

Structures and dynamics of the subcontinental lithospheric mantle over the central and eastern North American continent, constrained by numerical modeling based on tomography models.

by

Tochukwu Emmanuel Onyebum

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Approved by:

Major Professor
Dr. Claudia Adam

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Abstract

Despite being located within a continental plate, the central and eastern United States (CEUS) has been associated active seismicity, often concentrated along seismic zones such as the New Madrid, the East Tennessee, or the South Carolina Seismic Zones. The influence of mantle convection on the CEUS seismicity is still debated. In this study, we investigate the relationship between mantle dynamics and intraplate seismicity. We do this by modeling mantle convection and the induced tectonic regimes and maximal shear stresses, SH_{max} , from seismic tomography models. Such an approach has already been used by a few authors, but the results seem to be very sensitive to the parameters of the model, particularly the input tomography models, that depict different structures of the mantle and the lithosphere. In this study, we investigate the sensitivity of the model parameters in reproducing the observed seismicity pattern. The uniqueness of our approach is that we use several global (SEMUCB-WM1, TX2019slab, 3D2018_08Sv, and SL2013Sv) and regional (DNA13, CSEM North America, and CURSA2021) tomography models. Using recent conversion laws, $R_{p/v}$, we convert the seismic velocity anomalies provided by tomography models into density anomalies. The density anomalies are then used to compute the instantaneous mantle flow, by considering several rheologies. The modeled convection velocities are used to compute the stress tensor and assess the tectonic regimes and the maximal shear stress, SH_{max} . Our findings reveal that all the model parameters (i.e., the input tomography models, the conversion factor $R_{p/v}$, and viscosity laws) significantly influence the modeled tectonic regimes and SH_{max} directions, emphasizing the complexity of the problem. However, the SEMUCB-WM1 and TX2019slab global tomography models, provide a better fit to several observed patterns, such as the compression in the northeastern United States, the extension in the Rio Grande Rift and Basin & Range, and the compression reported in most of the seismic zones. The strike-slip deformation

evidenced by earthquake focal mechanism in the central US and extension in the region bound by longitudes 257° and 266° remain unexplained by mantle dynamics alone, warranting further investigation regarding alternative causative factors such as Glacial Isostatic Adjustment and Gravitational Potential Energy. Furthermore, our results show that the regional tomography models generally fail to capture observed tectonic regimes in CEUS, suggesting that the seismicity is created by a larger scale mantle flow than the one occurring beneath CEUS.

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Dedication

This work is dedicated to God and to everyone who has been a source of support and encouragement throughout this program.

Chapter 1 - Introduction

The central and eastern United States (CEUS) is a seismically active area (Fig. 1.1) (Heidbach et al., 2018; Levandowski et al., 2018). There is significant intraplate seismicity occurring in the region, which seems to be concentrated in belts called seismic zones (Fig. 1.1) (e. g., Saxena et al., 2021). Tectonic regimes refer to the deformational patterns observed across a zone which could be extension, compression, or strike-slip deformation as a result of tectonic activities. Earthquakes have been used to compute the tectonic regimes and the direction and orientation of maximum shear stress (SHmax) (Fig. 1.2) (Lund Snee & Zoback, 2020). There is great variability in the tectonic regimes, with reverse faults dominating the eastern parts (Sbar & Sykes, 1973), strike-slip faulting occurring in the central-eastern portion of the continent, and normal faulting occurring in the central and central-west parts of the continent (Fig. 1.2). The orientation of the maximum shear stress SHmax in the eastern part of the continent is mainly NE-SW while in the central part, the direction changes geographically.

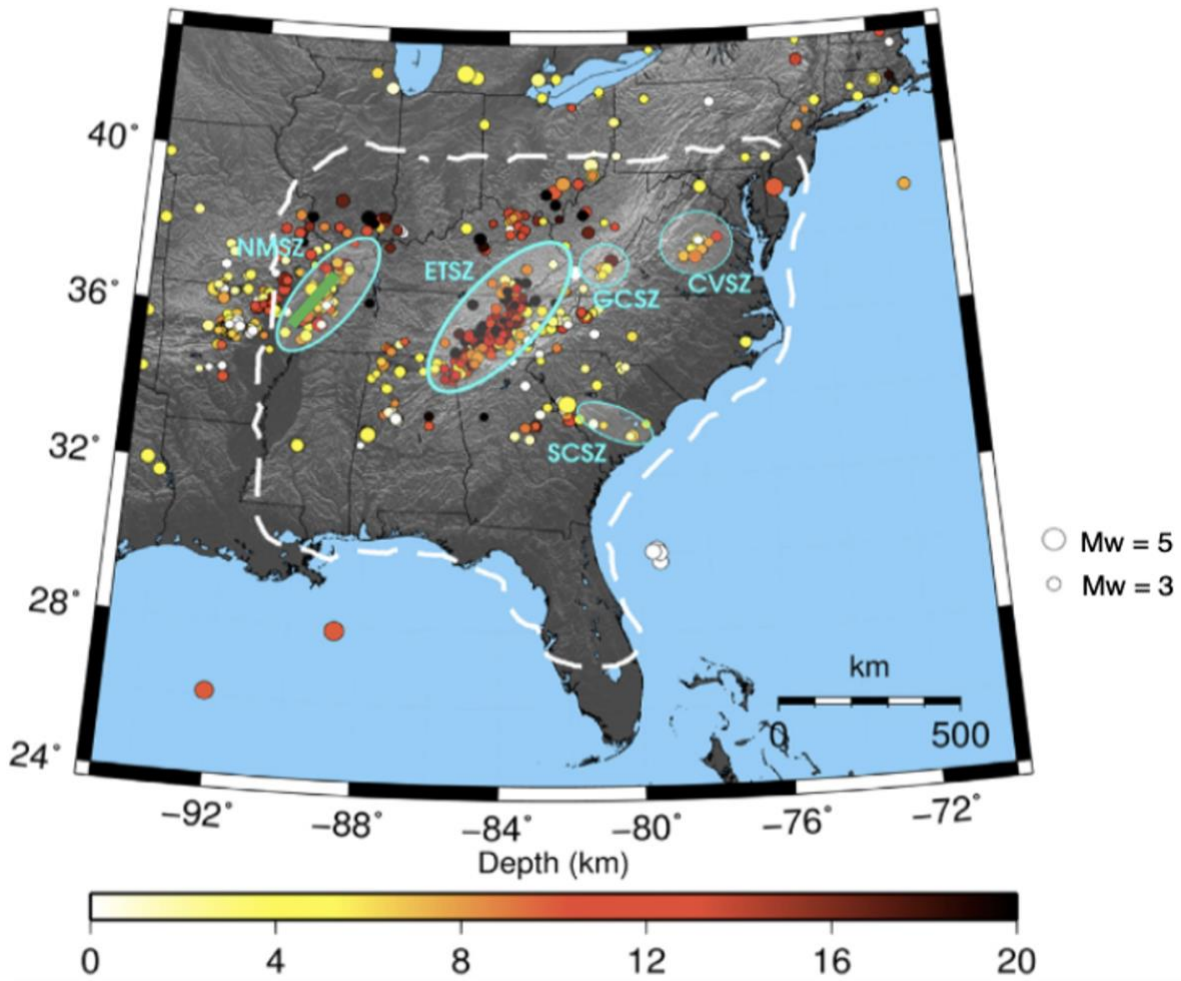
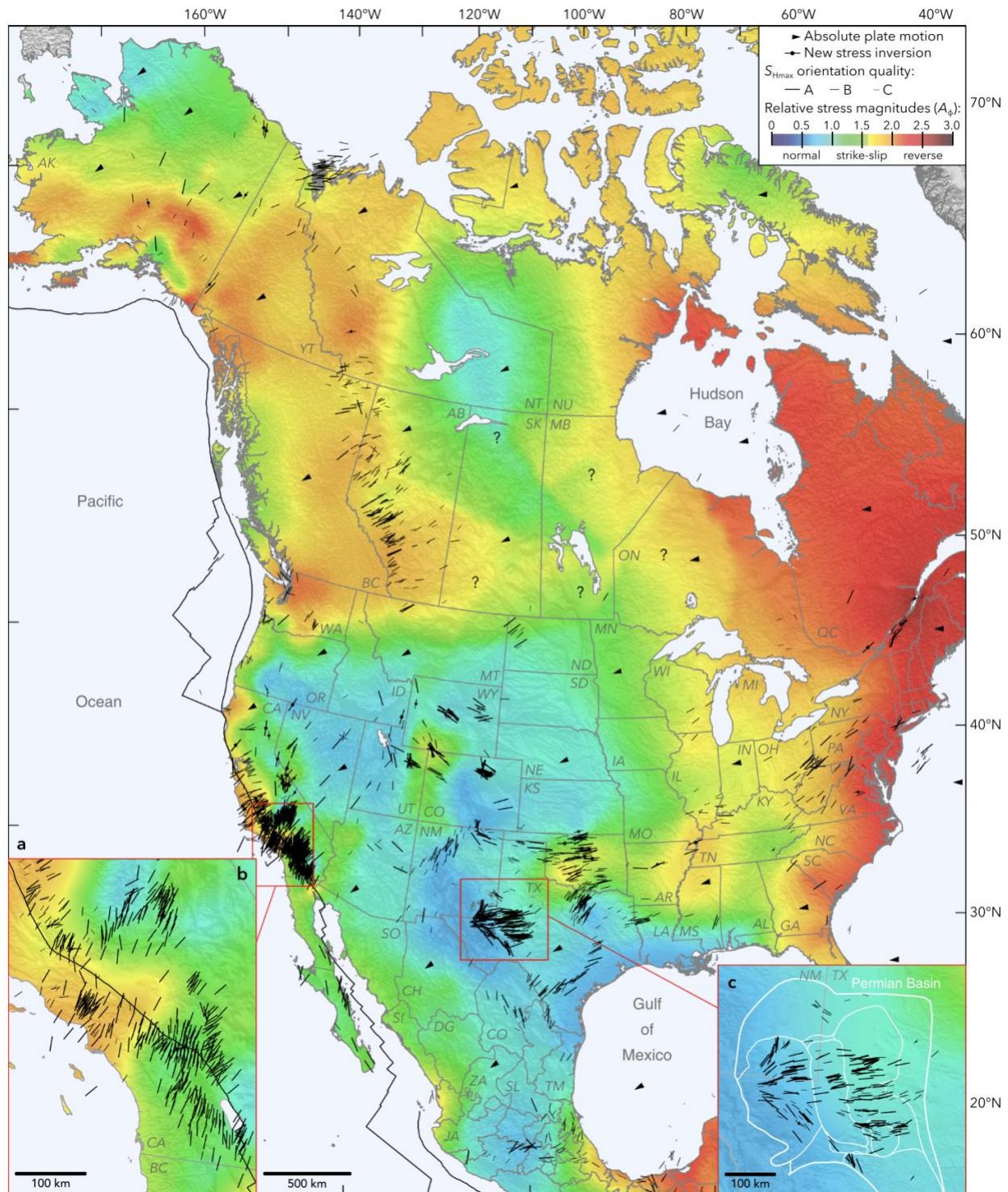


Figure 1.1: Map of the study area indicating the central and southeastern U.S. seismic zones: New Madrid Seismic Zone (NMSZ), eastern Tennessee Seismic Zone (ETSZ), South Carolina Seismic Zone (SCSZ), Giles County Seismic Zone (GCSZ) and Central Virginia Seismic Zone (CVSZ). The densely sampled area in the tomography results of Biryol et al. (2016) at 130 km is represented by the white dashed line. Earthquakes that occurred from December 2011 to December 2018 and had $M_w > 2.5$ are plotted as colored circles. Size and color of each circle represent event magnitude and depth, respectively (from Saxena et al., 2021). The green line is the approximate location of the Reelfoot Rift.

Many studies have been conducted on intraplate seismicity in the CEUS (e.g. Zoback, 1992; Wu & Johnston, 2000; Mazzotti & Townend, 2010; Chen et al., 2016; Brooks et al., 2017; Huang et al., 2017; Wang et al., 2019; Saxena et al., 2021). Several hypotheses have been proposed to

explain this seismicity. These include: the stresses induced by mantle convection (Forte et al., 2007; Saxena et al., 2021), stresses induced by forces acting at the plate boundaries (far field forces) (Humphreys & Coblentz, 2007), glacial isostatic adjustment (GIA) (Muir-Wood, 2000; Grollimund & Zoback, 2001; Mazzotti et al., 2005), variations in topography and lateral density heterogeneities in the crust and lithosphere (referred to as gravitational potential energy – hereafter referred to as GPE) (Levandowski et al., 2017, 2018; Ghosh et al., 2019; Saxena et al., 2021), variations in crustal density (Pollitz et al., 2001; Levandowski et al., 2016), anthropogenic factors (Bourne et al., 2014; Foulger et al., 2023; Wilson et al., 2017), and pre-existing weakness zones (Fig. 1.1) (Zhan et al., 2016; Saxena et al., 2021).

Authors have put forward varying views regarding the amount of intraplate deformation arising from plate boundary forces, mantle flow, and GPE in the lithosphere (e.g., Molnar et al., 2015; Levandowski et al., 2017; Zhou & Liu, 2019). A more precise and accurate image of both surface deformation and the structure of the subcontinental lithospheric mantle (SCLM) in terms of seismic velocity variations in the mantle and lithosphere has been provided using data from the mobile USArray (a seismic array that has imaged the SCLM over the North American continent during the last 19 years) (Fig. 1.3) and other supplementary networks in CEUS (Meltzer et al., 1999), such as the Advanced National Seismic System (Sipkin et al., 2006), the Southern Carolina Seismic Network (Hutton et al., 2006), etc. The information retrieved from these networks offers an excellent opportunity to study the connection between mantle dynamics and surface deformation over the CEUS.



Therefore, this research focused on the CEUS bringing new constraints to mantle/lithospheric coupling. We investigated the relationship between mantle dynamics and intraplate seismicity by modeling the mantle convection from global and regional tomography models. Modeled convection velocities were used to compute the stress tensor and assess the tectonic regimes and the shear stresses (SHmax and SHmin). These model outputs were compared to the observed deformation, i.e., the tectonic regimes derived from the earthquake's focal mechanisms (Heidbach et al., 2018; Levandowski et al., 2018; Lund Snee & Zoback, 2020).

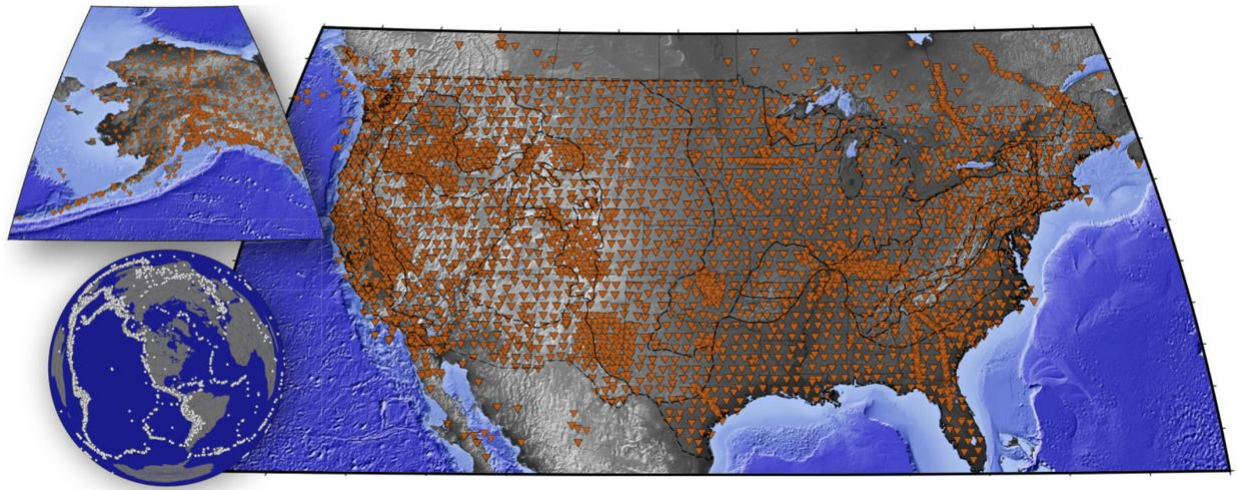


Figure 1.3: Distribution of the USArray stations in the contiguous United States and Alaska (upper inset). The lower inset illustrates event coverage focused on the contiguous United States (Porritt et al., 2021).

Chapter 2 - Background Information.

2.1 North American continent

The tectonic history of the North American continent is quite complex. The core of the continent (Canadian Shield) assembled in the Paleoproterozoic (2.0–1.8 Ga) via plate collisions of Archean continents and smaller Archean continental fragments (Kellett et al., 2020), resulting in the Trans-Hudson orogeny (Hoffman, 1988; Berman et al., 2007). Mobile belts (tectonically active regions, usually located at the boundaries of tectonic plates and characterized by tectonic activities such as earthquakes, volcanism, and orogenic processes) that have undergone many metasomatic processes, demarcate this assemblage of shields (Chen et al., 2009; Snyder et al., 2016). The subcontinental lithospheric mantle presently underneath North America, is primarily Archean in age and developed its current structure through collisional orogenesis (Whitmeyer and Karlstrom, 2007).

Several collisional events played major roles in the structural evolution of the Appalachian orogeny in the eastern United States, including the Taconic orogeny (Tull et al., 2018), the Acadian orogeny (Murphy et al., 2007), and the Alleghenian orogeny (Evans et al., 2014). The collision of the Ammonoosuc arc with the Laurentian craton occurred during the Upper Ordovician Taconic orogeny, with the edge of the continent being pulled beneath the arc. This was followed by slab failure of the descending amalgamated Laurentian-Ammonoosuc zone over New England and western New York (Hildebrand & Whalen, 2021), with the arc largely exposed at the Bronson Hill anticlinorium. This configuration led to the formation of other orogenic features such as the Dadeville Complex in the south Appalachian orogeny (Tull et al., 2018). The Acadian orogeny, a significant tectonic event during the Devonian period, occurred in the northern Appalachians

(Murphy & Keppie, 2005). This event resulted from the collision between Laurentia and Avalon terranes, leading to deformation, metamorphism, and magmatism. The orogeny, developing on an Andean-type margin, is associated with the collision of an oceanic plateau surrounding a plume (Murphy & Keppie, 2005). The succeeding Permian Alleghenian Orogeny has been associated with the subduction collision between North America and Africa (Hatcher et al., 1989; Rast, 1989).

The beginning of the Farallon subduction has been dated back to approximately 150 million years and played a significant role in the structural evolution of the western United States. Subduction is still going on in the northwestern part of the continent (Cascades subduction zone) (Atwater, 1970; Hawley & Allen, 2019). During the Sevier-Laramide orogenies (90 – 40 ma), the Farallon plate subducted beneath the western coast of the US at a shallow angle (Van der Lee & Nolet, 1997; Liu & Currie, 2016). This tectonic configuration resulted in a period of crustal compression, elevation, and modification of the crust and lithosphere (McCroory, 1989; Kelsey et al., 1994; van Hunen et al., 2002; DeCelles, 2004; O'Driscoll et al., 2009; Obrebski et al., 2011), like the formation of the Rocky Mountains (McMillan et al., 2006; Zhu et al., 2021). After the Laramide orogeny, the removal of the slab may have resulted in lithosphere-scale alterations that led to asthenospheric upwelling, uplift, extension, gravitational collapse in the Rio Grande Rift and Basin and Range, creation of the Cascadian subduction zone, and active magmatism (Christiansen et al., 2002; Camp & Ross, 2004; Obrebski et al., 2010, 2011). The intraplate seismicity in the New Madrid Seismic Zone (NMSZ) has been associated with the Farallon slab subduction (Chen et al., 2014; Forte et al., 2007; Liu & Stegman, 2011). The southern part of the slab, previously subducting under California, has detached and is now underneath CEUS (Chen et al., 2014; Forte et al., 2007; Liu et al., 2008). The seismicity pattern in CEUS is very complex. It may be accounted

for by the history of the North American continent, and/or by the present-day dynamics of the mantle, lithosphere, and crust.

2.2 Observed Seismicity Patterns and SHmax Orientation in CEUS Seismic Zones.

As intraplate earthquakes are concentrated along seismic zones, such as the New Madrid Seismic Zone, Eastern Tennessee Seismic Zone, South Carolina Seismic Zone, Central Virginia Seismic Zone, and Giles County Seismic Zone (Fig. 2.1), we focused our interpretation along them, as well as on other geographic regions, such as the Basin & Range, Rio Grande Rift (referred to as western CEUS features within this project), and the eastern US.

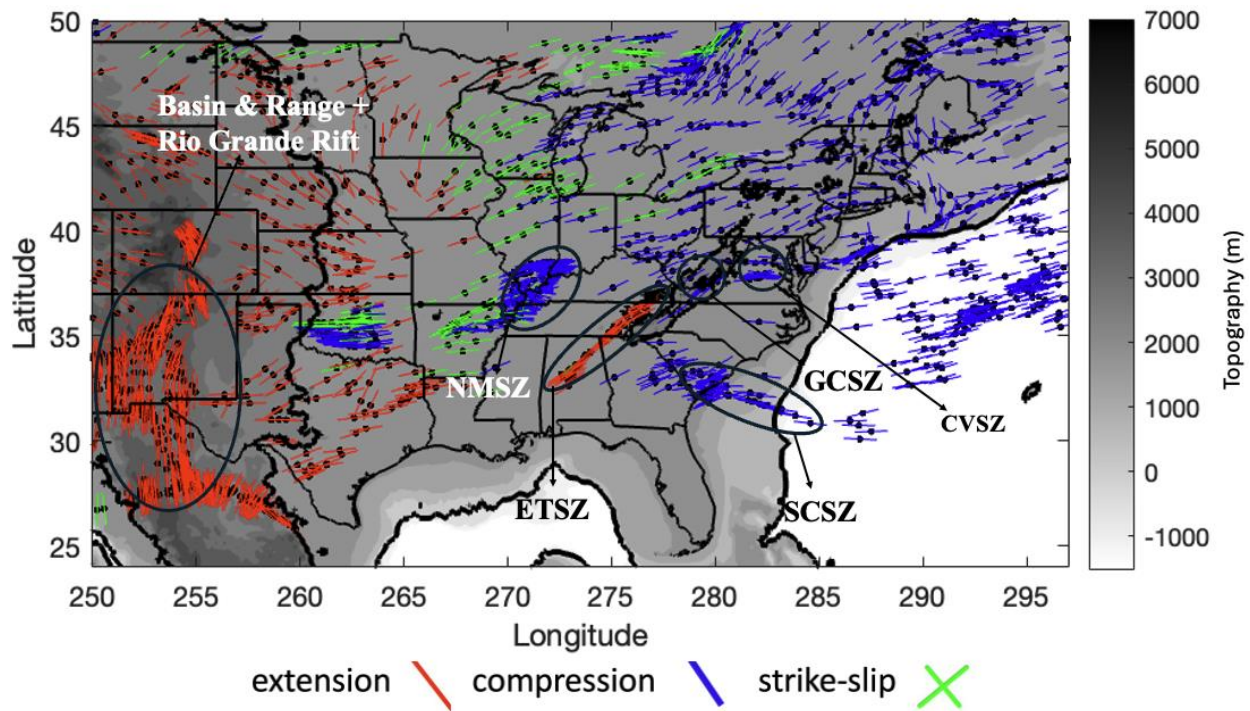


Figure 2.1: Observed tectonic regimes and orientations of SHmax computed from earthquake focal mechanics (Levandowski, et al., 2018), showing the locations of the NMSZ, ETSZ, SCSZ, GCSZ, and CVSZ (see Fig. 1.1 for details). The extension, compression, and strike slip regimes are represented in red, blue, and green, respectively. The black dots represent earthquake locations. The thick black lines are the 0 and 2500 m isolines (data from Tozer et al., 2019). The thin black lines are state's limits.

The observations depict compression over the eastern US, extension to the west of our study region, i.e., in the Basin & Range and Rio Grande Rift, while in the central part of our study, we observe a mix of strike slip, compressional and extensional regimes (Fig. 2.1).

2.2.1 The New Madrid Seismic Zone.

The New Madrid Seismic Zone (NMSZ) (Fig. 2.1) is an earthquake-prone region that stretches for 150 miles across the Midwestern United States, encompassing four states in the central Mississippi Valley. It extends from northeast Arkansas through southeast Missouri, western Tennessee, and western Kentucky to southern Illinois.

The deformation pattern shows a combination of compression and strike slip tectonic regimes (Fig. 2.1). The deformation style and slip directions in the southern part of the New Madrid seismic zone align with an overall ENE-WSW oriented compression and transform stress fields, whereas the northern part is predominantly characterized by compression (Fig. 2.1) (Liu & Zoback, 1997; McBride et al., 2003; Humphreys & Coblentz, 2007; Levandowski et al., 2016). The regional SHmax direction in the Central and Eastern US is $N64^{\circ}E \pm 6^{\circ}$, while the orientation in the New Madrid Seismic Zone is $N94^{\circ}E \pm 7^{\circ}$ with a 99.9% confidence limit (Levandowski et al. 2016) (Fig. 2.1).

The seismicity in the New Madrid seismic zone is associated with the Cambrian fault planes of the Reelfoot Rift (see Fig. 1.1 for location) (Braile et al., 1986; Pratt, 2012; Pratt et al., 2012). Earthquakes primarily occur between depths 4 and 14 km along the 30° dipping Reelfoot North and 44° dipping Reelfoot South faults, with the faults at depth being initially normal faults structurally inverted during the Paleozoic Appalachian-Ouachita Orogeny continental

compression (Csontos, 2008). Moment tensor analysis indicates distinct principal stress directions in the New Madrid Seismic Zone (NMSZ) compared to the surrounding region, with faults in the NMSZ aligning optimally in the local stress field rather than the regional stress field (Levandowski, 2016; Craig, 2017). This disparity has been primarily attributed to higher heat flow in the NMSZ than in the surrounding environment (Liu & Zoback, 1997; Pollitz, 2001; McKenna et al., 2007).

2.2.2 The Eastern Tennessee Seismic Zone

The Eastern Tennessee Seismic Zone (ETSZ) is the second most active seismic zone in the CEUS (Gallen & Thigpen, 2018; Powell et al., 2014; Powell & Thomas, 2016; Warrell et al., 2017) and stretches from the northeast of Alabama and the northwest of Georgia to the eastern part of Kentucky (Hatcher et al., 2012; Carpenter et al., 2014), covering a region that is more than 300 km long and 50 km wide (Warrell, et al., 2017). The tectonic regime across the ETSZ is mainly extension (Fig. 2.1), with the average SHmax orientation being N70°E (Hurd & Zoback, 2012; Warrell et al., 2017).

The Eastern Tennessee Seismic Zone (ETSZ) could represent the reactivation of a pre-existing shear zone that originated during the formation of the supercontinent Rodinia (Powell et al., 2014; Graw et al., 2015). Most of the zone's epicenters are located in the Valley and Ridge physiographic province in the southern Appalachians at depths ranging from 5km to 26km (Powell et al., 2014), occurring beneath the boundary that demarcates the Paleozoic sedimentary rocks of the Appalachian thrust belt from the underlying Grenville basement (Powell & Thomas, 2016).

2.2.3 South Carolina Seismic Zone

The South Carolina Seismic Zone is a complex region situated in the Southeastern portion of the North American continent and shows a predominant reverse deformational structure (Madabhusha & Talwani, 1993), with SHmax orientations ranging from NE in the northwestern portion to SE in the southeastern portion (Fig. 2.1). The seismic history of the South Carolina Seismic Zone is dominated by the 1886 Charleston earthquake, with NE trending modern reverse faults and ESE-oriented normal faults believed to have developed during Mesozoic rifting (Shah et al., 2023).

2.2.4 Central Virginia Seismic Zone

Virginia has historically experienced seismic activity concentrated in three main regions: the Central Virginia seismic zone (CVSZ), the Giles County seismic zone (GCSZ) (Fig. 2.1), and along the border with Tennessee, where it partially overlaps with the Eastern Tennessee seismic zone. The tectonic regime across the zone is mainly compressive with E-W oriented SHmax (Fig. 2.1).

In the regions surrounding these seismic zones, the tectonic regime is compression, while extension is observed along the southern border of the Eastern Tennessee Seismic Zone. The SHmax in these cases is generally oriented NE-SW, except in CVSZ, where it is oriented E-W.

2.2.5 Giles County Seismic Zone

This is one of the seismic zones in Virginia, geographically located between the ETSZ and CVSZ. Its deformation coincides with the regional compression observed towards the eastern portions of the study area (Fig. 2.1). The SHmax orientation is generally NE-SW.

2.3 Previous hypotheses on the origin of intraplate seismicity

Several hypotheses have been put forth to account for intraplate seismicity:

2.3.1 Glacial isostatic adjustment

In continental areas such as eastern North America and northern Europe, stress induced by glaciation and deglaciation dominates the crustal strain fields (Muir-Wood, 2000; Wu & Johnston, 2000; Marotta & Sabadini, 2004). For example, an investigation of the interactions between large-scale plate driving forces, lithospheric structure, and stresses caused by glacial loading and unloading-induced lithospheric bending revealed that the stress field in the areas around New Madrid was altered by the removal of the Laurentide ice sheet which covered much of the United States until ca. 20 ka. (Grollimund & Zoback, 2001). Mazzotti et al. (2005) show that massive earthquakes have been induced by post-glacial rebound.

2.3.2 Gravitational potential energy

GPE studies investigate the stresses induced by variations in topography and lateral density variations in the crust and lithosphere. The tomography model of Biryol et al. (2016) indicates that areas associated with high GPE have more earthquakes because they have a thin crust, which results in a thicker, high-density lithospheric mantle (Saxena et al., 2021). Topographic changes and lateral density heterogeneities in the lithosphere can create stresses both in the lithosphere and at sub-lithospheric levels (Pascal & Cloetingh, 2009). However, there is disagreement among studies looking at the impact of GPE regarding how lateral differences in lithospheric density affect intraplate seismicity (Levandowski et al., 2017; 2018, Ghosh et al., 2019).

2.3.3 Anthropogenic activities

Earthquakes have been artificially induced by human activities such as fluid injection during hydrocarbon production (Gan & Frohlich, 2013), mining activities (Li et al., 2020), and groundwater extraction (Shalev et al., 2023; Wetzler et al., 2019). A categorization system developed by Foulger et al. (2018) distinguishes 16 kinds of earthquakes that are caused by human activity, ranging from nuclear weapons testing to the withdrawal of groundwater for agricultural use. The seismicity rate in the CEUS has significantly increased because of human activities (Ellsworth et al., 2015; van der Baan & Calixto, 2017).

2.3.4 Mantle weakening

The Reelfoot Rift (see Fig. 1.1 for location) is an intracontinental rift which stretches from southern Illinois to northern Alabama. Chen et al. (2016) effectively imaged the fast-moving lower crust that lies beneath the Reelfoot Rift and its surroundings and conclude that areas associated with low shear velocity (i.e., weak mantle areas) correlate with intraplate seismic zones and concentrated crustal deformation. According to these authors, this shows that there is a relationship between mantle weakening and intraplate tectonics.

2.3.5 Mantle dynamics

The convection occurring in the mantle creates stresses that affect the surface topography (i.e., building of dynamic topography) and modify the stress field in the crust and lithosphere, which may create seismicity. Studying the mantle influence based on tomography provides new insights into the relationship between intraplate seismicity and mantle dynamics. For instance, Biryol et al. (2016) developed a high-resolution tomography model. They used it to image an area of the upper mantle that is seismically active and is between the New Madrid and East Tennessee Seismic

Zones (Fig. 1.1). According to their interpretation, the lithosphere is sinking and may be the cause of the earthquakes. Other authors (Chen et al., 2014; Nyamwandha et al., 2016) propose that the low P-wave velocity anomaly under the New Madrid Seismic Zone represents a weak zone that promotes stress transfer into the crust. A few studies have already investigated the effects of mantle dynamics on the observed stresses in CEUS (Forte et al., 2007; Becker et al., 2015; Saxena et al., 2021). However, these studies do not generally share the same conclusions regarding mantle/lithospheric coupling and its effects on intraplate seismicity. This is mainly due to the fact that each study is based on different tomography models, and the considered tomography models do not always image the same anomalies. The originality of the study we propose here is to consider several tomography models, in order to distinguish between “anecdotic” features and real features. Anecdotic refers to features that appear only in one or a few tomography models and therefore may be an artifact of the method or dataset.

2.4 Hypothesis

In this project, we investigate the hypothesis that mantle dynamics play an important role in CEUS intraplate seismicity. We achieved this by modeling instantaneous mantle flow induced by density anomalies which we derive from global and regional tomography models. Subsequently, we compare our geodynamic models to observations derived from earthquake focal mechanisms to see if they can be accounted for by mantle dynamics. The approach we use, based on regional and global tomography models, provides new insightful constraints on the influence of the mantle dynamics on the observed surface observations and a better fundamental understanding of mantle/lithosphere coupling.

2.5 Objectives

We evaluate the influence of mantle dynamics on the Central and Eastern United States (CEUS) intraplate seismicity through the following objectives:

1. Designing thermomechanical mantle structure by deriving density and temperature anomalies from the seismic velocity anomalies provided by regional and global tomography models.
2. Computation of instantaneous mantle flow induced by density anomalies via solving the conservation equations of mass and momentum, under the Boussinesq approximation, in a global or 3D spherical geometry.
3. Incorporating anelasticity and thermodynamic constraints in geodynamic models
4. Stress computations and derivation of modelled tectonic regimes.
5. Comparison of geodynamic models to observations from earthquake focal mechanism.

2.6 Potential Impact

This project generates new constraints on the lithosphere/mantle coupling over the North American continent, and in so doing provides new insights on what drives seismicity in regions where there are relatively no surface deformations. It could equally be a vital tool in risk assessment and understanding how underlying mechanisms might affect a region's seismogenic potential.

Chapter 3 - Methods

We model instantaneous mantle flow using global and regional tomography models which were developed with data from the USArray. To assess the influence of the model parameters on the geodynamic models, we vary these parameters including the seismic tomography models, viscosity laws, and conversion factors.

First, we design the thermomechanical mantle structure by deriving density and temperature anomalies from seismic velocity anomalies (provided by seismic tomography models) using conversion factors. We vary these conversion factors to test their influence on the geodynamic models. We then compute instantaneous mantle flow induced by these density anomalies by solving the conservation equations of mass and momentum under the Boussinesq approximation, in a global or regional 3D spherical shell geometry. We vary the viscosity laws used in the computation, to evaluate their influence on the geodynamic models. Using the modeled convection velocities and pressure induced by flow, we compute the stress tensor using the spatial derivatives of the convection velocities. On removing the pressure term (or hydrostatic stress), we obtain the deviatoric component of the stress tensor which we use in computing the principal stresses by solving for eigenvalues and eigenvectors at each grid point. We define the tectonic regimes using classical methods defined by Heidbach et al. (2010).

Finally, we compare our modeled stresses and tectonic regimes to observations obtained through computations of earthquake focal mechanism to assess which of these observations can be accounted for by mantle dynamics.

3.1 Geodynamic modelling

The seismic velocity anomalies provided by the tomography models are converted into density anomalies, using several recent conversion factors, discussed in Section 3.1.2. To compute the

instantaneous mantle flow induced by the density anomalies, we solve the conservation equations of mass and momentum, under the Boussinesq approximation, in a global or regional 3D spherical shell geometry using spherical polar coordinates (r, θ, ϕ) (Equations 1, 2 and 3).

When deriving the fundamental equations that govern the instantaneous mantle flow, the dimensions of length (L), velocity (v), and stress (σ) are normalized as follows:

$$\mathbf{L} = r_1 \mathbf{L}', \mathbf{v} = \frac{\kappa_0}{r_1} \mathbf{v}', \boldsymbol{\sigma} = \frac{\eta_0 \kappa_0}{r_1^2} \boldsymbol{\sigma}' \quad (\text{Equation 1})$$

Where r_1 is the Earth's radius, κ_0 is the reference thermal diffusivity, and η_0 is the reference viscosity. The numerical quantities of these values are listed in Table 3.1.

The mass and momentum conservation equations are:

$$\nabla \cdot \mathbf{v} = 0 \quad (\text{Equation 2})$$

$$-\nabla p + \nabla \cdot \eta (\nabla \mathbf{v} + \nabla \mathbf{v}^{tr}) + Ra_i \delta \rho \mathbf{e}_r = 0 \quad (\text{Equation 3})$$

Where p is the dynamic pressure, η is the viscosity, $\mathbf{v}$ is the velocity vector, the superscript tr represents the tensor transpose, $\delta \rho$ is the density anomaly, $\mathbf{e}_r$ is the unit vector in the radial direction, and Ra_i is the instantaneous Rayleigh number which is expressed as:

$$Ra_i = \frac{\rho_0 g r_e^3}{\kappa_0 \eta_0} \quad (\text{Equation 4})$$

Where ρ_0 is the reference density, g is the acceleration due to gravity, κ_0 is the reference thermal diffusivity, η_0 is the reference viscosity, and b is the mantle thickness being considered in the model. Table 3.1 enumerates the physical values of these parameters.

Table 3.1: Values of the physical parameters used in this study.

Symbol Definition	Value
Model thickness, b	660 or 2900 km

Earth's radius, r_l	6371 km
Gravitational acceleration, g	9.81 m s^{-2}
Reference density, ρ_0	3350 kg m^{-3}
Reference viscosity in the upper mantle, η_0	10^{21} Pa s
Reference thermal diffusivity, κ_0	$10^{-6} \text{ m}^2 \text{ s}^{-1}$
Instantaneous Rayleigh number, Ra_i	$9.45 * 10^7$ or $8.0 * 10^9$
Density contrast between the mantle and seawater, $\Delta\rho_s$	2320 kg m^{-3}
Density contrast at the core-mantle boundary, $\Delta\rho_c$	4337 kg m^{-3}
Gravitational constant, G	$6.66726 * 10^{11} \text{ N m}^2 \text{ kg}^{-2}$

The finite-volume (FV) based mantle convection code, ConvRS (Yoshida, 2008, 2010; Adam et al., 2010, 2021b), was used to calculate the instantaneous mantle flow.

The main model parameters tested in this project include the seismic tomography models, the seismic to density anomalies conversion factors, and viscosity laws.

3.1.1 Seismic Tomography Models

Several tomography models provide highly resolved images of the lithospheric and mantle structure of the North American continent. The mantle's influence on surface deformation is assessed using these tomography models. Considering several tomography models helps to evaluate the differences in features from seismic velocity anomalies observed in the models. Several global and US continent-scale seismic tomography models are tested within this project:

3.1.1.1 Global Tomography Models

The global tomography models whose influence is tested in the geodynamic models include the SEMUCB-WM1 model (French & Romanowicz, 2015), TX2019slab (Lu et al., 2019), 3D2018_08Sv (Debayle et al., 2016), and SL2013Sv (Schaeffer & Lebedev, 2013).

The development of the SEMUCB-WM1 model is based on the 'hybrid' waveform inversion approach, in which the accuracy and generality of the spectral finite element method (SEM) are combined for forward modeling of the global wavefield with nonlinear asymptotic coupling theory for efficient inverse modeling. TX2019slab was obtained via a P- and S-wave joint tomography inversion, which incorporates a three-dimensional thermal model of subducting slabs in the initial model (Lu et al., 2019). Consequently, the resultant tomography model shows slabs as higher amplitude subducting features. The 3D2018_08Sv tomography model is based on waveform modeling of Rayleigh waves recorded from 1976 and incorporates azimuthal anisotropy, achieving a lateral resolution of approximately 600km (Debayle et al., 2016). The SL2013Sv tomography model is a vertically polarized shear velocity model developed using a large set of waveform fits from approximately three-quarters of a million broadband seismograms (Schaeffer & Lebedev, 2013). The model captures surface expressions of deep structure, providing a favorable comparison benchmark. Depth cross sections along these tomography models are provided in Fig. 3.1.

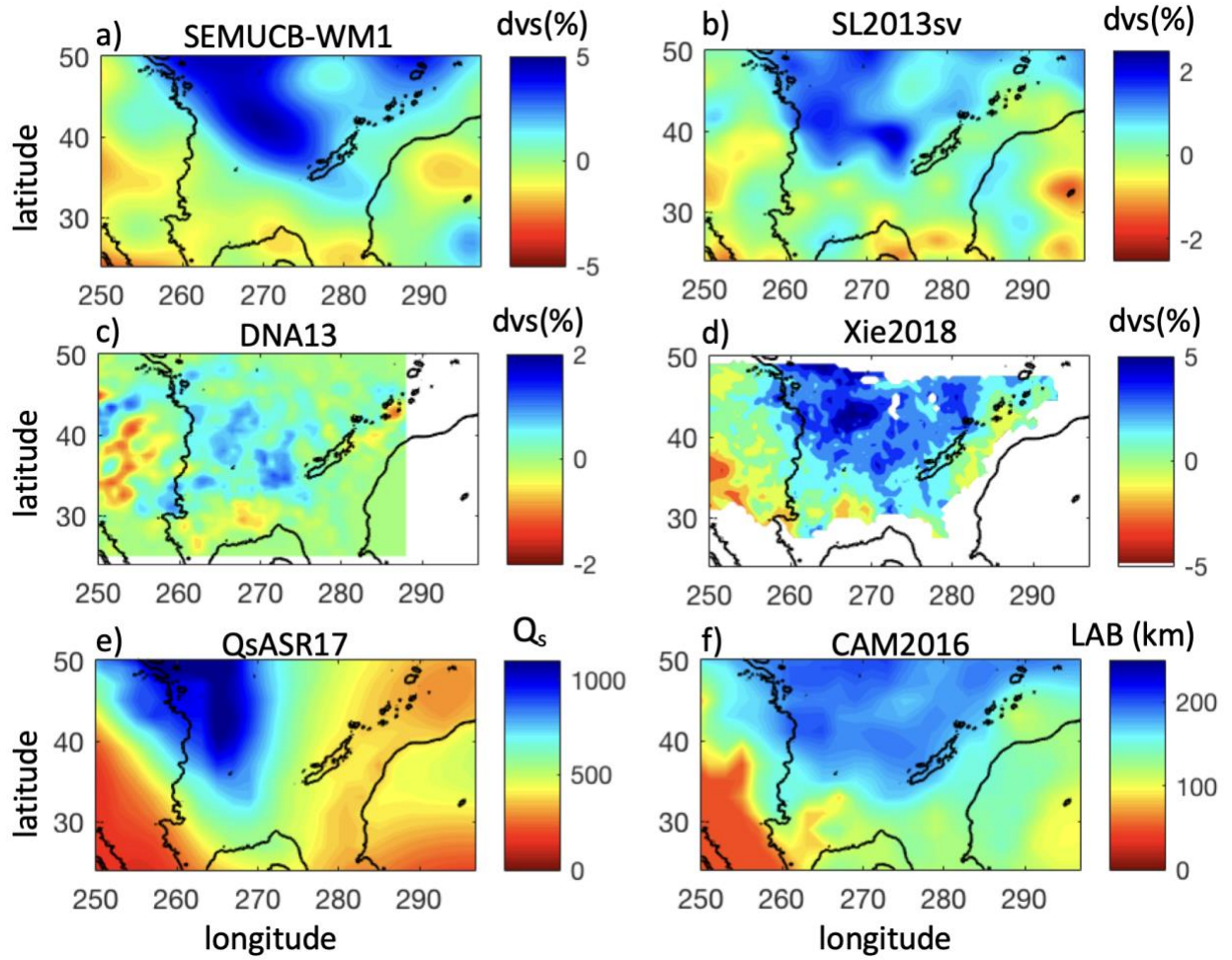


Figure 3.1: Variations in seismic velocity structure across several seismic tomography models considered in this study at 200 km depth. The black lines are the 0 and 2500 m isolines (data from Tozer et al., 2019).

3.1.1.2 Regional Tomography Models

Regional models designed over the North American continent were investigated, including the DNA13 model (Porritt et al., 2014), CSEM North America (Fichtner et al., 2018), and CUSRA2021 (Zhou et al., 2022).

The DNA13 model constrains the relative wave speed from the crust to the lower mantle. It achieves this by integrating teleseismic P observations, SH and SV observations, and surface-wave

phase velocities from teleseismic earthquakes and ambient noise. It highlights the upper mantle structure underlying the United States and refines images generated with USArray, providing insights into the deep structure of the Earth, including the subducting Farallon plate. The model highlights variations within the plate, such as high wave-speed anomalies in the eastern U.S.

The first-generation Collaborative Seismic Earth Model (CSEM) comprises 12 refinements from full seismic waveform inversion, ranging from regional crustal to continental-scale models (Fichtner et al. 2018). Through global full-waveform inversion, it ensures that regional refinements translate into the whole-Earth structure. The model evolution is based on a Bayesian updating scheme, which simplifies the process into a deterministic method that can be executed within present computational restrictions.

The CUSRA2021 (Zhou et al., 2022) model is a radially anisotropic shear wave speed model designed for the contiguous US and surrounding areas, employing full waveform tomography to image subsurface structures and characterize rock properties, including wave speeds, anisotropy, and attenuation. This model incorporates complex geodynamic processes such as hotspots and subduction, which have influenced the contiguous US region since the Mesozoic era. The deformation processes of the contiguous US, extensively studied across various disciplines, including geology, seismology, geochemistry, and geodynamics, are considered in the model. By utilizing seismic tomography as a direct method, the CUSRA2021 model aims to provide a comprehensive understanding of the subsurface, taking into account the radially anisotropic layering observed in the North America Craton region, which is associated with lithosphere accretion.

3.1.2 Conversion factors

In order to assess the role of the mantle using geodynamic models, the seismic velocity anomalies provided by the tomography models must be converted to density or temperature anomalies. The density to velocity heterogeneity ratio, $R_{\rho/v}$, which is used to convert seismic velocity anomalies ($\delta v/v_{ref}$), provided in %, to density anomalies, $\delta\rho$ is commonly expressed as;

$$\frac{\delta\rho}{\rho_{ref}} = R_{\rho/v} \frac{\delta v}{v_{ref}}, \text{ (Equation 5)}$$

where $\delta\rho$ is the density anomaly, and ρ_{ref} , the density along the 1600 K adiabat (Karato, 2008). In some studies, this factor has been presumed to be uniform across the mantle (Becker, 2006; Conrad et al., 2007; Conrad & Behn, 2010; Conrad & Lithgow-Bertelloni, 2006; Ghosh et al., 2017; Liu & Zhong, 2016; X. Wang et al., 2015). However, the assumption of a uniform conversion factor poses its own challenges. Most studies that employ this assumption in their approach usually exclude the upper 200–300 km of the seismic model. They do this because the large seismic velocities in this region result in unrealistically high temperatures and, consequently, buoyancy issues (Becker, 2006; Conrad & Lithgow-Bertelloni, 2006; King & Masters, 1992; Liu & Zhong, 2016; Yang & Gurnis, 2016). Other research has formulated a depth-dependent conversion by considering constraints from mineral physics, typically following the methodology outlined by Stixrude & Lithgow-Bertelloni (2005, 2007, 2011). This approach involves calculating phase equilibria and physical properties of multicomponent assemblages relevant to Earth's mantle using available thermodynamic data. Steinberger & Calderwood (2006) establish depth-dependent conversion factors with the assumption that both lateral density and seismic velocity variations have a purely thermal origin. Meanwhile, Simmons et al. (2009) determine $R_{\rho/v}$ by simultaneously

inverting seismic body wave travel times and geodynamic observations. This process involves incorporating realistic mineral physics parameters that connect wave speeds and density.

On the other hand, several studies employ a single seismic tomographic model and an assumed viscosity structure to derive a conversion factor that replicates the kinematic movement of lithospheric plates, dynamic topography, and/or geoid (Forte & Woodward, 1997; Simmons et al., 2009). Due to the peculiarity of the input parameters, the derived conversion factor may produce the best fit to the observations in the specific study, i.e., only for the tomography model considered in this study. However, this conversion factor will not be suitable when used with other tomographic models due to the variations between the tomographic models themselves (Shephard et al., 2017). Moreover, the geodynamic models rely on additional parameters, some of which are poorly constrained, with viscosity law being particularly problematic. The derived value of $R_{\rho/v}$ might lack robustness, as it hinges on the assumed viscosity law and the seismic tomography model utilized in the investigation.

Another approach consists of constraining the relationship between seismic velocity and density anomalies through the coupling of laboratory experiments and geophysical data (Steinberger & Calderwood, 2006; Karato, 2008). In this case, the conversion factor depends on numerous parameters, such as grain size and water concentration. Steinberger & Calderwood (2006) show a relatively uniform conversion factor in the upper mantle, while the lower mantle exhibits either a uniform or depth-decreasing velocity-density scaling (Fig. 3.2). Simmons et al. (2009) determine the velocity-density scaling by inversely calculating free-air gravity, kinematic velocities, and topography.

Studies such as Karato (2008) derive $R_{\rho/v}$ from laboratory experiments. Through the use of an anvil press, a sample is subjected to increasing pressures while several quantities, including the temperature derivative of seismic waves (A_{VT}) (equation 6), are measured. A_{VT} is used to compute the conversion factors for deriving density and mantle temperature anomalies from seismic velocity anomalies (equation 7).

$$A_{VT} = \frac{\partial v}{\partial T} \quad (\text{Equation 6})$$

$$R_{\rho/v} = \frac{-\alpha}{A_{VT}} \quad (\text{Equation 7})$$

Where α is the coefficient of thermal expansion along the 1600 K adiabat.

The temperature derivative of seismic waves, A_{VT} , can be expressed as the sum of its elastic and anelastic components:

$$A_{VT} = A_{VT_{el}} + A_{VT_{an}} \quad (\text{Equation 8})$$

With $A_{VT_{el}}$ and $A_{VT_{an}}$ being the elastic and anelastic components of A_{VT} respectively.

The elastic component of the temperature derivative of seismic waves (A_{VT}) of Karato (2008) can be measured from the high-pressure experiments. He computes the anelastic component from tomography models. The conversion factor of Adam et al. (2021a) employs a more sophisticated approach for implementing the anelastic component of the A_{VT} , based on the propagation of seismic waves rather than tomography models. To compute the elastic component, $A_{VT_{el}}$, Adam et al. (2021a) perform thermodynamic simulations. Fig. 3.2 shows the main conversion laws described above. We will study the effect of varying the conversion factor $R_{\rho/v}$.

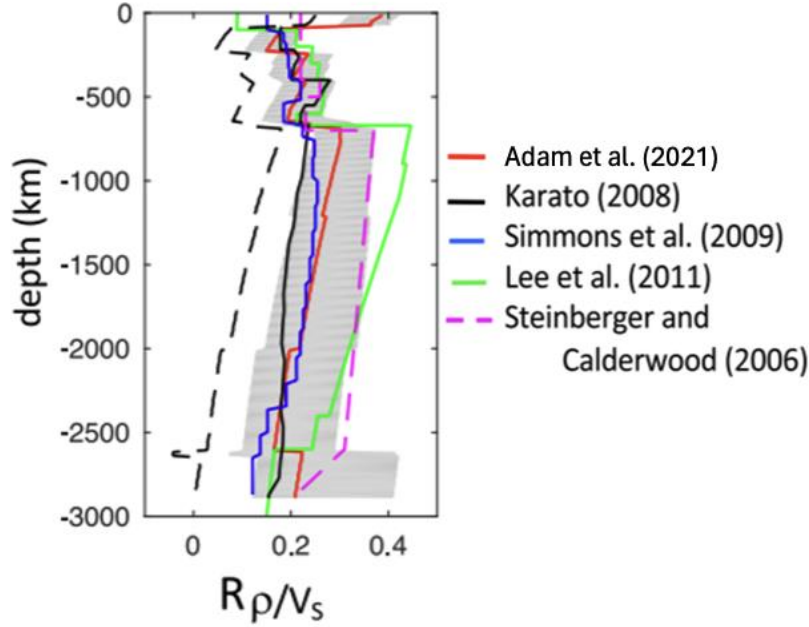


Figure 3.2: Conversion laws used in the study (Adam et al., 2021a). The grey area is the uncertainty associated with Adam et al. (2021a) conversion factor.

3.1.3 Viscosity models

Viscosity refers to the measure of a material's resistance to flow of deformation when subjected to external forces or stresses. It represents the strength of the material's internal friction and plays a crucial role in understanding the behavior of Earth materials such as rocks and magma.

We test the influence of depth and temperature dependent viscosities in the geodynamic models. In geodynamic models, the depth dependency generally consists in imposing a high-viscosity lithosphere ($\eta = 10^{24}$ Pa s) for depths 0 to 100 km, underlain by a low-viscosity asthenosphere ($\eta = 10^{19}$ Pa s) for depths 100 to 200 km (e.g., Adam et al., 2014), (green line in Fig. 3.3a). The upper mantle is associated with a 10^{21} Pa s viscosity, while the lower mantle is associated with a $30 \cdot 10^{21}$ Pa s or a $100 \cdot 10^{21}$ Pa s viscosity. Most of the models in this study are derived for the upper mantle.

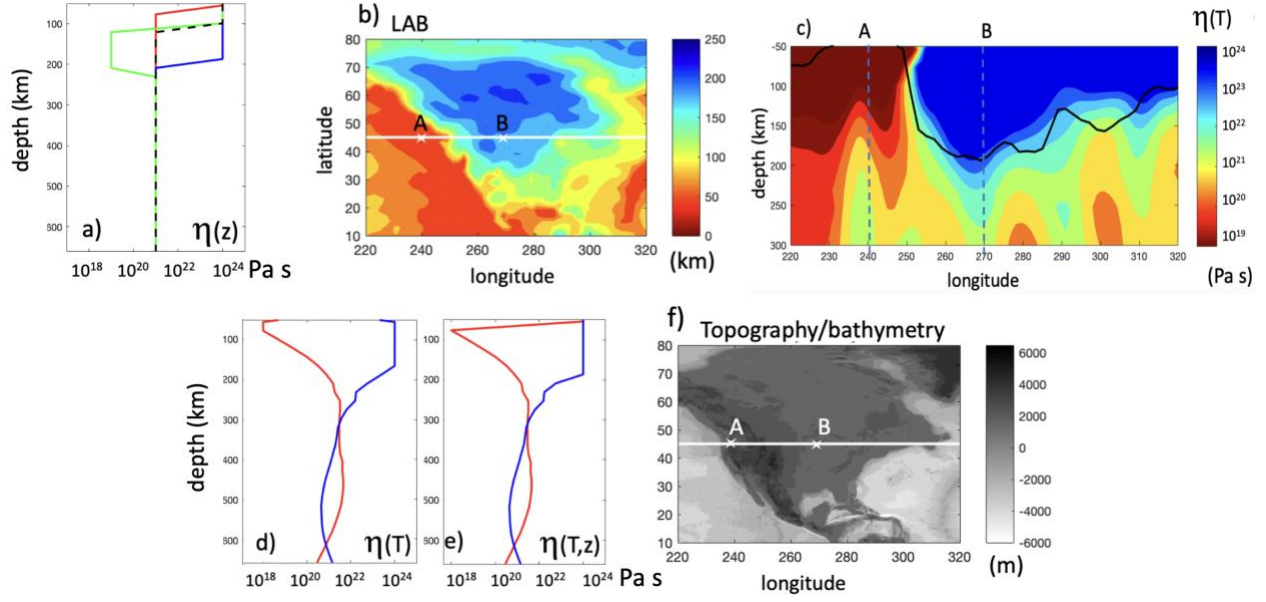


Figure 3.3: Viscosity laws considered in this study. a) depth dependent viscosities discussed in the text; b) LAB map (data from Priestley et al., 2019). The points A and B are located along thin and thick lithosphere, respectively; c) temperature dependent viscosity along the AB profile displayed in panel b; d) temperature dependent profiles along the points A (red lines) and B (blue line); e) depth and temperature dependent profiles along the points A (red lines) and B (blue line); f) bathymetry/topography map showing the AB profile location.

The previously described depth dependency is adequate for oceanic contexts. Within the North American continent, the base of the lithosphere also called the lithosphere-asthenosphere boundary, LAB, varies from 50 to 250 km (Fig. 3.3b) (CAM2016 model, Priestley et al., 2019). In order to provide a more realistic description of the viscosity under the North American continent, we will design models where we impose a high-viscosity lithosphere ($\eta = 10^{24}$ Pa s), underlain by a mantle with a with a 10^{21} Pa s viscosity (red and blue lines, Fig. 3.3a). Note that the low viscosity asthenosphere is mainly seen in oceanic contexts, so we remove it from the tests over North America.

We also test the temperature-dependent viscosity through an Arrhenius-type law (equation 9) with an activation enthalpy of 400 kJ/mol.

$$\eta = \eta_{ref}(d) \exp\left(\frac{H_a}{T} - \frac{H_a}{T_{ref}}\right) \quad (\text{Equation 9})$$

where H_a represents the activation enthalpy of mantle rock, R is the gas constant (8.31 kJ mol^{-1}), T and T_{ref} denote the dimensionless temperature and reference temperature, respectively, while $\eta_{ref}(d)$ is the dimensionless reference viscosity at $T = T_{ref}$. We use $H_a=400 \text{ KJ mol}^{-1}$ as in previous studies (e.g. Adam et al., 2021a; 2021b). The temperature dependency of the viscosity is displayed in Figs. 3.3c,d. We also design temperature and depth-dependent viscosities, where the depth dependency is imposed from the CAM2016 mode (i.e., $\eta = 10^{24} \text{ Pa s}$ within the lithosphere), and the temperature dependency is described above (Fig. 3.3e).

3.2 Principal stresses and tectonic regimes

We use the finite-volume (FV) mantle convection code, ConvRS (Adam et al., 2010, 2021b; Yoshida, 2008, 2010), to compute the instantaneous mantle flow. The inputs of the convection code are the density anomalies, the viscosity, and the boundary conditions. The outputs of the convection code are the modelled convection velocities and the pressure induced by the instantaneous mantle flow. These convection velocities include velocity variations along the radial axis (V_r), the latitudinal axis (V_θ), and the longitudinal axis (V_ϕ).

3.2.1 Stress Tensor

The stress tensor is computed using the modelled convection velocities derived from the instantaneous mantle flow. The convection code has been developed by Masaki Yoshida (2008) in Fortran 90, and it computes the stress tensor from the spatial derivatives of the convection velocities in spherical coordinates (equations 10-15):

$$\sigma_{rr} = -p + 2\mu \frac{\partial v_r}{\partial r} \quad (\text{Equation 10})$$

$$\sigma_{\theta\theta} = -p + 2\mu \left(\frac{1}{r} \frac{\partial v_\theta}{\partial \theta} + \frac{v_r}{r} \right) \text{ (Equation 11)}$$

$$\sigma_{\phi\phi} = -p + 2\mu \left(\frac{1}{r \sin \theta} \frac{\partial v_\phi}{\partial \phi} + \frac{v_r}{r} + \frac{\cot \theta}{r} v_\theta \right) \text{ (Equation 12)}$$

$$\sigma_{r\theta} = \sigma_{\theta r} = \mu \left(\frac{1}{r} \frac{\partial v_r}{\partial \theta} + \frac{\partial v_\theta}{\partial r} - \frac{v_\theta}{r} \right) \text{ (Equation 13)}$$

$$\sigma_{r\phi} = \sigma_{\phi r} = \mu \left(\frac{1}{r \sin \theta} \frac{\partial v_r}{\partial \phi} + \frac{\partial v_\phi}{\partial r} - \frac{v_\phi}{r} \right) \text{ (Equation 14)}$$

$$\sigma_{\theta\phi} = \sigma_{\phi\theta} = \mu \left(\frac{1}{r \sin \theta} \frac{\partial v_\theta}{\partial \phi} + \frac{1}{r} \frac{\partial v_\phi}{\partial \theta} - \frac{\cot \theta}{r} v_\phi \right) \text{ (Equation 15)}$$

Where σ are the stress components of the stress tensor (see equation 17), in the longitudinal ϕ , latitudinal θ , and radial r directions, p is pressure, μ is the viscosity, and v is the convection velocities in ϕ , θ , and r axis in a spherical coordinate system.

In this project, we translate the convection code from Fortran 90 to MATLAB.

3.2.2 Computation of eigenvalues and vectors

The deviatoric stress is the component of the stress tensor, which represents the part of the stress that causes deformation without altering the volume of the material. It excludes the hydrostatic stress, which is related to changes in volume. The stress tensor is a 3x3 matrix representing the stresses in three dimensions (equation 16). Mathematically, the deviatoric stress tensor is obtained by subtracting the mean (hydrostatic) stress (equation 17) from the total stress (equation 18).

$$\sigma = \begin{bmatrix} \sigma_{xx} & \sigma_{xy} & \sigma_{zx} \\ \sigma_{xy} & \sigma_{yy} & \sigma_{yz} \\ \sigma_{zx} & \sigma_{yz} & \sigma_{zz} \end{bmatrix} \quad \text{(Equation 16)}$$

$$\sigma_{mm} = \frac{\sigma_{xx} + \sigma_{yy} + \sigma_{zz}}{3} \quad \text{(Equation 17)}$$

$$\sigma_{dev} = \sigma - \begin{bmatrix} \sigma_{mm} & 0 & 0 \\ 0 & \sigma_{mm} & 0 \\ 0 & 0 & \sigma_{mm} \end{bmatrix} \quad (\text{Equation 18})$$

Where σ is the stress tensor, σ_{mm} is the mean or hydrostatic stress, and σ_{dev} is the deviatoric stress.

Going further, we use this deviatoric component to compute the principal stresses (Heidbach et al., 2010). The principal stresses are obtained by solving for the eigenvalues and eigenvectors in the deviatoric stress tensor.

The Fortran 90 convection code's subroutines collectively constitute a computational framework that determines the principal stresses and their corresponding eigenvectors within the stress tensor. The convection code starts by solving a system of linear equations to find direction cosines representing the orientation of a principal stress. The code uses complex arithmetic to identify three real solutions for a cubic polynomial corresponding to the stress tensor's eigenvalues. With these eigenvalues, it computes the maximum, intermediate, and minimum principal stresses and calculates the unnormalized direction cosines for the associated eigenvectors to ensure the proper orientation of the eigenvectors. The tectonic regime is determined based on the magnitudes and orientations of the principal stresses. This is explained in section 3.2.3.

However, we obtain the same results through a simplified approach in MATLAB (MATLAB R2016b, Mathworks®) by using its 'eig' function to compute the eigenvalues and eigenvectors of the stress tensor at each grid point. The eigenvalues τ_1 , τ_2 , and τ_3 , represent the principal stresses, while the corresponding eigenvectors, v_1 , v_2 , and v_3 , indicate the orientation of these principal stresses.

3.2.3 Definition of the tectonic regimes

Yoshida (2008) uses the classical method of defining tectonic regimes (Heidbach et al., 2010). In the following, the index 1 corresponds to the biggest eigenvalue τ_1 , the index 3 corresponds to the smallest eigenvalue τ_3 , and the index 2 corresponds to the intermediary eigenvalue τ_2 . Their associated eigenvectors are v_1 , v_2 , and v_3 , respectively. If the most vertical eigenvector is v_1 , i.e., the eigenvector associated with τ_1 , then the tectonic regime is extensional. If v_3 is the most vertical eigenvector, then the tectonic regime is compressional. If the most vertical eigenvector is v_2 , then the tectonic regime is strike-slip.

Other studies consider different approaches to defining the tectonic regimes, including the one previously described, as well as the methods developed by Fitzenz & Miller (2004) and Lund Snee & Zoback (2020). Angelier (1979) defines a stress ratio ϕ (equation 19) to quantify the ratio between principal stresses.

$$\phi = \frac{(\sigma_2 - \sigma_3)}{(\sigma_1 - \sigma_3)} \quad (\text{Equation 19})$$

However, this parameter does not define the tectonic regimes. Consequently, a new parameter A_ϕ (equation 20) is used to incorporate all tectonic regimes (Simpson, 1997; Levandowski et al., 2018; Lund Snee & Zoback, 2020).

$$A_\phi = (n + 0.5) + (-1)^n(\phi - 0.5) \quad (\text{Equation 20})$$

In this equation, n is 0 for normal faulting ($SH_{\max} = \sigma_1$), 1 for strike-slip faulting ($SH_{\max} = \sigma_2$), and 2 for reverse faulting ($SH_{\max} = \sigma_3$). A_ϕ varies from 0 to 1 for extensional regimes, 1 to 2 for strike slip regimes, and 2 to 3 for compressional regimes (Fig. 3.4). The value of n can only be

imposed when the focal mechanism of an earthquake is known. Therefore, equation 20 is mainly used by seismologists to map the tectonic regimes deduced from earthquakes.

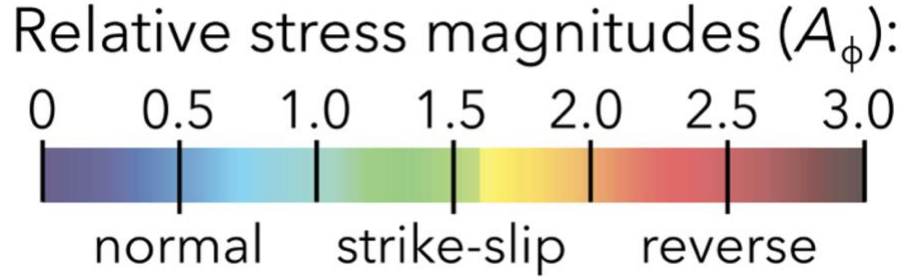


Figure 3.4: Tectonic regime classification based on A_ϕ values.

Fitzenz & Miller (2004) employ another approach, which combines the most vertical principal stress with the stress ratio, ϕ (equation 20) to define nine end-member tectonic regimes (Table 3.2).

Table 3.2: Tectonic regimes of Fitzenz & Miller (2004)

VERTICAL PRINCIPAL STRESS (σ_v)	$\phi = 0$ ($\sigma_2 = \sigma_3$)	$\phi = 0.5$ ($\sigma_2 = (\sigma_1 + \sigma_3)/2$)	$\phi = 1$ ($\sigma_1 = \sigma_2$)
$\sigma_v = \sigma_3$	Reverse strike-slip	Pure reverse	Constrictive
$\sigma_v = \sigma_1$	Radial extensive	Pure normal	Normal strike-slip
$\sigma_v = \sigma_2$	Reverse strike-slip	Pure strike-slip	Normal strike skip

When modeling the stresses from the instantaneous mantle flow, there is no possibility of knowing prior to the tectonic regime: the n in equation 20. Therefore, to model the tectonic regimes, we will

use the definition provided by Heidbach et al. (2010) and Fitzenz & Miller (2004). The observed stresses are generally provided and expressed by the A_ϕ variable. In this case, we will consider that we have extension for A_ϕ values encompassed between 0 and 1, strike-slip for A_ϕ values encompassed between 1 and 2, and compression for A_ϕ values encompassed between 2 and 3.

3.3 Comparison with Surface Observations

The modelled tectonic regimes and maximum horizontal stress are compared to the ones computed from earthquake focal mechanisms (Fig. 3.5) (e.g., Levandowski et al., 2018; Lund Snee & Zoback, 2020). We consider the regions where tectonic regimes are recovered in the models to have experienced substantial influences from mantle dynamics.

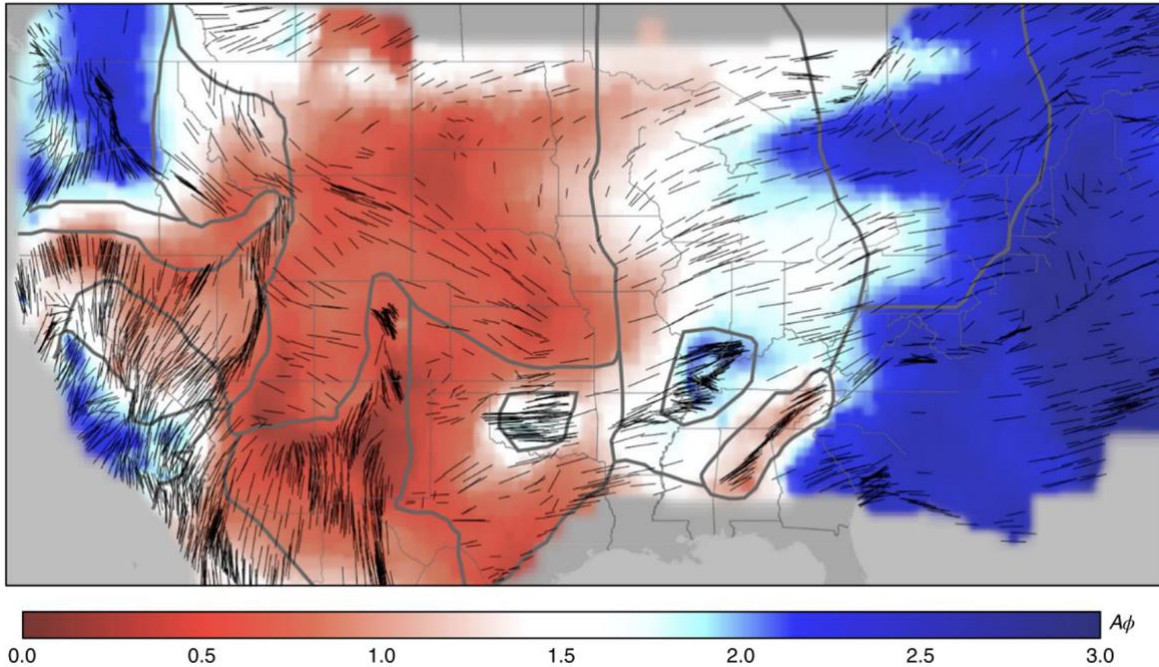


Figure 3.5: Stress map showing observed tectonic regimes across the United States based on A_ϕ classification (Levandowski et al., 2018). Extension regimes are in red, compression regimes are in blue, while strike slip is in white. The black lines indicate the SHmax orientation.

3.4 Quantitative assessment of the geodynamic models.

We also provide a quantification of the fit between the modeled tectonic regimes and the observed ones. We made a 1degree by one degree grid over our study area. At each node, we impose the value 0 if the computed and observed tectonic regimes are the same. We impose the value 1 if they differ. We then compute the RMS (root mean square) of this “difference” grid. The RMS values are reported in Fig. 4.8 and Table 4.1

Chapter 4 - Results

Several studies have used A_ϕ classification (equation 20) to compute the tectonic regimes (discussed in section 3.2.3) and the orientation of SHmax from earthquake focal mechanisms (e.g. Levandowski et al., 2018; Lund Snee & Zoback, 2020). Within this project, we compare the outputs of our geodynamic models to the observed regimes (discussed in Section 2.2) obtained by Levandowski et al. (2018) (Fig. 2.1) across the CEUS. In the following sections, we test the influence of varying the model parameters including the conversion factors, viscosity laws, and seismic tomography models. We show the observations of Levandowski et al. (2018) on each figure where we compare our modeled regimes.

4.2 Conversion Factor Tests

To assess the influence of the conversion factors used for converting seismic velocity anomalies to density anomalies, $R_{\rho/vs}$, we test some of the $R_{\rho/vs}$ discussed in section 3.1.2 while considering one tomography model (SEMUCB-WM1, French & Romanowicz, 2014) and the temperature viscosity dependent viscosity (discussed in section 3.1.3). Specifically, we test the conversion factors developed by Adam et al. (2021a), Simmons et al. (2009), and Steinberger & Calderwood (2006). We then investigate how varying $R_{\rho/vs}$ influences the tectonic regimes and SHmax orientation (Fig. 4.1).

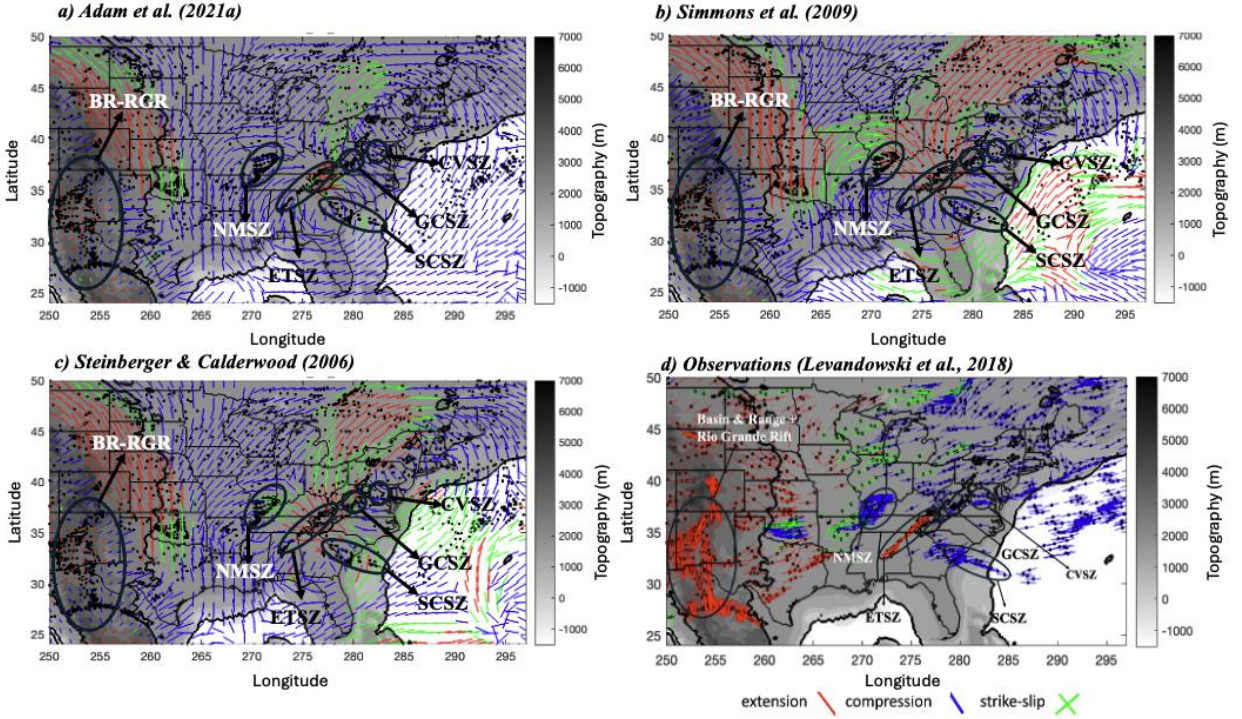


Figure 4.1: Variations of tectonic regimes and SHmax orientations at 77 km depth, obtained using the conversion factor of (a) Adam et al. (2021a), (b) Simmons et al. (2009), and (c) Steinberger & Calderwood (2006), (d) Observations retrieved from earthquake focal mechanism. The instantaneous flow is modeled from the SEMUCB-WM1 tomography model (French & Romanowicz, 2014), while using a temperature dependent viscosity, $\eta(T)$. The extension, compression, and strike slip regimes are represented in red, blue, and green, respectively. The black dots represent earthquake locations. The circled regions are the New Madrid Seismic Zone (NMSZ), eastern Tennessee Seismic Zone (ETSZ), South Carolina Seismic Zone (SCSZ), Giles County Seismic Zone (GCSZ) and Central Virginia Seismic Zone (CVSZ). The thick black lines are the 0 and 2500 m isolines (data from Tozer et al., 2019). The thin black lines are state's limits.

All models partially recover the observed extension over the western CEUS features (i.e., Basin & Range and Rio Grande Rift) (Fig. 4.1d) and the compression in the northeastern US, although they generally fail to recover the strike-slip deformation observed in the central part. They also recover the observed extension at the northern part of the Eastern Tennessee Seismic Zone.

The Adam et al. (2021a) conversion factor is more realistic than the other conversion factors, given that it incorporates anelasticity and thermodynamics constraints. Additionally, it majorly recovers the regional compressive regime in the eastern United States better than the other conversion

factors, albeit not in the regional NE-SW orientation observed in Levandowski et al. (2018) observations (Fig. 4.1d). It recovers the extension observed towards the western CEUS features (Rio Grande Rift and Basin & Range) roughly the same extent as the other conversion factors (Figs 4.1b and c) and retrieves the compression and SHmax orientation along the New Madrid Seismic Zone better than the other tested conversion factors. The conversion factors of Simmons et al. (2009) (Fig. 4.1b) and Steinberger & Calderwood (2006) (Fig. 4.1c) appear to exhibit a random mix of all three tectonic regimes (extension, compression, and strike-slip) towards the eastern edge of the continent, which is contrary to Levandowski et al. (2018) observations. Consequently, they do not recover the regimes across the South Carolina Seismic Zone, Central Virginia Seismic Zone, and Giles County Seismic Zone. The compression across these seismic zones is retrieved with the conversion factor of Adam et al. (2021a).

To test the other parameters, we fix the conversion factor of Adam et al. (2021a). It is more realistic due to the incorporation of thermodynamics and anelasticity constraints, in addition to better recovery of the observed regimes.

4.3 Viscosity Tests

Varying the viscosity significantly should in theory affect the mantle and lithosphere rheology, and then affect the instantaneous mantle flow and the derived tectonic regimes and SHmax directions. Therefore, we test the influence of the viscosity laws using the depth dependent viscosities derived from the LAB model of CAM2016 (Priestley et al., 2019) and at 100 km depth. We also test temperature dependent viscosities. The instantaneous flow is computed with the SEMUCB-WM1 tomography model (French & Romanowicz, 2014), and the conversion law derived by Adam et al. (2021a).

The resultant tectonic regime and SHmax are displayed in Fig. 4.2.

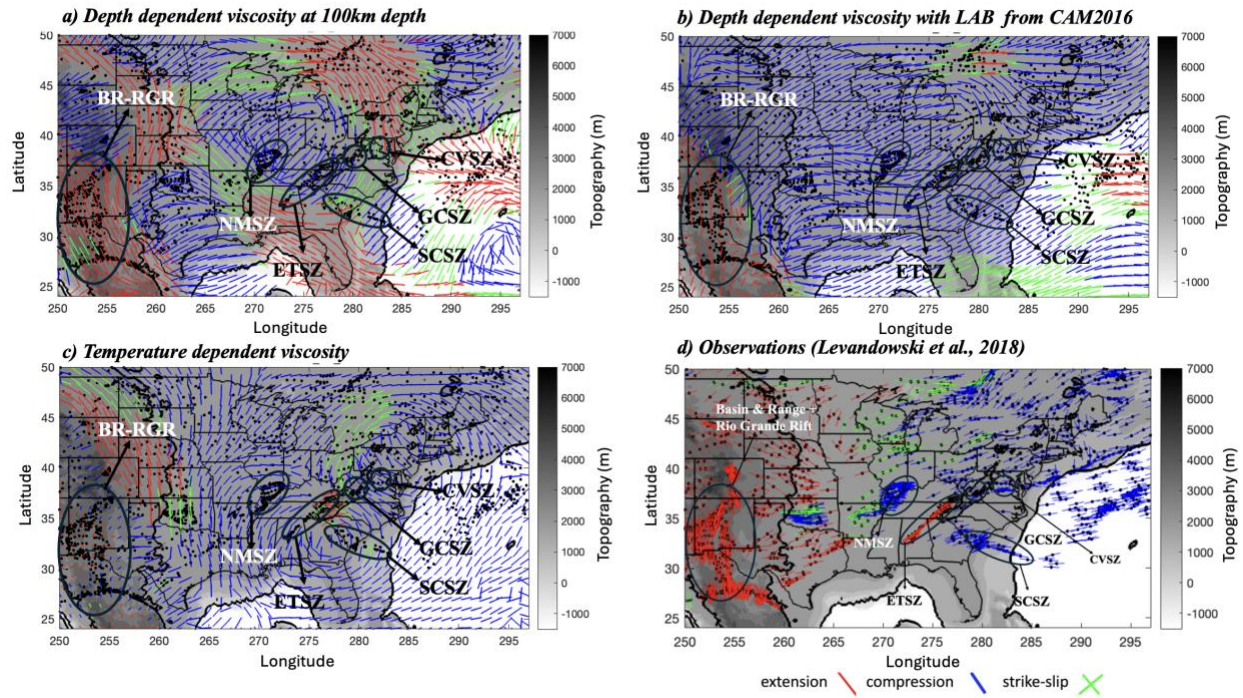


Figure 4.2: Variations of tectonic regimes and SHmax orientations obtained from imposing (a) depth-dependent viscosity including only a 100 km thick lithosphere (b) depth-dependent viscosity, where the base of the lithosphere is taken from the CAM2016 (Priestly et al., 2019), (c) temperature-dependent viscosity discussed in Section 3.1.3. Observations are reported in panel d. The models are based on the SEMUCB-WM1 tomography model (French & Romanowicz, 2015) and the conversion factor developed by Adam et al. (2021a). The black dots represent earthquake locations. See Fig. 4.1 for details.

All four viscosity laws are significantly different (Fig. 4.2), retrieving mainly the compression across the New Madrid Seismic Zone along with the regional NE-SW compressive SHmax orientation, excluding the depth dependent viscosity at 100 km depth (Fig. 4.2a) which retrieves NW-SE compressive SHmax orientation. The temperature dependent viscosity tests (Figs. 4.2c and d), and the depth-dependent viscosity from CAM2016 (Priestly et al., 2019) (Fig. 4.2b) recover well the compressive tectonic regimes and SHmax orientations across the eastern parts of CEUS, thus retrieving the compression across the South Carolina, Giles County and Central Virginia Seismic Zones. The depth dependent viscosities retrieve the extension and general N-S oriented SHmax in the earthquake regions of the western CEUS features, especially the viscosity from the

CAM2016 model (Fig. 4.2b). While this configuration seems to retrieve most of the observations across the seismic zones, it partially recovers the extension in the western CEUS features and does not retrieve the regional shear zone at the centre of the study area and the extension at the Eastern Tennessee Seismic Zone.

To test the other model parameters, we use the depth dependent viscosity derived from CAM2016 (Fig. 4.2b). CAM2016 model (Fig. 3.3b) accounts for the variations in lithospheric thickness and retrieves the regimes in most Seismic Zones, excluding the Eastern Tennessee Seismic Zone.

4.3.1 Weak Plate Boundary

Several authors came to impose weak plate boundaries, WPB, in their geodynamical models (e.g. Ghosh et al., 2008; 2010), as they represent plate boundaries in a more realistic way. To include WPB, we have to impose a viscosity drop along the plate margins, such as mid-oceanic ridges and subduction zones. We superimpose the WPB on the depth-dependent viscosity with thickness from CAM2016 model. The recovered tectonic regime and SHmax direction (Fig. 4.3a) does not differ from the structure obtained without the boundary conditions (Fig. 4.3b). Our results show that the boundary conditions do not have a significant influence on the varying tectonic regimes and SHmax orientations in the geodynamic models.

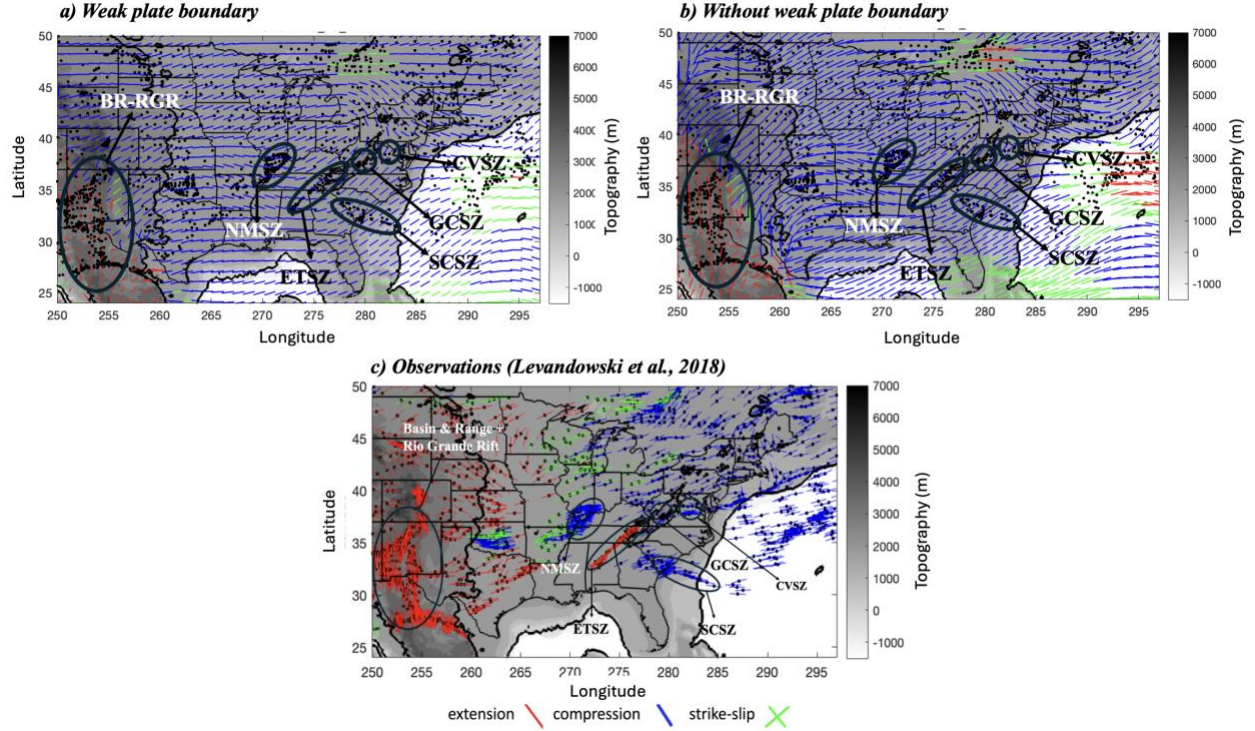


Figure 4.3: Tectonic regimes and SHmax orientations at 77 km depth, obtained (a) by imposing a weak plate boundary, WPB and considering the SEMUCB-WM1 tomography model, with the temperature-dependent viscosity and Adam et al. (2021a) scaling factor; (b) same parameters but without the WPB; (c) Observations. The black dots represent earthquake locations. See Fig. 4.1 for details.

4.4 Tomography Model Tests

Several regional and global tomography models have been developed using different approaches and datasets (discussed in section 3.1.1). Within this project, we test the influence of varying the input tomography models. Several global and regional tomography models have been considered. As discussed above (section 3.1.2), we convert the seismic velocity anomaly provided by these models into density anomalies using the conversion factor $R_{\rho/v}$, derived by Adam et al. (2021a). We then compute the instantaneous mantle flow using a depth-dependent viscosity model, where the base of the lithosphere is taken from the CAM2016 (Priestley et al., 2019). We also implement WPB, as several authors show this is a more realistic boundary (e.g., Ghosh et al., 2021). We then

compute the derived tectonic regimes and SHmax directions and compare them to the ones derived from focal mechanisms.

4.4.1 Global Tomography Models

The global tomography models we test here are some of the latest and most widely used models, including SEMUCB-WM1 (French & Romanowicz, 2014), 3D2018_08Sv (Debayle et al., 2016), TX2019slab (Lu et al., 2019), and SL2013sv (Schaeffer and Lebedev, 2013). We test these global models using weak plate boundary depth-dependent viscosity derived from CAM2016 (Priestley et al., 2019) and Adam et al. (2021a) conversion factor. The tectonic regimes and SHmax orientations at 77km depth, retrieved from the geodynamic models using SEMUCB-WM1 (French & Romanowicz, 2014), 3D2018_08Sv (Debayle et al., 2016), TX2019slab (Lu et al., 2019), and SL2013sv (Schaeffer and Lebedev, 2013) are as shown in Figs. 4.4 a, b, c, and d, respectively.

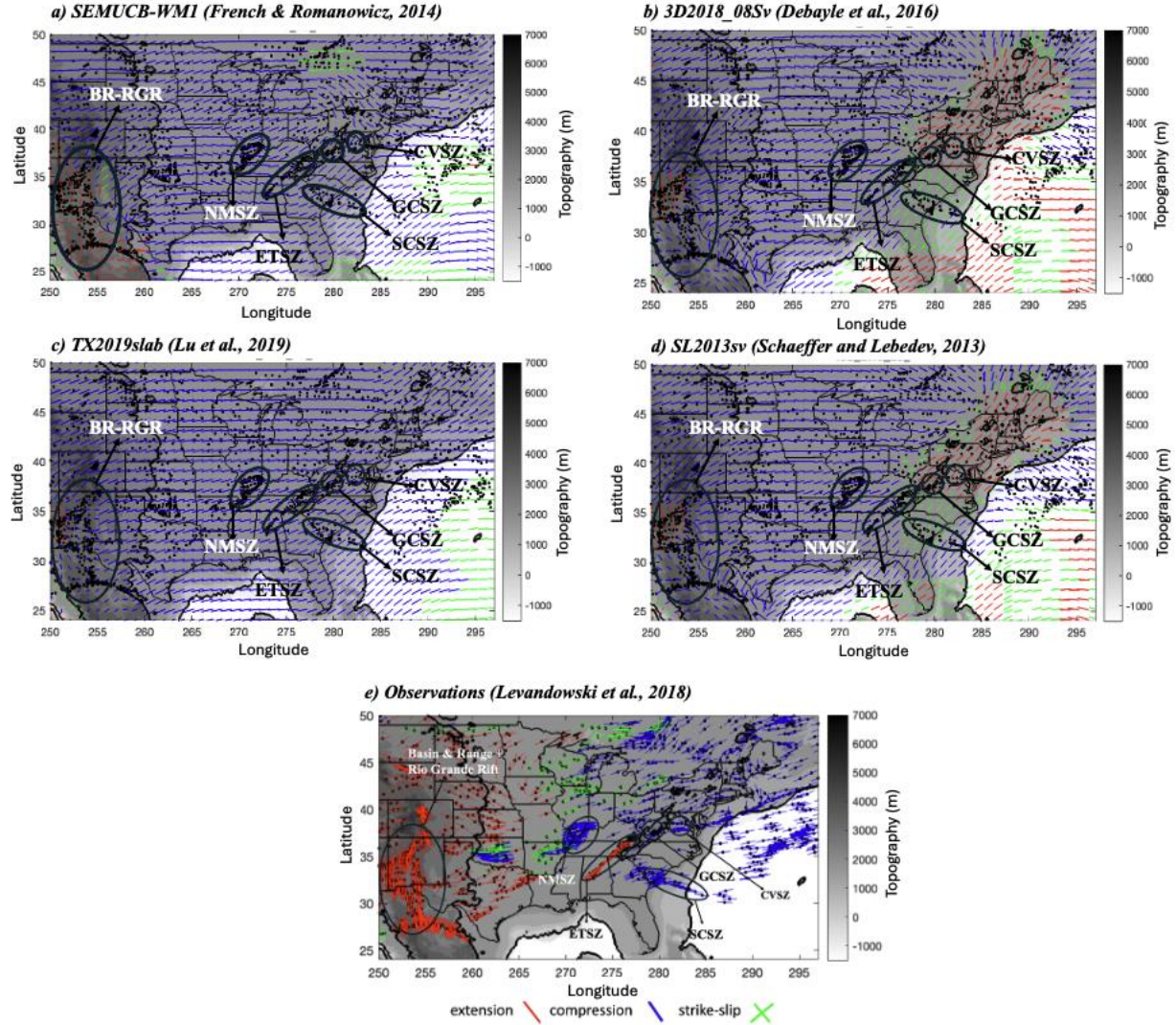


Figure 4.4: Variations of tectonic regimes and SHmax orientations at 77 km depth, obtained using (a) SEMUCB-WM1 (French & Romanowicz, 2014), (b) 3D2018_08Sv (Debayle et al., 2016), (c) TX2019slab (Lu et al., 2019), (d) SL2013sv (Schaeffer and Lebedev, 2013) seismic tomography models with weak plate boundary conditions imposed on depth dependent viscosity, $\eta(z)$, where the base of the lithosphere is taken from the CAM2016 (Priestley et al., 2021), WPB, and Adam et al. (2021a) conversion factor, and (e) Observations (Levandowski et al., 2018). The black dots represent earthquake locations. See Fig. 4.1 for details.

All global models retrieve the observed compression across the New Madrid Seismic Zone with the NE-SW oriented SHmax. They also retrieve the observed extension over the western CEUS features (Rio Grande Rift and Basin & Range) either partially (3D2018_08Sv, TX2019Slab, and SL2013sv, Figs. 4.4 b, c, d respectively) or completely (SEMUCB-WM1, Fig. 4.4a). The

geodynamic models using the SEMUCB-WM1 (French & Romanowicz, 2014) (Fig. 4.4a) and TX2019slab (Lu et al., 2019) (Fig. 4.4c) tomography models retrieve the regional compression observed across northeastern CEUS. Consequently, they retrieve the observed compression over the South Carolina, Giles County, and Central Virginia Seismic Zones. They both display a combination of compression and strike-slip regimes over the oceanic crust towards the eastern margin of the tectonic plate, a pattern that is not observed (Fig. 4.4e).

Furthermore, the geodynamic models where we use 3D2018_08Sv (Debayle et al., 2016) (Fig. 4.4b) and SL2013sv (Schaeffer and Lebedev, 2013) (Fig. 4.4d) seismic tomography models show a mix of extension, compression, and strike-slip regimes over the continental and oceanic crusts towards the eastern margin of the North American plate (latitudes 25° – 50° , and longitudes 277° – 295°) which does not exist in the observations (Fig. 4.4e). All the geodynamic models (Figs. 4.4 a-d) find compression across the Eastern Tennessee Seismic Zone, which contradicts the observed extension over the zone.

In Fig. 4.5 below, we show the modelled mantle convection patterns along latitude 40° , bound by longitudes 220° and 350° (across the United States) using the SEMUCB-WM1 model (Fig. 4.4a). The model shows upwellings (red regions) in the Mid Atlantic Ridge towards eastern US, and other ridge systems towards western US. It shows downwelling between longitudes 255° and 320° (blue regions) which corresponds to the thick lithospheric keel. This type of convection pattern (Fig. 4.5) creates compressive regimes in eastern US and suggests that the stresses are caused by far-field forces.

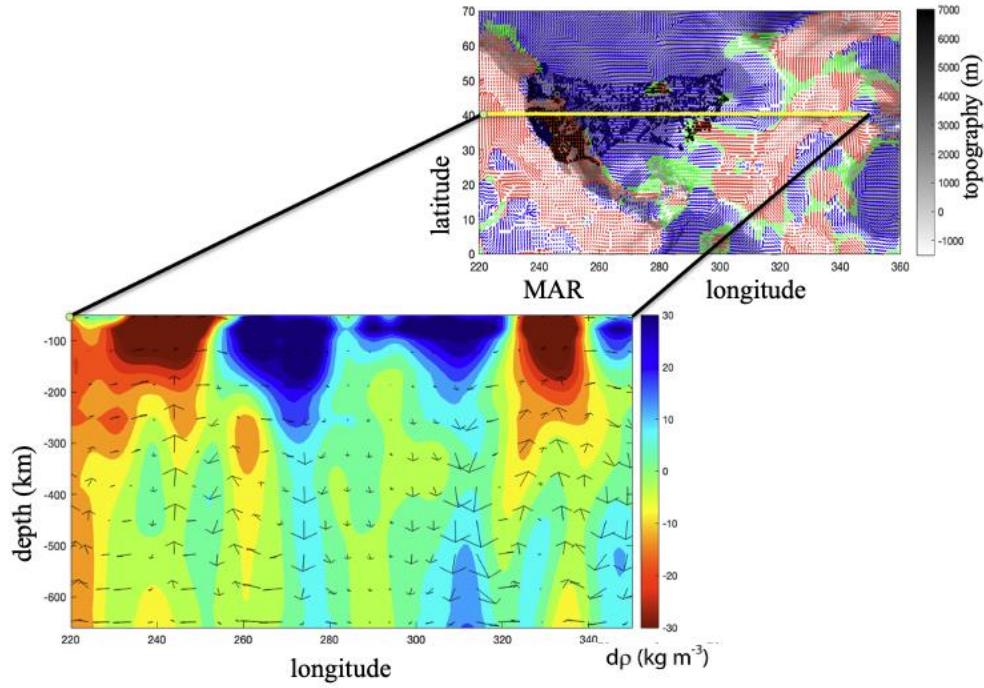


Figure 4.5: Mantle convection patterns retrieved with the SEMUCB-WM1 global model. The upper panel shows the tectonic regimes and SHmax directions at the 77 km depth. The extension, compression, and strike slip regimes are represented in red, blue, and green, respectively. The lower panel shows the density anomaly along the yellow profile, located at latitude 40°N. The arrows represent the instantaneous mantle flow.

4.4.2 Regional Tomography Models

To assess how variation of the input regional tomography models impacts the modeled tectonic regimes and SHmax, we examine some of the newly obtained tomography models, including CUSRA2021 (Zhou, et al., 2022), DNA13 (Porritt et al., 2014), CSEM_North_America (Fichtner et al., 2018). The instantaneous flow is computed by considering a depth dependent viscosity, $\eta(z)$, where the base of the lithosphere is taken from the CAM2016 (Priestley et al., 2019), and with weak plate boundaries, WPB. The conversion factor is from Adam et al. (2021a).

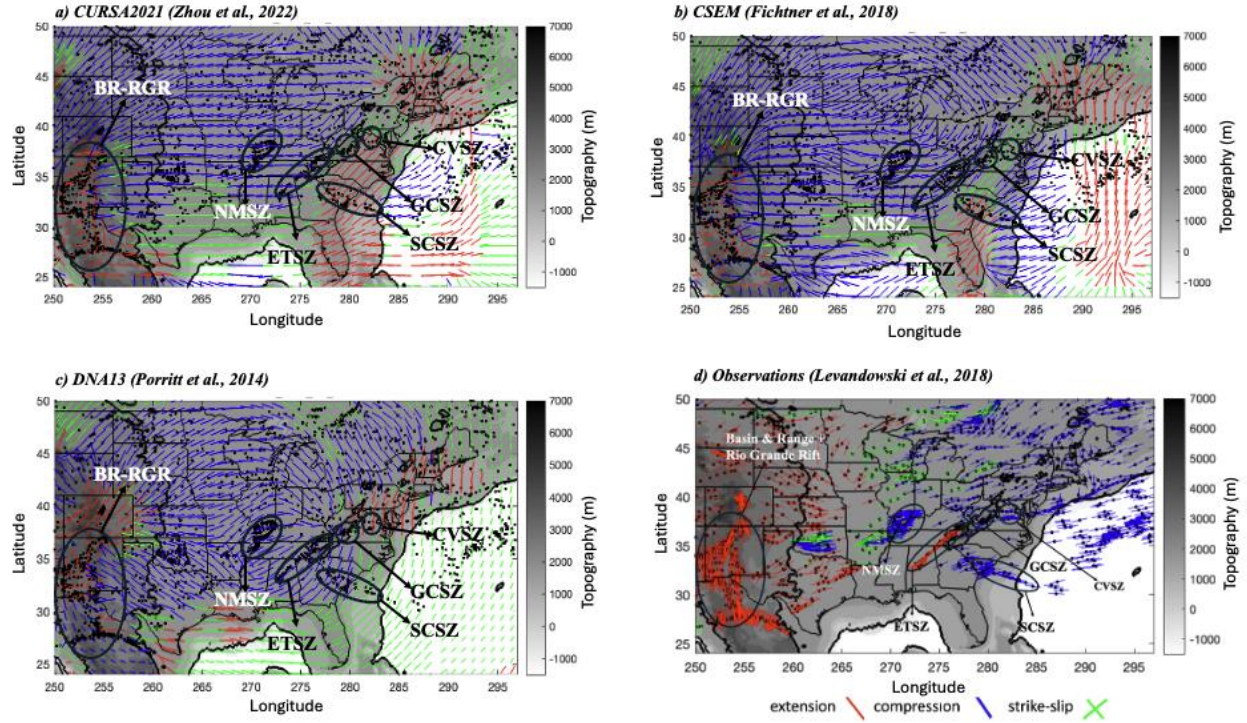


Figure 4.6: Variations of tectonic regimes and SHmax orientations at 77 km depth, obtained using (a) CURSA2021 (Zhou et al., 2022), (b) CSEM-North America (Fichtner et al., 2018), and (c) DNA13 (Porritt et al., 2014) seismic tomography models using depth dependent viscosity, $\eta(z)$, with WPB, where the base of the lithosphere is taken from the CAM2016 (Priestley et al., 2019), and Adam et al. (2021a) conversion factor, and (d) Observations. See Fig. 4.1 for details.

All the tested regional tomography models, including CURSA2021 (Zhou et al., 2022), CSEM-North America (Fichtner et al., 2018), and DNA13 (Porritt et al., 2014) (Figs. 4.6a, b, and c) retrieve compression over the New Madrid Seismic Zone with the regional NE-SW SHmax orientation, as well as extension across the western CEUS features (Rio Grande Rift and Basin & Range). However, the extensional regime over the western CEUS features is only partially retrieved with DNA13 (Fig. 4.6c). All three regional models (Figs. 4.6 a, b, and c) show a mix of extension, compression, and strike slip regimes over the continental and oceanic crusts towards the eastern margin of the continent, a feature that is not observed when we compare to observations from earthquake focal mechanisms (Fig. 4.6d). The compression observed over the Giles County Seismic Zone is retrieved with the CSEM and DNA13 models (Figs. 4.6b and c), while the

compressional regime over the South Carolina and Central Virginia Seismic Zones is not fully retrieved by any of the regional models.

4.5 Variations of tectonic regimes and SHmax with depth

To investigate how the stresses and tectonic regime vary with depth in a global geodynamic model, we use the SEMUCB-WM1 (French & Romanowicz, 2014) model with depth dependent viscosity, $\eta(z)$, where the base of the lithosphere is taken from the CAM2016 (Priestley et al., 2021) where we impose a weak plate boundary, and Adam et al. (2021a) conversion factor to show the variations in tectonic regimes and SHmax orientations at depths ranging from 77 km to 165 km (Figs. 4.7a – e).

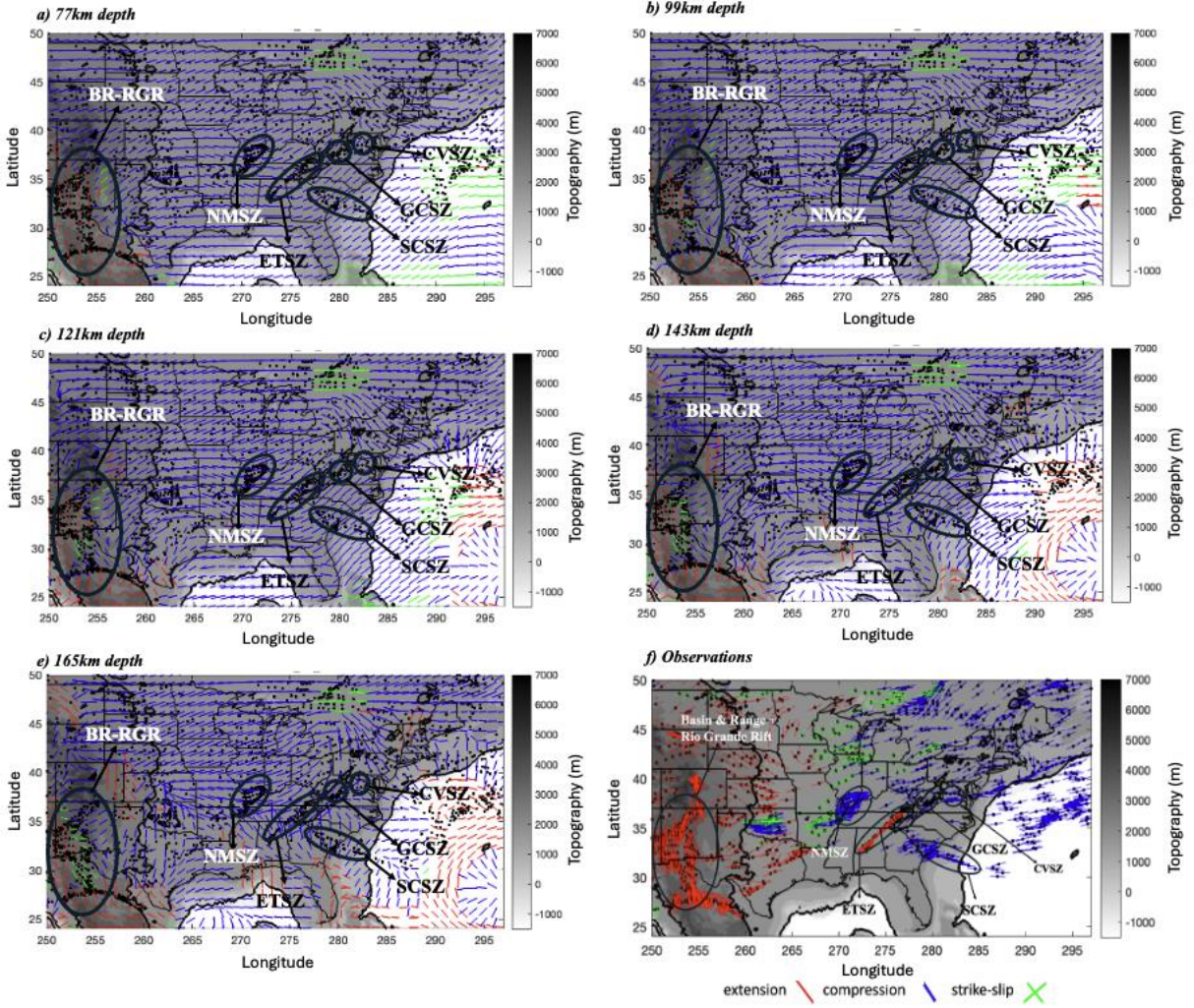


Figure 4.7: Variations of tectonic regimes and SHmax orientations with depth, obtained using SEMUCB-WM1 (French & Romanowicz, 2014) seismic tomography model with depth dependent viscosity, $\eta(z)$, where the base of the lithosphere is taken from the CAM2016 (Priestley et al., 2021), and Adam et al. (2021a) conversion factor. The black dots represent earthquake locations. See Fig. 4.1 for details.

SEMUCB-WM1 (French & Romanowicz, 2014) majorly retrieves compression in the central parts of CEUS from 77 km to 165 km depths (Figs. 4.7 a-e), as well as the extension over the Rio Grande Rift and Basin & Range. From 121 km depth (Fig. 4.7c), the model shows extensional regime at the eastern part of the study area, which widens in extent down to 165 km depth (Fig. 4.7e), corresponding with the upwelling convection patterns we see in Fig. 4.5.

4.6 Quantitative comparison of the geodynamic models.

To discuss the influence of the model parameters, we provide a qualitative and a quantitative comparison between the model outputs and the observations. For the quantitative fit, we have made a 1degree by one degree grid over our study area. At each node, we impose the value 0 if the computed and observed tectonic regimes are the same. We impose the value 1 if they differ. We then compute the RMS (root mean square) of this “difference” grid, as we described in Section 3.4. We computed the RMS for each depth, but in Figure 4.8 we report a value which is the median of the RMS found for depths between 33 km and 253 km. Best performing models are expected to have the smallest RMS.

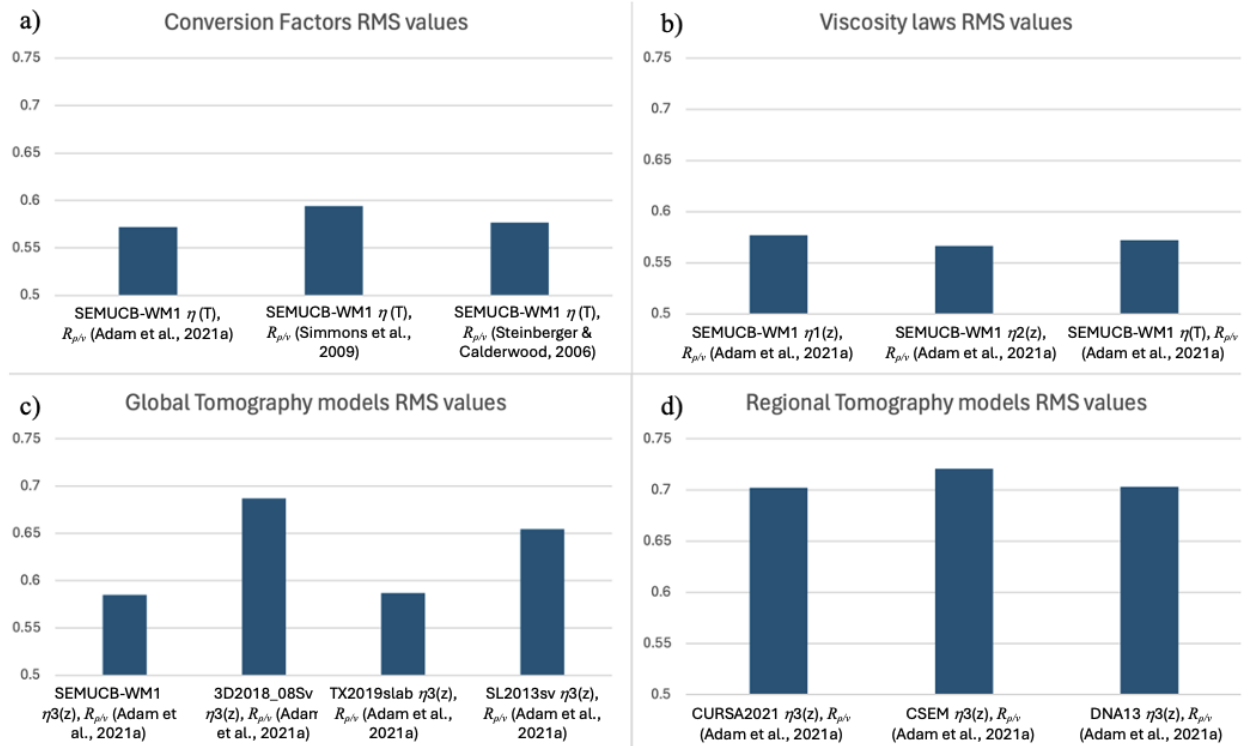


Figure 4.8: Comparison of RMS values for tested model parameters, including a) conversion factors, b) viscosity laws, c) global tomography models, and d) regional tomography models. See text for details.

The comparison we made in Fig. 4.8a shows that Adam et al. (2021a) conversion factor (0.572 RMS) retrieves the observed seismicity patterns better than other conversion factors. We expect this performance because it incorporates constraints from thermodynamics and anelasticity. Also, Fig. 4.8c shows that the global SEMUCB-WM1 (French & Romanowicz, 2014) and TX2019slab (Lu et al., 2019) performs better in retrieving the patterns better than other models. We notice however, that the RMS values vary between 0.56 and 0.72 so they are in a relatively similar range.

Chapter 5 - Discussion

In this project, we compute the instantaneous mantle flow induced by the density anomalies derived from seismic tomography models. We vary several model parameters, such as the input tomography models, the conversion factors used to convert the seismic velocity anomalies into density anomalies, $R_{\rho/v}$, and the viscosity laws, η . We then compute the stress tensor using the modelled convection velocities, and derive the tectonic regimes induced by the instantaneous mantle flow, as well as the direction of the maximal shear stress, SHmax. We investigate if the modeled tectonic regimes and SHmax account for the ones deduced from earthquake focal mechanism.

5.1 Sensitivity of geodynamic models to tested parameters.

One of the main conclusions of our study is that all the parameters we vary (i.e., the input tomography models, $R_{\rho/v}$, and η) have a significant influence on the modelled tectonic regimes and SHmax directions. This problem is then very complex. There are however some features that are retrieved for several model parameters. For example, using the SEMUCB-WM1 global tomography model (French & Romanowicz, 2014), the compression in the northeastern United States and the extension in the Rio Grande Rift and Basin & Range are retrieved for several $R_{\rho/v}$, and η . In this case, we need to incorporate other independent information for choosing “our favorite model”. For instance, one of the conversion factors we test in this project, $R_{\rho/v}$ (Adam et al., 2021a) incorporates constraints from thermodynamics and anelasticity, which makes it more pertinent than the others.

This project is among the first to conduct detailed tests on all geodynamic model parameters, providing a sensitivity study of these model parameters. So far, the geodynamic model with the global SEMUCB-WMI (French & Romanowicz, 2014), $R_{\rho/v}$ (Adam et al., 2021a), and depth dependent viscosity, $\eta(z)$, where the base of the lithosphere is taken from the CAM2016 (Priestley et al., 2021) retrieves the observed regimes in all CEUS seismic zones, except the Eastern Tennessee Seismic Zone whose tectonic regime is not agreed on by authors (e.g. Levandowski et al., 2019; Lund Snee and Zoback, 2020). Table 5.1 gives a summary of the comparison between the observed and modeled tectonic regimes in all the CEUS regions previously discussed. A quantitative comparison is provided in table 5.2, which shows the RMS between the models and the observations for depths 33-253 km, as well as the mean and median of these RMS. We provide both qualitative and quantitative estimations of the fits, as the qualitative comparison allow us to discuss the fits and lack of in more details, and the quantitative measures provide an unbiased estimate of “how good the fit is”. In general, both estimates agree with each other which is rather encouraging.

Table 5.1: Summary of geodynamic model retrieval of CEUS seismic zones based on the tested parameters. The notations for the viscosity laws are outlined in table 3.2 ($\eta_1(z)$ - depth-dependent viscosity including only a 100 km thick lithosphere, $\eta_2(z)$ - depth-dependent viscosity, where the base of the lithosphere is taken from the CAM2016 (Priestly et al., 2019), $\eta_3(z)$ - $\eta_2(z)$ with WPB, $\eta(T)$ - temperature dependent viscosity). C, T, and SS are used to denote Reverse, Normal, and Strike slip regimes, respectively.

Models/Regions	NE-US	Basin & Range Rio Grande Rift	NMSZ	ETSZ	SCSZ	CVSZ	GCSZ
Observation	C	T	C/SS	T	C	C	C
Tests on Conversion Factors							
SEMUCB $R_{p/v}$ -Adam $\eta(T)$	C 55-143 km	T 55 - 99 km	C – all depths	T partially – 55 – 77 km	C 77-99 km	C 77-187 km	C 77-187 km
SEMUCB $R_{p/v}$ -Simmons $\eta(T)$	C partially – all depths	T partially – all depths	C 121-209km	T partially – 33 – 99km, 143 - 209 km	C 55-143 km	C – all depths	C 121–209 km
SEMUCB $R_{p/v}$ -Steinberger & Calderwood $\eta(T)$	C partially -77 – 209 km	T partially 55 – 209km	C partially - 55 – 77 km Fully 99 – 209 km	T partially – all depths	C 99-121km	C 77-99km	C 121–209km
Tests on Viscosity							
SEMUCB $R_{p/v}$ -Adam $\eta_1(z)$	C partially 55 – 209 km	T – all depths	C – all depths C/SS 77 – 143km	T partially – all depths	C Nil	C 143 – 209 km	C – all depths
SEMUCB $R_{p/v}$ -Adam $\eta_2(z)$	C 55 – 143 km C partially 165 – 209 km	T 55 – 77 km T partially 99 – 209 km	C – all depths SS 209 – 231 km	T Nil	C 55 – 187 km	C - all depths	C - all depths
SEMUCB $R_{p/v}$ -Adam $\eta(T)$	C partially – all depths Fully at 99 km	T partially – all depths Fully at 55 km	C – all depths	T partially 165 – 209 km	C 77 – 121 km	C – 77 km C Partially all depths	C – all depths
WPB Tests							
SEMUCB $R_{p/v}$ -Adam $\eta_2(z)$	C 55 – 143 km C partially 165 – 209 km	T 55 – 77 km T partially 99 – 209 km	C – all depths SS 209 – 231 km	T Nil	C 55 – 187 km	C - all depths	C - all depths
SEMUCB $R_{p/v}$ -Adam $\eta_3(z)$ – WPB	C 55 – 143 km C partially 165 – 209 km	T 55 – 77 km T partially 99 – 209 km	C – all depths SS 209 – 231 km	T Nil	C 55 – 187 km	C - all depths	C - all depths
Tests on global tomography models							
SEMUCB $R_{p/v}$ -Adam $\eta_3(z)$	C 55 – 143 km C partially 165 – 209 km	T 55 – 77 km T partially 99 – 209 km	C – all depths SS 209 – 231 km	T Nil	C 55 – 187 km	C - all depths	C - all depths
3D2018 $R_{p/v}$ -Adam $\eta_3(z)$	C Nil	T partially 55 – 143 km	C 55-187 km	T Nil	C Nil	C Nil	C Nil
TX2019 $R_{p/v}$ -Adam $\eta_3(z)$	C 55-187 km	T partially 121-209 km	C – all depths	T Nil	C 55-187 km	C 55-165 km	C 55-165 km
SL2013sv $R_{p/v}$ -Adam $\eta_3(z)$	C Nil	T partially 55 – 143 km	C 55-187 km	T Nil	C partially 55-143 km	C Nil	C Nil
Tests on regional tomography models							
CURSA21 $R_{p/v}$ -Adam $\eta_3(z)$	C Nil	T – all depths	C 55-187 km SS Nil	T Nil	C Nil	C Nil	C Nil
CSEM $R_{p/v}$ -Adam $\eta_3(z)$	C Nil	T 55 – 143 km	C 55-187 km SS Nil	T Nil	C 143 – 187 km partially	C 143-187 km	C 55-187 km
DNA13 $R_{p/v}$ -Adam $\eta_3(z)$	C Nil	T 55 – 143 km	C 55-187 km SS Nil	T Nil	C 55-187 km partially	C Nil	C 77-165 km

Table 5.2: Summary of statistical values (RMS, mean and median) of the geodynamic models at various depths. The notations for the viscosity laws are outlined in table 3.2 ($\eta_1(z)$ - depth-dependent viscosity including only a 100 km thick lithosphere, $\eta_2(z)$ - depth-dependent viscosity, where the base of the lithosphere is taken from the CAM2016 (Priestly et al., 2019), $\eta_3(z)$ - $\eta_2(z)$ with WPB, $\eta(T)$ - temperature dependent viscosity).

Geodynamic models	33km	55km	77km	99km	121km	143km	165km	187km	209km	231km	253km	Mean RMS	Median RMS
SEMUCB-WM1 $\eta(T)$, $R_{p/v}$ (Adam et al., 2021a)	0.572	0.5478	0.481	0.5627	0.5831	0.5896	0.6045	0.6128	0.5831	0.5673	0.5543	0.5689	0.572
SEMUCB-WM1 $\eta(T)$, $R_{p/v}$ (Simmons et al., 2009)	0.6267	0.5915	0.5933	0.6091	0.6258	0.6156	0.5942	0.5961	0.5803	0.5636	0.5515	0.5953	0.5942
SEMUCB-WM1 $\eta(T)$, $R_{p/v}$ (Steinberger & Calderwood, 2006)	0.662	0.611	0.5562	0.5506	0.5942	0.5822	0.5803	0.5766	0.5645	0.5617	0.545	0.5804	0.5766
SEMUCB-WM1 $\eta_1(z)$, $R_{p/v}$ (Adam et al., 2021a)	0.5302	0.5515	0.546	0.5153	0.5376	0.5785	0.5766	0.584	0.5859	0.5905	0.5868	0.5621	0.5766
SEMUCB-WM1 $\eta_2(z)$, $R_{p/v}$ (Adam et al., 2021a)	0.5274	0.5534	0.5441	0.5757	0.5664	0.5552	0.5627	0.5766	0.5822	0.585	0.5859	0.565	0.5664
SEMUCB-WM1 $\eta_3(z)$, $R_{p/v}$ (Adam et al., 2021a)	0.5357	0.5395	0.5469	0.5868	0.5738	0.5692	0.585	0.5933	0.598	0.598	0.5952	0.5747	0.585
3D2018_08Sv $\eta_3(z)$, $R_{p/v}$ (Adam et al., 2021a)	0.7623	0.7484	0.7289	0.7447	0.7177	0.6871	0.6732	0.637	0.6221	0.6305	0.6332	0.6895	0.6871
TX2019slab $\eta_3(z)$, $R_{p/v}$ (Adam et al., 2021a)	0.5877	0.5915	0.5915	0.5868	0.5822	0.5655	0.5673	0.5785	0.585	0.5877	0.6128	0.5851	0.5868
SL2013sv $\eta_3(z)$, $R_{p/v}$ (Adam et al., 2021a)	0.7038	0.701	0.6825	0.7019	0.6852	0.6546	0.6527	0.6332	0.6091	0.6295	0.6286	0.662	0.6546
CURSA2021 $\eta_3(z)$, $R_{p/v}$ (Adam et al., 2021a)	0.7372	0.7363	0.7512	0.7688	0.753	0.7019	0.6472	0.5942	0.5952	0.6156	0.6314	0.6847	0.7019
CSEM $\eta_3(z)$, $R_{p/v}$ (Adam et al., 2021a)	0.7252	0.7242	0.7335	0.7502	0.7372	0.6778	0.6388	0.6258	0.6592	0.7001	0.7205	0.6993	0.7205
DNA13 $\eta_3(z)$, $R_{p/v}$ (Adam et al., 2021a)	0.7057	0.7103	0.7159	0.7409	0.7261	0.7029	0.636	0.6351	0.6472	0.6583	0.6518	0.6846	0.7029

When tested with several parameters, the global model SEMUCB-WM1 (French & Romanowicz, 2014) retrieves several features such as compression in the northeastern United States, extension

in the Rio Grande Rift and Basin & Range, and compression in the New Madrid, South Carolina, Central Virginia, and Giles County Seismic Zones for depths which represent roughly the base of the lithosphere. The global seismic tomography model TX2019slab (Lu et al., 2019) also retrieves the observed tectonic regimes across all the seismic zones we investigate within CEUS, excluding the Eastern Tennessee Seismic Zone.

However, the global models fail to retrieve the observed strike-slip deformation observed in the central part of CEUS within longitudes 270° and 280° , and latitudes 30° and 50° , as well as the observed extension at the eastern portions of the Rio Grande Rift and Basin & Range (longitudes 257° and 265° , and latitudes 30° and 50°). These non-retrieved deformation patterns must have been created by other phenomena, not mantle dynamics. In future studies, we will investigate which other phenomena are possibly responsible for these observed patterns.

Other considered global tomography models, including 3D2018_08Sv (Debayle et al., 2016) (Fig. 4.6b) and SL2013sv (Schaeffer and Lebedev, 2013) (Fig. 4.6d), do not retrieve most of the observed deformational patterns (see global tomography model tests section of table 4.1), such as Rio Grande Rift and Basin & Range, Giles County, Central Virginia, Eastern Tennessee Seismic Zones. We will investigate whether these discrepancies stem from variations in the used datasets and methods.

The regional tomography models do not generally retrieve the observed tectonic regimes in the seismic zones within the study area. This indicates that the deformation occurring in the central and eastern United States is dominantly controlled by mantle flow occurring on a larger scale than the CEUS regional scale (see Fig. 4.5).

Our results agree with some previous studies (e.g., Ghosh et al., 2019; Saxena et al., 2021a), which show that mantle flow can produce stresses as the origin of the intraplate seismicity observed in CEUS. However, we provide a more detailed analysis of the model parameters here.

5.2 Disparity in observed deformation over ETSZ

While comparing our models with observations obtained through computations of earthquake focal mechanics, we discovered a disagreement among authors regarding the deformation across the Eastern Tennessee Seismic Zone. The studies of Levandowski et al., (2018) show an extension regime across the ETSZ (Fig. 5.1), while that of Lund Snee & Zoback (2020) show a strike slip regime across the same seismic zone (Fig. 5.2). Other studies suggest a compressive regime over the ETSZ (e.g. Hurd & Zoback, 2012; Warrell et al., 2017) as retrieved by some of our models. Therefore, there is need to consider other synthesis of tectonic regimes in further studies.

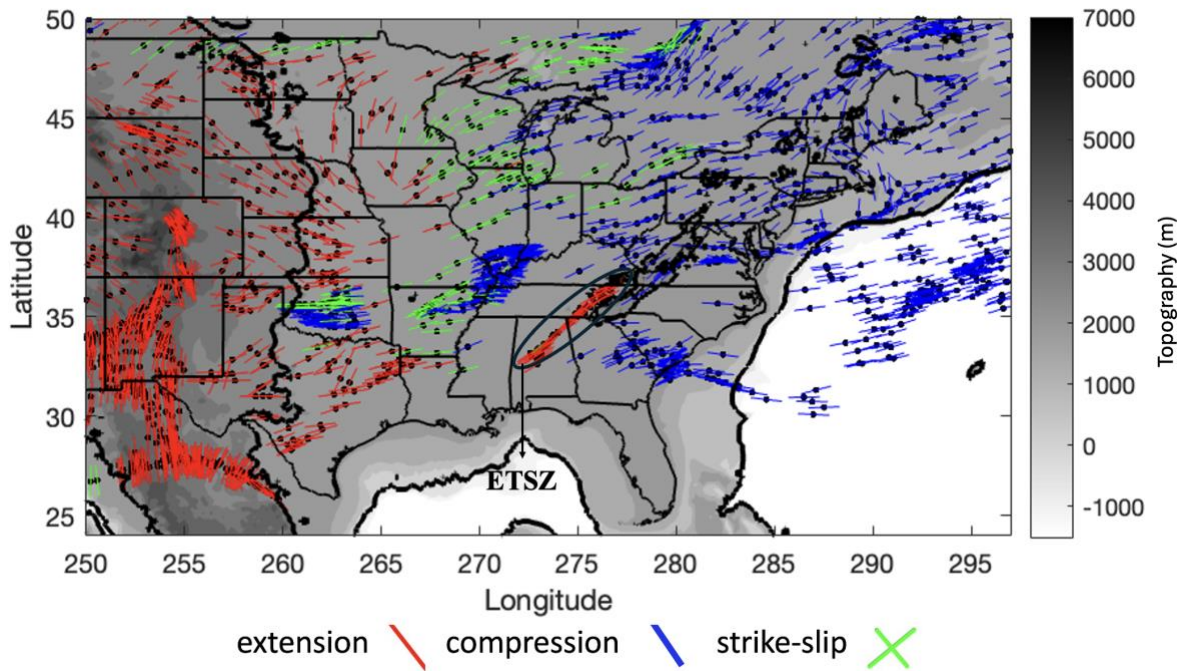


Figure 5.1: Levandowski et al. (2018) observed tectonic regime across the Eastern Tennessee Seismic Zone

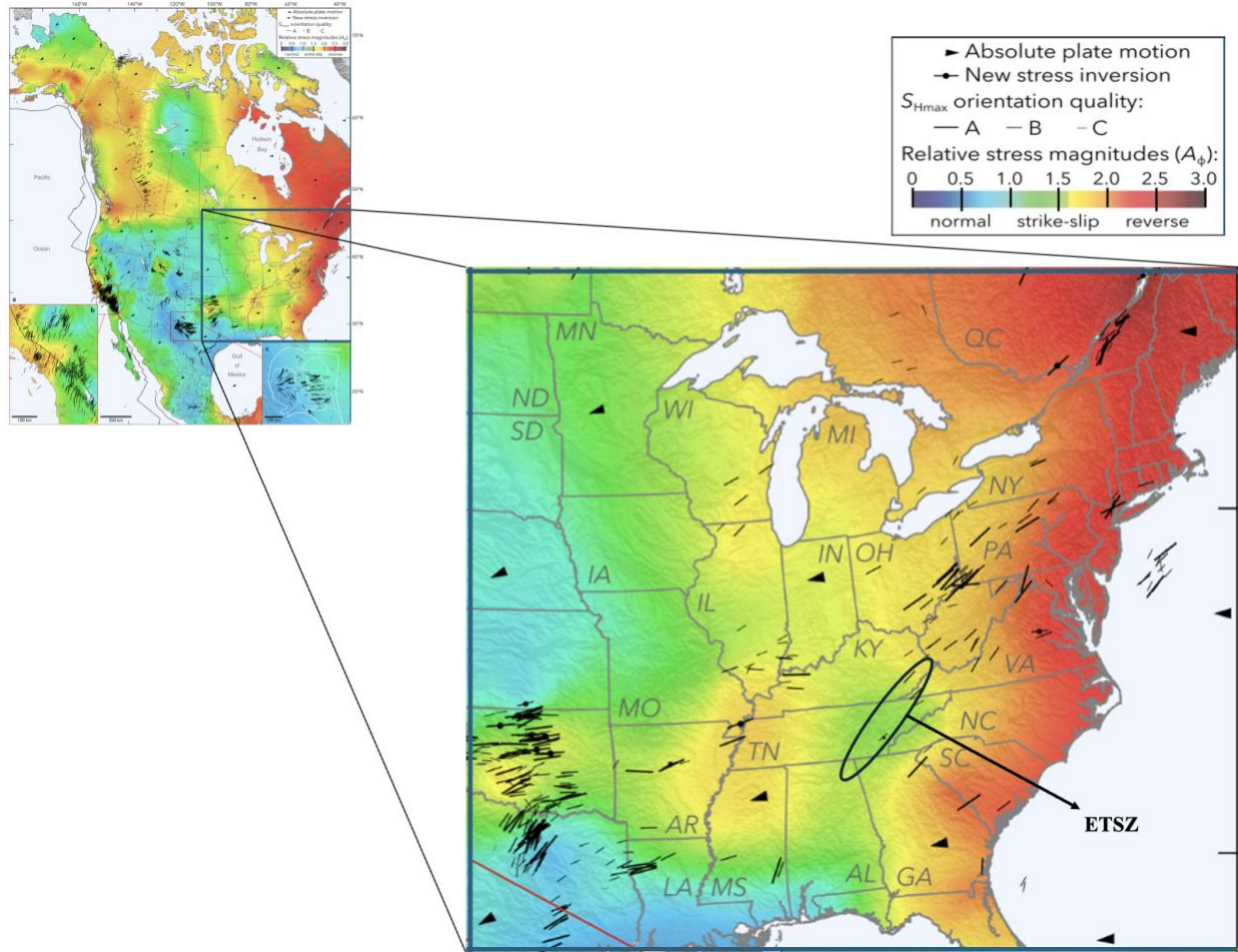


Figure 5.2: Lund Snee & Zoback (2020) observed tectonic regime across the Eastern Tennessee Seismic Zone.

Our models do not retrieve the observed extension regime over the ETSZ. However, this extension is only reported by Levandowski et al., (2018), while other authors disagree (Warrell et al., 2017; Hurd & Zoback, 2012; Lund Snee & Zoback, 2020). Before investigating this seismic zone further, we have to assess which is the focal mechanism (or focal mechanisms) representative of this region from seismic data.

Finally, if we had to pick a “favorite model”, the geodynamic model designed with the global SEMUCB-WMI (French & Romanowicz, 2014), the conversion factor $R_{p/v}$ provided by Adam et al. (2021a), and depth dependent viscosity, $\eta_3(z)$, retrieves most of the observed regimes in all

CEUS seismic zones, except the extension over the Eastern Tennessee Seismic Zone, the existence of which is actually debated (e.g. Levandowski et al., 2019; Lund Snee and Zoback, 2020). Further studies should consider the dynamic topography and the geoid created by the different models of structure and dynamics of the mantle and lithosphere.

Chapter 6 - Conclusion

We investigate the relationship between mantle dynamics and intraplate seismicity by modeling mantle convection from seismic tomography models. Within the project, we test the effects of several model parameters, such as (i) the conversion factor between seismic velocity anomalies and the density anomalies, (ii) the viscosity used in the geodynamical model, and (iii) the input tomography models. Unlike previous similar studies that consider one or two tomography models, we consider four global and three regional, recently published seismic tomography models. The results indicate that most of the model parameters, specifically the conversion factors, viscosity laws, and the input tomography models exerts significant influence on the modeled tectonic regimes and SHmax directions. The SEMUCB-WM1 (French & Romanowicz, 2014) and TX2019slab (Lu et al., 2019) global tomography models retrieve the tectonic regimes evidenced by the focal mechanisms across most of CEUS seismic zones (except Eastern Tennessee Seismic Zone) with most of the tested parameters. Most of the regional tomography models do not recover the observed seismicity pattern, except for the extension under the Basin & Range and Rio Grande Rift. This means that the seismicity occurring in CEUS is created by a mantle flow occurring at a scale bigger than the CEUS or the North American continent one.

Our “favorite model” is the geodynamic model designed with the global SEMUCB-WMI (French & Romanowicz, 2014), the conversion factor $R_{\rho/v}$ provided by Adam et al. (2021a), and depth dependent viscosity where the base of the lithosphere is taken from the CAM2016 model (Priestley et al., 2019), and weak plate boundaries, as it retrieves most the observed regimes in all CEUS seismic zones. In the future, we will be considering the constraints provided by the geoid to help assessing which is the most realistic model.

Going further, we will also incorporate the effect of glacial isostatic adjustment (GIA) and gravitational potential energy (GPE) and will also consider other synthesis of observed deformation patterns to harmonize the disparity in the deformation pattern over the Eastern Tennessee Seismic Zone.

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