSZ is located in a transition zone wherein velocity is high to the east and low to the west. The densest seismicity in the ETSZ occurs in the high-velocity region to the east of the NY-AL lineament, which could be interpreted as the reactivation of an ancient shear zone (Powell et al., 2014).

5.1.3 Seismicity Relationship to Surface and Ground Water

Another possible explanation for the existence of the ETSZ is due to erosion from rivers. Erosion from the Upper Tennessee drainage basin has removed approximately 3,500 cubed km of rock over the last 9 Ma (Gallen and Thigpen, 2018). This has increased the local stress field through redistribution of surface loads, potentially driving the mechanism for ETSZ seismicity. A similar mechanism has been invoked to explain the seismicity in the NMSZ (due to erosion of the Mississippi rivers) (e.g., Calais et al. 2010; Craig et al., 2017).

Finally, Costain and Bollinger (2010) hypothesized that most intraplate and near-intraplate earthquakes are associated with the dynamics of hydrologic cycles, including both short-term (rainfalls, hurricanes) and long-term (annual or decadal changes in ground waters) processes (Gupta, 2002; Talwani et al., 2007; D’Agostino et al, 2018). However, as most earthquakes occur within the Grenvillian basement rock, below the Appalachian decollement, this suggests that the ETSZ seismicity is too deep and is unlikely to be related to the surface processes (Bollinger et al., 1991, Powell et al., 1994).
5.1.5 Potential for Seismic Hazard

As mentioned before, no clear evidence for surface rupture has been reported in the ETSZ (Powell and Beavers, 2009). Intraplate earthquakes are not well understood, so the nature and origins of seismicity remain ambiguous.

Paleoseismicity suggests that at least two M>6.5 earthquakes occurred in the ETSZ during the late Quaternary (Hatcher et al., 2012). Likewise, additional evidence shows a history of three M>6 earthquakes in the ETSZ in the last 200,000 years, with two during the last 25,000 years (Warrell et al., 2017), but this is not conclusive. Although historical recordings of moderate-size events in this region are available only back to about 200 years, estimates based on the geologic and tectonic setting of the ETSZ suggest that a future M>6 earthquake might be possible in this region (Hatcher et al., 2012).

Understanding why the ETSZ is seismically active is important for seismic hazard assessment. The region in which the ETSZ is situated contains vital structures such as government research laboratories, hydroelectric dams, nuclear power plants, and population centers such as Knoxville and Chattanooga, that risk adverse effects of potentially damaging earthquakes. It is unclear if present-day seismicity can be related to basement structure generated during ancient orogenic events, and the extent of the faults on which seismicity occurs. More detailed analysis on the locations and orientations of earthquake faults in this seismic zone are required to determine the source and relation of seismicity to nearby geophysical features.

5.2 Matched Filter Detection and Refinement

The quality of the detected earthquakes over the 15-year study period varies. Several Duration Magnitude (Md) 2 to 3 earthquakes are detected, but there are far more small
earthquakes, possible anthropogenic noise, mine blasts and false detections that are also detected in the raw matched filter detections. In general, larger magnitude earthquakes are detected on more stations (in part because the energy travels farther), with clearer signals (and therefore larger SNR). However, it is more challenging to disentangle false detections from small magnitude earthquakes, and to then use them for relocation and magnitude estimation.

I ran matched filter detection using computation time on the XSEDE Bridges supercomputer. This took approximately 30,000 GPU-hours, which is almost four years of computing time. I ran all the continuous time detections in under a year and a half, however, by running multiple jobs in parallel.

The original matched filter detection threshold was set to $9 \times \text{MAD}$, which is the same for the 8-week study period in the time around the Mw 4.4 Decatur, Tn earthquake. However, there is considerable temporal and spatial variation in the quality of detections in the much longer time period of over 15 years, compared to the time and area around the Mw 4.4 earthquake. Around some times and areas, there appeared to be a prevalence of false detections and/or anthropogenic noise. This resulted in a high percentage of false detections in the overall final catalogs for the over 15-year period. Therefore, I raise the threshold for matched filter to be $10 \times \text{MAD}$ for the 15-year study period.

Earthquake relocation generally requires arrivals on at least two or three stations. Many of the small detected earthquakes in the ETSZ lack clear arrivals on more than one or two stations. Therefore, I create two distinct final data products. The first is a more inclusive catalog of detections, including many of the smaller earthquakes for which it is not feasible to find relocations and many of which don’t meet the minimum qualifications to compute magnitudes.
The second dataset is stricter in its criteria, and therefore contains less events but of greater quality for input to relocation.

To create the first, more inclusive catalog, I require detections to have a minimum CC of 0.3 between template and detection waveforms for P arrivals and 0.4 for S arrivals. Arrivals must have a minimum SNR of 3 on at least two stations. This reduces the chance that the catalog contains detections which are anthropogenic noise, or local phenomena that only one station recorded. This catalog contains 13,671 events, including templates (Table S5.1).

The cumulative distribution of these earthquakes with time is shown in Figure 5.2a. There are several months that have a below average number of earthquakes. The most obvious such time period is a several-month period in 2009, approximately from February to October, during which there are few detections that are detected and meet the minimum criteria. 122 of the 13,671 detections occur in 2009, which represents about 0.9% of the events in this catalog.
Figure 5.2 (a) Cumulative distribution of detected (and template) earthquakes with time. (b) Cumulative distribution of the 1,421 earthquakes input to relocation.

Part of the reason for which there are less templates -- and somewhat consequently, less detections -- during several months in 2009 is that there was a drop in station data availability
during much of 2009 (Figure 5.3). Several months in the year of 2009 saw a drop from 60 stations with available data to about 40 stations, which corresponds to a roughly 30% drop in available data. The ETSZ has sparse station coverage compared to more studied seismic regions, especially in California and other interplate seismic zones. Any drop in station coverage or data issues on stations in the ETSZ can result in much poorer spatial coverage, making earthquake detection on quality signals challenging. The positive side is that the station coverage in the ETSZ appears to increase with time. More stations added over time biases detections towards the later times in the 15-year study, but this is a positive trend for this seismic zone, as the current station spacing is farther apart than well-studies seismic zones.

Figure 5.3 This graph was made by requesting 30 seconds of continuous data each week for the 15+ year study for each station that has template arrival times. Any location codes for stations are requested. If an error occurred, or no data was available, the station was presumed to not have available data that week. If at least one of the 3 components is downloaded, I deem that the station is available that week. It is important to note that some of the stations on which templates
have picks are not in the ETSZ, but near to it, as larger magnitude earthquakes have arrivals on stations that are farther away.

The second and stricter catalog, which is used as input to relocation, requires the same criteria, as well as the additional constraints that detections must have at least 2 stations that have both a P and S arrival that meet all previous criteria. These requirements not only help to create reliable relocations, but also decrease the chance that the catalog contains events which are less likely to be earthquakes or have imprecise arrival times. This catalog contains 1,421 events, including templates (Figure 5.2b, Table S5.2). This catalog of earthquakes seems much more uniform over time, which is a further indication that it likely contains very few false detections.

5.3 Earthquake Relocation

As aforementioned, this more selective catalog of 1,421 earthquakes with clear phase arrivals is used as input for relocation, including 967 templates and 454 detections. I apply two methods of relocation, XCORLOC and hypoDD, to compare features of the relocated seismicity.

5.3.1 XCORLOC Relocation

The input parameters to XCORLOC are given in Table S5.3. 1,414 earthquakes are relocated using both waveform-based cross-correlations and phase differential times, including 952 templates and 462 detections. For the 405 CC-based relocated events, the median estimated horizontal uncertainty is 0.06 km, and the median estimated vertical uncertainty is 0.3 km. Figure 5.4 below shows the distribution of the relocated earthquakes, along with a more detailed view of the central part of the ETSZ. Figure 5.4a also shows which relocated earthquakes are templates and which are new detections. The depths of relocated seismicity are compared to the original
locations in Figure 5.5.
Figure 5.4 Earthquakes relocated using XCORLOC. (a-c) show original locations of earthquakes relocated. (d-f) show locations of earthquakes relocated using XCORLOC (a) and (b) show templates (blue circles) and detections (orange circles). (b) and (e) show an enlarged view of the main part of the ETSZ, for detail. Red star is the 2018 Mw 4.4 mainshock.

![Earthquake Depths](image)

Figure 5.5 Comparison of relocated event depths using XCORLOC with original catalogued depths of events relocated.

5.3.2 hypoDD Relocation

The input parameters to hypoDD are given in Table S5.4. The DAMP parameter used for individual relocation of each cluster in given in Table S5.5. 1,212 earthquakes are relocated using both waveform-based cross-correlations and phase differential times, including 819 templates and 393 detections. The plots below in Figure 5.6 show the distribution of the relocated earthquakes, along with a more detailed view of the central part of the ETSZ. Figure
5.6a also shows which relocated earthquakes are templates and which are new detections. The depths of relocated seismicity are compared to the original locations in Figure 5.7.
Figure 5.6 Earthquakes relocated using hypoDD. (a-c) show original locations of earthquakes relocated. (d-f) show locations of earthquakes relocated using hypoDD (a) and (b) show templates (blue circles) and detections (orange circles). (b) and (e) show an enlarged view of the main part of the ETSZ, for detail. Red star is the 2018 Mw 4.4 mainshock.

![Earthquake Depths](image)

Figure 5.7 Comparison of relocated event depths using hypoDD with original catalogued depths of events relocated.

### 5.4 Relocation Comparison and Comparison with High Gradient in Magnetic Anomaly

Both XCORLOC and hypoDD show clearer clustering/linear features, especially along the NY-AL lineament and to its east. There are several areas, which when viewed in greater detail, show clearer structure than before. The linear features may suggest the location of several faults along which the seismicity is concentrated, which are highlighted in Figure 5.8. The
original locations of earthquakes relocated using XCORLOC, plotted alongside the relocated
XCORLOC events, is provided in Figure 5.9 for comparison.

For cross section A-A’, in the northeastern section of the ETSZ, the results of
XCORLOC and hypoDD look the most similar at depth. There is one shallow, steeply dipping
plane of seismicity, but also deeper planes at different angles. In the central cross section, the
relocated seismicity is perhaps the most difficult to interpret. HypoDD suggests a long plane of
seismicity at about 10 to 15 km depth, but XCORLOC has much more spread in the depths. Both
methods seem to suggest that the seismicity in this middle section may dip away from the NY-
AL lineament at greater depths. HypoDD again finds a plane of seismicity around 10 to 15 km
depth in the southwestern part of the ETSZ. While XCORLOC does not show this plane, both
methods show that seismicity in the southwestern part of the ETSZ are shallower in general than
the central and northeastern sections. There may also be one or two nearly vertical planes in this
section.

It is noteworthy that hypoDD relocates less events, especially not events which are not
well clustered with others, and therefore provides a more simplistic overview than the results of
XCORLOC. Earthquake depths are also not as constrained in their depths when relocated, so the
exact depths of features are not always well constrained as their epicenters. Although my
relocations reveal more linear features, there is no clear single one or two planes of seismicity
that would explain the majority of seismicity. Hence I argue that my results support the
hypothesis that earthquakes occur on small, reactivated faults rather than large, continuous ones
(Bollinger et al., 1991).
Figure 5.8 Comparison of earthquake depths using hypoDD and XCORLOC, with 3 cross-sections across the ETSZ, denoted A-A’, B-B’, and C-C’. (a-d): HypoDD relocations. (e-h) XCORLOC relocations. Vertical black dashed line in (b-d) and (f-h) denotes approximate crossing point of the NY-AL Lineament. Red dashed line in (b) and (f) shows the approximate depth to the decollement. Short red dashed lines represent interpreted planes and structures that are present in both hypoDD and XCORLOC relocations of detected seismicity.
Figure 5.9 Comparison of earthquake depths using (a-d): Original locations before relocation by XCORLOC and (e-h) Locations using XCORLOC, with 3 cross-sections across the ETSZ, denoted A-A’, B-B’, and C-C’. Vertical black dashed line in (b-d) and (f-h) denotes approximate crossing point of the NY-AL Lineament. Red dashed line in (b) and (f) shows the approximate depth to the decollement.
Many of the new earthquakes detected lie near the NY-AL Lineament, or slightly to the east, forming an even stronger “front” of seismicity along this feature, compared to the results using only catalogued seismicity. This provides stronger evidence that earthquakes are concentrating in areas correlated to the magnetic anomaly, and in particular the zone of high gradient change (which defines the NY-AL lineament).

Magnetic and gravity observations provide constraints on poorly understood lithospheric patterns of intraplate seismicity, as they can map large-scale contrasts in density related to the structural and geological properties of the region (von Frese et al., 2007; Lei et al., 2022). A magnetic anomaly refers to the change in magnitude of the Earth's magnetic field with respect to the expected value for a particular location. The anomaly is calculated as follows, with units of nanotesla (nT), or gamma:

\[
\text{Anomaly} = \frac{\text{Measured Value}}{} - \frac{\text{Expected Value}}{}
\]

As the magnetic anomaly reflects variation in geology, this could help explain the spatial distribution of earthquakes in the ETSZ. For example, large quantities of magnetic materials will increase the intensity of the Earth’s magnetic field. The magnetic anomaly can therefore delineate major tectonic evolutionary events of a region, although there is some controversy in the exact tectonic meaning (Lei et al., 2022). Figure 5.10 shows the original catalogued seismicity and the relocated XCORLOC seismicity using XCORLOC, overlaid with the global lithospheric magnetic field EMAG2 (Earth Magnetic Anomaly Grid). This dataset has a 2-arc-minute resolution which was compiled from satellite, ship, and airborne magnetic measurements (Maus, S. 2009). The magnetic anomaly data can be downloaded at:

Regardless of the exact significant and origin of the NY-AL Lineament, it is clear that earthquakes have the strongest concentrations directly to the east of the NY-AL Lineament and in the Ocoee block.
Figure 5.10 Total magnetic anomaly (nT) in the study area. Black dashed lines denote the New York–Alabama (NY-AL) and Clingman Lineaments. (a) and (c) show catalogued seismicity from January 1st, 2005, to July 31st, 2020 as black circles, in a large picture and zoomed-in version, respectively, while (b) and (d) show seismicity relocated by XCORLOC.

5.5 Magnitude Calculations

I compute magnitudes for many of the new detections in the inclusive catalog. There are 3,230 detections which meet the criteria for magnitude estimation using the method of Shelly et al. (2016), as shown in Figure 5.11. I use a minimum CC between template and detection waveform of 0.4 when computing the amplitude ratios. Lastly, I also impose a threshold to remove events with relatively large magnitudes and relatively low CC values, which is based on the expectation that a larger magnitude earthquake should be observed on a greater number of stations and components. A detected earthquake with magnitude, $M$, is accepted if it has at least $n$ components that meet all previous criteria for magnitude calculation (Zhai et al., 2021), according to Equation 5.2.

$$n = 3(M - 1)$$  \hspace{1cm} 5.2

Otherwise, I remove the event from the catalog. The value of $n$ is limited to a maximum of 15 components, corresponding to a minimum of 3 stations. For example, a magnitude 3 event would require 9 components and a magnitude 4 and above would require 15 components.
Figure 5.11 Magnitudes (Md) of templates (blue circles) and calculated magnitudes of detections (red circles) plotted as a function of time in between January 1, 2005 and July 31, 2020 in the ETSZ.

There are several detections which have moderate magnitudes, one of which is over 3. The detection with the largest magnitude is a Md 3.22 that occurred at 7/6/2020 at 3:30:38.1 UTC. This earthquake has clear signals with high SNR and as such, is observed on many stations and is likely to have a moderate magnitude. This earthquake was not a catalogued event, (as of August 1st 2020, when I first downloaded catalogued events). However, upon a new search in the ANSS catalog, this earthquake is now in the catalog with magnitude Md 2.5. It is likely that the earthquake was not yet in the catalog less than a month after it occurred, but has since been added. The template that detected it is a Md 2.38 that occurred on 7/1/2019.
The next largest magnitude detection is a Md 2.67 that occurred on 11/30/2016 at 11:44:53.4 UTC. It was detected by a Md 1.9 template from 4/14/2015, and is not in the initial catalog nor in the current-day ANSS ComCat catalog. These two detections are plotted below in Figure 5.12.

Figure 5.12 Waveforms of the template (red) are overlaid on top of continuous waveforms (black) around the time of detection for (a) the July 6, 2020 Md 3.22 and (b) November 30, 2016 Md 2.67 detections.
5.6 Magnitude of Completeness

There are 1,086 earthquakes in the original ANSS ComCat catalog between January 1st, 2005 and July 31st, 2020, in the ETSZ. However, there were missing events from this catalog, due to the difficulty of detecting events with low magnitudes. I calculate the Magnitude of Completeness (Mc) for the ETSZ using the Maximum Curvature Technique (MAXC) (Wiemer and Wyss, 2000). The Magnitude of Completeness Mc is a term used to describe the smallest earthquake magnitude that can be reliably detected and depends on the station coverage and detection sensitivity in a seismic zone.

The Gutenberg–Richter (GR) law states that earthquake magnitudes in a region are distributed exponentially by the relation in Equation 5.3,

\[
\log_{10} N(m) = a - bm
\]

where \( N(m) \) is the number of earthquakes having a magnitude greater than or equal to \( m \), and \( a \) and \( b \) are constants (Gutenberg and Richter 1944). The b-value is the slope of a fitted line for the power law. This statistic describes the relative proportion of small to large earthquakes in a region and is calculated only using earthquakes above the Mc.

For the 15-year catalog, the Gutenberg-Richter plot is shown in Figure 5.13a. The estimated Mc, using the MAXC technique, is 2.1 and the b-value is estimated at 1.48. The b-value of 1.48 using the catalogued seismicity initially suggests that the ETSZ has a lack of large magnitude earthquakes compared to small ones.

After detecting new earthquakes using the matched filter method, I can fill in the catalog with those detections for which there are calculated magnitudes. This results in 4,313
earthquakes with a much lower Mc of 0.8 (Figure 5.13b), which suggests that I detected most earthquakes of magnitude greater than 0.8, thus greatly filling in the catalog of ETSZ seismicity.

The updated b-value using my catalog of detections with magnitudes is 1.16, which is smaller than the original estimate of 1.48, but still above 1. This statistic suggests that the relative numbers of large and small magnitude earthquakes is about what is expected for a typical b-value of around 1, with a slight skew towards the ETSZ having a relative abundance of small magnitude earthquakes compared to large ones.

The typical b-value for earthquakes in tectonic regions is around 1, though this varies with region and time. Comparatively, studies of volcanic and induced seismicity typically result in higher b-values (McNutt, 2015). A high b-value is associated with high heterogeneity and weakness in the crustal rocks, which can reflect the geologic complexity of a region (Mogi, 1962, Montuori et al., 2010, Scholz, 2015). A high b-value can also indicate that a region has a relatively low level of stress accumulation (Mogi, 1962), as in the case of intraplate regions like the ETSZ.

There is also an association of higher b-values in areas with normal faults, intermediate b-values in strike-slip fault regions, and lower b-values in thrust fault regions (Schorlemmer et al., 2005). As the ETSZ predominantly hosts strike-slip earthquakes with some normal faulting, it is expected that a b-value slightly above 1 would be observed, and my b-value of 1.16 is in line with this model.

The only previous estimate of the b-value of the ETSZ was done by Bockholt et al. (2015), which used a dataset of 690 events of magnitude 0.3-4.6 between 2003 and 2014. They had a higher Mc of 1.3, and upon fitting events above this magnitude, found a b-value of 0.9.
Figure 5.13 (a) Gutenberg-Richter plot of catalogued seismicity in the ETSZ between January 2005 and July 31st 2020, using the rectangular bounding region in Figure 5.1. (b) Gutenberg-Richter plot of catalogued and detected seismicity that have magnitudes estimated. Vertical black line denotes magnitude of completeness (Mc).

5.7 ETAS De-clustering

Earthquakes tend to cluster temporally and spatially in sequences, for example in the often-observed mainshock-aftershock relationship. It is possible to separate events that occur independently of other earthquakes in a catalog, resulting in a declustered catalog. Independent, or “background” earthquakes are assumed to be mostly caused by tectonic loading or external stress transients that are not caused by previous earthquakes.

Comparatively, “triggered” earthquakes are triggered by factors such as static or dynamic stress changes, fluid changes due to previous earthquakes, and after-slip (Prejean and Hill, 2009; Hardebeck and Okada, 2018). To avoid biasing the temporal distribution of seismicity, the catalog is declustered. This allows for a more reliable assessment of the spatio-temporal
earthquake and background properties, which can lead to greater earthquake hazard assessment for a region.

One way earthquake catalogs are typically declustered is by the Epidemic-Type Aftershock Sequence (ETAS) model, which is based on the concept that every earthquake can trigger aftershocks, regardless of its magnitude (Ogata, 1998). The ETAS model assumes that an earthquake sequence contains aftershocks, which are “triggered” by other earthquakes, and “background” earthquakes, which occur independently of other events.

I can further refine an apparent trend in seismicity by including only background earthquakes that are statistically independent of others, and exclude triggered earthquakes. I use earthquakes that are above the magnitude of completeness, $M_c$ of 0.8, as input to the ETAS model and stochastic declustering algorithm (Zhuang, 2006).

There are 3,094 events which have magnitudes above the $M_c$ of 0.8, including catalogued events and detections. However, not all of these events are relocated by XCORLOC. Therefore, for those detections which do not have relocations, I use a random perturbation between -0.04 and 0.04 degrees in latitude and longitude compared to the location of the template that detected the event. This shift is necessary because the ETAS program does not accept a catalog in which some earthquakes have identical location. The average epicentral distance between detections with magnitudes and the templates that detected them is 2.7 km. The average absolute difference in latitude is 0.017 degrees, and 0.0174 degrees for longitude. Therefore, I choose a random absolute perturbation of approximately twice this value, at 0.04 degrees, for both latitude and longitude. The distributions of differences in epicentral distance, latitude, and longitude used to empirically determine a reasonable location perturbation are shown in Figure 5.14.
Figure 5.14 Distributions of differences in (a) epicentral distance, (b) latitude, and (c) longitude between detections and templates that detected them. The averages of the latitude and longitude differences are used to empirically determine a reasonable location perturbation for detections whose relocations were not determined.

The ETAS package determines that there are 1,166 background events and 1,928 triggered events in the study region. Figure 15.15 depicts a spatial and temporal perspective of the resulting background and triggered earthquakes, showing that several clusters of seismicity in time and space are triggered by other earthquakes.
Figure 5.15 Locations and times of earthquakes in the ETSZ, above the Mc of 0.8. Locations for earthquakes whose locations were not constrained by relocation are assigned a location equal to that of the template that detected it, plus an epicentral perturbation. (a) Epicentral locations. Blue circles represent background events, while red circles denote triggered earthquakes. (b)
Epicentral latitudes against earthquake times.

5.8 Hydrologically Driven Seismicity in the ETSZ

5.8.1 Reservoir Induced Seismicity (RIS) around the Watts Bar Reservoir

The December 12, 2018 Mw 4.4 mainshock occurred 4.6 km from the nearby Watts Bar Dam. The relative locations of the mainshock, dam, and nuclear power plant are shown in Figure 5.16a. Seismicity around the reservoir indicates that there have been several earthquakes near the reservoir over the last 15 years. Due to the proximity of the mainshock to the Watts Bar Reservoir, I examine the mainshock and compare it with other earthquakes that occurred within 30 km of the dam in order to evaluate whether this may be a case of reservoir induced seismicity (RIS) (Simpson et al., 1988; McGarr et al., 2002).

One situation under which RIS occurs is related to the initial filling of a reservoir, resulting in an increased stress due to loading, or a delayed increase in pore pressure, which decreases the effective normal stress on a fault (Talwani, 1997; Simpson et al., 1988). As the Watts Bar Reservoir was completed in 1942, there is likely no effect between the initial loading and present-day seismicity. A second situation which accounts for RIS is seasonal changes in water level, wherein a delayed increase in pore pressure at depth triggers earthquakes after the seasonal maxima of water level has been attained (Michas et al., 2020). An increase in seismicity after reservoir loading can also be associated with a response of the fluid saturated ground to the additional weight of the water (Smirnov et al., 2017). The Mw 4.4 mainshock occurred about a month and a half after the reservoir level started to drop in 2018, which may be consistent with the model of delayed pore fluid pressure diffusion (Michas et al., 2020).
In order to analyze whether there is a significant relationship between reservoir water drawdown and seismicity surrounding the reservoir, I plot in Figure 5.16b the weekly water level and number of earthquakes, averaged over 15 years between January 1st 2005 and July 31st 2020. I plot my catalog of all catalogue and detected seismicity which have magnitudes, as well as this same catalog above the Mc of 0.8, and also only background events.
Figure 5.16 (a) There are 67 background events (blue circles) within 30 km of the Watts Bar Dam, as determined by ETAS de-clustering using the method of Zhuang (2006). Red circles represent triggered events, which are not used in analyzing potential RIS. Note that the locations of some earthquakes are not by relocation, but by random perturbation compared to the template
location that detected the event, as input to ETAS. The black square is the location of the Watts Bar Dam. The black diamond is the Watts Bar Nuclear Power Plant. The Blue star is the approximate location of the 2018 Mw 4.4 earthquake. The large black circle shows the boundary region for the 30 km area around the dam. (b) Black line shows the weekly dam head water elevation, in feet, at the Watts Bar Reservoir, averaged over January 1, 2005 to July 31, 2020, for the 53 calendar weeks of a year. The other lines show earthquakes within 30 km of the Watts Bar Dam. The blue line shows the average weekly number of catalogued earthquakes and detections with magnitudes (210). The green line shows catalogued earthquakes and detections with magnitudes that are above the Mc of 0.8 (120). The red line shows just background earthquakes (63).

At the Watts Bar Reservoir, the TVA keeps the water level high during summer months for recreation, then draws down the water level in preparation for the management of increased rainfall in the winter months. Looking at the catalogued and detected seismicity, there is a strong trend of increased seismicity in the weeks following decreasing water levels in fall and winter, as well as some increased seismicity in the weeks after the reservoir level increases in spring. As mentioned before, some earthquakes occur with a delay of several weeks after the loading or unloading of the reservoir (McGarr et al., 2002).

It appears that there is increased seismicity with a delay after reservoir water level loading or unloading, when looking at all seismicity. This may suggest that RIS does play a role in some of the seismicity around the Watts Bar Reservoir, including the Mw 4.4 mainshock.

However, it is also possible to look at background events as determined by ETAS, to exclude triggered earthquakes near the reservoir. Upon removing seismicity in the vicinity of the
reservoir that are not background events, there is not as strong an increase in seismic activity around loading or unloading of the reservoir. This does not preclude the possibility of the Mw 4.4 mainshock being a case of RIS, especially as it was an unusually shallow earthquake (5 to 6 km) for this study region. The declustered background seismicity may not be the most reliable dataset to examine in determining the possibility RIS, however, as parameter choices for ETAS and the small sample size of earthquakes near the reservoir may not yield consistent results.

Alternately, instead of examining stacked monthly earthquake rates averaged over the 15-year study, it is also helpful to examine changes throughout the 15-year period for patterns exhibited by earthquakes within 25 km of the reservoir. Figure 5.17a shows the seismic rate as a function of time, for both the full catalog of events near the reservoir and those above the Mc. The data shows several clear increases in the seismic rate either following loading or unloading of the reservoir, especially for the full catalog. It appears that many of the potentially reservoir induced earthquakes are below the Mc of 0.8, which would suggest that small magnitude events are most easily triggered by changes in the reservoir level.

Figure 5.17b shows the cumulative seismic moment of events within 25 km of the reservoir. The seismic moment for each earthquake is calculated using the empirical relation between duration magnitude (Md) and seismic moment (Mo), which is derived in Bora (2016):

$$\log(Mo) = 1.39Md + 9.54$$

The cumulative seismic moment curve is dominated by the December 12, 2018 Mw 4.4 earthquake. However, it is also possible to see other years in which an increase in the seismic moment occurred following loading, and especially, unloading of the reservoir. This pattern
suggests that the Watts Bar Reservoir plays a role in triggering some of the nearby seismicity over the 15-year study period.

Figure 5.17 (a) Hourly (blue line) and weekly (orange line) water level height at the Watts Bar Reservoir over the 15-year study period, compared to the weekly seismic rate of all events within 25 km of the reservoir (black line) and events above the Mc within 25 km of the reservoir (green line). (b) Hourly (blue line) and weekly (orange line) water level height at the Watts Bar Reservoir over the 15-year study period, compared to the cumulative seismic moment of events.
within 25 km of the reservoir (black line).

5.8.2 Seasonal surface water changes across the ETSZ

The lithosphere is continuously in a state of deformation as surface or subsurface hydrologic processes change surface loads, change surface stresses, and cause variations in fluid pore pressure (Costain and Bollinger, 2010; Johnson et al., 2017). Crustal stresses from surface processes, including rainfall and groundwater, has been observed to modulate seismicity rates in certain regions with seasonal patterns (Perry and Bendick, 2021; Craig et al., 2017; Johnson et al., 2017; Hsu et al., 2021; Montgomery-Brown et al., 2019). A previous study that examined river discharge, precipitation, and groundwater well level data found only minimal to moderate annual and long-term influences of hydrological processes on seismicity in the ETSZ (Cameron, et al., 2016).

Here I search for a significant variation in the rate of earthquakes in the ETSZ at annual and multi-annual timescale to determine if there is a pattern that coincides with hydrological loading in this region, similar to Craig et al. (2017). Figure 5.18 shows the distribution of seismicity in the ETSZ by month, to better examine whether there is a seasonal pattern to the events. When using all events in the inclusive catalog of 13,671 that meet the minimum detection criteria listed in previous sections, there appear to be two large increases in the rate of seismicity, during the months of June and November. However, when only using events above the Mc, with magnitudes calculated, after applying ETAS declustering, it is clear that there is a fairly uniform distribution of background events by month, which does not suggest modulation by seasonal rainfall changes.
Figure 5.18 Comparison of different catalogs revealing monthly changes in seismicity rate, plotted as a fraction of events that fall within that month. Blue line is the entire catalog of 13,671 catalogued and detected events. Orange line is same catalog, but only detections with magnitudes, for a total of 4,197 events. Green line is catalogued events and detections with magnitudes that are above the Mc of 0.8, for 3,094 events. Red line is the catalog of 1,166 background events (also above the Mc).

It is also possible to look at earthquakes in the ETSZ that have depths of less than approximately 10 km. If there is a modulation of seismicity by seasonal rainfall changes, I would expect the strongest changes in seismicity to occur in the shallow portion of the seismic zone, where fluids may migrate down to, or the influence of poroelastic stress change may reach to. Figure 5.19 shows the results on this new dataset. When only looking at the background events in the top 10 km, there is a slightly pronounced higher rate of seismicity in the month of November, and a minimum in March. In eastern Tennessee, the highest rainfall is in March, and
the driest month is in October, one month before November. Therefore, there may be some minor modulation of seismicity in the shallowest part of the ETSZ from seasonal rainfall variations.

Figure 5.19 Same as 5.12, but only for events shallower than 10 km. Blue line is the entire catalog of 7,782 catalogued and detected events, shallower than 10 km. Orange line is same catalog, but only detections with magnitudes, for a total of 2,299 events. Green line is catalogued events and detections with magnitudes that are above the Mc of 0.8, for 1,580 events. Red line is the catalog of 424 background events (also above the Mc).

To further examine shallow seismicity with depths of less than 10 km in the ETSZ and its potential modulation by hydrology, I plot the seismic rate against the mean groundwater level in the ETSZ. I use the average of 4 different groundwater stations with available continuous data in the ETSZ (Sv:E-002, Hm:O-019, 03PP01, and CE-029), the data for which can be accessed at https://nwis.waterdata.usgs.gov/tn/nwis/gwlevels. The largest data gap is on the order of a few
weeks, and missing data is linearly interpolated, similar to Hsu et al., 2021. The weekly average groundwater level, measured as the depth to the water level in feet below the surface, is plotted in Figure 5.20a. A higher depth to the groundwater level implies a period of drought, while lower depth to the groundwater level implies the ground is more fully saturated. During a time of low depth to the groundwater level, faults would be expected to be more lubricated with less friction and a higher potential for slip (Montgomery-Brown et al., 2019).

The seismic rate of all shallow earthquakes as well as for those only above the Mc do not appear to show a clear correlation with groundwater level (Figure 5.20a). Weekly earthquake rates, plotted as a function of mean weekly groundwater level, also do not show a peak at low groundwater levels, which would correspond to more earthquakes occurring during times of higher groundwater saturation (Figure 5.20b). Likewise, when seismic rates are correlated with the groundwater levels, there is no clear peak in the cross-correlation function which would indicate a time lag of several days to weeks between earthquakes and groundwater levels (Figure 5.20c). The normalized cross-correlation coefficients are less than 0.05, further indicating that there is not a clear delay between seismic rates and groundwater level changes for shallow earthquakes in the ETSZ.
Figure 5.20 Comparison of groundwater level and shallow (less than 10 km depth) seismicity in the ETSZ. (a) Weekly earthquake counts of all ETSZ seismicity (black line), ETSZ seismicity above the Mc (green line) and mean ground water level in the ETSZ (blue line). (b) Earthquake rates, plotted as a function of mean weekly groundwater level. Blue line uses all events, while orange line only considers those above the Mc. (c) Cross correlation coefficient versus time lag between the earthquakes and groundwater level. Positive time lag indicates earthquakes lag groundwater level, while negative time lag indicates groundwater level lags earthquakes. Blue line uses all events, while orange line only considers those above the Mc.
Another method of searching for periodicity in the declustered earthquake catalog is to examine the Schuster spectra (Ader and Avouc, 2013). The Schuster spectra is built on the Schuster test (Schuster, 1897) and tests whether a catalog contains intervals between earthquakes that are more common than expected by random chance. I only use events in the top 10 km, as before, and earthquakes classified as background events by ETAS declustering.

Figure 5.21 shows the results of the Schuster spectra. The results show a single peak above the 99% interval, although it is barely above this level. This peak occurs at approximately 0.50383 years, or 183.9 days. This corresponds to a half-year periodicity. Although November experiences the highest seismicity rate, June and February follow closely behind. The time between the peaks sometime during June and November may contribute to a half-year periodicity. The results of the Schuster spectra therefore do not offer any new support for a yearly, or sub-yearly periodicity related to rainfall in the ETSZ.

Figure 5.21 Schuster spectra using background events in the ETSZ (January 1, 2005 to July 31, 2020) located at less than 10 km depth.
5.9 Conclusions

I detected microseismicity in the ETSZ using over 15 years of continuous data, yielding the most detailed complete catalog yet for this seismic zone, as well as magnitude estimations and more defined structure. I found the greatest concentration along or to the east of the NY-AL Lineament, as defined by the magnetic anomaly, supporting the evidence that this feature’s origin is linked to seismicity in the ETSZ. I also examined seismicity in the vicinity of the Watts Bar Reservoir, which is located about 5 km from the epicenter of the Mw 4.4 December 12, 2018 Decatur, Tn earthquake, to search for Reservoir Induced Seismicity (RIS). I find evidence for triggering of earthquakes within 25 km of the reservoir following water level changes in the reservoir. I also examined seismicity in the entire ETSZ to search for a correlation between shallow earthquakes and seasonal hydrologic changes. The groundwater level does not appear to modulate much of shallow seismicity in the ETSZ, suggesting a different driving mechanism for most shallow seismicity than hydrologic cycles.
CHAPTER 6. Middleton Place-Summerville Seismic Zone (MPSSZ)

In the MPSSZ, near Summerville, South Carolina, I focus on detecting and relocating previously uncatalogued earthquakes near the source area of the M~7 1886 earthquake. I take advantage of a recent temporary seismic deployment and initial detections in 2011-2012 on which I run matched filter detection. The locations of newly relocated seismicity are compared to a previous study which hypothesized an extrapolated westward-dipping major fault plane. These results will be combined with future efforts by the Georgia Tech group, which will further analyze an ongoing temporary deployment during 2021-2023 in the same region.

6.1 Introduction

On August 31, 1886, a large M~7 earthquake near Summerville, South Carolina caused massive destruction in the city of Charleston, South Carolina, in what is known as the Middleton Place-Summerville Seismic Zone (MPSSZ). This was the largest historical earthquake on the east coast of the United States (Neely et al., 2018). Magnitude estimates range from a Mw 6.9 (Bakun and Hopper, 2004) to Mw 7.3 (Johnston, 1996). The strongest shaking was felt about 25 km northwest of Charleston and in areas to the east and south (Duton, 1889). The lack of obvious surface rupture at the time of the mainshock and the lack of modern geophysical instrumentation has left few indications of the location, orientation, and fault type of this M~7 earthquake.

To this day, aftershocks of the M~7 mainshock are felt in the area around Summerville, South Carolina. It is thought that seismicity occurs along the fault, or fault structure, that is responsible for the M~7 1886 earthquake (Chapman et al., 2016; Madabhushi and Talwani, 1993), therefore presenting an opportunity to better locate the fault for future seismic hazard
analysis. A previous deployment during 2011-2012 provides the station coverage needed to reliably detect and relocate present-day seismicity and is described in greater detail in the next section (Chapman and Beale, 2011).

Previous studies have attempted to determine the fault structure in the MPSSZ. Madabhushi and Talwani, 1993, suggest that the Ashley River Fault (ARF) is not a planar feature, but is comprised of short segments with varying strikes and dips, but generally striking NW and dipping SW. Chapman et al. (2016) found that much of current-day seismicity occurs on a west-dipping, NNE-SSW-trending reverse fault. Cramer et al. (2020) compared predicted ground motion accelerations for three different potential source models to macroseismic observations from 1886 (Dutton, 1889) and found a better correlation with strike-slip source models (e.g., Dura-Gomez and Talwani, 2009) than a westward dipping source.

6.1.1 The 2011-2012 XY Network Deployment

A temporary deployment of 8 short-period stations was carried out starting on 7 August 2011, and ending on 30 August, 2012 (Chapman and Beale, 2011). The deployment lasted about a year, although several stations were relocated early to record aftershocks of the 23 August 2011 Mineral, Virginia earthquake. The data is archived at the IRIS Data Center under seismic network code XY. The network resulted in the detection of 134 earthquakes with magnitudes of Md -1.8 to 2.6 in the vicinity of Summerville, South Carolina (Chapman and Beale, 2011). A subsequent study by Chapman et al., 2016 used both the temporary stations and nearby permanent stations for a total of 12 stations during the period of the temporary deployment, shown in Figure 6.1. The 134 earthquakes were input to relocation using hypoDD. Initial locations were found using a single-event location program, HYPOELLIPSE (Lahr, 1999). 123
earthquakes were relocated using hypoDD. The greatest concentration of recent seismicity is along the Ashley River valley.

Figure 6.1 Study region of the 2011-2012 XY temporary deployment. Temporary stations are indicated as triangles, and permanent stations are drawn as inverted triangles. Earthquakes detected during this time as shown as green circles. Source: Chapman et al., 2016.

Chapman et al., 2016 suggests that there is a south-striking, west-dipping seismic zone that the seismicity delineates, as shown in Figure 6.2. However, there are not many earthquakes that lie in the deeper part of this inferred plane. I wish to detect more earthquakes with the hope
that some of the newly detected earthquakes will show a clear pattern at the deeper (>10 km) depths, to either support or challenge this inferred plane.

Figure 6.2 Locations of earthquakes (green circles) during 2011-2012 detected in Chapman et al., 2011. Upright triangles are the 2011-2012 temporary seismic stations, and inverted triangles are permanent stations. (a) Initial epicenters, (b) hypocenters located using hypoDD in Chapman et al., 2011, (c) initial hypocenters projected onto vertical profile D-D’, and (d) hypocenters located using hypoDD projected onto vertical profile D-D’. Dashed line delineates the inferred south-striking, west-dipping seismic zone. Taken from Chapman et al., 2016.

6.2 Seismic Data and Pre-Processing

To create a comprehensive catalog of earthquakes in the MPSSZ around the Summerville area between August 7, 2011 and August 30, 2012, I use catalogued events from 2 sources: the
123 relocated earthquakes from Chapman et al., 2016, and those listed in the USGS Comprehensive Earthquake catalog (ComCat).

I select earthquakes in the ComCat catalog that occur between August 7, 2011 and August 30, 2012, and are located within the study region box of Figure 6.3. There are 10 events that meet these criteria. This makes the total earthquakes in my starting catalog equal to 133.

![Figure 6.3 Cumulative distribution of catalogued events (blue line) and template events (orange line) with time.](image)

We originally had our IRIS intern from summer 2021, Mandy Jackson, try EQTransformer to detect new events that I could use as templates. However, the results were not very good. Of the detections made, the false detection rate was very high, and there were not many picks made for actual earthquakes.

To better understand whether there aren’t many new events to detect, or whether the deep learning method wasn’t working well, Mandy ran EQTransformer around the times of the catalogued Chapman earthquakes, from 1 hour before to 1 hour after each catalogued event. At a
lower threshold than is typical (20% likelihood of being an earthquake instead of 30%), the results only detected 56.7% of the catalogued earthquakes, while simultaneously detecting many events which do not look like earthquakes in these 2-hour windows around the catalogued events, suggesting the method does not work well at this study region (Figure 6.4). Therefore, these results are not used in this study.

Figure 6.4 Performance of catalogued event detection using EQTransformer, at two different thresholds: 0.3 and 0.2. Credit: Mandy Jackson
I use the 133 ComCat and Chapman catalogued earthquakes to create templates for a matched filter detection. A plot of the cumulative catalogued seismicity with time is shown in Figure 6.3.

I cut the waveforms 0.5 seconds before a P or S arrival time to 3 seconds after the arrival time. Each earthquake must be recorded on at least 3 stations with a minimum signal to noise ratio (SNR) of 5 on the waveforms filtered from 0.5 to 30 Hz, using a noise window of 3.5 seconds before phase picks. Waveforms are sampled at 100 Hz, as this was the lowest common denominator of the networks, including the temporary XY network as well as stations which are part of the permanent NM and CO networks. These criteria result in 127 templates, which are shown in Figure 6.3.

6.2.1 Velocity Models

Chapman et al., 2016 used a velocity model based on Chapmen et al., 2003 in order to relocate the 123 earthquakes with hypoDD. I use the same velocity model as input to hypoDD, as shown in Figure 6.5. I adapt this velocity model to that of a gradient model, as XCORLOC does not accept a “cake layer” model as input.
Figure 6.5 Velocity models for Charleston study. The input velocity model to hypoDD is taken from Chapman et al., 2016, using a Vp/Vs ratio of 1.732. The input velocity model to XCORLOC is an adapted gradient version of the same model.

6.3 Matched Filter Detection and Refinement

There are **11,719** raw matched filter Detections at 9*MAD, however this number includes duplicate detections. After requiring a preliminary minimum CC of 0 between waveforms of the detection with waveforms of the template that detected it, on at least 2 components, and removing duplicate detections within 5 seconds of one another, there are **6,125** candidate detections. However, many of these detections are of low quality, or the earthquake signals cannot be clearly seen.

Therefore, I next require that detection to have a minimum CC of 0.3 between template and detection waveforms for P arrivals and 0.4 for S arrivals. Figure 6.6 shows the distribution of CC values between waveforms of detections and the templates that detected the events, in order
to determine appropriate CC thresholds. The CC of 0.3 for P arrivals and 0.4 for S arrivals was determined to be the best, empirically, for keeping high and medium quality arrivals while rejecting low quality and false arrivals.

Figure 6.6 Distribution of CC values between waveforms of detections and templates that detected the events. Used in determining CC threshold for P and S arrivals between detection and template waveforms that detected them, for removing false or low-quality detections in the MPSSZ study.

Arrivals must have a minimum SNR of 3 on at least two stations, with at least one station that has a P-S pair. This reduces the chance that the catalog contains detections which are anthropogenic noise, or local phenomena that only one station recorded. This catalog contains 269 events, including 127 templates and 142 detections. Figure 6.7 is an example of such a detection, which is not in the catalog of Chapman or ANSS earthquakes.
Figure 6.7 Example of a detection of an earthquake that was not in the catalog previously using the matched filter method.

6.4 Earthquake Relocation

I apply two separate relocation packages, XCORLOC (Lin, 2018) and hypoDD (Waldhauser, 2001), to relocate seismicity in the MPSSZ so as to compare the relocation results and check for robustness of the solution.

6.4.1 XCORLOC Relocation

The input parameters to XCORLOC are given in Table S6.1. Of the 269 events input to XCORLOC relocation, 186 are relocated (127 templates and 59 detections). The median estimated horizontal uncertainty is 0.13 km, and the median estimated vertical uncertainty is 0.2 km. The results on a map view are plotted on Figure 6.8.
Figure 6.8. Earthquakes relocated using XCORLOC. (a) Original locations of earthquakes relocated. (b) Locations of earthquakes relocated using XCORLOC (a) and (b) show templates (blue circles) and detections (orange circles).

6.4.2 HypoDD Relocation

The input parameters to hypoDD relocations are given in Table S6.2. No individual relocation of clusters and adjustment of the DAMP parameter is needed because there is only 1 cluster of seismicity that hypoDD determines. Of the 269 events input to hypoDD relocation, 214 are relocated (126 templates and 88 detections). The median estimated horizontal uncertainty is 1.2 km, and the median estimated vertical uncertainty is 0.8 km. The results on a map view are plotted on Figure 6.9.
Figure 6.9 Earthquakes relocated using hypoDD. (a) Original locations of earthquakes relocated. (b) Locations of earthquakes relocated using hypoDD (a) and (b) show templates (blue circles) and detections (orange circles).

6.5 Comparison of Relocation Results with Chapman et al., (2016)

XCORLOC does not perform as well as hypoDD in this case, as there seems to be more spread in the earthquake relocations. Aside from differences in the method, XCORLOC also has an input velocity model which is adapted from the best known velocity model for the region. This introduces some bias in the relocation.

I can compare both of these solutions with the original relocated earthquakes from Chapman et al., 2016, as shown in Figures 6.10 and 6.11 below.
Figure 6.10 N/NE to S/SW cross-section comparison of different relocated seismicity datasets. The black line labeled ‘A’ to ‘B’ is the line along which the cross sections are made. Black dots are locations of earthquakes that occurred during the 2011-2012 deployment: (a) the 126 events relocated by hypoDD from Chapman et al., 2016 as well as 10 catalogued ANSS ComCat events (b) 186 earthquakes (templates & detections from matched filter) relocated using XCORLOC (c) 214 earthquakes (templates & detections from matched filter) relocated using hypoDD.
Figure 6.11 W/NW to E/SE cross-section comparison of different relocated seismicity datasets. The black line labeled ‘A’ to ‘B’ is the line along which the cross sections are made. Black dots are locations of earthquakes that occurred during the 2011-2012 deployment: (a) the 126 events relocated by hypoDD from Chapman et al., 2016 as well as 10 catalogued ANSS ComCat events (b) 186 earthquakes (templates & detections from matched filter) relocated using hypoDD (c) 214 earthquakes (templates & detections from matched filter) relocated using hypoDD. The red dashed lines depict my interpreted fault planes for the seismicity at depth (>10 km).

As with the results of Chapman et al., 2016, I also observe a gap in seismicity between the shallower hypocenters (<7 km) and the deeper hypocenters (>10 km). There are one or two clear lines of seismicity when viewing the deeper section of the W/NW to E/SE cross-section. Two events relocated in the deep clusters are chosen randomly and their waveforms are plotted in Figure 6.12. Although the new detection chosen at random does seem to contain earthquake signals, it only meets the minimum criteria to be relocated. This suggests that the seismicity in these deep clusters may contain some smaller magnitude events, which do not have as constrained a solution for their locations.
Figure 6.12 (a) Earthquakes relocated using hypoDD. (left) Original locations of earthquakes relocated. (right) Locations of earthquakes relocated using hypoDD. Stars denote locations of events chosen in deep cluster of seismicity. Red star shows the epicentral location of event plotted in (b) relocated in a deep cluster. Green star shows the epicentral location of event plotted in (c) relocated in a deep cluster. This event is a template, and not a new detection.

XCORLOC shows one nearly horizontal line of seismicity, while hypoDD shows two lines of seismicity that are somewhat horizontal, one dipping roughly to the west, the other to the east. However, these trends are not clearly in line with the inferred south-striking, west-dipping seismic zone of Chapman et al., 2016, suggesting the fault structure may be more complex.

6.6 Magnitude Calculations

Of the 127 templates I used as input to matched filter detection, 117 were events from Chapman et al., 2016 and are not listed in the ANSS ComCat catalog. There are no associated magnitudes with these 117 earthquakes. In order to calculate the magnitudes of the new earthquakes from matched filter detection, I need to calculate the magnitudes of the templates that detected them.

I use the method of Shelly et al., 2014 to calculate magnitudes, with the same requirements as for my study of 15 years of continuous data in the ETSZ. Namely, I use a minimum CC between waveforms of 0.4 when computing the amplitude ratios. I also impose a threshold to remove events with relatively large magnitudes and relatively low CC values, which is based on the expectation that a larger magnitude earthquake should be observed on a greater number of stations and components. A detected earthquake with magnitude, $M$, is accepted if it
has at least \( n \) component components that meet all previous criteria for magnitude calculation (Zhai et al., 2021), according to Equation 6.1.

\[
n = 3^{(M-1)} \tag{6.1}
\]

Otherwise, I remove the event from the catalog. The value of \( n \) is limited to a maximum of 15 components, corresponding to a minimum of 3 stations. For example, a magnitude 3 event would require 9 components and a magnitude 4 and above would require 15 components.

For each Chapman earthquake, I sort the 10 ANSS ComCat templates in order of distance to the earthquake. It is a reasonable assumption that the closest ANSS earthquake to the Chapman event will have the most similar waveforms. However, it is possible that the closest earthquake is small and does not have many phase arrivals, or has a different focal mechanism, reducing the CC between waveforms. Therefore, I iteratively attempt to compute the relative magnitude of the Chapman event, using the closest ANSS template first, and continuing through the sorted list of the 10 ANSS events until the magnitude is calculated. Using this method, I calculate the magnitudes of 63 of the 117 Chapman events, giving a total of 73 templates with magnitudes (Figure 6.13).

Then, I calculate the magnitudes of newly detected earthquakes, also using the method of Shelly et al., 2014. This results in the magnitudes of 44 newly detected events (Figure 6.13). my results show a relatively even distribution of earthquakes with time between 7 August, 2011 and 30 August, 2012. This suggests that the time period of the temporary deployment coincided with background seismicity and that no clear earthquake sequences occurred in this time frame.
Figure 6.13 Magnitudes (Md) of ANSS ComCat templates (solid blue circles), calculated earthquakes in the Chapman et al., 2016 catalog (open blue circles) and calculated magnitudes of detections (open red circles) plotted as a function of time in between 7 August, 2011 and 30 August, 2012 in the MPSSZ near Summerville, SC.

6.7 Conclusions

In the MPSSZ, near Summerville, South Carolina, I focus on detecting and relocating previously uncatalogued earthquakes near the source area of the M~7 1886 earthquake. I take advantage of a recent temporary seismic deployment and initial detections in 2011-2012 on which I run matched filter detection. The locations of newly relocated seismicity are compared to a previous study which hypothesized an extrapolated westward-dipping major fault plane. I found deep clusters and linear features which are not clearly in line with the currently hypothesized extrapolation of the major fault plane in this region at depth. There is also a section between the deep and shallow clusters which does not exhibit seismicity during the study period.
This has implications for the fault structures which are responsible for the 1886 M-7 Summerville, SC earthquake to be compared with future studies.
Appendices

Supplementary Tables

Table S3.1 Event detections of the 2014 Mw 4.1 Edgefield, South Carolina earthquake study listing origin time (UTC), estimated magnitude, and mean cross-correlation value (cc).

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Table S4.1 (Included as a text file, located at http://geophysics.eas.gatech.edu/zpeng/Catalog/CDaniels_PhD_2022_Table_4.1.txt) Catalog of templates and detections (before relocation) during the 8 weeks around the mainshock recorded on a minimum of 3 stations. Columns are: event number, origin time (UTC), relocated latitude, relocated longitude, relocated depth, estimated (or catalogued) magnitude (Md), origin time of template that detected the event, original latitude of template that detected the event, original longitude of template that detected the event, original depth of template that detected the event, Md magnitude of template that detected the event.

Table S4.2 XCORLOC Input Parameters for 8 week study around the 2018 Mw 4.4 Decatur, Tennessee mainshock.
Table S4.3 HypoDD Input Parameters for 8 week study around the 2018 Mw 4.4 Decatur, Tennessee mainshock.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Data Type</td>
<td>CC</td>
</tr>
<tr>
<td>Phase Type</td>
<td>P and S</td>
</tr>
<tr>
<td>DIST</td>
<td>400 km</td>
</tr>
<tr>
<td>min # of obs/pair for cc data</td>
<td>4</td>
</tr>
<tr>
<td>Inversion Type</td>
<td>LSQR</td>
</tr>
<tr>
<td>Sets of Iterations</td>
<td>2</td>
</tr>
<tr>
<td>DAMP</td>
<td>Depends on cluster</td>
</tr>
<tr>
<td>Number of iterations (first set)</td>
<td>5</td>
</tr>
<tr>
<td>P cc weight</td>
<td>0.01</td>
</tr>
<tr>
<td>S cc weight</td>
<td>0.01</td>
</tr>
<tr>
<td>cc residual threshold (as factor of s.d.)</td>
<td>N/A</td>
</tr>
<tr>
<td>Max dist between cc linked pairs</td>
<td>N/A</td>
</tr>
<tr>
<td>P phase weight</td>
<td>1</td>
</tr>
<tr>
<td>S phase weight</td>
<td>1</td>
</tr>
<tr>
<td>Phase residual threshold (as factor of s.d.)</td>
<td>6</td>
</tr>
<tr>
<td>Max dist between phase linked pairs</td>
<td>10</td>
</tr>
<tr>
<td>Number of iterations (second set)</td>
<td>5</td>
</tr>
</tbody>
</table>
Table S4.4 Catalog of templates and detections during the 8 weeks around the mainshock that have Md magnitudes calculated, and have relocated latitudes, longitudes, and depths calculated using XCORLOC.

<table>
<thead>
<tr>
<th>Event Origin Time (UTC)</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth (km)</th>
<th>Magnitude (Md)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2018-12-12T09:14:43.316000Z</td>
<td>35.599</td>
<td>-84.737</td>
<td>4.04</td>
<td>3.88</td>
</tr>
<tr>
<td>2018-12-12T09:27:07.929000Z</td>
<td>35.603</td>
<td>-84.729</td>
<td>4.29</td>
<td>2.96</td>
</tr>
<tr>
<td>2018-12-12T10:41:26.906000Z</td>
<td>35.603</td>
<td>-84.73</td>
<td>4.51</td>
<td>2.1</td>
</tr>
<tr>
<td>2018-12-12T14:17:02.630000Z</td>
<td>35.603</td>
<td>-84.727</td>
<td>4.62</td>
<td>1.4</td>
</tr>
<tr>
<td>2018-12-13T06:09:31.253000Z</td>
<td>35.603</td>
<td>-84.727</td>
<td>4.51</td>
<td>1.22</td>
</tr>
<tr>
<td>2018-12-14T00:24:06.187000Z</td>
<td>35.602</td>
<td>-84.731</td>
<td>4.38</td>
<td>2.47</td>
</tr>
<tr>
<td>2018-12-14T01:28:20.098000Z</td>
<td>35.603</td>
<td>-84.727</td>
<td>4.51</td>
<td>1.78</td>
</tr>
<tr>
<td>2018-12-18T20:02:49.939000Z</td>
<td>35.603</td>
<td>-84.734</td>
<td>4.29</td>
<td>2.06</td>
</tr>
<tr>
<td>2018-12-19T18:11:02.527000Z</td>
<td>35.606</td>
<td>-84.717</td>
<td>5.09</td>
<td>2.12</td>
</tr>
<tr>
<td>2018-12-24T06:24:34.323000Z</td>
<td>35.607</td>
<td>-84.718</td>
<td>4.49</td>
<td>1.45</td>
</tr>
</tbody>
</table>
Table S4.5 Nodal Plane solutions for the 4 events for which HASHpy returned focal mechanisms.

<table>
<thead>
<tr>
<th>Origin Time (UTC)</th>
<th>Magnitude (Md)</th>
<th>Strike (°)</th>
<th>Dip (°)</th>
<th>Rake (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>12/12/2018 9:14:43.61</td>
<td>3.88 (mainshock)</td>
<td>352.1</td>
<td>69.4</td>
<td>175.6</td>
</tr>
<tr>
<td>6/10/2018 11:51:57.85</td>
<td>2.65</td>
<td>312.4</td>
<td>89.1</td>
<td>-138.1</td>
</tr>
<tr>
<td>12/12/2018 9:27:07.94</td>
<td>2.96 (aftershock)</td>
<td>159</td>
<td>75.7</td>
<td>-170.4</td>
</tr>
<tr>
<td>12/14/2018 0:24:06.17</td>
<td>2.47</td>
<td>10.6</td>
<td>76.5</td>
<td>161.6</td>
</tr>
</tbody>
</table>

Table S5.1 (Included as a text file, located at http://geophysics.eas.gatech.edu/zpeng/Catalog/CDaniels_PhD_2022_Table_5.1.txt) Templates and detections of the inclusive catalog of 13,671 events. Columns are: Origin time (UTC), relocated latitude, relocated longitude, relocated depth, estimated (or catalogued) magnitude (Md), origin time of template that detected the event, original latitude of template that detected the event, original longitude of template that detected the event, original depth of template that detected the event, Md magnitude of template that detected the event.

Table S5.2 (Included as a text file, located at http://geophysics.eas.gatech.edu/zpeng/Catalog/CDaniels_PhD_2022_Table_5.2.txt) 1,421 event detections and templates of the 15+ year study in the ETSZ earthquake study listing origin times (UTC).
Table S5.3 XCORLOC Input Parameters for 15 year study in the ETSZ.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Data Type</td>
<td>CC and phase</td>
</tr>
<tr>
<td>Phase Type</td>
<td>P and S</td>
</tr>
<tr>
<td>Max. epicentral distance</td>
<td>150 km</td>
</tr>
<tr>
<td>Number of Iterations for static station terms</td>
<td>5</td>
</tr>
<tr>
<td>Number of Iteration for SSST</td>
<td>10</td>
</tr>
<tr>
<td>Starting nmed, dmax for SSST</td>
<td>3300, 50</td>
</tr>
<tr>
<td>Ending nmed, dmax for SSST</td>
<td>20, 5</td>
</tr>
<tr>
<td>Minimum # of obs/pair for CC data</td>
<td>4</td>
</tr>
<tr>
<td># Iterations for Phase relocation</td>
<td>10</td>
</tr>
<tr>
<td>Initial box size</td>
<td>5 km</td>
</tr>
<tr>
<td>Fraction of shrinking box</td>
<td>0.67</td>
</tr>
<tr>
<td># Iterations for CC relocation</td>
<td>10</td>
</tr>
<tr>
<td>Initial box size for</td>
<td>2 km</td>
</tr>
<tr>
<td>Fraction of shrinking box</td>
<td>0.67</td>
</tr>
</tbody>
</table>

Table S5.4 HypoDD Input Parameters for 15 year study in the ETSZ.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Data Type</td>
<td>CC and phase</td>
</tr>
<tr>
<td>Phase Type</td>
<td>P and S</td>
</tr>
<tr>
<td>DIST</td>
<td>400 km</td>
</tr>
<tr>
<td>min # of obs/pair for cc/phase data</td>
<td>4</td>
</tr>
<tr>
<td>Inversion Type</td>
<td>LSQR</td>
</tr>
</tbody>
</table>
Sets of Iterations | 2  
---|---  
DAMP | Depends on cluster  
**Number of iterations (first set)** | 5  
P cc weight | 0.01  
S cc weight | 0.01  
cc residual threshold (as factor of s.d.) | N/A  
Max dist between cc linked pairs | N/A  
P phase weight | 1  
S phase weight | 1  
Phase residual threshold (as factor of s.d.) | 6  
Max dist between phase linked pairs | 10  
**Number of iterations (second set)** | 5  
P cc weight | 1  
S cc weight | 1  
cc residual threshold (as factor of s.d.) | 6  
Max dist between cc linked pairs | 10  
P phase weight | N/A  
S phase weight | N/A  
Phase residual threshold (as factor of s.d.) | 6  
Max dist between phase linked pairs | 10  

Table S5.5 Breakdown of Clusters Analyzed using hypoDD.

<table>
<thead>
<tr>
<th>Cluster ID</th>
<th># Events</th>
<th>DAMP</th>
<th>CND (start/end)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>867</td>
<td>150</td>
<td>64/71</td>
</tr>
<tr>
<td>2</td>
<td>94</td>
<td>35</td>
<td>80/45</td>
</tr>
<tr>
<td>3</td>
<td>73</td>
<td>60</td>
<td>59/64</td>
</tr>
<tr>
<td>4</td>
<td>58</td>
<td>25</td>
<td>58/58</td>
</tr>
</tbody>
</table>
Table S6.1 XCORLOC Input Parameters for area around Charleston in MPSSZ.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Data Type</td>
<td>CC and phase</td>
</tr>
<tr>
<td>Phase Type</td>
<td>P and S</td>
</tr>
<tr>
<td>Max. epicentral distance</td>
<td>150 km</td>
</tr>
<tr>
<td>Number of Iterations for static station terms</td>
<td>5</td>
</tr>
<tr>
<td>Number of Iteration for SSST</td>
<td>10</td>
</tr>
<tr>
<td>Starting nmed, dmax for SSST</td>
<td>3300, 50</td>
</tr>
<tr>
<td>Ending nmed, dmax for SSST</td>
<td>20, 5</td>
</tr>
<tr>
<td>Minimum # of obs/pair for CC data</td>
<td>4</td>
</tr>
<tr>
<td><strong># Iterations for Phase relocation</strong></td>
<td>10</td>
</tr>
<tr>
<td>Initial box size</td>
<td>3 km</td>
</tr>
<tr>
<td>Fraction of shrinking box</td>
<td>0.67</td>
</tr>
<tr>
<td><strong># Iterations for CC relocation</strong></td>
<td>10</td>
</tr>
</tbody>
</table>
Initial box size for | 1 km
---|---
Fraction of shrinking box | 0.67

Table S6.2 HypoDD Input Parameters for area around Charleston in MPSSZ.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Data Type</td>
<td>CC and phase</td>
</tr>
<tr>
<td>Phase Type</td>
<td>P and S</td>
</tr>
<tr>
<td>DIST</td>
<td>400 km</td>
</tr>
<tr>
<td>min # of obs/pair for cc/phase data</td>
<td>4</td>
</tr>
<tr>
<td>Inversion Type</td>
<td>LSQR</td>
</tr>
<tr>
<td>Sets of Iterations</td>
<td>2</td>
</tr>
<tr>
<td>DAMP</td>
<td>220</td>
</tr>
<tr>
<td>CND (start/end)</td>
<td>54/55</td>
</tr>
<tr>
<td><strong>Number of iterations (first set)</strong></td>
<td>5</td>
</tr>
<tr>
<td>P cc weight</td>
<td>0.01</td>
</tr>
<tr>
<td>S cc weight</td>
<td>0.01</td>
</tr>
<tr>
<td>cc residual threshold (as factor of s.d.)</td>
<td>N/A</td>
</tr>
<tr>
<td>Max dist between cc linked pairs</td>
<td>N/A</td>
</tr>
<tr>
<td>P phase weight</td>
<td>1</td>
</tr>
<tr>
<td>S phase weight</td>
<td>1</td>
</tr>
<tr>
<td>Phase residual threshold (as factor of s.d.)</td>
<td>6</td>
</tr>
<tr>
<td>Max dist between phase linked pairs</td>
<td>10</td>
</tr>
<tr>
<td><strong>Number of iterations (second set)</strong></td>
<td>5</td>
</tr>
<tr>
<td>P cc weight</td>
<td>1</td>
</tr>
<tr>
<td>Feature</td>
<td>Value</td>
</tr>
<tr>
<td>---------------------------------</td>
<td>-------</td>
</tr>
<tr>
<td>S cc weight</td>
<td>1</td>
</tr>
<tr>
<td>cc residual threshold (as factor of s.d.)</td>
<td>6</td>
</tr>
<tr>
<td>Max dist between cc linked pairs</td>
<td>10</td>
</tr>
<tr>
<td>P phase weight</td>
<td>N/A</td>
</tr>
<tr>
<td>S phase weight</td>
<td>N/A</td>
</tr>
<tr>
<td>Phase residual threshold (as factor of s.d.)</td>
<td>6</td>
</tr>
<tr>
<td>Max dist between phase linked pairs</td>
<td>10</td>
</tr>
</tbody>
</table>
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https://digitalcommons.memphis.edu/etd/1244


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https://doi.org/10.1186/BF03352815


