

ABSTRACT

Title of Dissertation: Earth's Radiogenic Heat Production,
and Models for the Composition of
the Deep Continental Crust

Laura Sammon
Doctor of Philosophy, 2022

Dissertation Directed by: Professor William F. McDonough
Department of Geology

Much of the continental crust, the 40+ km thick plates of rock that make up the outer shell of our planet, is inaccessible to us living on its surface. Thus its composition is a mystery. We lack the technology to sample it directly at depths past 5 km, aside from a few deep (expensive) drill holes, so we must come up with a clever alternative for establishing its composition.

The deep crust, the lower two-thirds of the continent, serves as a supporting root. When continents collide, they make mountain ranges, or when pulled apart they make rift valleys and basins. The composition of the deep crust, and specifically its silica (SiO_2), molecular water, and heat producing element (HPE: K, Th, U) contents, directly influence the crust's rheology during tectonic events and its potential for deadly earthquakes. Its chemical makeup is the sum of 4.5 billion years of crustal evolutionary processes that continuously shape and reshape the platform upon which society sits. An accurate description of the deep crust, however, requires careful integration of many different data sources.

My research combines geochemistry with thermodynamics, geophysics, mineral-physics, seismology, and even particle physics to produce self-consistent models for the crust's composition. Using thermodynamic calculations, I generate densities and seismic sound wave speeds from a range of chemical compositions. Matching these forecasted models to Earth's seismic and gravity data allows me to translate the deep crust's physical properties into chemical compositions on both the regional and the global scale. Importantly, by quantifying not only the compositions, but also the uncertainties and the misfit in these results, I can better define the differences between competing models for crust deformation and evolution.

Charting the distribution of Earth's geochemical resources has led to our collaborations with particle physicists, who need our expertise to determine the frequency of radioactive decay and therefore the amount of HPE decay emissions (known as geoneutrinos) in the crust; this geoneutrino flux is the background signal in their nuclear physics experiments. Their global flux measurements constrain our models for heat production and the amount of radiogenic energy that heats the Earth – which provides power to mantle convection, plate tectonics, and the destruction and creation of more continental crust. Our main sources of data are threefold. First, we have critically compiled geochemical analyses of >10,000 rock samples from pre-existing literature (Earthchem.org and affiliates). Second, we use geophysical data provided by sources such as the United States Geological Survey, the Earthscope USArray, and others to determine which of our geochemical samples could produce Earth's observed seismic and density signals. Third, we partner with particle physicists in the United States, Canada, Italy, Japan, and China to jointly interpret data from three international geoneutrino detectors.

By focusing on Earth as a whole system we seek a comprehensive understanding of its natural hazards and resources. Using multidisciplinary constraints, my goal is to build compositional models of the continental crust, with quantifiable uncertainties, that can be applied regionally and at larger scales. These findings will provide predictive insights on the strength and response of the continents when subjected to the dynamic processes of plate tectonics.

Earth's Radiogenic Heat Production,
and the Composition of
the Deep Continental Crust

by

Laura Sammon

Dissertation submitted to the Faculty of the Graduate School of the
University of Maryland, College Park in partial fulfillment
of the requirements for the degree of
Doctor of Philosophy
2022

Advisory Committee:

Professor William F. McDonough, Chair/Advisor

Associate Professor Nicholas C. Schmerr

Assistant Professor Megan Newcombe

Dr. Ondřej Šrámek

Professor Gregory Sullivan, Dean's Representative

© Copyright by
Laura Sammon
2022

Dedication

To my friends, my family at Maryland - every success you made sure was celebrated, and every failure you made sure was not in vain.

Acknowledgments

I was told at the beginning of graduate school (I cannot remember the source) that you can only get two out of the three options: a good advisor, good research, or good coworkers. Luckily, I never really had to choose.

Though Bill started out as my academic advisor, he accumulated many titles over the years. He has been a graduate school mentor and colleague, yes, but also someone to confide in, to share my hopes and fears with. He was, at times, a PR representative, a sous chef, a taxi driver, an emergency contact, a travel agent, and a supplier of absolutely terrible “dad jokes”.

Many other collaborators helped push this research forward as well. I would like to acknowledge especially Walter Mooney at the USGS for his insight, data, and financial support for a trip to California that acted as a turning point in my research projects and my morale. I also want to thank many others: Ingrida Semenec from the SNO+ team, an incredible physicist and persistent friend; Wiesen Shen and Wolfgang Szwilius, who taught me enough so that I could pretend to be a geophysicist; Scott Wipperfurth, who was an invaluable mentor in the beginning of my program, especially when my advisor was halfway across the globe; and everyone who has read and reviewed my papers over the years, especially Roberta Rudnick, Derek Schutt, Derrick Hasterok, and Jamie Connolly, all of whom I bothered on multiple occasions.

I cannot even begin to name all of the graduate students and professors in the Geology Department at Maryland who have had such an incredible impact on my life. Never did I expect such congeniality from professors, though it is not surprising to find among geologists. The geology graduate student body at UMD was more of a dream team. Every new year, I made more friends, and it became harder and harder to watch old friends graduate and move on. I would not trade you all for the shiniest rock or biggest pot of coffee in the world.

And thank you so much to my loving family and support team. Throughout the PhD process you have stayed by my side, *on my side* when I needed it most. You believed when I doubted, you held me while I cried, and you listened (or tried very, very hard to) when I rambled about neutrinos, codes, presentations, and rocks in general.

My research was financially supported by NSF grants, the Ann G. Wylie Fellowship, a GSA graduate student research grant, ESSIC travel grants, the UMD Geology Department and Graduate School.

Table of Contents

Dedication	ii
Acknowledgements	iii
Table of Contents	v
List of Tables	viii
List of Figures	x
Chapter 1: Introduction	1
1.1 Neutrino Geoscience and Crust Composition	2
1.1.1 Geoneutrino Detection	5
1.2 Deep Crust Composition Models	6
1.2.1 Relating Chemical Composition to Seismic Observations	7
Chapter 2: Quantifying Earth's the radiogenic heat budget	11
2.1 Overview	11
2.2 Abstract	12
2.3 Introduction	13
2.4 Background	15
2.5 Lithospheric Modeling	19
2.5.1 Near-Field and Far-Field Lithosphere	20
2.5.2 Numerical Model	23
2.6 Importance of the Near-Field Lithosphere Model	25
2.7 Heat Flux Constraints on Lithospheric Models	28
2.8 The Future of Neutrino Geoscience	31
2.9 Conclusion	33
2.10 Acknowledgments	35
Chapter 3: A Geochemical Review of Amphibolite, Granulite, and Eclogite Facies Lithologies: Perspectives on the Deep Continental Crust	36
3.1 Overview	36
3.2 Abstract	37
3.3 Introduction	38
3.4 The Art and Science of Deep Crustal Modeling	40
3.5 Datasets	43

3.5.1	Amphibolite, Granulite, and Eclogite	43
3.5.2	Potential Biases	45
3.6	Major Element Compositions	50
3.6.1	SiO ₂ , MgO, FeO, and the Daly Gap	50
3.6.2	The Constancy of Al and Ga	58
3.7	Minor and Trace Element Composition	62
3.7.1	Rare Earth Elements	63
3.7.2	Heat Producing Elements	66
3.8	Distributions that Trend, Periodically	69
3.9	A Basalt, by Any Other Name	73
3.10	Constructing the Continental Crust	76
3.11	Conclusion	85
Chapter 4: Localized Lower Crustal Composition from Joint-Inversion of Geochemical and Geophysical Datasets		88
4.1	Overview	88
4.2	Abstract	89
4.3	Introduction	90
4.4	Background	93
4.4.1	Compositional Modeling of the Lower Crust	93
4.4.2	Geologic Setting	94
4.5	Methods	96
4.6	Results	99
4.7	Discussion	104
4.7.1	Lower Crust Composition	104
4.8	Conclusion	114
Chapter 5: The composition of the deep continental crust inferred from geochemical and geophysical data		115
5.1	Overview	115
5.2	Abstract	116
5.3	Introduction	117
5.4	Methods	118
5.4.1	Model Inputs	123
5.4.2	Model Uncertainties	126
5.4.3	Quality, Expense, and Time: Global vs. Local Models	128
5.5	Results	129
5.5.1	Empirical Composition-Velocity Trends	129
5.5.2	Deep Crustal Density	133
5.5.3	Composition	135
5.6	Discussion	139
5.6.1	SiO ₂ and Overall Deep Crustal Composition	139
5.6.2	CaO and Sr	146
5.6.3	Heat Production and Moho Heat Flux	148
5.7	Conclusions	158

Chapter 6: Conclusions and Future Work	160
6.1 Key conclusions	160
6.2 Future geoneutrino project locations	163
Appendix A: Supporting Information for Chapter 2	166
A.1 Full electron antineutrino flux equation	166
A.2 Heat production from K, Th, and U decay	167
Appendix B: Supporting Information for Chapter 3	168
B.1 Eu Anomalies	168
B.2 Fluid Mobile Elements	170
B.2.1 K, Rb, and Cs	170
B.2.2 La/Th and Th/U	173
B.2.3 Alkalis and Alkaline Earth Metals, in General	173
B.3 High Field-Strength Elements	174
B.4 Transition Metals	176
B.5 Halogens	179
B.6 Supplementary Tables	180
Appendix C: Supporting Information for Chapter 4	208
C.1 Compositional Maps and Uncertainties	208

List of Tables

2.1	Borexino Models for the upper crust in the NFL, bulk calculated Signal, and Radiogenic Power	34
3.1	Major element compositions for amphibolite, granulite, and eclogite . . .	55
3.2	Major oxide compositions from amphibolites, granulites, and eclogites compared to other basalts	74
3.3	Continental crust elemental abundances	77
3.4	Comparison of expected deep crustal physical properties	85
4.1	Uncertainties associated with deep crustal modeling parameters	99
4.2	Colorado Plateau SiO ₂ content	106
4.3	Northern Basin and Range SiO ₂ content	106
4.4	Southern Basin and Range SiO ₂ content	107
4.5	Major element concentrations in the lower crust of the southwestern US .	107
4.6	Summary of lower crustal seismic properties and mineralogies	108
5.1	Tectonic regimes by surface area	125
5.2	SiO ₂ concentration for different tectonic regimes	137
5.3	Calculated middle and lower crust bulk composition	139
5.4	Comparison of continental crust composition estimates	147
5.5	Heat production calculation parameters	153
5.6	Heat production in the continental crust	154
A.1	Heat production and geoneutrino flux results	166
A.2	Radionuclide heat production	167
B.1	Amphibolite Facies Lithologies	180
B.2	Granulite Facies Xenoliths	183
B.3	Post-Archean Granulite Terrains	186
B.4	Archean Granulite Terrains	189
B.5	Eclogite Facies Xenoliths	192
B.6	Eclogite Facies Terrains	195
B.7	Amphibolite Facies Lithologies	198
B.8	Granulite Facies Xenoliths	199
B.9	Post Archean Granulite Facies Terrains	200
B.10	Archean Granulite Facies Terrains	201

B.11 Eclogite Facies Xenoliths	202
B.12 Eclogite Facies Terrains	203

List of Figures

2.1	Geoneutrino signal strength vs. distance from Italian detector	16
2.2	U concentration in common crustal rocks	21
2.3	Schematic drawing of the location of the Borexino experiment and its Near-Field lithosphere	22
2.4	Near-Field Lithosphere vs. Mantle geoneutrino signal trade-off	23
2.5	Two different Near-Field Lithosphere models	24
2.6	Total geoneutrino signal vs. lithospheric signal	26
2.7	Geological and surface heat flow maps of the central Italian peninsula . .	27
2.8	Map of geoneutrino detectors	32
3.1	Mg# vs. SiO ₂ for medium- to high- grade metamorphics	52
3.2	Al Saturation Index for metamorphic lithologies	61
3.3	Rare earth element median values	64
3.4	Rare Earth element abundances in amphibolite, granulite, and eclogite . .	65
3.5	Elemental abundances in amphibolites	71
3.6	Elemental abundances in granulites	72
3.7	Major element abundances compared to basalt	74
3.8	MORB normalized rare earth element abundances in amphibolite, gran- ulite, and eclogite	75
3.9	Bulk crustal elemental abundances from different models	83
4.1	Sample collection and seismometer locations for this study	91
4.2	Perple_X predicted velocities and densities vs. SiO ₂	95
4.3	Modeling process flow chart	97
4.4	Geophysical properties of the southwestern US	102
4.5	Composition probability histograms for three southwestern US geological provinces	102
4.6	SiO ₂ abundance in the southwestern US	103
4.7	Seismic velocities vs. isochemical velocity prediction	105
4.8	Schematic of geological structure and composition in the southwestern US	113
5.1	Weighted area proportion of tectonic regimes	119
5.2	Global tectonic regimes and seismic velocity profile locations	121
5.3	Conceptual illustration of how velocity distributions are used to predict crust compositions	122
5.4	Vp vs. SiO ₂ for amphibolite and granulite facies lithologies	130

5.5	Vs vs. SiO ₂ for amphibolite and granulite facies lithologies	131
5.6	Vp/Vs vs. SiO ₂ for amphibolite and granulite facies lithologies	132
5.7	Density vs. depth for different types of continental crust	134
5.8	Global SiO ₂ concentration in the lower crust	136
5.9	Global SiO ₂ concentration at various depths	138
5.10	Global CaO concentration and uncertainty in the lower crust	140
5.11	Global Sr concentration and uncertainty in the lower crust	141
5.12	Global U abundances and uncertainties in the lower crust	149
5.13	Heat production in the layers of the continental crust	151
5.14	Moho heat fluxes for different tectonic provinces	155
5.15	Global Moho heat flux	156
5.16	Crustal U concentration vs. Moho heat flux	157
B.1	Eu anomalies in amphibolite, granulite, and eclogite	169
B.2	Fluid mobile element concentrations, amphibolite and granulite	171
B.3	La/Th vs. Th/U	172
B.4	Fluid mobile elements	175
B.5	Location map and SiO ₂ content of the metamorphic sample datasets	204
B.6	Location map and Mg# of the metamorphic sample datasets	205
B.7	Nb vs. Ta	206
B.8	Ln(Th) vs. Ln(U)	207
C.1	Vs as a function of temperature	208
C.2	Uncertainty from SiO ₂ calculations	209
C.3	Distribution of mafic oxides MgO + FeO	210
C.4	Uncertainty in MgO + FeO	211
C.5	Relative uncertainty in deep crustal composition	212
C.6	Median Vp from the joint model	213
C.7	Median Vs from the joint model	214
C.8	Median Vs from seismic data	215
C.9	Example seismic inversions	215

Chapter 1: Introduction

The processes that make Earth a unique and habitable planet are tied to its composition, past and present. The starting materials Earth inherited during its formation were crucial to early planetary differentiation ([McDonough et al., 2020](#)), and the resources that are left today still fuel its mantle convection, plate tectonics, and the geodynamo. This fuel is primarily expressed as heat. Earth’s total heat budget of 46 ± 3 terrawatts (TW, 10^{12} watts) ([Jaupart et al., 2016](#)), is a combination of primordial (kinetic) heat from planetary accretion and of radiogenic heat. With two sources of fuel, Earth’s internal engine is like that of a hybrid car. Unfortunately, we do not have a fuel gauge to tell us what proportion of heat is primordial or radiogenic. While humans are unlikely to be around when both of Earth’s fuel tanks reach “empty”, knowing Earth’s heat evolution and future trajectory tells us about the abundance of valuable elements, and helps us understand the cycles of heating and cooling in other rocky planets – and their potential for hosting life.

The radioactive decay of heat producing elements (HPEs: U, Th, K) contributes anywhere between 10 and 30 TW of energy to Earth ([Javoy, 1995](#); [McDonough and Sun, 1995](#); [Turcotte and Schubert, 2014](#)), and is the only process that has added significant heat to the Earth’s interior since roughly 4.4 billion years ago. Th and U in particular occupy unique positions in the study of Earth’s heat production and composition. First,

their radioactive isotopes, ^{232}Th , ^{238}U , and trivial amounts of ^{235}U , account for $>80\%$ of Earth's radiogenic heat generation (McDonough et al., 2020). Second, Th and U are *refractory* elements which condensed out of the solar nebula in conserved ratios to 26 other refractory elements (Bellini et al., 2013). Accurately and precisely determining the concentration of Th and/or U in the bulk Earth therefore allows us to calculate the abundances of those other refractory elements as well, which include Ca, Al, Nb, and the economically valuable rare earth elements.

Because of the chemical differentiation of Earth's layers, the highest concentration of HPEs is locked in Earth's continental crust. Unfortunately, we who live on the surface can only directly access the uppermost parts of the crust (the deepest drill hole, Kola Superdeep Borehole, was abandoned at 12 km depth). The deep continental crust and the mantle are inaccessible to traditional geological sampling methods. This dissertation advances the study of Earth's composition by reconciling *geoneutrino* fluxes measured by particle physicists with geochemical, geophysical, and seismological observations, to produce an internally consistent compositional model of Earth. The overarching goal of this research is to build an Earth compositional model with quantifiable uncertainties using an interdisciplinary approach, and to highlight the need for further collaboration among the geosciences.

1.1 Neutrino Geoscience and Crust Composition

Geoneutrinos are a direct probe into Earth's interior. Geoneutrinos, abbreviated $\bar{\nu}_e$, are electron antineutrinos emitted during the radioactive decay of Earth's natural isotope

systems. These antineutrinos carry with them unique energy tags, allowing us to identify each $\bar{\nu}_e$'s radioactive parent isotope. Measurements of geoneutrino fluxes can therefore directly, quantitatively constrain the number of Th and U decays occurring inside Earth.

Geoneutrinos are nearly massless, chargeless particles. Once generated, these ghost-like particles leave Earth to traverse the universe. Due to their elusive nature (interaction cross-section of $< 10^{-47} \text{m}^{-2}$), we only detect about one $\bar{\nu}_e$ per 10^{19} that pass through a one kiloton detector, or roughly 10 $\bar{\nu}_e$ per year. Modern geoneutrino detectors have the capability to measure $\bar{\nu}_e$ from Th and U isotopes, which carry enough energy to pass the instruments' 1.8 MeV reaction threshold. Because the decay rates of ^{232}Th and ^{238}U are well-known (McDonough et al., 2020), and the ratio of Th to U is (relatively) constant (Wipperfurth et al., 2018), we can calculate the absolute abundances of these HPEs from geoneutrino measurements.

Geoneutrino detectors, however, cannot measure directional information, meaning that the $\bar{\nu}_e$ from HPE decays in Earth's deep mantle look the same as $\bar{\nu}_e$'s from the crust. Earth's total geoneutrino emission signal is a combination of geoneutrinos produced from all of Earth's silicate layers. Negligible geoneutrinos are emitted from Earth's core (Wipperfurth et al., 2018). The number of $\bar{\nu}_e$ observed by physicists is therefore:

$$S_{total} = S_{crust} + S_{mantle} \quad (1.1)$$

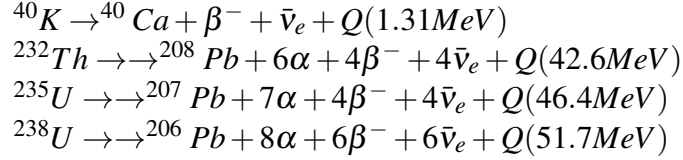
The total signal, S_{total} , is reported in Terrestrial Neutrino Units (TNU) to normalize measurements among different detectors. One TNU equals one antineutrino per one kiloton of scintillation fluid (10^{18} free protons) per year. S_{total} is proportional to the con-

centrations of Th and U divided by the square of their distance (r^{-2}) from the detector. Although the mantle is Earth's largest silicate reservoir, variations in the local crust's concentrations of Th and U have the strongest effect on the $\bar{\nu}_e$ signal simply because the local crust is closer to the detector (smaller r).

To understand how many geoneutrinos are coming from the mantle, and therefore how much radioactive heat is left to power mantle convection, we must first determine S_{crust} , a crustal correction for the abundance of Th and U. Chapter 2 of this dissertation shows that slight differences in the modeled upper crustal correction vastly impact the final calculation of radiogenic heat in the mantle. Chapter 2's discussion highlights the importance of understanding the geology local to each geoneutrino detector and demonstrates why geoscientists need to take an active role in the calculation of the geoneutrino signal. Chapters 3 – 5 quantify the composition and uncertainties of the deeper portions of the continental crust through a combination of geochemical data analyses and geophysical observations. Due to its inaccessibility and poorly constrained major and trace element abundances, the composition of the deep continental crust has been a point of debate among geoscientists for decades (see [Dumond et al. \(2018\)](#); [Hacker et al. \(2015\)](#) for summaries of these debates). However, we combine independent information from geochemical studies (Chapter 3) with petrological, geophysical, and seismological information (Chapters 4, 5) to construct a more precise model for deep crustal composition. Our joint geochemical-geophysical approach to determining the major element and HPE content of the deep crust is effective on a local (Chapter 4) and a global (Chapter 5) scale, both of which are necessary for calculating the local (Near-Field) and global (Far-Field) contribution of Th and U to the total geoneutrino signal.

1.1.1 Geoneutrino Detection

Geoneutrinos are emitted during the decay of four primary nuclides, along with energy, reported as heat (Q) in MeV:



As stated earlier, though, only $\bar{\nu}_e$ from the Th and U decay chains are energetic enough to accomplish the inverse beta decay reaction (IBD, equation 1.2). In kiloton-sized detectors of scintillating fluid, $\bar{\nu}_e$ collides with a free proton (p^+ , hydrogen nucleus) transforming it into a positron (e^+) and a neutron (n). This reaction results in two flashes of light that are correlated in time, space, and energy.

$$\bar{\nu}_e + p^+ \rightarrow e^+ + n \quad (1.2)$$

The emitted positron promptly annihilates with an electron, producing two photons summing to 1.022 MeV as the first detectable flash of light. About ~ 200 microseconds later, the neutron thermalizes and is captured by a nearby proton to form a ${}^2\text{H}^+$ deuterium nucleus. This reaction produces another flash of light with a discrete energy of 2.2 MeV, (the nuclear binding energy for deuterium). The two flashes of light and the separation timer of 200 μs allow for the unique identification of $\bar{\nu}_e$ with low background noise.

1.2 Deep Crust Composition Models

The deep continental crust's chemical composition is central to the debate of crustal formation, evolution, strength, temperature, and bulk composition. The inaccessible nature of the deep crust forces reliance on analogue samples and modeling results to determine its bulk composition. Medium- to high-grade continental metamorphic lithologies serve as geochemical markers for the deep crust. These amphibolite, granulite, and eclogite facies rocks are sampled through tectonically emplaced or eroded exposures of metamorphic rock (terrains) or through deep crustal xenoliths that are rapidly carried to the surface through volcanic eruptions. The difficulty remains in determining whether these isolated occurrences of deep crustal material are representative of the deep crust as a whole. On the other hand, geophysical surveying techniques, such as seismic velocity measurements, provide a means of measuring large portions of the deep crust *in situ*, but only report on the physical properties of the deep crust rather than its chemical state. However, a combination of both geochemical and geophysical analyses can place constraints on deep crustal composition. The physical properties of small-scale geochemical samples, measured in laboratory experiments, can be compared to large scale observations of the crust.

[Christensen and Fountain \(1975\)](#) were among the first to formally predict lower crustal compositions from laboratory derived Vp velocities in granulite. By comparing laboratory experiments to large seismic datasets, the authors identified average seismic and compositional trends in the deep crust. [Holbrook et al. \(1992\)](#), [Rudnick and Fountain \(1995\)](#), and successive publications offered interpretations of suites of seismic sur-

veys in the context of global crustal structure, incorporating even more petrological and geochemical analyses of deep crustal samples. Following the work of Holbrook et al., [Christensen and Mooney \(1995\)](#) developed a model which addressed the global crustal structure through synthesizing large seismic data sets. To combat sparse data coverage in remote areas, such as Antarctica or central Africa, the models extrapolate seismic profiles of various tectonic provinces to regions of similar crustal structure (e.g., island arcs, orogens, Archean crust, etc.). The CRUST and LITHO family of models ([Bassin et al., 2000](#); [Mooney et al., 1998](#); [Pasyanos et al., 2014](#)) has applied the same extrapolation method to generate what are now the most prevalent of global seismic models for the lithosphere. CRUST models define V_p , V_s , density, and thickness structures for different tectonic settings by relating V_p field measurements to V_s and density through empirical seismic observations. Recent iterations of similar techniques (e.g., [Szwillus et al., 2019](#)) more explicitly outline their methods and quantify the statistical uncertainty in their models.

1.2.1 Relating Chemical Composition to Seismic Observations

Comparing geochemical measurements to geophysical observations requires translation between bulk rock composition and seismic velocities. A rock's composition controls its mineralogy, which in turn dictates its physical properties, such as its density, compressibility, and shearability. The bulk compressibility modulus (K), shear modulus (μ), and density of a rock define the velocities at which seismic waves travel through it:

$$V_p = \sqrt{\frac{K + \frac{4}{3}\mu}{\rho}} \quad (1.3)$$

$$V_s = \sqrt{\frac{\mu}{\rho}} \quad (1.4)$$

where V_p is compressional wave velocity and V_s is shear wave velocity.

Thermodynamic modeling softwares can use Voigt¹-Reuss²-Hill³ (VRH) averaging to build bulk rock compositions from individual mineral constituents. One such software, *Perple_X* (Connolly, 2005), predicts equilibrium mineral assemblages from bulk rock compositions using Gibbs free energy minimizations. From these mineral assemblages, VRH averaging is used to generate bulk rock properties. *Perple_X* derives mineral phase relationships from experimental and empirical input databases, e.g. from the experiments conducted by Holland and Powell (2004). Connolly (2005) offers an overview of the software's free energy minimization technique for calculating mineral assemblages. Chapter 3 of this dissertation explains in detail how V_p and V_s are predicted for different deep crustal lithologies, the temperature and pressure variations considered, and the uncertainties associated with converting major oxide compositions to seismic velocities. Chapters 4 and 5 then compare *Perple_X*-derived seismic velocities to field observations of V_p and V_s .

Trace elements, such as Th and U, cannot be directly derived from seismic properties. By definition, they are not major rock forming constituents and therefore cannot be modelled as thermodynamic rock components. However, many trace elements have quantifiable relationships to one or more major oxides. Chapter 3 analyzes the predictable

¹W. Voigt, *Lehrbuch der Kristallphysik* (B. B. Teubner, Leipzig, 1928), p. 739.

²A. Reuss, *Z. Angew. Math. Mech.* 9, 49 (1929)

³R. Hill, *Proc. Phys. Soc. (London)* 65,349 (1952)

behaviors between HPEs and SiO_2 . Chapter 5 then generates trace element maps as a function of major oxide abundance using a simple bivariate probability analysis.

The goal of this project was to investigate Earth's radiogenic heat production by placing multiple, independent constraints on its composition. The research presented here focuses specifically on creating joint geochemical-geophysical models for the deep continental crust. The inaccessibility and heterogeneity of the deepest portions of the continental crust compels us to reconcile datasets from different fields of geology and physics. By incorporating geochemical, petrological, and geophysical datasets, and accounting for the uncertainties within each, this dissertation lays one more brick in the foundation of a a holistic Earth model.

In addition, our models of continental crust composition assess the concentration of heat producing elements (K, Th, and U) in areas local to geoneutrino detectors and throughout the Earth. The detection of geoneutrinos, $\bar{\nu}_e$, emitted from within Earth is a tool for directly measuring Earth's heat production capacity and ultimately the abundance of > 25 elements. These geoneutrino measurements, however, require accurate crustal composition corrections. This document outlines a multidisciplinary approach to solving for crustal composition, of the deep crust in particular, first on the local, then on the global scale.

Each new chapter's specific undertakings depend on observations and techniques developed in previous parts of the project. Chapter 2 outlines the importance of accurate crustal composition models for geoneutrino and radiogenic heat production calculations.

In-depth geochemical characterization of amphibolite and granulite samples (Chapter 3), including major element and HPE correlations, builds an extensive library of possible deep crustal compositions. Thermodynamic modeling translates these compositions to seismic velocities, which are then compared to both regional (Chapter 4) and global (Chapter 5) seismic data. The conclusion to this dissertation (Chapter 6) discusses the future of neutrino geoscience and crustal compositional modelling, including which avenues should be pursued next.

Chapter 2: Quantifying Earth's the radiogenic heat budget

2.1 Overview

Earth's internal heat drives mantle convection and plate tectonics. Measurements of geoneutrino fluxes show what portion of that heat is radiogenic, as opposed to primordial. For a geoneutrino flux measurement to be interpreted into total radiogenic heat, though, a lithospheric correction must be applied. An accurate geochemical understanding of the rocks surrounding geoneutrino detectors is of the utmost importance. Here, we show how two different local geological models around the Borexino detector in Italy lead to two very different radiogenic heat results. We show that Borexino's results agree with other geoneutrino experiments when Italy's regional geology is properly taken into account.

[1] This chapter has been submitted to *Earth and Planetary Science Letters* as:

Sammon, L. G., & McDonough, W. F. (2022). Quantifying Earth's radiogenic heat budget.

[2] LGS and WFM contributed to the conceptualization of this project. LGS constructed the synthetic models. LGS and WFM wrote and edited this manuscript together.

2.2 Abstract

Earth’s internal heat drives its dynamic engine, causing mantle convection, plate tectonics, and the geodynamo. These renewing and protective processes, which make Earth habitable, are fueled by a primordial and radiogenic heat. For the past two decades, particle physicists have measured the flux of geoneutrinos, electron antineutrinos emitted during β^- decay. These ghost-like particles provide a direct measure of the amount of heat producing elements (HPE: Th & U) in the Earth and in turn define the planet’s absolute concentration of the refractory elements. The geoneutrino flux has contributions from the lithosphere and mantle. Detector sensitivity follows a $1/r^2$ (source detector separation distance) dependence. Accordingly, an accurate geologic model of the Near-Field Lithosphere (NFL, closest 500 km) surrounding each experiment is required to define the mantle’s contribution. Because of its proximity to the detector and enrichment in HPEs, the local lithosphere contributes $\sim 50\%$ of the signal and has the greatest effect on interpreting the mantle’s signal.

We re-analyzed the upper crustal compositional model used by [Agostini et al. \(2020\)](#) for the Borexino experiment. We documented the geology of the western Near-Field region as rich in potassic volcanism, including some centers within 50 km of the detector. In contrast, the Agostini study did not include these lithologies and used only a HPE-poor, carbonate-rich, model for upper crustal rocks in the surrounding ~ 150 km of the Borexino experiment. Consequently, we report $3\times$ higher U content for the local upper crust, which produces a 50% decrease in Earth’s radiogenic heat budget, when compared to their study. Results from the KamLAND and Borexino geoneutrino experiments are at

odds with one another and predict mantle compositional heterogeneity that is untenable. Combined analyses of the KamLAND and Borexino experiments using our revised local models strongly favor an Earth with ~ 20 TW present-day total radiogenic power. The next generation of geoneutrino detectors (SNO+, counting; and JUNO, under construction) will better constrain the HPE budget of the Earth.

2.3 Introduction

A combination of primordial and radiogenic energy drives Earth’s engine, with the former coming from planetary accretion and the latter from decay of K, Th, and U. Our planetary vehicle lacks a fuel gauge to define the amount of fuel left to power plate tectonics, mantle convection, and the geodynamo. Defining the thermal evolution of the planet gives insights into the cooling and crystallization history of the core, the temporal variation in mantle viscosity, and the nature of the cosmic building blocks of the Earth. With the dawn of geoneutrino detection ([Araki et al., 2005](#)), we now have the opportunity to define the Earth’s radiogenic fuel budget, which in turn can specify the proportional contribution of these heat producing elements (K, Th, U) in the crust and mantle.

Twenty years have passed since particle physicists began detecting the Earth’s emission of geoneutrinos (chargeless and near-massless particles emitted during β^- decay) ([Araki et al., 2005](#)). The first generation of detectors (KamLAND in Japan and Borexino in Italy) have reported their flux measurements and interpreted their data in the context of an assumed geological model. The precision of the flux measurement (σ) continues to improve with exposure time, as it follow counting statistics ($\sigma \sim 1/\sqrt{N}$, N =number

of observed events). The accuracy of the interpretation and its uncertainties depends on the assumed geological model. To interpret the geoneutrino flux measurement, one uses a detailed assessment of the Th and U abundances and distribution in the lithosphere surrounding the detector (closest ~ 500 km, which typically contributes 40 to 50% of the measured signal). A reference model is assumed for contributions from the remaining global lithosphere and mantle, with the Earth's core having negligible quantities of K, Th, and U, and no significant contribution to the signal. Combined analyses of the results from the KamLAND and Borexino experiments favor an Earth with ~ 20 TW present-day total radiogenic power (or a ~ 16 TW Earth for just Th and U power) ([Bellini et al., 2021](#); [McDonough et al., 2020](#)). This finding indicates that $\sim 40\%$ of the Earth's estimated power of 46 ± 3 TW ([Jaupart et al., 2015](#)) comes from radiogenic sources.

Controversy remains, however, regarding the assumed geological model used to describe the local lithospheric contribution to the geoneutrino flux. For the lithosphere surrounding the KamLAND detector the various geological models predicting the local 3D distribution of Th and U differ by a factor of 1.4, based on their reported geoneutrino fluxes ([Enomoto, 2006](#); [Huang et al., 2013](#); [Wipperfurth et al., 2020](#)). In contrast, for the Borexino detector the various predictions differ by a factor of 3 ([Agostini et al., 2020](#); [Coltorti et al., 2011](#); [Huang et al., 2013](#); [Wipperfurth et al., 2020](#)). The interpretation of the regional geology is important for geoneutrino studies as it fundamentally influences the final result, and the global abundances of Th and U.

The latest interpretation of geoneutrino data from the Borexino experiment ([Agostini et al., 2020](#)) predicts a low contribution from their local crust to the overall geoneutrino signal. Consequently, their inferred mantle geoneutrino signal is high (~ 25 TW

from Th+U), as well as their calculation for the bulk Earth’s radiogenic power (~ 38 TW from K+Th+U), with model uncertainties at $\sim 34\%$ (Agostini et al., 2020). This prediction contrasts with other geoneutrino experiments (Agostini et al., 2015; Gando et al., 2013) and numerous geochemical (Huang et al., 2013; Javoy et al., 2010; McDonough and Sun, 1995, e.g.) and geophysical (Jaupart et al., 2016; Turcotte and Schubert, 2014, e.g.) models for Earth. Agostini et al. (2020) places their upper limit of uncertainty at 51 TW of radiogenic heat production, which is outside of all geological observations.

Here we review the data for constructing a local geological model for the lithosphere immediately surrounding the Borexino detector. We evaluate the local geological model used in Agostini et al. (2020) and compare it with competing models. We then test whether such models are consistent with the known regional geology and heat flux constraints. Using these findings, we identify the best local lithospheric models for the Borexino experiment. Relying on the same principles, we discuss the competing local lithospheric models for the next generation of geoneutrino experiments.

2.4 Background

Neutrinos are weakly-interacting fundamental particles that stream freely through matter, carrying information about their decay source. Detection of electron antineutrinos ($\bar{\nu}_e$) is accomplished via the Inverse Beta Decay (IBD) reaction with a free protons (p): $\bar{\nu}_e + p \rightarrow e^+ + n$ [n , neutron; e^+ , positron] with an energy threshold of $E_{\bar{\nu}_e}^{thr} = 1.8$ MeV. This restriction allows detection of only the highest energy antineutrinos produced during some of the β^- decays in the ^{238}U and ^{232}Th decay chains (Araki et al., 2005).

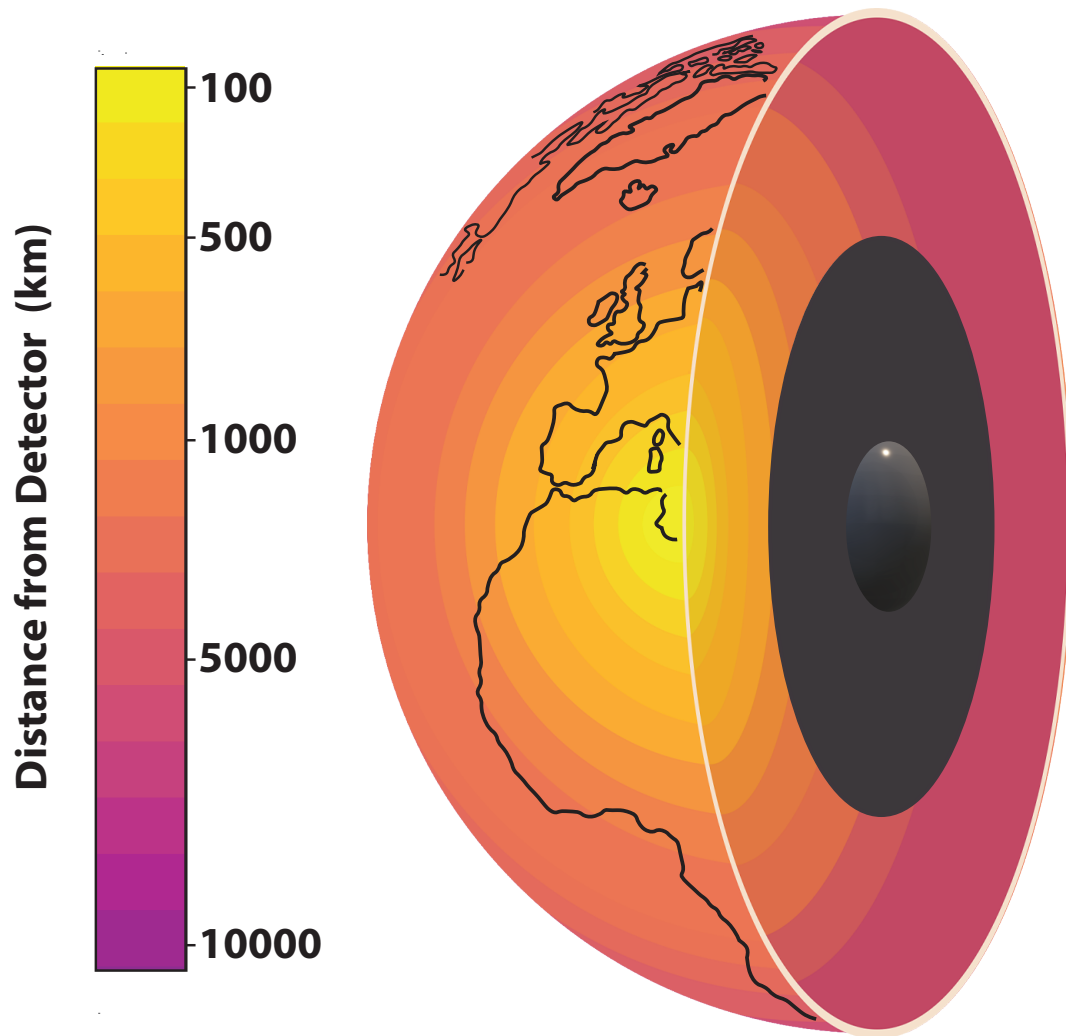


Figure 2.1: The strength of a geoneutrino signal depends on the abundance of the emitter (Th or U), and the $1/\text{distance}^2$ from the emitted to the detector, regardless of direction. A detector in central Italy (Borexino) sees the strongest signal (yellow) from its immediate surrounding geology and the weakest signal from the opposite side of Earth (pink). The outer and inner core do not contribute to the geoneutrino signal and are grayed-out.

Earth's total geoneutrino emission comes from the lithosphere and mantle, with the number of $\bar{\nu}_e$ observed (i.e., S , signal) by physicists is therefore:

$$S_{total} = S_{lithosphere} + S_{mantle} \quad (2.1)$$

S_{total} is reported in Terrestrial Neutrino Units (TNU) to normalize between detectors of different sizes; 1 TNU equals 1 antineutrino detection per 1 kiloton of scintillation fluid (10^{32} free protons) per year of exposure in a 100% efficient detector. S_{total} is proportional to the concentration of U times its sensitivity factor (α) and the same for Th, divided by the square of their distance (r) from the detector:

$$S_{total} \propto \frac{\alpha[U] + \beta[Th]}{r^2} \quad (2.2)$$

Sensitivity factors account for a neutrino's interaction cross-section, which scales with its discrete energy, and accounts for the higher energy neutrino from U being more detectable than those from Th, despite $4\times$ lower concentration levels of the former.

Figure 2.1 shows the sensitivity of S_{total} relative to distance from the detector in central Italy. At a known decay rate, a relatively constant $(^{232}\text{Th}/^{238}\text{U})_{molar}$ value (Wipperfurth et al., 2018), and an assumed K/U value, we calculate the abundance of the heat producing elements (K, Th, and U; HPEs). Please refer to Supplementary equation S1-Eq1 for the full calculation of the total $\bar{\nu}_e$ signal.

Compositional variations in the local lithosphere have the strongest effect on the geoneutrino signal because the lithosphere is closer to the detector (smaller r) and is 100-fold enriched in HPE relative to the mantle. Although the Earth's mantle is largest silicate

reservoir, its low U concentration (≤ 10 ng/g) and distance (greater r) causes its signal to be muted.

To determine the contribution of geoneutrinos from the mantle, and therefore how much radioactive heat is left to power mantle convection, plate tectonics, or the geodynamo, we must first determine the U and Th concentrations in the lithosphere surrounding the detector. Subtracting the lithospheric signal from the total signal is done to establish the mantle value and its Th and U content. The $S_{lithosphere}$ has Near-Field Lithospheric (NFL) and Far-Field Lithospheric (FFL) contributions. Thus, the mantle geoneutrino signal is:

$$S_{mantle} = S_{total} - (S_{NFL} + S_{FFL}) \quad (2.3)$$

The relative contributions of these components are: Near-Field lithosphere (40 to 50%), Far-Field lithosphere (30 to 40%, i.e., global lithospheric signal), and mantle ($\leq 25\%$) (Wipperfurth et al., 2020). The lithosphere includes the mechanically coupled, underlying lithospheric mantle, which has limited compositional variation (McDonough, 1990) and contributes little (order ~ 1 TNU, $< 10\%$ of the signal) to the lithospheric signals (Huang et al., 2013). Araki et al. (2005) observed that the first 50 km and 500 km from KamLAND contributes $\sim 25\%$ and $\sim 50\%$, respectively, of the total signal.

Modeling uncertainties: The relative uncertainties on the flux measurement at KamLAND and Borexino experiments improve over time; KamLAND went from $\sim 54\%$ to $\sim 15\%$ uncertainty for its measured flux, while Borexino went from $\sim 42\%$ to $\sim 19\%$. The modern mantle with depleted and enrich domains is predicted to show only $\sim 10\%$ total

variation in its geoneutrino signal ([Šrámek et al., 2013](#)). Likewise, only $\sim 10\%$ relative variation is observed in estimates of the Far-Field lithospheric signal. Typically, the upper crust (i.e., the top 1/3 of the crust) contributes $\sim 70\%$ of the geoneutrino signal from the lithosphere. Hence, the greatest impact on interpreting the mantle signal comes from accurately predicting the upper crustal composition, that is, the S_{NFL} .

2.5 Lithospheric Modeling

Disentangling the mantle’s contribution to S_{total} is a major goal of geoneutrino studies. Doing so requires accurate models for $S_{Lithosphere}$. Importantly, uncertainties (statistical and systematic) in the NFL model contribute most significantly to uncertainties in the modern mantle and global results.

Given the limited ($\pm 10\%$) variation in the mantle’s signal, one expects its predicted values from different geoneutrino experiments to agree at this level. However, the local estimates of the modern mantle S_{mantle} range from $\sim 30 \pm 13$ TNU (power from K, Th, and U) by the Borexino team ([Agostini et al., 2020](#)) to $\sim 7 \pm 1.6$ TNU by the KamLAND team ([Gando et al., 2013](#)). Consequently, the disparate nature of these findings either means (1) the mantle is grossly heterogeneous (i.e., beyond scales envisaged by geology), or (2) there are substantial inaccuracies in lithospheric modeling.

The distribution, volume, composition (HPE content), and petrology of the formations surrounding a detector must be accurately determined for its contribution to S_{NFL} . Shales and granites are enriched in HPEs, whereas peridotites and carbonates normally are not. However, the degree of HPE enrichment is variable even within a given rock type.

HPE concentrations differ among igneous, metamorphic, and sedimentary rocks, and between silicate and carbonate lithologies (Figure 2.2). It is therefore crucial to model accurately the proportional contribution of each geological formation and its HPE content near a detector.

The Borexino geoneutrino experiment at Gran Sasso National Laboratory was located outside of L'Aquila, Italy, in the central Italian peninsula (Figure 2.3, 13.57°E, 42.45°N, with 1.4 km of rock overburden). The Apennines consist primarily of fore-land basin sediments and siliciclastic foredeep basin sediments, covered by Middle Pleistocene to Recent volcanics (on the western side) and continental shelf and marine deposits (Cosentino et al., 2010; Vignaroli et al., 2019). The marine deposits are mainly dolomitic (marble, where metamorphosed). Extensional forces from mantle spreading to the west of the Apennines have led to a fault-block system of grabens filled with terrigenous sediments in a region known as the Tyrrhenian Extensional Zone (Cosentino et al., 2010). As a result, the uppermost crust near the detector could contain a mixture lithologies ranging from < 1 ppm to > 4 ppm U (Figure 2.2).

2.5.1 Near-Field and Far-Field Lithosphere

The Near-Field Lithosphere (NFL) is oftentimes, for the sake of computational ease, treated as the $4^\circ\text{latitude} \times 6^\circ\text{longitude}$ area centered on the detector (Huang et al., 2013), rather than a circle with a 500 km diameter. The Far-Field Lithosphere (FFL) consists of the rest of the Earth's lithosphere (oceanic and continental). The crucial step, which requires geoscientific expertise, is determining the concentration and distribution of HPEs

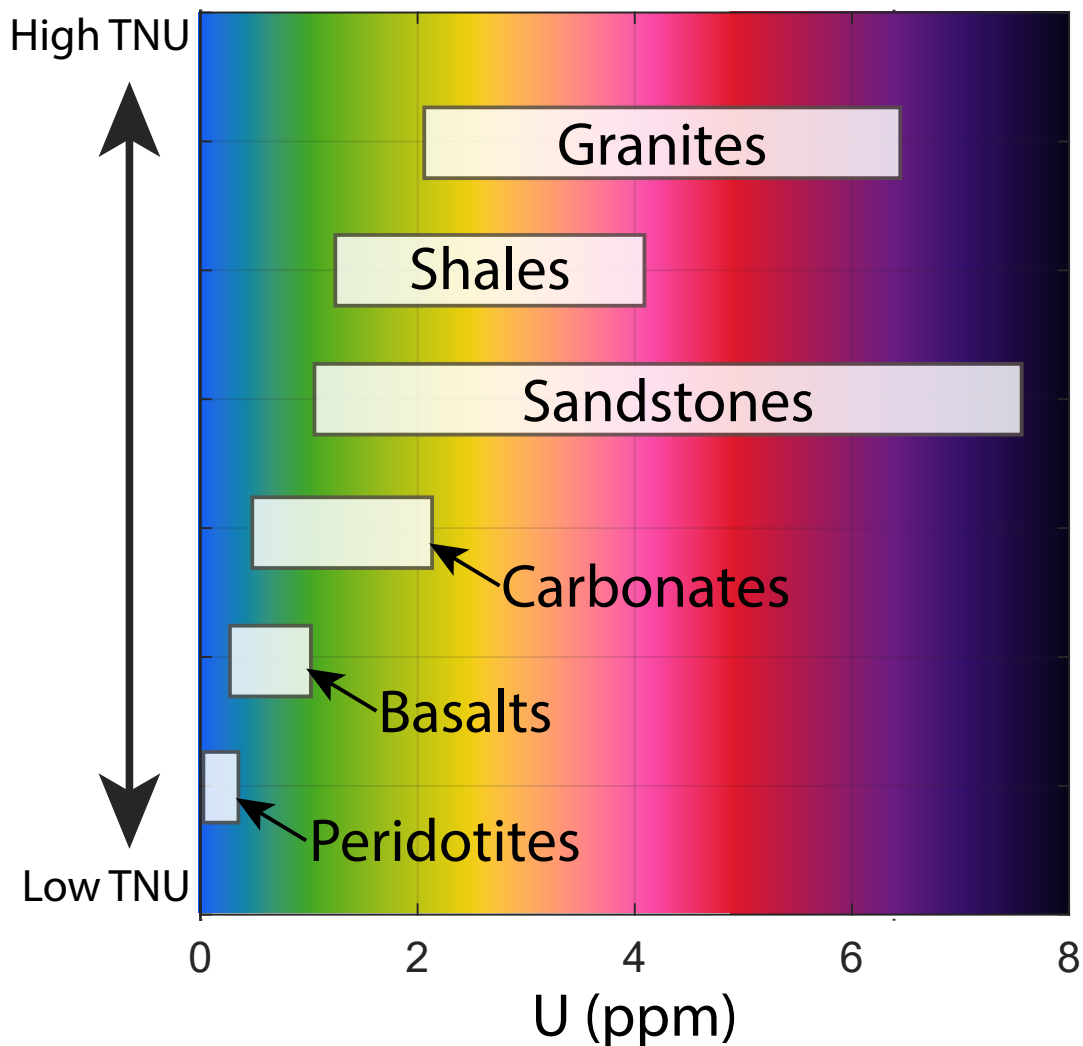


Figure 2.2: The average and range of U (and Th) depends on rock lithology. Granites tend to have higher average HPE content while carbonates and mafic rocks have lower averages. Sandstones, on the other hand, can have a wide range of U content depending on their formation and surrounding rocks. The white bar for each rock type shows the interquartile range of U concentrations from the Earthchem.org Database <https://www.earthchem.org>. See text for the definition of TNU.

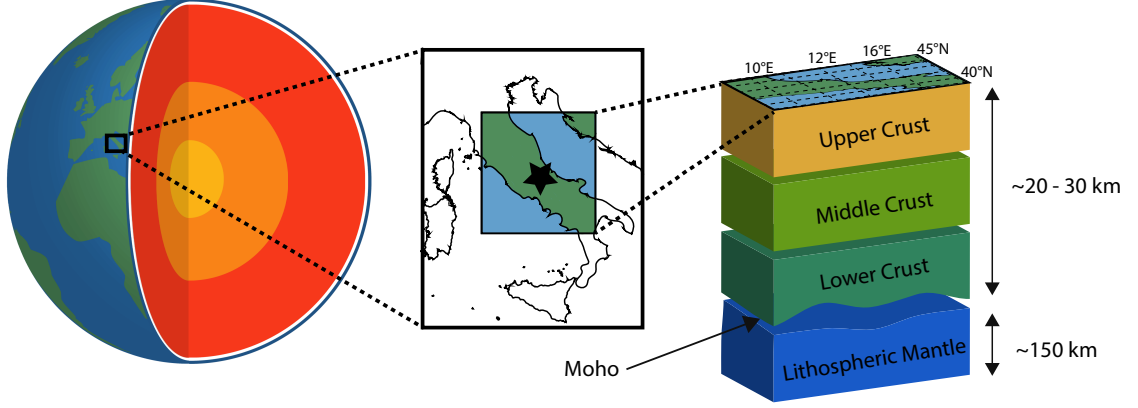


Figure 2.3: A schematic drawing of the location of the Borexino experiment and its Near-Field lithosphere (NFL; highlighted colored map in the center). Though the global abundance of U and Th contribute to the measured geoneutrino signal, the (continental) crust immediately surrounding the detector has the strongest effect on the signal.

in the lithologies of the Near-Field Lithosphere.

S_{FFL} is a global average of the continental and oceanic lithospheric contribution to a detector's farfield geoneutrino flux. Model predictions for the S_{FFL} at existing and future planned detector sites are consistent, with estimates agreeing at better than the $\pm 20\%$ level. The competing predictions for $Borexino S_{FFL}$ agree at 16 ± 1 TNU (Agostini et al., 2020; Coltorti et al., 2011; Huang et al., 2013; Wipperfurth et al., 2020).

Whether a signal is from a moderate source of heat producing elements in the lithosphere near the detector or from a more concentrated mantle source is where discrepancies are introduced. To illustrate this point, and to highlight the need for accurate lithospheric models for the area surrounding geoneutrino detectors, we walk through the impacts of two different scenarios of upper crustal concentrations for Th and U near the Borexino geoneutrino detector.

Figure 2.4 illustrates the signal trade-off between HPE content of the Near-Field Lithosphere and mantle. S_{total} depends on the total mass of HPEs and their distance from

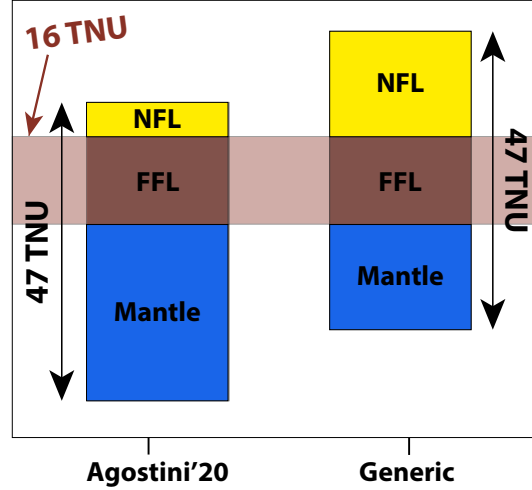


Figure 2.4: The total geoneutrino signal, S_{total} (length of the boxes in the figure) measured at a given detector remains relatively constant over time; the uncertainty decreases as the number of geoneutrino events detected increases. The amount of signal attributed to the Near-Field Lithosphere (NFL, yellow) determines how much signal must come from the mantle (blue). The average signal of the Far-Field Lithosphere (FFL, brown) generally stays the same (i.e., 16 ± 1 TNU for Borexino).

the detector. The non-uniqueness of the modeling drives us to construct more accurate 3D descriptions of the HPE contents of the Near-Field Lithosphere, to evaluate better the mantle HPE concentrations.

2.5.2 Numerical Model

Figure 2.5 presents two NFL models used to analyze the effects of vastly different abundances of Th and U in the upper crust surrounding the Borexino detector: (1) a low Th+U content (e.g., dominantly carbonate) and (2) medium Th+U content (e.g., shale-like, or averaged carbonate + siliciclastics + volcanic). These idealized models are comparable to those reported in (1) [Agostini et al. \(2020\)](#) and [Coltorti et al. \(2011\)](#), and (2) [Huang et al. \(2013\)](#), [Wipperfurth et al. \(2020\)](#), and [McDonough et al. \(2020\)](#).

Using Monte Carlo numerical modeling ([Wipperfurth et al., 2020](#)), we determined

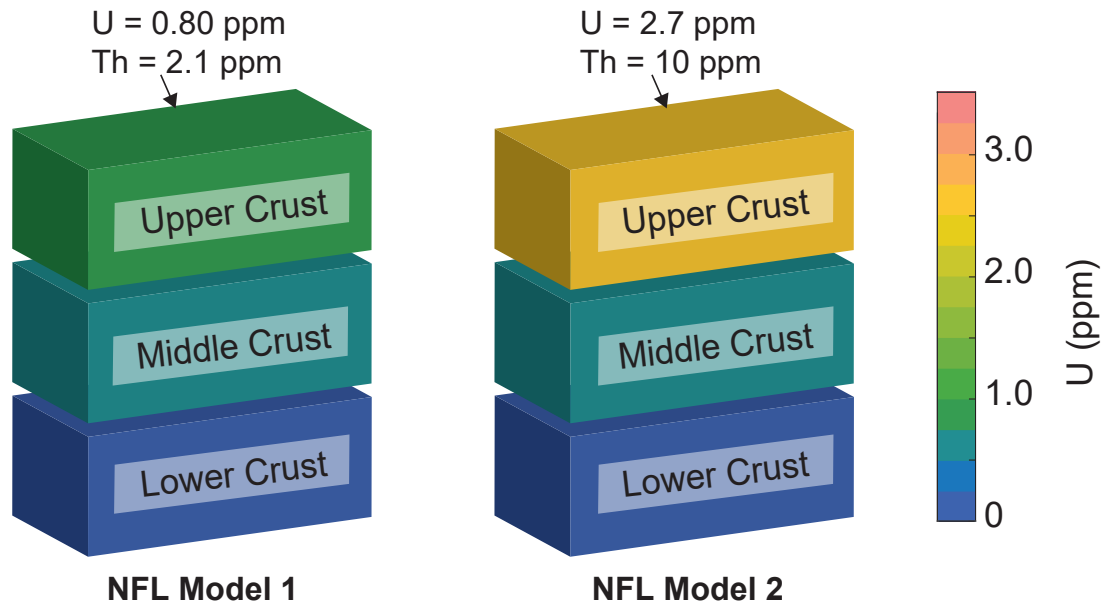


Figure 2.5: Two different Near-Field Lithosphere models illustrate low (Agostini'20) and medium (Generic) U and Th scenarios in the uppermost crust near our geoneutrino detector. The middle and lower crust are kept the same among the three models since we are primarily interested in the effects of upper crustal compositional changes. See [Huang et al. \(2013\)](#) for discussions on middle/lower/deep crustal geoneutrino contributions.

the expected $^{Borexino}S_{NFL}$ assuming two different scenarios: low and medium HPE contents for the upper crust. The HPE content of the middle, and lower crust and lithospheric mantle are taken from [Sammon and McDonough \(2021\)](#); [Sammon et al. \(2022\)](#). For the physical description of the local lithosphere, we use the LITHO1.0 model ([Pasyanos et al., 2014](#)) (i.e. density, distance from detector) with 1°latitude x 1°longitude horizontal resolution for the upper, middle, and lower crust and lithospheric mantle. Table 2.1 lists the compositional model parameters for the NFL, its signal, and that for the total lithosphere and mantle. This table also reveals the predicted power of the mantle and bulk Earth for these two different upper crustal models and thus NFL models.

A factor of three difference in the HPE budget of the upper crust for these two NFL models produces a factor of ~ 2 difference in both the estimated mantle and bulk Earth radiogenic power (Figure 2.6). These gross differences in the predicted radiogenic power demonstrate the significance of producing an accurate NFL model.

2.6 Importance of the Near-Field Lithosphere Model

The Apennines of the central Italian peninsula exposes a geological paradox across its eastern and western divide. Its Adriatic eastern side is composed of a compressional fold and thrust belt, whereas its Tyrrhenian western side is composed of extensional fault-block mountains. The paradox of this mountain belt is the juxtaposition of both compressional and extensional tectonic forces over a relatively narrowed (~ 150 km) east-west traverse.

Figure 2.7 shows that carbonate sediments surround the Borexino detector, whereas

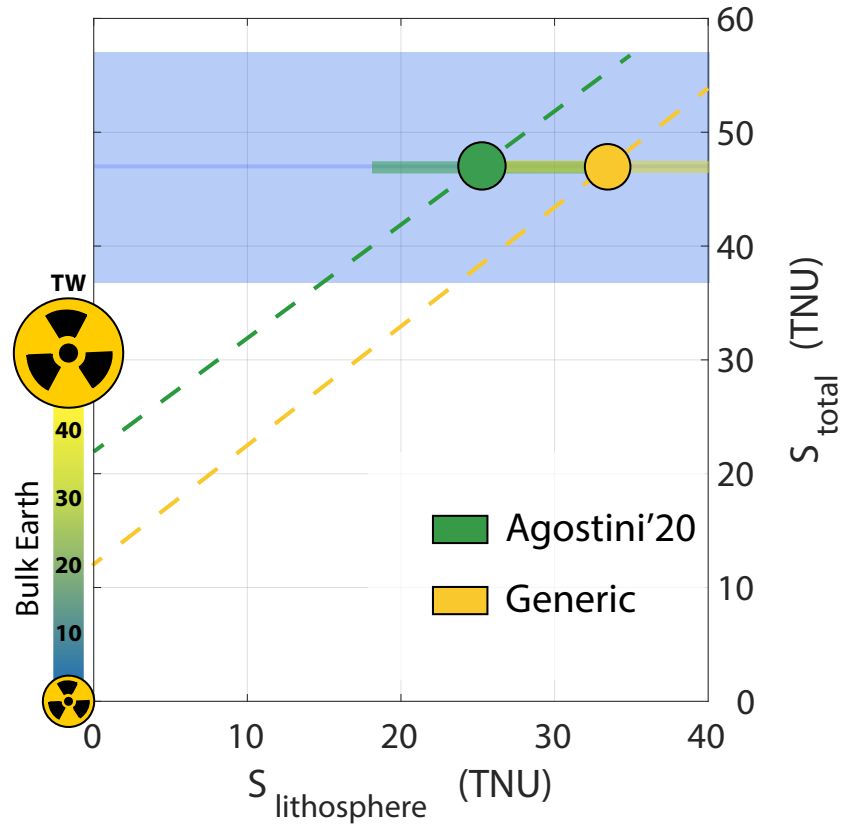


Figure 2.6: The lithospheric geoneutrino signal (predicted, $S_{\text{lithosphere}}$) vs. the measured geoneutrino signal (S_{total}) for the Agostini et al. model and Generic Model introduced in Table 2.1. The Agostini et al. Model has a smaller predicted lithospheric signal, attributing $21.3^{+7.3}_{-10}$ TNU to lithospheric U and Th. The Generic model has a higher concentration of U and Th in the upper crust of the NFL, and therefore a greater lithospheric flux, $32.3^{+7.9}_{-6.4}$ TNU. The dotted lines with slopes = 1 show the y-intercept for each model. The y-intercept is the S_{mantle} . The blue-shaded area shows the measured S_{total} of $47^{+10.8}_{-9.6}$ TNU

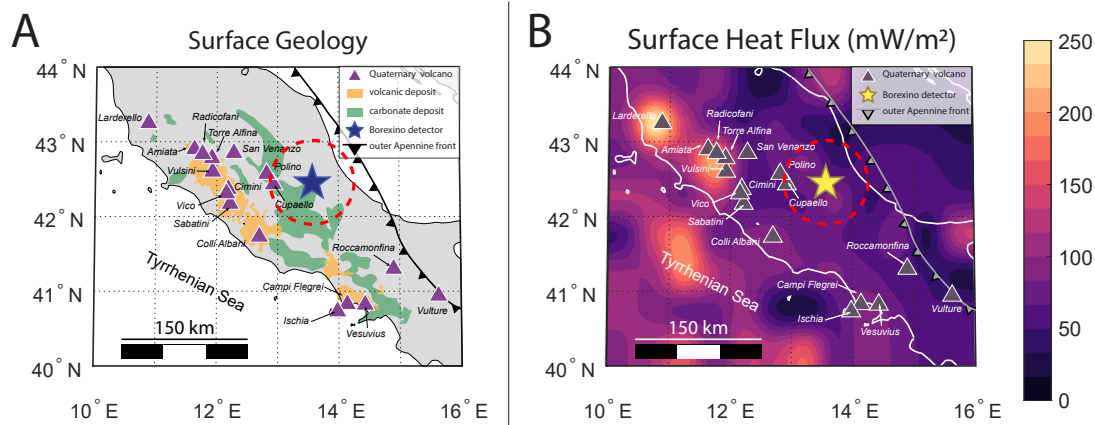


Figure 2.7: A simplified geological map (A) of the central Italian peninsula showing extensive volcanism on the western portion and carbonate platforms to the east. The red dashed line circles the Borexino detector (blue star) at a radius of 50 km. Quaternary volcanic deposits in the west coincide with high surface heat flux (B).

the western half of the Near-Field region exposes extensive deposits of Neogene to Quaternary igneous rocks (Lima et al., 2005; Xhixha et al., 2014). The Tuscan and Roman magmatic provinces are exposed all throughout the Tyrrhenian side of the Apennines and coastal plains. This western portion of the Italian peninsula is enriched in K, Th, and U, with some rocks containing as much as $25 \mu\text{g/g}$ U (Conticelli et al., 2007), which is slightly less than 10 times enriched over average upper crustal rocks (Rudnick and Gao, 2014).

These western Tuscan and Roman magmatic rocks are HPE-enriched and make up a significant portion of the upper crust of the NFL. Some of these rocks are within 50 km of the Borexino detector and need to be incorporated into any NFL model, but unfortunately these lithologies were not discussed by (Agostini et al., 2020; Coltorti et al., 2011). Agostini et al. (2020) highlighted the central tile, which includes the area within ~ 100 km of the Borexino detector and noted “Up to a distance of ~ 150 km from Borexino, 100% of the geoneutrino signal is generated from the LOC [local lithosphere].” Nearly all of

the volcanoes identified in Figure 2.7, some of which are enormous volcanic centers, are within 150 km of the Borexino detector. In addition, the CROP 11 seismic refraction line that the Agostini et al. (2020) model cites as evidence for 13 km of carbonate sediments shows thick layers of siliciclastic sediments as well (e.g., Di Luzio et al. (2009); Patacca et al. (2008)).

To develop our alternative model of the *Borexino* NFL, we followed the practices of Huang et al. (2013) and McDonough et al. (2020) and used a generic, average upper crust composition (Rudnick and Gao, 2014). Using such a generic model for the upper crust of the NFL results in a mantle and bulk Earth model that is consistent with studies that favor a 20 TW radiogenic Earth (Bellini et al., 2021; McDonough et al., 2020).

Disparities between the predicted HPE concentrations in the upper crust for the NFL cause the greatest systematic uncertainties in calculated radiogenic heat production. Constructing a purely carbonate versus a generic upper crust around the detector changes the expected mantle radiogenic heat budget from 30 TW to 13 TW, respectively. These contrasting models illustrate the consequences of modeling different proportions of HPE lithologies for the NFL. Consequently, inaccurate estimates of the subsurface composition near a detector vastly change the implications of the observed geoneutrino signal S_{total} .

2.7 Heat Flux Constraints on Lithospheric Models

To further assess the upper crustal model of the *Borexino* NFL we turned to the available heat flux data for the central Apennines (Pauselli et al., 2019; Verdoya et al., 2021). Given the regional tectonic setting discussed above, it is not surprising to observe a clear

distinction between the western, high heat flux ($>150 \text{ mW/m}^2$) and the eastern low heat flux ($<70 \text{ mW/m}^2$) provinces (Pauselli et al., 2019) (Figure 2.7). Moreover, using observable crustal radiogenic heat production data, Verdoya et al. (2021) concluded that low surface heat flux estimates (e.g., values $<45 \text{ mW/m}^2$) are unreliable in the Apennines. This study also concluded that the central Apennines region has an average heat flux of $\sim 70 \text{ mW/m}^2$ (with eastern and western limbs being approximately 55 and 150 mW/m^2 , respectively). On average, the *Borexino* NFL has a relatively normal continental surface heat flux value (e.g., $\sim 63 \text{ mW/m}^2$, (Lucazeau, 2019)).

Surface heat flux is the sum of contributions from heat production in the crust and the heat flux across the Moho. The total surface heat flux ($Total_{HF}$) can be expressed as the sum of crustal and Moho heat fluxes:

$$Total_{HF} \equiv Crust_{HF} + Moho_{HF} \quad (2.4)$$

Normally, a regionally averaged surface heat flux (e.g., $\sim 63 \text{ mW/m}^2$) is dominated by an upper crustal fraction (i.e., 50 to 60%) and, less so, by a $\sim 1/3$ contribution from the Moho heat flux (i.e., $21 \pm 10 \text{ mW/m}^2$) (Sammon et al., 2022). If we assume a generic crustal compositional model (Table 2.1), the regional $Total_{HF}$ for the Italian peninsula appears normal in terms of its heat production and surface heat flux (i.e., $\sim 70 \text{ mW/m}^2$). In contrast, assuming the compositional model for the NFL adopted by Agostini'20 (Coltorti et al., 2011) puts the $Crust_{HF}$ contribution at 24 mW/m^2 and a $Moho_{HF}$ of 46 mW/m^2 – more than double the global average. While this level of Moho heat flux is possible, it is only observed in areas of recent volcanism, which contradicts the low HPE carbonate

shelf model.

The [Agostini et al. \(2020\)](#) model for the mantle’s radiogenic heat (30 TW) is also inconsistent with their choice of a 8.1 TW global lithosphere model. The Earth has 46 ± 3 TW of heat ([Jaupart et al., 2016](#)), which is both radiogenic and primordial in origin, with other contributions including 3 TW from oceanic hot spots ([Jaupart et al., 2007](#); [Labrosse et al., 2001](#)), 0.4 TW from tidal heating, crust-mantle differentiation, and thermal contraction ([Jaupart et al., 2007](#)), and a minimum of 6 TW to 12 TW from secular cooling of the mantle ([Labrosse, 2007](#)). Consequently, for [Agostini et al. \(2020\)](#)’s accounting to be correct, it leaves anywhere from -2 to -8 TW for the core-mantle boundary (CMB) heat flux, meaning that the mantle is radiating 2 to 8 TW of heat into Earth’s core as it heats up over time. Our alternative model has $7.6^{+2.1}_{-1.6}$ TW for the global lithosphere ([Wipperfurth et al., 2020](#)) and 12.9 TW in the mantle. This model yields a CMB heat flux of 10 to 16 TW, in agreement with estimates from previous studies ([Korenaga, 2008](#); [Labrosse, 2002](#); [Labrosse et al., 2001](#); [Nakagawa and Tackley, 2010](#); [Roberts et al., 2003](#)).

The first experiment to detect geoneutrinos, KamLAND, in Kamioka, Japan, predicts a low radiogenic power Earth, $11.2^{+7.9}_{-5.1}$ TW for Th and U only, or 14 TW when including the decay of other isotopes ([Gando et al., 2013](#)). This result is intermediate between the low H (H = heat production) estimates for the Earth ([Javoy et al., 2010](#)) and middle H estimates ([McDonough and Sun, 1995](#); [Palme and O’Neill, 2014](#)). The NFL model used by the KamLAND team ([Enomoto et al., 2007](#)) predicts an Earth with a low radiogenic power, whereas that proposed by [Wipperfurth et al. \(2020\)](#) predicts an Earth with 20 TW of total radiogenic power.

These KamLAND results challenge the Earth model of [Agostini et al. \(2020\)](#) that

predicts 38 TW of radiogenic power. Either (1) the geological compositions of the KamLAND and/or the Borexino models need to be thoroughly re-investigated, or (2) one would have to predict a hemispherical dichotomy in the mantle’s composition. The latter hypothesis is, of course, unsupported by empirical data on the composition of mid-ocean ridge basalts and ocean island basalts. The second hypothesis seems completely untenable.

In summary, we document the significance of geology’s input into interpreting the particle physics flux data. The combined results for KamLAND and Borexino experiments strongly favor a 20 TW radiogenic Earth model. Moreover, these results confirm that the bulk Earth has a $1.9\times$ enrichment in refractory elements over a CI chondritic composition ([Yoshizaki and McDonough, 2021](#)).

2.8 The Future of Neutrino Geoscience

High resolution crustal models accounting for the specific types and proportions of lithologies surrounding each geoneutrino detector must be constructed to interpret geoneutrino flux measurement. The geology underlying active geoneutrino detectors (Figure 2.8) in Gran Sasso, Italy, Kamioka, Japan, and Sudbury, Canada, reveal complicated tectonic features (e.g., (paleo-)subduction and synorogenic extension, ocean-continent subduction zone, large impact structure). Geoneutrino data already exists from two of these locations, but these crustal models are either low resolution or in conflict with one another. We must reconcile the geoneutrino signal at each location with improved local and regional geology. We must use a wide range of independent geoscientific data to

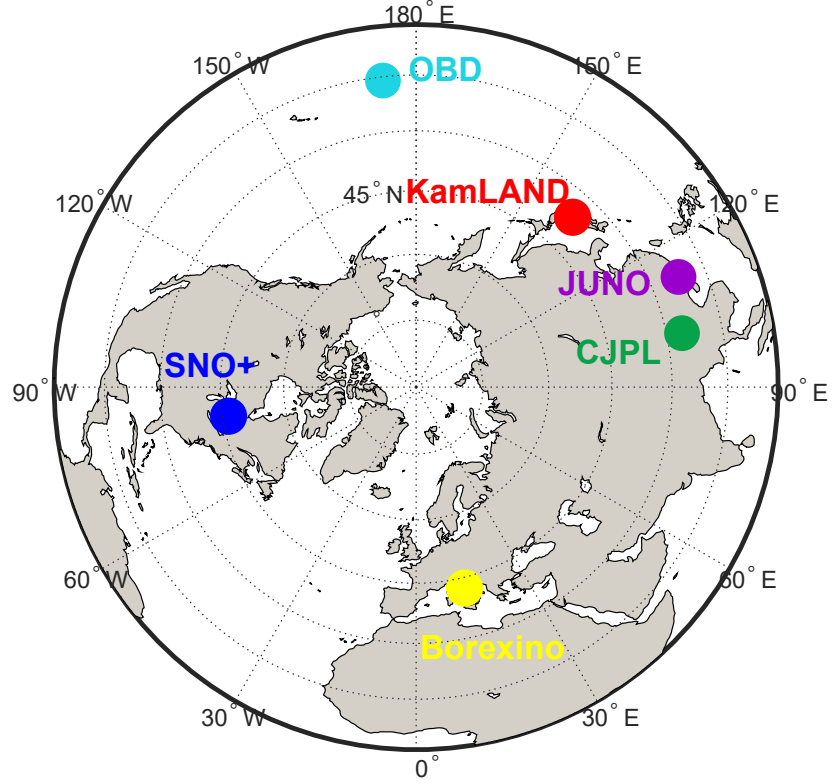


Figure 2.8: Borexino and KamLAND will be joined by the next generation of geoneutrino detectors, including SNO+, which is already counting, and JUNO, which is under construction. The under-development CJPL detector next to the Himalayas marks the fifth detector in the northern hemisphere, allowing for unprecedented mantle resolution. The OBD (ocean bottom detector) experiment is a mobile device and its position can be optimized as being 3000 km away from South America, Australia, and the core mantle boundary.

constrain the composition of the NFL. Moreover, our compositional models needs to be internally consistent with available heat flow, geochemistry/petrology, structural geology, and seismology data to reduce the systematic uncertainties on Earth's HPE content and thermal budget.

There are three more geoneutrino projects under construction or development: Jiangmen Underground Neutrino Observatory (JUNO, Figure 2.8 purple dot) in southeastern China, which will be 20x larger than any existing detector (An et al., 2016); China Jin-

ping Underground Laboratory (CJPL, Figure 2.8 green dot) sited on the eastern slope of the Tibetan plateau and Himalayan ramp and at 2.4 km depth (Beacom et al., 2017); and OBD, a movable, Ocean Bottom Detector (Figure 2.8 teal dot) proposed by a team of scientists and engineers working with JAMSTEC (Sakai et al., 2022). These projects each represent massive feats of engineering and decades-long data collection experiments and require substantial geoscientific input.

The decay of HPEs contribute substantially to Earth’s internal heat. By quantifying Earth’s geoneutrino flux, we can precisely establish how much fuel from HPEs is left to power mantle convection and the recycling processes of plate tectonics. Geoneutrinos studies use modern physics technology to measure directly and instantaneously the current compositional properties of the inaccessible mantle. Th and U exist in Earth in constant, chondritic ratios to 26 other elements (McDonough and Sun, 1995); if we constrain the abundance of HPEs, we can establish Earth’s concentrations of Ca, Al, Nb, and the economically valuable rare earth elements. With the second generation of geoneutrino detectors on the horizon, geoscientists and physicists are poised to unravel Earth’s heat budget from the tallest mountains to the bottom of the oceans.

2.9 Conclusion

The power of geoneutrino studies lies in directly quantifying the amount of heat producing elements in the bulk Earth. Deep reservoirs in Earth that before were unreachable are being sampled by particle physicists, but these studies have not reached a consensus on what their results mean for mantle heat production. The geoneutrino sig-

		Agostini'20	Generic	Units
UC [†]	K	9,600	23,200	$\mu\text{g/g}$
	Th	3.0	10.5	$\mu\text{g/g}$
	U	0.8	2.7	$\mu\text{g/g}$
	HP [‡]	0.19	0.62	nW/kg
Signal	S_{NFL}	9.7	16.6	TNU
	$S_{Lithosphere}$	25.7	32.3	TNU
	S_{Mantle}	21.3	14.7	TNU
R* Heat	Mantle	30	13	TW
	Total	38	20	TW

UC[†] local model for the Upper Continental Crust. NFL = Near-Field Lithosphere (i.e., closest ~ 500 km to a detector). Units: $\mu\text{g/g}$ (10^{-6} kg/kg); TNU (Terrestrial Neutrino Unit, see text for details); TW (Terra Watts, 10^{12} watts). R* radiogenic power. HP[‡] Heat Production.

Table 2.1: Borexino Models for the upper crust in the NFL, bulk calculated Signal, and Radiogenic Power

nal at a given detector is a combination of crust-sourced and mantle-sourced Th and U decays. Since geoneutrinos do not carry directional information, the lithospheric signal must be constrained to quantify the mantle's abundances of Th and U.

Approximately 50% of the geoneutrino signal is produced from the Near-Field Lithosphere (NFL), with 25% of the signal coming from the HPEs within 50 km of the detector. Conflicting Near-Field Lithospheric compositional models lead to profoundly different consequences for the predicted HPE content in the mantle and Earth's thermal evolution.

The Borexino particle physics team ([Agostini et al., 2020](#)) modeled the NFL surrounding their detector as predominantly carbonate, with low concentrations of Th and U. Their model therefore requires most of the geoneutrino signal to come from the distant mantle, implying a 30 TW of mantle radiogenic heat production. Consequently, $>80\%$ of

all of the Earth's internal heat is radiogenic. This high heat production mantle is inconsistent with measurements from the detector at KamLAND and with heat flux observations.

Alternatively, the inclusion of Neogene to Recent, HPE-rich volcanic deposits in the Borexino NFL region results in a more normal average upper crustal composition for Th and U. Using this upper crustal model (versus a low HPE model) can explain the Borexino signal, resulting in 13 TW of radiogenic power in the mantle or a 20 TW radiogenic Earth. It is therefore imperative to produce high-resolution NFL maps with accurate proportions of each HPE lithology.

The direct measurement of geoneutrinos can provide crucial insights into the sources and distribution of heat producing elements in the Earth. When paired with accurate geological knowledge, these high-energy antineutrinos emitted from HPE decays within Earth helps establish the composition of the planet's building blocks as well as the fuel left to power Earth's dynamic interior.

2.10 Acknowledgments

LGS and WFM gratefully acknowledge the NSF for support (grants EAR1650365 and EAR2050374) and University of Maryland for a Wiley fellowship. Both authors would also like to acknowledge Jordan Goldstein, who worked with them to create the figures and schematics in this paper.

Chapter 3: A Geochemical Review of Amphibolite, Granulite, and Eclogite Facies Lithologies: Perspectives on the Deep Continental Crust

3.1 Overview

The composition and origins of the bottom $\frac{2}{3}$ of the continental crust has been a topic of geologic debate for many years. Because of the inaccessible depths of these middle and lower sections of the continents, we cannot sample them directly. We must rely on rocks brought to the surface through mountain building and magma entrainment processes. Deep crustal rocks delivered via these processes come from a wide variety of depths and encompass many different chemical compositions. This chapter seeks to characterize the average composition of the deep crust (typically from 15 to 40 km beneath the surface) and identify the processes that produced the crust's present-day, chemically layered structure.

[1] This chapter has been published as:

Sammon, L. G., & McDonough, W. F. (2021). A Geochemical Review of Amphibolite, Granulite, and Eclogite Facies Lithologies: Perspectives on the Deep Continental Crust.

[2] LGS wrote this text and curated the deep crustal dataset with significant insights and text additions from WFM. LGS would like to acknowledge Caerwyn Hartten for her assistance in curating the metamorphic datasets.

3.2 Abstract

Debate abounds regarding the composition of the deep (middle + lower) continental crust. Exhumed medium- and high-grade metamorphic rocks, which range in composition from mafic to felsic, provide information about the bulk composition of the deep crust. This study presents a global compilation of geochemical data on amphibolite ($n = 6500$), granulite ($n = 4000$), and eclogite ($n = 200$) facies lithologies and quantifies trends, uncertainties, and sources of bias in the deep crust sampling. The continental crust's Daly Gap is well documented in amphibolite and most granulite facies lithologies. Igneous differentiation processes likely control the compositional layering in the crust. Al_2O_3 , Lu, and Yb vary little from top to bottom of the crust. In contrast, SiO_2 , light rare earth elements, Th, and U show a wider range of abundances throughout. Because of oversampling of mafic lithologies, our predictions are a lower bound on middle crustal composition. Additionally, the distinction between granulite facies terrains (intermediate SiO_2 , high heat production, high incompatibles) or granulite facies xenoliths (low SiO_2 , low heat production, low incompatibles) as being the best analogs of the deep crust remains disputable. We have incorporated both rock types, along with amphibolite facies lithologies, to define a deep crustal composition that approaches 57.6 wt.% SiO_2 . This number, however, represents a compositional middle ground, as seismological studies indicate a general in-

crease in density and V_p and V_s velocity with increasing depth. Future studies should analyze more closely the depth dependent trends in deep crustal composition to develop composition models that are not limited to a three-layer crust.

3.3 Introduction

The composition of the deep continental crust has been the subject of many studies for the past half century because of its importance in crustal evolution and the lack of consensus on its composition. The combined middle and lower continental crust (referred to here as the “deep crust”) are the integrated chemical products of billions of years of crust formation and deformation, yet their inaccessibility (deeper than 10 km) has led to a poorly constrained compositional model for the lower two-thirds of the continent. The deep continental crust can be sampled through tectonically emplaced or eroded exposures of high-grade metamorphic rock (referred to here as “terrains”) or deep crustal xenoliths that are rapidly carried to the surface through volcanic eruptions. The composition of these deep crustal analogues ranges widely, encompassing lithologies from metamorphosed basalt to granite. Varied tectonic regimes and widespread crustal heterogeneity have led to numerous geochemical and geophysical models that help to explain local phenomena, but struggle to produce a coherent global picture. Attempts to resolve the debate are limited by nonunique solutions and poorly quantified uncertainties. Defining the bulk compositional properties of the deep continental crust and describing its depth dependent changes endures as a long-standing challenge. Thus, the deep crustal composition puzzle remains, troublingly, unsolved.

[Rudnick \(1995\)](#) posited the paradox of the continental crust: the continental crust has an andesitic composition, however melts from the mantle are basaltic. In doing so, she identified that the formation of continental crust, as compared to making oceanic crust, must be an open system process involving, to different extents, weathering, intra-crustal melting (leaving behind a dense residue), and delamination as some of the operating processes. Consequently, the geochemical uncertainty associated with deep crust composition has led to competing models for crust formation ([Bürgmann and Dresen, 2008](#); [Hacker et al., 2015](#); [Rudnick and Gao, 2014](#)).

In developing their model, [Hacker et al. \(2015\)](#) outlines two processes that they envisage as shaping crustal evolution: delamination and relamination. Delamination occurs when gravitationally unstable material in the deep crust, such as eclogite and other garnet-rich lithologies, separates and flows into the less dense underlying mantle. This process leads to a dense, mafic deep crust as eclogitization occurs but before the lower crust delaminates. In contrast, the process of relamination thrusts subducting sediment under the continental crust, resulting in a more felsic, less dense lower crust. While individual examples can be found to support each of these processes, the difficulty remains in determining the dominant pattern of crust evolution.

The continental crust is conventionally split into upper, $\pm$ middle, and lower layers, though distinct seismic or petrological/geochemical boundaries are not always evident ([Holbrook et al., 1992](#)). Petrological and geochemical studies of the deep continental crust have therefore sought to define its composition through analysis of various high grade metamorphic lithologies. It is difficult to gauge, however, if isolated metamorphic samples are representative of the entire deep crust. Temperature and pressure, and there-

fore metamorphic grade, increase with increasing depth in the crust, though the geothermal gradient varies by up to a factor of ~ 3 depending on continental crust type and tectonic regime ([Christensen and Mooney, 1995](#)). If a pressure of 1 GPa is reached at 35 km (assuming an average crustal density of $2,900 \text{ kg/m}^3$ ([Wipperfurth et al., 2020](#))), the deep crust could plausibly be composed of greenschist, amphibolite, granulite, and/or eclogite facies lithologies. However, amphibolite and granulite facies material dominate what are interpreted as deep crustal cross-sections (such as metamorphosed terrains exhumed in the Ivrea-Verbano Zone, Italy), with minimal evidence for greenschist facies lithologies ([Rudnick and Gao, 2014](#)). Eclogite facies lithologies likely contribute to orogenic regions with thickened deep crust (pressures up to 1.5-2 GPa following the same density scheme as above)([Leech, 2001](#); [Lombardo and Rolfo, 2000](#)). For these reasons, this study focuses on amphibolite, granulite, and eclogite facies lithologies as potential major components of the deep crust. We report on an expanded database developed by [Rudnick and Presper \(1990\)](#) and added to by [Hacker et al. \(2015\)](#), including data sourced from Earthchem.org; we examine the chemical trends among various medium to high grade metamorphic lithologies to understand and better characterize what is the average composition of the deep continental crust and follow this with implications for crustal differentiation and evolution processes.

3.4 The Art and Science of Deep Crustal Modeling

In many ways, predicting the composition of the deep continental crust is as much an art as it is a science, because deep crustal models depend not only on the input data,

but also the approach each modeler takes to interpreting said data. The definition of the deep crust depends on the question each researcher is trying to address, and is therefore neither a static nor universal term. These differing can sometimes lead to confusion and produce seemingly contradictory models of the crust when in fact, each model is simply looking at the crust through a different lens.

How many layers should we split the crust into? What is the scale of lateral variations in the crust? The answers differ based on the model. This fundamental question is the crux of the disagreement between popular composition models ([Hacker et al., 2015](#); [Rudnick and Gao, 2014](#)). While some split the crust into two layers ([Hacker et al., 2015](#)), upper and lower, shallow and deep, others split it into three (upper, middle, and lower) or more sections ([Christensen and Mooney, 1995](#); [Mooney et al., 1998](#)). Thus, debates about compositional models need to be clear about their specific crustal mass fractions. Much geophysical effort has gone into determining the layering and seismic structure of the continental crust. Such topics are beyond the scope of this study, but we want to bring the concept of model resolution to the readers' attention so that they can appreciate the complexity of the task of modeling deep crust composition and be mindful that we are taking but one approach. Nevertheless, averaging geochemical data on large and diverse sample sets has merits. An average deep crustal composition provides an integrated view of how the crust has evolved through time, informing us of the dominant crust formation pathways, though unique tectonic regions may still require specialized compositional models. Bulk crust and silicate Earth compositions are also useful when comparing Earth to other planetary bodies, such as Mars or Super Earths. On a more concrete level, crustal corrections are required for geophysical applications, such as tomographic models of the mantle

or geoneutrino studies, and an average crustal composition is detailed enough to provide further resolution than PREM ([Dziewonski and Anderson, 1981](#)) but simple enough to implement without significant detriment to calculation time. As such, we have decided to press forward with a global-scale deep crustal model.

Classically, two approaches have been taken to assess deep crustal composition: sample driven modeling and process driven modeling. Sample-driven models base their conclusions on the premise that deep crustal analogue samples, such as mafic high grade metamorphic xenoliths, are by and large representative of the composition of the deep crust. Empirical analyses are the main source of data for this type of model. This geochemical inverse model takes measured element concentrations from exhumed rocks and derives the conditions under which they formed. A second approach considers physical processes and constraints that build the deep crust and the effects of crust formation and evolution. A variety of mafic and felsic compositions can satisfy the geophysical observables, such as V_p or viscosity ([Hacker et al., 2015](#); [Shinevar et al., 2018](#)). These forward process models consider all possible geochemical solutions, avoiding the potential bias of xenoliths, which may be sampling a restricted portion of the deep crust, or whose chemistry has been influenced by their limited eruption environments. Both approaches have their strengths, by being based on petrological and geochemical observation or by reducing sampling bias, and in the end, both methods are valid.

This study more closely resembles the first approach, using samples to infer deep crustal composition. We are mindful of the potential biases this leads to (refer to [Section 3.5.2](#)). For the sake of comparison to other models, we operate under the assumption of a three-layer crust, though we advocate for embracing the potential for vertical and lateral

compositional variation by analyzing the full spectrum of available data. Future studies should move beyond bisecting or trisecting the crust, taking advantage of the quality and resolution of both geochemical and geophysical data currently being produced.

3.5 Datasets

3.5.1 Amphibolite, Granulite, and Eclogite

For the rest of this study, “amphibolite”, “granulite”, and “eclogite” will refer to rocks in those metamorphic facies, without imposing constraint on composition. Both amphibolite and granulite facies lithologies range from mafic (< 52 wt.% SiO_2) to felsic (>68 wt.% SiO_2) in composition, and can have Mg#’s (molar $\frac{\text{Mg}}{\text{Mg}+\text{Fe}}$) that resemble the mantle (Mg# ~ 89), the upper continental crust (Mg# ~ 30), or any number in between. Eclogite facies lithologies are less heterogeneous than amphibolite or granulite facies. Eclogite facies mineral assemblages are dominated by (clino)pyroxene and garnet, leaving less room for variations in silica content. Please note that “eclogite” as a metamorphic facies is less restrictive in composition than the largely garnet-omphacite, bi-mineralic rock, eclogite, and can include arclogite samples.

The medium pressure (e.g., 0.2-0.8 GPa) and temperature (e.g., 200-600°C) equilibrium assemblages of amphibolite facies lithologies presumably reflects the conditions of the middle continental crust. Granulite facies lithologies are widely held to comprise the lower continental crust, with its base being defined seismically by the Moho. Amphibolite facies lithologies are generally sampled through exhumed terrains and are more rarely sampled through xenoliths. Granulite facies lithologies can be sampled via terrains

or xenoliths. Granulite facies xenoliths have predominantly mafic to intermediate-mafic (45 - 55 wt.% SiO₂) silica content, while granulite facies terrains span the range of mafic to felsic. Granulite facies rocks are distinguished from amphibolite facies rocks by the dehydration of hydrous mineral phases ([Rudnick and Fountain, 1995](#)). The water-rich minerals that can occur in amphibolite, such as amphiboles and micas, break down into pyroxenes in the granulite stability field due to higher temperatures. Granulite facies metamorphism can initiate at 600°C, meaning that any granulite facies rocks present in areas where the crust is thin and/or the lower crust is at temperatures < 600°C are likely in thermal disequilibrium. Granulite facies lithologies, however, are only expected to undergo retrograde metamorphism under limited circumstances due to the kinetic barrier of rehydration ([Semprich and Simon, 2014](#)). Thus, many studies still use metastable granulite as a lower crustal analogue.

The eclogite facies is traditionally bounded by the pressures and temperatures required to transform basaltic mineral assemblages into high pressure clinopyroxene and garnet ± rutile ± accessory minerals. Though it can be difficult to achieve the pressures required to form eclogite in average continental crustal settings (crustal thicknesses < 40 km), eclogite facies materials may be a significant component of modern and paleo-orogenic belts ([Leech, 2001](#); [Lombardo and Rolfo, 2000](#)). Eclogite facies lithologies, however, may not be preserved through time even in this thicker crust because, unlike granulite, they are more susceptible to retrograde metamorphism ([Abbott and Greenwood, 2001](#); [Carswell and Cuthbert, 1986](#); [Dokukina and Mints, 2019](#); [Krogh et al., 1994](#)). Despite their occurrence being potentially localized to orogenic regions in the crust, eclogite facies lithologies are considered germane to the discussion of deep continental crust be-

cause eclogite is commonly found at subduction zones, one of the main tectonic settings where continental crust is generated (Hawkesworth and Kemp, 2006; Rudnick, 1995).

3.5.2 Potential Biases

The first step in analyzing a dataset is to admit that it is potentially biased. Throughout this paper, we scrutinize the statistical uncertainty of deep crust compositions. Systematic uncertainties, however, are not so easily quantified. This section offers what limited insight we have on the potential for systematic bias in our deep crust sample set. The analyses and conclusions in the rest of this paper are generally founded upon the assumption that the following systematic biases have a limited effect on our dataset, and if any datasets do fall prey to bias, they can be amended without significantly changing the overall picture of deep crust composition.

Our compendium of deep crustal samples, available in the supplemental information of this paper, consists of published data from various sources, most of which are available on Earthchem.org (www.earthchem.org). We used a subset of the data available, limiting our calculations and analyses to samples whose major oxide content is reported and totals to $100 \pm 10\%$. Because of the numerous opportunities for bias in our dataset, we only limited samples by metamorphic grade and major oxide totals. Removing the oxide totals filter does not substantially change the distributions of most elements, but tends to increase the data scatter. The filtered and unfiltered data sheets are available as supplemental information.

3.5.2.1 Location Bias

The global distribution of medium and high grade metamorphic samples shows little correlation between composition and location (Figures S1 and S2). In fact, samples of mafic and felsic compositions are often found within the same region. We are, of course, limited to areas where terrains and/or xenoliths have been exposed at Earth's surface, but our data include samples from all seven continents. Amphibolite facies lithologies have been extensively studied in crust of various ages. Granulite facies lithologies are also widely sampled, though the xenoliths are relegated to areas that have experienced uncommon eruptions of mafic, xenolith-bearing magmas. In addition, Archean granulite facies terrains are generally restricted to cratonic regions, as active tectonic environments tend not to preserve Archean-aged rocks. Eclogite facies xenoliths and terrains are our most limited datasets, but >200 samples are still available for study. Many eclogite facies samples are from the western United States, potentially biasing the dataset towards the eclogites of the Franciscan Complex and eclogites formed from oceanic crust subduction (Tsujimori et al., 2006). South America and Antarctica are not represented in the eclogite facies dataset.

3.5.2.2 Buoyancy and Transport Mechanism Bias

The deep crust may not be fully represented by the *analogue* samples that have reached Earth's surface. Medium and high grade metamorphic lithologies that have survived surface transport do not necessarily reflect the full distribution, abundance, or composition of the deep crust. Felsic terrains could be over-represented at the surface due

to their lower densities. Buoyancy is a significant dynamical force that may play a critical role in determining what types of metamorphic terrains outcrop at the surface (Gerya et al., 2002; Kelemen and Behn, 2016).

On the other hand, eruption type and location may likewise bias xenolith compositions (Jaupart and Mareschal, 2003), including contaminating them with the basaltic lavas. (Rogers and Hawkesworth, 1982; Rudnick and Presper, 1990; Rudnick and Taylor, 1987). Studies have also found that felsic xenoliths often cannot withstand the frequently hot, violent eruptions that transport samples to the surface and tend to be re-assimilated (Halliday et al., 1993; Rudnick and Fountain, 1995). Granulite facies xenoliths in particular could be biased by location and/or eruption method: they tend to be co-located with cratonic crust and they are often carried by kimberlite eruptions and fast-erupting alkali basaltic volcanism (Rudnick and Presper, 1990; Russell et al., 2012). Cratonic xenoliths, which presumably have not experienced recent tectonic activity aside from their transport to Earth's surface, may be sampling the composition of an older deep crust.

3.5.2.3 Preservation and Exposure Bias

We recognize also the potential of sample preservation bias. Recent studies outline different weathering rates for different metamorphic rock compositions (Ohta and Arai, 2007; Price and Velbel, 2003). Age and weathering rate, along with protolith composition, may affect the current metamorphic sample population.

Metastable conditions in the deep crust are another concern. Granulitic lithologies would not be in thermal equilibrium under most projected geotherms (Kusznir and Park,

1987). On the other hand, it has been proposed that felsic granulite facies lithologies will not transform into eclogite facies under high pressures but remain as metastable granulite (Hacker et al., 2010). By contrast, the middle continental crust should be stable in the greenschist facies and the lower crust in amphibolite facies. This lower grade combination is not often observed in exposed cross-sections, which, on average, display high temperature/pressure metamorphism and have a mean pressure of 0.8 GPa (i.e., 25-30 km depth)(Brown and Johnson, 2019). The abundance of amphibolite and granulite facies material in (what we deem to be) deep crustal cross-sections suggests that the deep crust reached peak metamorphic conditions some time in the past and has since cooled off. Additionally, metamorphic terrains may undergo retrogression during the exhumation process. We classify these observations as a “preservation bias” because we are preferentially preserving metastable mineral assemblages.

3.5.2.4 Sample Collection and Naming Bias

Lastly we face the bias that we as scientists impose ourselves: collection and classification bias. Unique localities can be over-sampled for their novelty, and thus, overly abundant in the dataset. Common andesitic rocks, with their lack of attractive phenocrysts and dull grayish-pink hue, may unfortunately be glossed over in favor of more attractive samples (apologies to Dr. J. Blundy and colleagues). Oversampling the same locations seems to plague the amphibolite facies dataset most, with many nearly identical samples in Japan, Alaska, the western United States, and the Appalachian region of the eastern United States. We look more closely at the consequences of this redundant sampling in

Section 3.10. However, for the main analysis of amphibolite facies lithologies in this paper, we keep all amphibolite facies samples in the datasets so that we can see the full span of available data.

Metamorphic lithologies can be categorized by texture (e.g., schist, gneiss) or pressure-temperature (P-T) grade (e.g., amphibolite, granulite), and incomplete changes in lithology can lead to subjective naming. Unfortunately, >6000 gneisses, schists, and metaigneous samples were excluded from this study because they were not accompanied by mineral assemblage information. Metamorphic texture on its own cannot be correlated to precise P-T conditions.

To mitigate the oversampling of individual geologic formations, we averaged all samples of the same facies collected within $0.2^\circ \times 0.2^\circ$ latitude x longitude of each other. This averaging did not change the median composition of granulite facies xenoliths. The median composition of eclogite facies xenoliths and terrains, and granulite facies terrains of all ages changed by $< 4\%$. For amphibolite facies lithologies, however, the median composition, especially SiO_2 , increased drastically by $>10\%$.

While an unknown amount of bias plagues our dataset, over 10,000 samples contribute to our understanding of deep crustal composition. Systematic differences among the different metamorphic lithologies are discussed in the appropriate sections. These differences, where quantifiable, serve as markers for different possible deep crustal compositions. This study focuses on a contextualized overview of compositional aspects; it does not delve deeply into metamorphic processes. Should any systematic errors fundamentally shift our understanding of the deep crust, that in and of itself would be worthy of future assays.

3.6 Major Element Compositions

3.6.1 SiO₂, MgO, FeO, and the Daly Gap

The abundance of major oxides in deep crustal analogue samples is difficult to summarize with a single value and uncertainty. Elemental distributions are not always well defined by the convenient-to-describe Gaussian, normal, log normal, or gamma distribution functions. Table 3.1 reports summary statistics for amphibolite, granulite, and eclogite facies lithologies major oxide content, but are by no means the most comprehensive descriptions of these complex distributions. Unless indicated otherwise, we will reference the median $\pm \frac{1}{2}$ the interquartile ranges because of their resistance to skewness and outliers.

A discussion of the significance of median versus mean should also include the practice of evaluating element ratios. Should one consider the representative element ratio to be represented by a ratio of the means or a mean of the ratios? (Likewise, be represented by a ratio of the medians or a median of the ratios?) There is no simple answer to this question; it has been debated extensively without reconciliation. The fundamental question asks – how representative is one’s data set of the geological domain being evaluated? For the deep crust, there are many unknowns including unknown unknowns. Hence our preference is to use median values and a median of the ratio, as these resist the influence of skewness and outliers.

Because the distributions of major oxides tend to be skewed and/or multi-modal, our oxide totals are between 90 and 95% when summing the means or medians. We

delve further into assessing the modality of our distributions in Appendix B. Appendix tables B1 - B6 list distribution parameters and summary statistics for all elements. We acknowledge our hubris in attempting to parameterize non-parametric distributions, but we are condemned to using bite-sized descriptions of data in the somewhat Sisyphean task of quantifying the chemical composition of a crust we cannot easily access.

The most noticeable data trend is the bimodal distribution of primitive and evolved samples, illustrated by Figure 3.1. The phenomenon published by Rudnick and Presper (1990) persists in this dataset of over 4,000 granulite facies samples and is also present in over 6,000 amphibolite facies samples. Granulite facies xenoliths are dominantly mafic, having <55 wt.% SiO_2 and ranging from mantle-like Mg#'s ~ 89 to Mg#'s of 45-50. Granulite facies terrains encompass both mafic and felsic compositions. The felsic samples follow a Fe-enrichment/Mg-depletion trend, leading to a double-peaked structure, resembling a chair, when plotted in Mg# vs. SiO_2 space. There is also an age-dependent trend in composition within the granulite facies terrains dataset: older, Archean samples are more evolved than Post-Archean samples. Amphibolite facies lithologies show the same chair-like structure, but with a greater concentration of mafic samples. No distinction is made between amphibolite facies terrains and xenoliths in the dataset because of the scarcity of amphibolite facies xenolith data.

The corollary to this bimodality is the “missing” intermediate samples between 53 and 68 wt.% SiO_2 . The well-documented Daly Gap (Daly, 1914) describes an apparent dearth of intermediate composition rocks, and is collectively observed in *all* 586,000 metamorphic and igneous samples in the *Earthchem.org* database. Thermodynamic instability of intermediate compositions (Daly, 1914; Dufek and Bachmann, 2010) and liquid

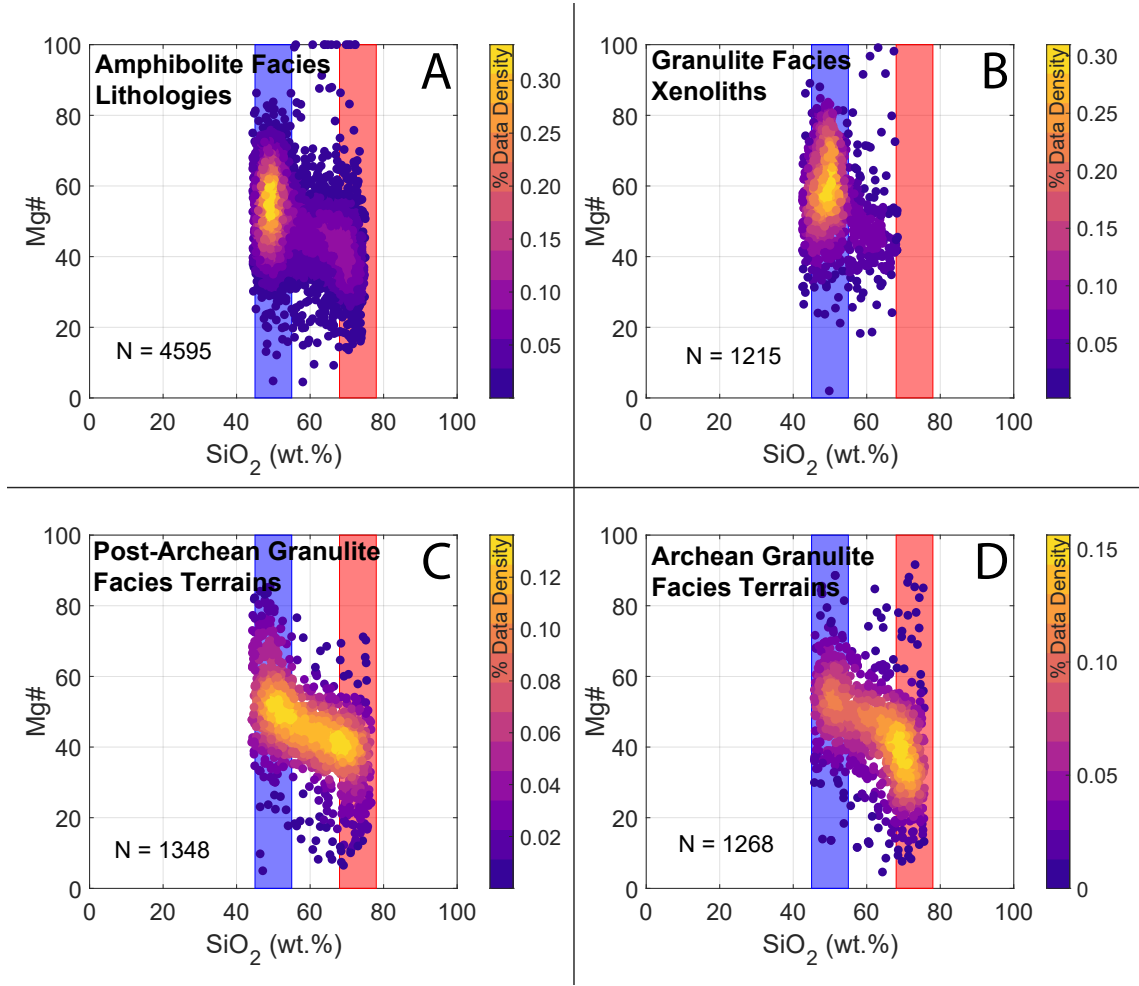


Figure 3.1: Mg# vs. SiO₂ for A) amphibolite lithologies, and granulite facies B) xenoliths, C) Post-Archean terrains, and D) Archean terrains. Color indicates relative data point density. Blue and red fields mark mafic and felsic SiO₂ abundances. Mg# is calculated as molar $\frac{[Mg]}{[Mg]+[Fe]}$. All show high concentrations of mafic and/or felsic compositions and comparatively few compositions of intermediate SiO₂.

immiscibility (Charlier et al., 2011; Reubi and Blundy, 2009), among other hypotheses (Jackson et al., 2018; Yamasaki, 2018), have been proposed to explain the gap. While it is possible that these rocks are not representative of the crust, we conclude that this is dubious given its coherence across multiple lithologies.

The systematically mafic composition of granulite facies xenoliths was noted by Rudnick and Presper (1990) along with many other studies thereafter. Among the proposed explanations for the relative abundance mafic xenoliths are that felsic xenoliths are less likely to survive the eruption process (Halliday et al., 1993; Rudnick and Fountain, 1995) and that xenoliths might sample deeper regions of the crust than terrains (Bohlen and Mezger, 1989; Rudnick and Fountain, 1995). Terrains, on the other hand might be biased towards sampling shallower or more felsic regions because mafic terrains are less buoyant and less likely to reach the surface (Gerya et al., 2008). Granulite facies terrains also show aged-based compositional biases, with Archean samples (61.5 ± 8.5 wt.% SiO_2) being, in general, more evolved than Post-Archean samples (61.5 ± 8.5 wt.% SiO_2) despite having similar median values. Studies have suggested that the ages recorded in these high grade metamorphic samples have been affected by open system behavior (Ashwal et al., 1999) or, as more traditionally argued, hotter temperatures in the Archean allowed for greater amounts of delamination of mafic material, leaving the Archean crust enriched in felsic components (Martin, 1986).

There is no discernible compositional difference between granulite facies terrains and xenoliths of comparable SiO_2 . Most other compositional trends, such as elevated median CaO in granulite facies xenoliths or rare earth element enrichment in terrains (discussed later), correlate to the sample's silica content. The composition of granulite

facies lithologies seems to have little dependence on location (other than the fact that xenoliths are generally most accessible in regions that have experienced volcanism); if surface transport mechanisms are affecting the composition of these granulite samples, then they are not doing so beyond preferentially selecting for certain SiO₂.

The strong preference for mafic compositions in amphibolite facies lithologies is likely biased by mineralogy and geologic naming conventions. Amphibolite facies lithologies unsurprisingly contain amphibole minerals, which generally form in mafic rock compositions. Felsic rocks of similar metamorphic grade seem to be categorized as schists, gneisses, or even metapelites. It is likely that many amphibolite facies samples were excluded from our study because they were given a textural metamorphic grade designation. Thousands of intermediate and felsic gneisses could not be assigned to amphibolite or granulite facies because of insufficient metadata.

The eclogite facies xenoliths and terrains are limited to 46.2 ± 1.2 and 47.2 ± 2.2 wt.% SiO₂, respectively. This is likely due to the stricter definition of “eclogite”, which can refer to a bi-mineralic rock or require basaltic mineral assemblages to reach high pressure. Eclogite facies lithologies have Mg# of 30 to 90, with no correlation to location or method of surface transport.

Table 3.1: Major Element Compositions

	Mean	Median	Geo- Mean	γ - Mean	STD	IQR	Geo- STD	γ - STD	N (fil- tered)	N (origi- nal)
Amphibolite Facies Lithologies										
SiO ₂	59.1	57.1	58.4	59.1	9.43	18.2	1.17	9.37	2240	5490
TiO ₂	0.830	0.730	0.684	0.830	0.494	0.653	1.92	0.50	2240	5490
Al ₂ O ₃	14.8	14.8	14.7	14.8	1.57	2.15	1.11	1.58	2240	5490
FeO _T	5.31	5.26	4.33	5.31	2.88	5.03	2.03	3.29	847	2510
MnO	0.130	0.138	0.111	0.130	0.0650	0.120	1.85	0.07	2220	5430
MgO	4.13	3.70	2.93	4.13	2.81	5.03	2.57	3.27	2240	5490
CaO	5.94	5.69	4.58	5.94	3.61	6.95	2.23	4.13	2230	5480
Na ₂ O	2.96	2.97	2.78	2.96	0.985	1.46	1.46	1.06	2230	5480
K ₂ O	1.72	1.32	1.18	1.72	1.33	2.05	2.60	1.42	2230	5470
P ₂ O ₅	0.148	0.138	0.125	0.148	0.0830	0.110	1.84	0.084	2157	5284
Mg#	46.4	46.8	45.0	46.4	10.6	15.1	1.28	11.2	2220	5430
Granulite Facies Xenoliths										
SiO ₂	51.6	50.2	51.3	51.6	5.05	4.96	1.10	4.88	147	1490
TiO ₂	0.975	0.923	0.879	0.975	0.415	0.581	1.62	0.436	144	1480
Al ₂ O ₃	16.2	16.2	16.1	16.2	1.74	2.31	1.11	1.75	147	1490
FeO _T	6.76	6.71	6.42	6.76	2.15	2.75	1.39	2.16	81	723
MnO	0.150	0.147	0.143	0.150	0.043	0.056	1.36	0.045	143	1440

MgO	7.05	7.19	6.60	7.05	2.29	3.14	1.48	2.55	145	1480
CaO	9.23	9.56	8.83	9.23	2.48	3.27	1.37	2.73	147	1490
Na ₂ O	2.56	2.48	2.44	2.56	0.744	1.14	1.38	0.783	147	1490
K ₂ O	0.779	0.658	0.604	0.779	0.543	0.716	2.10	0.535	147	1480
P ₂ O ₅	0.163	0.153	0.130	0.163	0.101	0.132	2.08	0.106	143	1434
Mg#	56.9	57.1	56.2	56.9	9.05	12.9	1.18	9.16	145	1480

Post Archean Granulite Facies Terrains

SiO ₂	59.3	61.5	58.6	59.3	8.73	16.10	1.16	8.83	145	1660
TiO ₂	0.871	0.763	0.752	0.871	0.438	0.650	1.78	0.462	122	1630
Al ₂ O ₃	15.8	15.5	15.6	15.8	2.05	2.88	1.14	2.04	122	1630
FeO _T	5.75	5.59	5.33	5.75	2.10	3.33	1.50	2.21	81	758
MnO	0.129	0.130	0.120	0.129	0.0447	0.0711	1.50	0.049	120	1600
MgO	5.06	4.10	3.79	5.06	3.64	5.06	2.25	3.70	123	1630
CaO	6.11	5.41	4.83	6.11	3.76	7.17	2.08	4.05	123	1630
Na ₂ O	2.51	2.50	2.33	2.51	0.860	1.32	1.51	0.954	122	1620
K ₂ O	1.87	1.73	1.30	1.87	1.36	2.09	2.66	1.53	155	1700
P ₂ O ₅	0.150	0.122	0.119	0.150	0.100	0.111	2.05	0.100	112	1454
Mg#	48.0	45.8	46.5	48.0	12.0	15.7	1.28	11.9	119	1610

Archean Granulite Facies Terrains

SiO ₂	60.1	61.5	59.5	60.1	8.44	16.9	1.15	8.52	123	1530
TiO ₂	0.655	0.609	0.575	0.655	0.338	0.422	1.68	0.327	122	1490
Al ₂ O ₃	15.3	14.9	15.2	15.3	2.10	2.36	1.14	2.02	122	1510

FeO _T	5.12	4.13	4.06	5.12	3.28	5.08	2.05	3.37	78	902
MnO	0.117	0.101	0.101	0.117	0.061	0.094	1.79	0.063	112	1440
MgO	4.36	3.72	3.40	4.36	2.88	4.53	2.10	2.96	123	1520
CaO	5.33	4.49	4.25	5.33	3.24	5.65	2.08	3.48	122	1510
Na ₂ O	3.05	3.24	2.83	3.05	1.06	1.63	1.51	1.17	120	1500
K ₂ O	1.50	1.20	1.14	1.50	1.06	1.67	2.18	1.07	125	1630
P ₂ O ₅	0.163	0.156	0.144	0.163	0.082	0.087	1.67	0.080	105	1426
Mg#	48.3	47.0	47.3	48.3	9.81	12.2	1.22	9.56	122	1500

Eclogite Facies Xenoliths

SiO ₂	46.2	46.2	46.1	46.2	1.91	2.46	1.04	1.93	15	173
TiO ₂	0.617	0.607	0.490	0.617	0.344	0.621	2.15	0.406	15	173
Al ₂ O ₃	16.1	15.7	15.9	16.1	2.79	2.76	1.17	2.61	15	173
FeO _T	8.56	8.74	8.39	8.56	1.64	2.63	1.23	1.72	6	46
MnO	0.186	0.176	0.183	0.186	0.037	0.035	1.20	0.034	15	172
MgO	11.7	11.6	11.3	11.7	3.02	4.34	1.31	3.07	15	173
CaO	11.4	11.3	11.1	11.4	2.49	4.99	1.26	2.55	15	173
Na ₂ O	2.06	1.58	1.61	2.06	1.49	1.31	2.06	1.39	15	173
K ₂ O	0.375	0.200	0.211	0.375	0.404	0.380	3.09	0.374	14	131
P ₂ O ₅	0.065	0.063	0.055	0.065	0.030	0.035	1.97	0.037	12	86
Mg#	51.8	47.9	50.8	51.8	9.90	15.2	1.21	9.71	13	123

Eclogite Facies Terrains

SiO ₂	47.5	47.2	47.4	47.5	2.64	5.33	1.06	2.65	14	60
------------------	------	------	------	------	------	------	------	------	----	----

TiO ₂	1.33	1.09	1.14	1.33	0.715	1.23	1.79	0.732	14	60
Al ₂ O ₃	14.9	15.0	14.8	14.9	2.14	3.26	1.16	2.16	14	60
FeO _T	9.06	8.59	8.48	9.06	3.27	6.30	1.44	3.27	11	31
MnO	0.227	0.192	0.212	0.227	0.089	0.115	1.41	0.080	14	60
MgO	8.11	7.98	7.80	8.11	2.42	2.80	1.31	2.25	14	60
CaO	11.1	11.1	10.9	11.1	2.13	3.85	1.22	2.18	14	60
Na ₂ O	2.74	2.56	2.36	2.74	1.57	2.25	1.72	1.46	14	59
K ₂ O	0.454	0.428	0.314	0.454	0.334	0.691	2.54	0.371	12	57
P ₂ O ₅	0.134	0.108	0.097	0.134	0.106	0.117	2.27	0.102	14	57
Mg#	41.7	40.8	40.8	41.7	8.84	9.60	1.24	8.95	14	60

STD - standard deviation

IQR - interquartile range

Geo-Mean - geometric mean

Geo-STD - geometric standard deviation (reported in log units)

γ - gamma function mean, $\kappa * \sigma$ (scale parameter * shape parameter)

$\gamma - STD$ - gamma function standard deviation, $\sqrt{\kappa * \sigma^2}$

3.6.2 The Constancy of Al and Ga

Notably, Al₂O₃ content remains relatively constant (i.e., ~12% variation) throughout all samples. Though eclogite facies lithologies have slightly elevated Al₂O₃ content compared to the other samples (Table 3.1), estimates for Al₂O₃ only range from 14-17

wt.%. The Al_2O_3 values of granulite facies lithologies are roughly 5-15% lower than the commonly accepted lower crustal Al_2O_3 values of [Rudnick and Gao \(2014\)](#) though still within the study's given error. Our estimated Al_2O_3 content in granulite facies lithologies are more in line with [Wedepohl \(1995\)](#) and [Gao et al. \(1998\)](#) lower crustal values.

Elements of the same group in the periodic table tend to behave similarly. For example, the abundance of Ga tracks with Al and Ge tracks with Si ([De Argollo and Schilling, 1978](#)). Comparable to Al, Ga concentrations are nearly constant in amphibolite and granulite facies lithologies: median abundance ranges from 17.3 to 19.5 ppm. Eclogite facies samples again behave differently. Due to the significantly smaller sample sizes of the eclogite lithologies. There is little or no data reported for Ge and so we predict its concentration in the deep crust to be relatively invariable at about 1.3-1.4 ppm, based on chemical trends for igneous rocks ([De Argollo and Schilling, 1978](#)).

3.6.2.1 Understanding Protolith Populations

A comparison of the molar abundances of Al to alkali metals and alkaline earths provides a potential provenance indicator for the origin of deep crustal rocks. Sedimentary rocks typically have Al_2O_3 contents of ~ 20 wt.% ([Taylor and McLennan, 1985](#)), whereas most igneous rocks vary from 12 to 19 wt.% Al_2O_3 ([De La Roche et al., 1980](#)). Anorthosites and other plagioclase-rich cumulate rocks, however, can have much higher Al_2O_3 contents.

Al content and Aluminum Saturation Index have in the past been used to help infer the protolith of deep crustal samples. When a rock's Aluminum Saturation Index (ASI;

molar $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}) > 1$, it is classified as peraluminous (Zen, 1988), but with no characterization of its source of origin, so caution is needed. Though sedimentary rocks tend to have higher Al contents (ASI = 1.12, excluding carbonates), Keller et al. (2015) found that common crustal cumulates are also often Al_2O_3 -rich and. Zen (1988) reported granites having ASI values between 1 and 1.4 and noted that these rocks can be derived from a variety of source lithologies, with the proviso that for large bodies of strongly peraluminous granitic rocks peraluminous sources seem necessary. Chappell et al. (2012) observed that many I(igneous)-type granites are peraluminous and owe their origins to partial melting of more mafic source rocks. They also noted that gradations from peraluminous felsic granites to metaluminous igneous compositions are seen for rock suites that have a shared, closed isotopic system.

Unknowns remain significant regarding the abundance of sedimentary lithologies the deep crust. Our samples have median ASI values ranging from 0.65 to 1.06, yet the distribution of aluminous indices is wide and sometimes asymmetrical (Figure 3.2). Amphibolite facies lithologies and granulite facies xenoliths have ASI values comparable to igneous lithologies (Earthchem.org data, median ASI = 0.76), but eclogite facies lithologies have a median ASI values lower than either. The eclogite facies distributions lack the long tail of higher ASI values which will skew their medians somewhat despite its robustness to outliers. If eclogite facies lithologies were not sufficiently sampled to encounter high ASI samples, then it is possible that their actual median ASI value is higher. The change in eclogite ASI distribution shape would lend the appearance of a log-normal trend among the datasets.

It is difficult to distinguish between metaigneous and metasedimentary protoliths

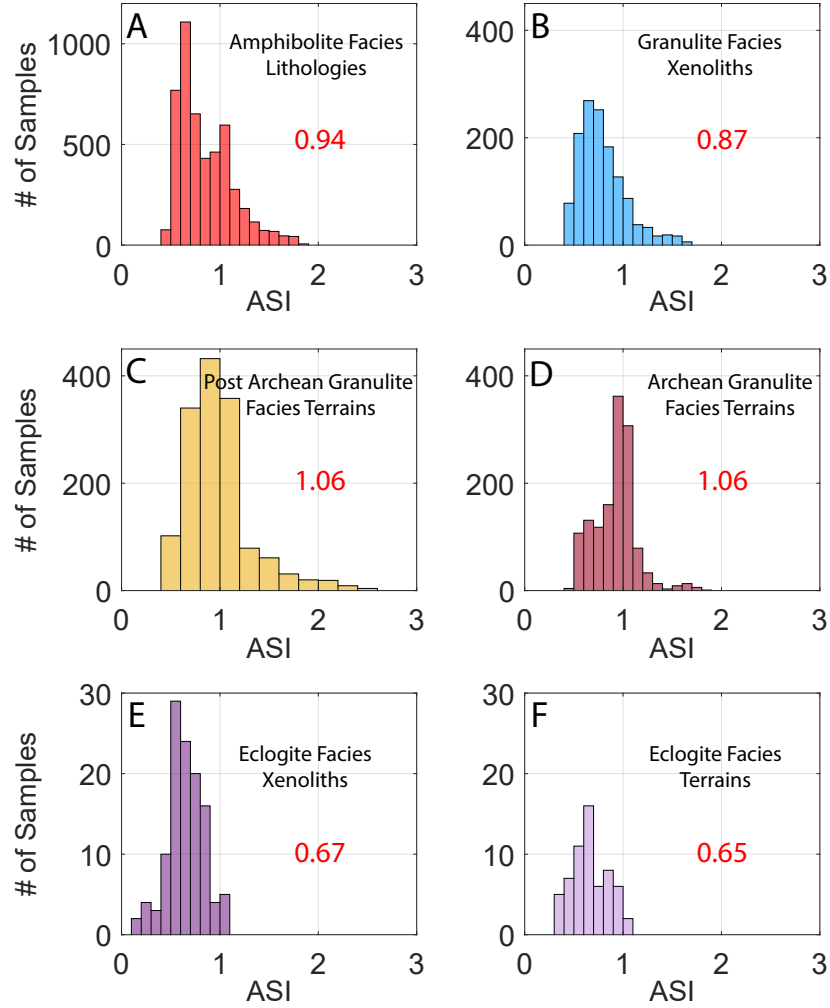


Figure 3.2: Natural log of the aluminum saturation index index (ASI), molar $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$. Red numbers indicate the median value for each sample type. The ASI value is an ambiguous indicator of a rock's protolith; rocks with $\text{ASI} > 1$ can be from igneous or metasedimentary protoliths can be an indicator of metasedimentary contributions to sample populations since sediments are typically more Al rich than igneous sources. The more mafic datasets have smaller ASI values while the granulite facies terrains can more often have values around 1.

for deep crustal lithologies with certainty ([Wilkinson et al., 2009](#)). [Hacker et al. \(2015\)](#) identified 44% Archean and Post-Archean granulite-facies rocks as peraluminous and noted that they may be metasedimentary; going further, they suggested that amphibolite-facies terrains have similar statistics and that 16% granulite-facies xenoliths that are peraluminous may be metasedimentary. We do not find compelling evidence that the ASI value provides unambiguous indication of what is a metasediment. In fact, as cautioned by [Chappell et al. \(2012\)](#), many peraluminous rocks are igneous, including those derived from the remelting of igneous rocks. Given that the term peraluminous does not effectively identify what might be a metasediment, we turned to a machine learning algorithm to predict a metamorphic protolith from major element chemistry ([Hasterok et al., 2019b](#)). This method, however, also produced unclear results. The algorithm showed low confidence in whether the protoliths were igneous or sedimentary, with ratings close to 0 instead of -1 (confidently igneous) or 1 (confidently sedimentary). A broader view of factors must be considered to determine the formation and evolution processes of the deep crust.

3.7 Minor and Trace Element Composition

Here we discuss key geochemical trends seen in incompatible elements; other observations are not covered here for the sake of brevity. The rest of the data is addressed in more detail in Appendix B, which reviews our findings for fluid mobile elements, high field strength elements (HFSE), transition metals, and other important groups of elements. Regardless of surface transport mechanism (eruption as xenoliths or tectonic emplace-

ment as terrains), there are no differences in trace element content between granulites of similar SiO₂ content. Therefore, granulite facies xenoliths and terrains can be treated as one lithology when discussing silica-correlated compositional trends. Eclogite facies xenoliths and terrains have fewer data points, so it remains unclear whether or not they should be given the same treatment.

3.7.1 Rare Earth Elements

Figures 3.3 and 3.4 illustrate that the rare earth element patterns of all of the samples are congruent, having greater variation in the light rare earths (LREE) and than in the heavy rare earth elements (HREE). The amphibolite and granulite data show LREE enrichments and their variability is comparable to that shown by [Rudnick and Gao \(2014\)](#), with granulite facies terrains having the highest median concentrations of La through Nd. Igneous processes – rather than metamorphic changes or chemical weathering – control the relative enrichment in LREEs seen in granulite facies terrains compared to granulite facies xenoliths or amphibolite lithologies. The greatest abundance of La and Ce is not seen in the most hydrated samples (amphibolites) but in the most evolved samples (granulite facies terrains). Eclogite facies lithologies are relatively depleted in LREE compared to amphibolites, yet they are more enriched than granulite facies xenoliths. The standard deviation of the REE distributions narrows from La to Lu. Eclogite facies samples surprisingly show no relative enrichments in HREEs, which would be typical of rocks with more abundant garnet. The HREEs concentrations, especially Tm, Yb, and Lu are identical for all of the metamorphic facies in question.

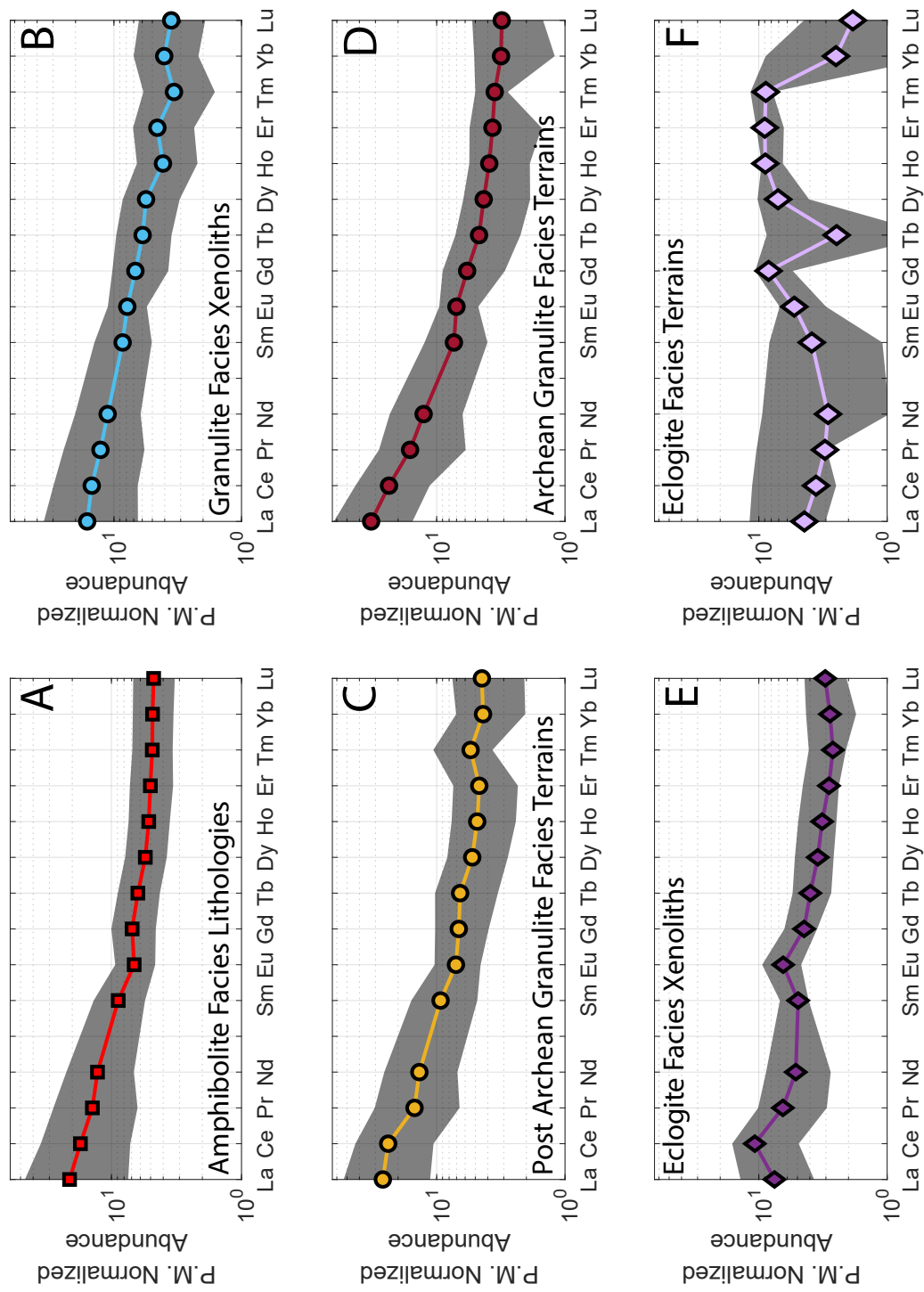


Figure 3.3: Rare earth element median values. The gray shaded bands represent the 1st and 3rd quartile values for each REE.

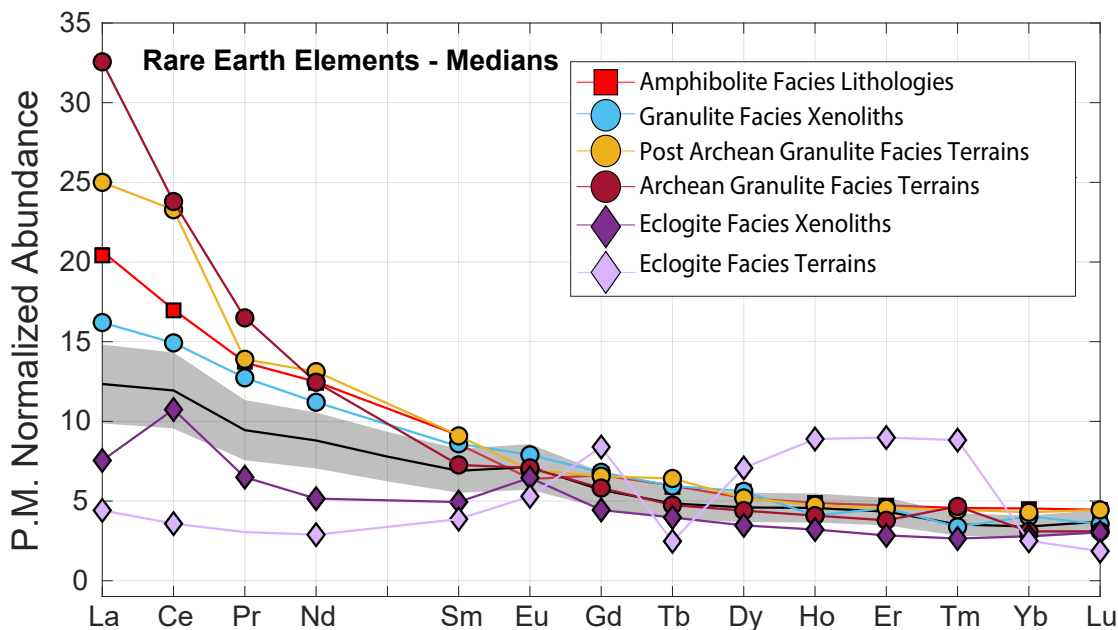


Figure 3.4: While the most variation is seen in the light rare earth elements, the heavy rare earth elements are tightly knit around 0.25 ppm (5x primitive mantle). Note the linear scale. The black line and gray shaded region surrounding it is the Rudnick and Gao (2014) lower crustal composition $\pm 15\%$.

Amphibolite facies lithologies and granulite facies xenoliths span similar ranges of La/Yb, 8.16 ± 6.5 and 7.52 ± 3.2 , respectively. Archean granulite facies terrains have a much higher median value ($\text{La/Yb} = 16.0 \pm 10.2$) than Post-Archean terrains ($\text{La/Yb} = 10.1 \pm 4.50$) despite having similar SiO_2 content, yet La/Yb and is correlated to SiO_2 and forms the same chair-like structure when in natural log La/Yb space as SiO_2 vs. Mg#. The bimodal structure suggests that the Daly Gap affects La/Yb, and that the ratio reflects the original igneous processes that formed the rock rather than metamorphism or weathering.

Flatter rare earth patterns (lower La/Yb ratios), common among the eclogite facies xenoliths and terrains, do not seem to be relegated to specific regions. It is possible that eclogite facies terrains are biased by alteration and subduction processes despite how closely they resemble xenoliths because of our limited dataset (Tsuji-mori et al., 2006).

As mentioned earlier, many of our eclogite facies terrains samples originate from areas that have evidence of significant subduction, such as the western United States. Eclogite facies terrains from the Caledonides in Norway ([Rockow et al., 1997](#); [Svensen et al., 2001](#)) also show REE variation even within the same formation. Sampling exposures of eclogite facies terrains in regions that have not been subjected to significant amounts of subduction would provide more clarity – if such terrains exist. The apparent enrichment in Gd, Dy, Ho, Er, and Tm in eclogite facies terrains is due to our limited dataset, with those elements having only 1 to 5 datapoints each.

On average the crust shows a systematic vertical concentration gradient in REE abundances (i.e., UC $\rightarrow$ MC $\rightarrow$ LC, showing La 36 $\rightarrow$ 18 $\rightarrow$ 12; Yb 3.1 $\rightarrow$ 2.2 $\rightarrow$ 1.7 ppm), with a mildly fractionated downward decrease in the LREE (factor of 3) and HREE (factor of 2). Likewise, Eu/Eu* changes from a 30% negative anomaly in the upper crust to essentially no anomaly in the lower crust. These compositional gradient are most likely products of intracrustal differentiation, with granite magmas moving upward and residuals being stored in the lower crust or lost to the mantle via gravitational processes.

3.7.2 Heat Producing Elements

Heat producing elements (HPEs: K, Th and U) are of particular interest because they are crucial to understanding Earth's radioactive heat budget (these three elements produce 99.5% on the radiogenic heat) through time as well as the temperature and strength of the crust. [Rudnick and Gao \(2014\)](#) estimate the continents host 35 to 40% of the Earth's budget of HPE. Constraining Earth's HPE abundances (especially the abun-

dance of the refractory lithophiles, e.g., U and Th) also constrains ~ 26 other elements ([McDonough and Sun, 1995](#)) that are in conserved, chondritic ratios relative to U and Th.

3.7.2.1 Th, U, and K

The behavior HPE can often be understood through comparisons of elemental ratios. About 80% of the Earth's total heat production comes from Th and U and thus the Th/U ratio is key. [Wipperfurth et al. \(2018\)](#) recently reviewed Th/U values for $\sim 150,000$ crustal rocks and sediments and found that the median values for igneous and metamorphic rocks were close to the bulk Earth's value of 3.8. We find that amphibolites have a median Th/U of 3.7, whereas granulite terrains appear to have lost U (see below) relative Th (median Th/U for Archean and Post-Archean granulite terrains are 7.3 and 6.6, respectively). In contrast, the median Th/U of granulite xenoliths (3.4) appears to be normal (i.e., no U loss). There is little correlation between K content and K/U, with median K/U values for all metamorphic lithologies ranging upwards of one to three times that of upper continental crust. We observe K/U values from 10,000 to 100,000 with uncertainties on the order of 60%. Whether K behaves as a trace element or thermodynamic component (i.e., mineral) controls the K abundance in our samples. That is, K abundance is high if K-feldspar is present in the system, whereas K is low if K-feldspar is not present. The K/Th value is relatively constant in deep crustal lithologies and similar to that of the upper crust ([Alessio et al., 2018](#); [Rudnick and Gao, 2014](#)), implying K/U fractionation is due to U loss, not K loss, and supporting the earlier finding of [Rudnick and Presper \(1990\)](#).

A question that remains is, how long ago did this uranium loss occur in the granulite terrains? To address this issue we combined the two separate measures of Th/U. The isotopic ratio of $^{232}\text{Th}/^{238}\text{U}$ value is referred to as κ ($\kappa \sim \text{Th}/\text{U} \times 1.033$), while the time-integrated Pb isotopic ratio ($^{208}\text{Pb}^*/^{206}\text{Pb}^*$, the decay products of $^{232}\text{Th}/^{238}\text{U}$) is κ_{Pb} , which serves as a proxy for Th/U. [κ_{Pb} values are calculated from the measured lead isotopic composition of the sample minus its primordial lead contribution; see details in [Wipperfurth et al. \(2018\)](#).] The κ_{Pb} provides a measure of the time-integrated Th/U value and is resistant to recent resetting. The average (and median) κ_{Pb} for the amphibolites, granulite xenoliths, and Archean and Post-Archean granulite terrains is 4.1 ± 0.1 (4.0 ± 0.1 ; $n=165, 357, 33$, and 4 , respectively), while eclogite xenoliths are 5.5 (5.8 ; $n=21$) and no data for eclogite terrains; see Table S3 for further details. The $\sim 70\%$ difference between Th/U and κ_{Pb} values for granulite facies terrains is consistent with a relatively recent (< 450 Ma) uranium loss. On average it appears that exhumation or surficial weathering processes result in the loss of U from the granulite and less so due to dehydration metamorphism.

In general, U and Th show positive correlations with SiO_2 . Mean and median U and Th values increase with increasing SiO_2 abundance for amphibolite and granulite facies rocks. While the relationship between SiO_2 and U or Th is log-normal within uncertainties, the concentration of U and Th could potentially also be derived from SiO_2 through a probability analysis ([Gard et al., 2019](#); [Hasterok et al., 2019a](#)).

Deep crustal heat production is but a fraction of upper crustal heat production. The median heat production in the deep crustal lithologies ranges from 0.04 to 0.41 nW/kg (roughly 0.1 to 1.2 $\mu\text{W}/\text{m}^3$, assuming a density of 2900 kg/m^3). Post Archean granulite

facies terrains have the highest heat production to 0.4 ± 0.5 nW/kg ($\sim 1.2 \mu\text{W}/\text{m}^3$). How we calculate our heat production is significant: the *mean of the averages* does not equal the *average of the means*. Using the median K₂O, Th, and U abundances, we calculate heat production for Post-Archean and Archean granulite facies terrains to be 0.14 and 0.15 nW/kg (0.41 and $0.42 \mu\text{W}/\text{m}^3$), respectively. The answer lies in the shape of the distributions, which are neither normal nor log-normal. In this case, our simplified statistics are sub par descriptors of the datasets. Yet, we see that median deep crustal heat production should be minimal unless there is significant incorporation of granulite facies terrain material.

3.8 Distributions that Trend, Periodically

The periodic table is a wonderful tool to use when analyzing elemental trends because of its structure and organization. It shows that elements of similar valence states and radii behave predictably and highlights anomalies caused by specific minerals or sampling methods. We expect to see more skewed distributions for elements that change abundance with depth, since our dataset includes samples from a range of depths ([Bohlen and Mezger, 1989](#)). The difference between mean and median is one metric for quickly assessing the shape of non-normal distributions. [McDonough \(1990\)](#) found that major and compatible trace elements have similar average and median values, whereas median values are systematically lower than average values for the incompatible trace elements (e.g., LREE, K, Rb), with differences between average and median values increasing with increasing incompatibility. Figures [3.5](#) and [3.6](#) are color coded to show the % difference

between mean and median for amphibolite and granulite facies lithologies. The same methodology can be applied to eclogite facies lithologies, but the distributions are more discontinuous, and trends are less clear due to having an order of magnitude fewer samples. Even in the amphibolite and granulite datasets, elements with relatively few data points (such as the highly siderophile elements) appear highly skewed.

Al and Ga are unimodal, with little variation in their abundances compared to other elements, and Na is relatively constant, with some possible bimodality. The mean and median values for Si are similar because of its bimodal distribution, but we do not find conclusive evidence for bimodality in other oxides. Fe, Mg, and Ca show some degree of multi-modality. Most of other elements in the table are unimodal. The rare earth elements exhibit consistent patterns between amphibolite and granulite facies lithologies, though amphibolites have greater differences between mean and medium in the light rare earth elements. The homogeneity of the rare earth element distributions underscores their predictable behavior, with the greatest skewness occurring in the light rare earths which tapers off to a steady $\sim 10\%$ difference between mean and median towards the heavy rare earths.

Both Th and U have highly skewed distributions that verge on log-normal for both amphibolite and granulite facies lithologies. The distributions of U and Th in granulite facies terrains are indistinguishable from both a gamma and a log-normal distribution (using a Wilcoxon rank sum test of median values). The distributions of U and Th in granulite facies xenoliths and amphibolite facies lithologies, however are more accurately described by log-normal distributions. That is to say, the misfit between either the Th or U distribution and log-normal distributions with the same μ and σ is negligible according to

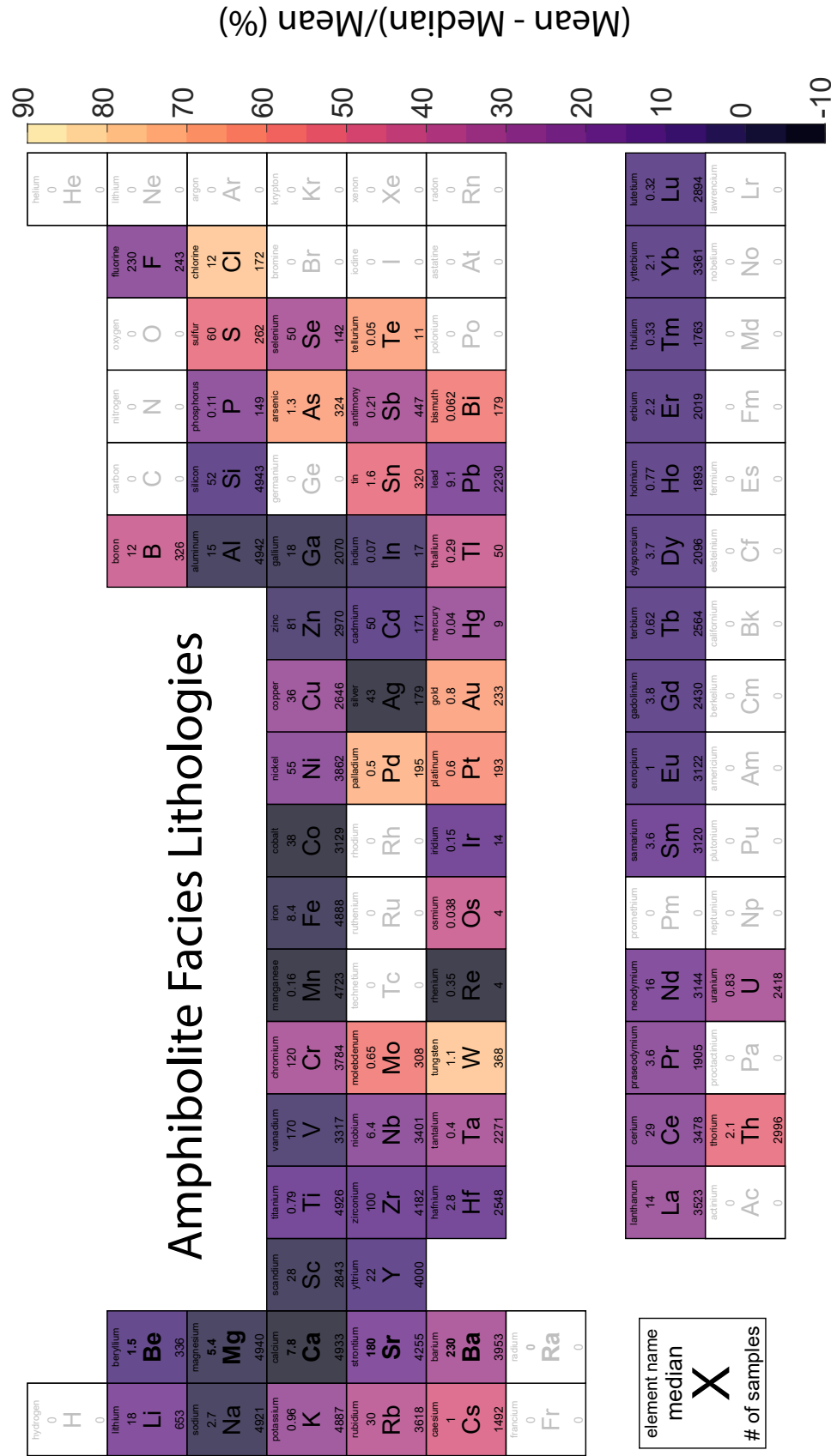


Figure 3.5: The percent difference between the mean and median for elemental distributions is an indicator of the symmetry of the distribution. On this periodic table, color correlates to that percent difference. Elements that are grayed out have insufficient data. The number at the bottom of each element box is the number of available samples, and the top number is the median concentration. Median concentrations are given in wt.% oxide for Na, Mg, Al, Si, K, Ca, Ti, Mn, and Fe. All other elements are reported in ppm. Similar trends emerge among elements of the same family and among the lathanides. Heavier, larger elements vary more in amphibolite facies lithologies than in granulite facies lithologies.

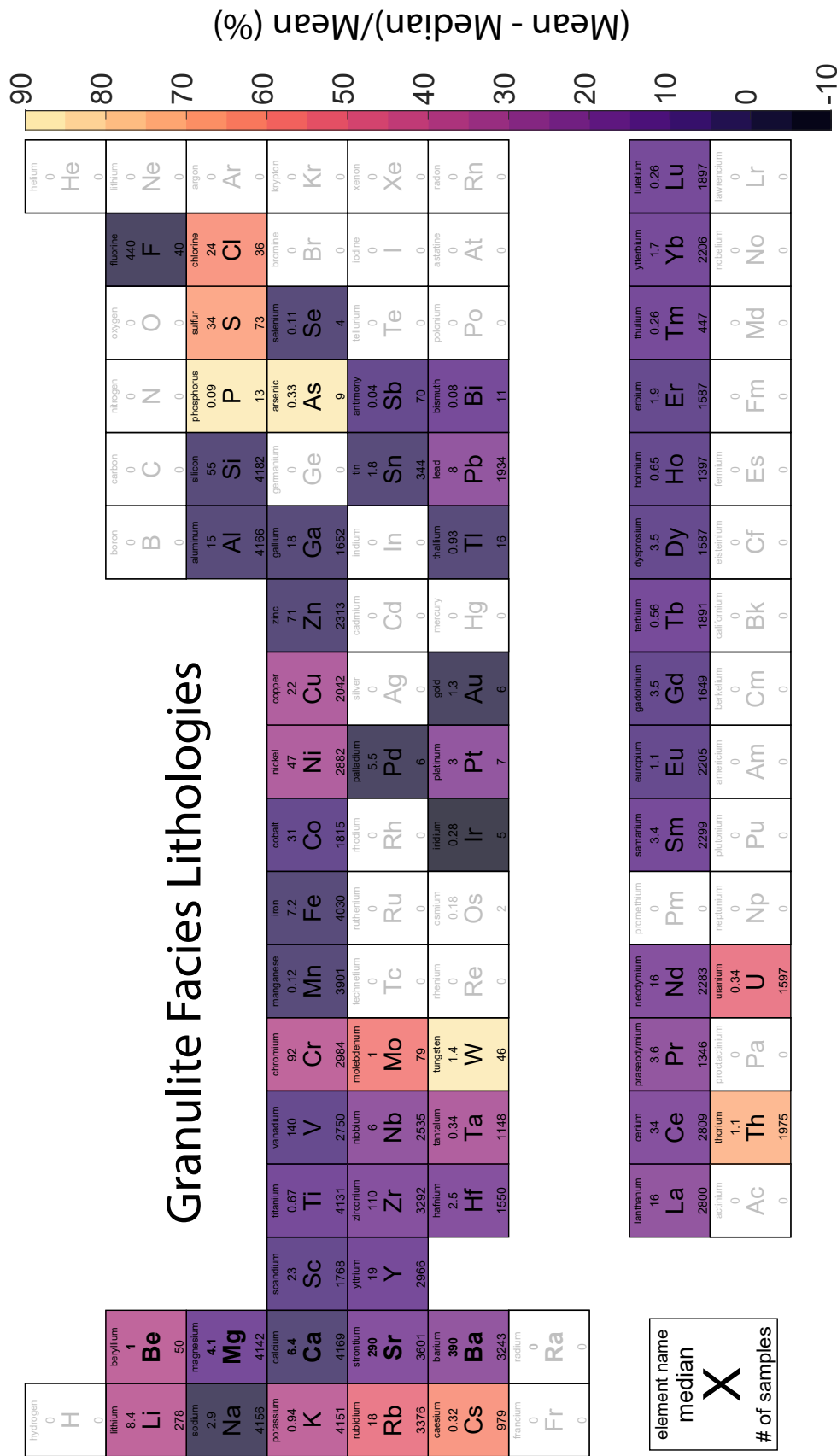


Figure 3.6: Color correlates to the percent difference between the mean and median concentration of a given element in granulite facies lithologies. Elements that are grayed out have insufficient data. The number at the bottom of each element box is the number of available samples, and the top number is the median concentration. Median concentrations are given in wt.% oxide for Na, Mg, Al, Si, K, Ca, Ti, Mn, and Fe. All other elements are reported in ppm. Terrains and xenoliths are treated as one dataset since there is no observable difference between terrains and xenoliths of comparable SiO₂. As with amphibolite facies lithologies, Al and Ga are the narrowest, least variable distributions. Similar trends emerge among elements of the same family and among the lanthanides.

the (admittedly simplistic) statistical test mentioned above. On the other hand, the misfit between Th or U and the corresponding gamma distributions with the same shape and rate parameters is significant according to the same statistical test. Th and U are expected to be skewed because their abundance changes rapidly as a function of depth (Huang et al., 2013; Rudnick and Fountain, 1995; Rudnick and Gao, 2014).

3.9 A Basalt, by Any Other Name

If the deep (particularly the lower) continental crust looks like granulite facies xenoliths, or to an extent amphibolite facies lithologies, then it looks like a basalt. Similarities between these deep crustal samples and mid ocean ridge basalt (MORB) and ocean island basalt (OIB) span SiO_2 , Al_2O_3 , MgO , and MnO (Fig. 3.7, Table 3.2). CaO and Na_2O differ by about 10% among the three different basalts, with MORBs having the highest concentration of both. An important note, though, is that all of our deep crustal analogues are depleted in Ti compared to MORB or OIB. Since there is no complementary Ti enrichment in the upper crust (Rudnick and Gao, 2014; Taylor and McLennan, 1985; Wedepohl, 1995), there may exist an unsampled Ti reservoir on Earth (McDonough, 1991; Rudnick et al., 2000).

If the deep crust resembles granulite facies terrains, then it differs more substantially from MORB and OIB (Fig. 3.8). Granulite facies terrains have 16-27% higher SiO_2 content than MORB and 3 to 4 times higher concentrations of incompatible elements as a result. However, while comparable in major element space, both granulite facies xenoliths and terrains have on average $\frac{1}{3}$ to $\frac{1}{2}$ the concentration of LREEs of OIBs. The undepleted

	Amph. Lith.	Gran. Xen.	PA Gran. Ter.	A Gran. Ter.	Ecg. Xen.	Ecg. Ter.	MORB	OIB
SiO ₂	62.2	53.3	63.18	65.4	48.0	50.1	50.7	50.1
TiO ₂	0.79	0.98	0.78	0.65	0.63	1.16	1.69	2.53
Al ₂ O ₃	16.1	17.2	15.9	15.8	16.3	15.9	14.8	14.1
FeO _T	5.73	7.12	5.74	4.39	9.09	9.11	10.47	11.37
MnO	0.15	0.16	0.13	0.11	0.18	0.20	0.18	0.16
MgO	4.03	7.63	4.21	3.96	12.06	8.47	7.61	8.53
CaO	6.20	10.15	5.56	4.77	11.75	11.78	11.44	10.44
Na ₂ O	3.23	2.63	2.57	3.45	1.64	2.72	2.80	2.16
K ₂ O	1.44	0.70	1.78	1.28	0.21	0.45	0.16	0.38
P ₂ O ₅	0.15	0.16	0.12	0.17	0.07	0.11	0.18	0.25
Mg#	46.8	57.1	45.8	47.0	47.9	40.8	-	-
Mg# Calc.	55.6	65.6	56.7	61.6	70.3	62.4	56.4	57.2

Table 3.2: Median major oxide compositions for our sample sets, mid ocean ridge basalts (MORB, (Gale et al., 2013)), and ocean island basalts (OIB, (MacDonald and Katsura, 1964))

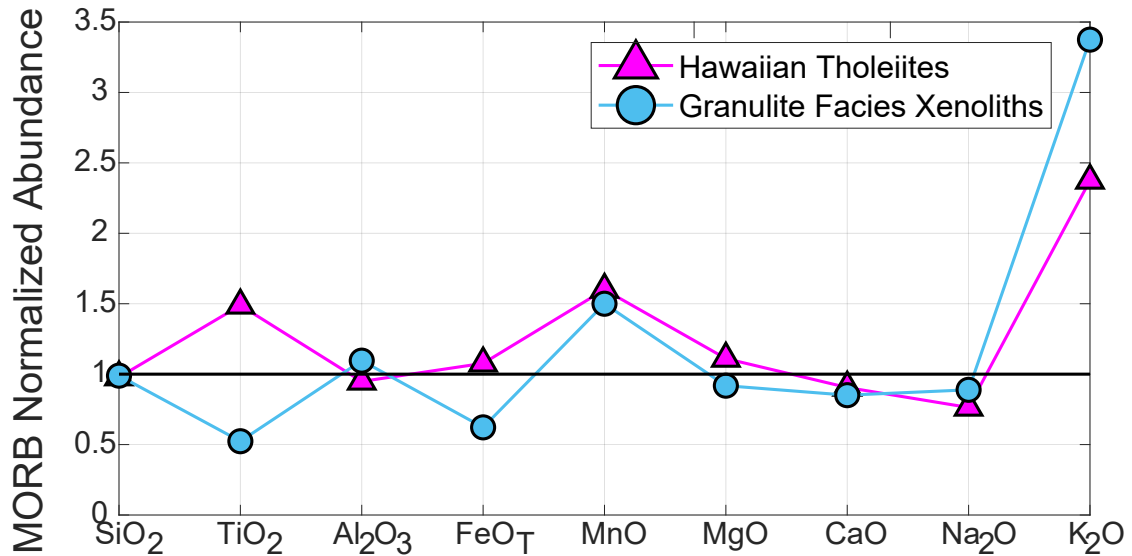


Figure 3.7: Ocean island basalt Hawaiian tholeiites (MacDonald and Katsura, 1964) and granulite facies xenoliths have similar major element patterns except in TiO₂. Both resemble mid ocean ridge basalts (Gale et al., 2013) except in TiO₂ and K₂O.

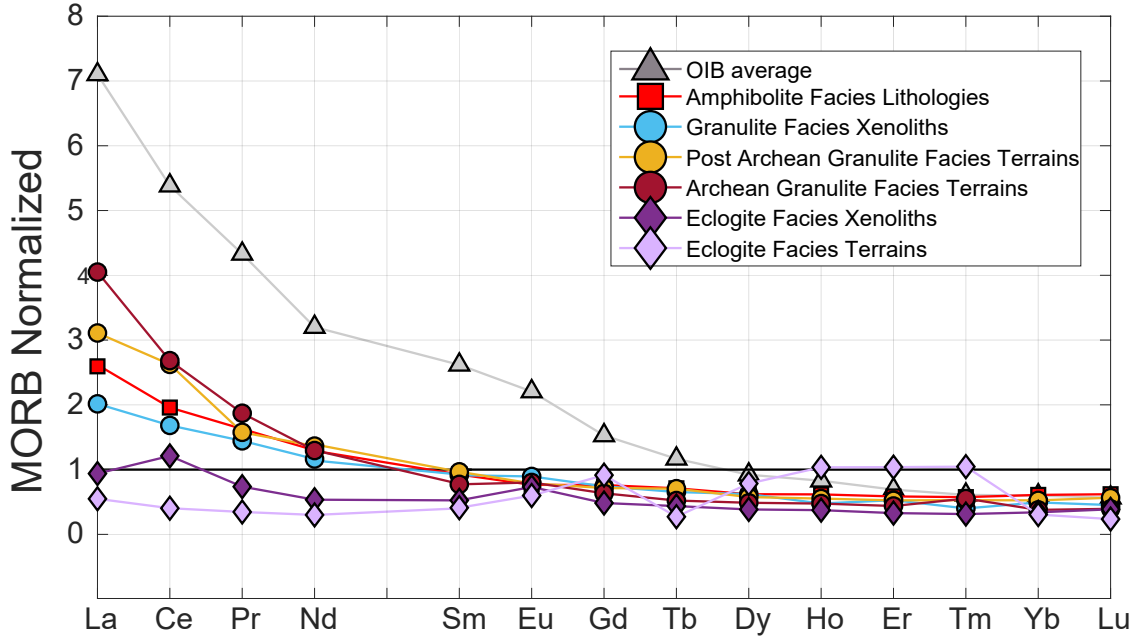


Figure 3.8: Rare earth element concentrations for granulite and amphibolite facies lithologies fall between MORB (Gale et al., 2013) and OIB (Arevalo and McDonough, 2010) abundances. The shape of their REE patterns, however, resemble OIB moreso than MORB. Amphibolite, granulite, eclogite, and OIB rare earth element patterns converge towards the heavy rare earths.

or otherwise uniquely enriched source of OIB material is not reflected in crustal basalts.

Eclogite facies lithologies are essentially basaltic in their bulk compositions and comparable to that of MORBs & OIBs (Table 3.2) when normalized to 100 wt.%. With 16 wt.% Al_2O_3 , 9 wt.% FeO_T , and varying amounts of MgO , however, we expect eclogite facies lithologies to maintain their traditionally-higher-than-basaltic densities. Differences in expected deep crustal densities, discussed below, suggest that eclogites will by and large be gravitationally unstable in most lower crustal models.

3.10 Constructing the Continental Crust

The composition of the deep continental crust is a direct result of the many processes through which it was constructed and evolved. The abundance of incompatible elements in our dataset provides a tool for analyzing the probability of different deep crustal compositions.

Using a simplified, 3-layer modeling approach, we identify the elements that stand out as markers of different crustal compositions and formation processes, and which contribute significantly to estimates of bulk silicate Earth (BSE) composition. Our bulk continental crust composition is calculated by weighting the elemental abundances from the upper, middle, and lower crust by each layer’s mass fraction (Table 3.3). The mass fractions of these crustal layers are from [Wipperfurth et al. \(2020\)](#). The upper crust’s composition uses [Gaschnig et al. \(2016\)](#)’s concentrations for all elements for the top $\frac{1}{3}$ except Sr and Mo ([Rudnick and Gao, 2014](#)), and [Rudnick and Gao \(2014\)](#) for the bottom $\frac{2}{3}$ of the upper crust. We take the upper crust HSEs from ([Chen et al., 2016](#)), and Ag from ([Chen et al., 2020](#)). We use amphibolite facies lithologies as representatives of a middle crustal composition and split the lower crust 50/50 between granulite facies xenoliths and terrains. Debate remains regarding models for the composition of the lower crust ([Hacker et al., 2015](#); [Rudnick and Gao, 2014](#)); there is no obvious solution to determining the composition of the lower crust. Our solution for deciding on a model for the lower crust’s composition recognizes that granulite facies terrains come from on average ~ 0.8 GPa ([Brown and Johnson, 2019](#)), approximately the upper portion of the lower crust, whereas granulite facies xenoliths appear to dominate the bottom of the lower crust ([Rudnick and](#)

Gao, 2014).

Table 3.3: Recommended Continental Crust Composition

	Upper Crust	Middle Crust	Lower Crust	Deep Crust	Bulk Crust
SiO ₂	68.0	62.2	53.3	57.6	61.1
TiO ₂	0.663	0.795	0.980	0.889	0.812
Al ₂ O ₃	15.1	16.1	17.2	16.66	16.1
FeO _T	5.21	5.73	7.12	6.44	6.02
MnO	0.100	0.150	0.156	0.153	0.135
MgO	2.29	4.03	7.63	5.87	4.66
CaO	2.75	6.20	10.15	8.21	6.36
Na ₂ O	2.63	3.23	2.63	2.93	2.82
K ₂ O	3.11	1.44	0.698	1.06	1.75
P ₂ O ₅	0.169	0.150	0.162	0.156	0.161
Li	26.8	15.0	6.89	10.9	16.0
Be	2.18	1.44	0.478	0.956	1.35
B	10.6	9.00	-	-	-
N	51.1	-	-	-	-
F	343	399	-	-	-
S	382	22.0	140	81	178
Cl**	181	29.3	151	90.5	120
Sc	13.3	21.0	28.0	24.5	20.9
V	88.1	134	186	160	137

Cr	77.2	81.0	168	125	109
Co	15.2	29.9	46.8	38.4	30.9
Ni	39.2	39.7	100	70.1	60.1
Cu	26.3	30.0	37.8	33.9	31.5
Zn	69.3	78.0	81.1	79.6	76.2
Ga	17.8	18.0	17.3	17.6	17.7
Ge	1.53	-	-	-	-
As	2.92	1.30	-	-	-
Se	0.058	0.0530	-	-	-
Br	0.98	-	-	-	-
Rb	92.8	43.8	10.6	27.1	48.3
Sr [†]	320	201	465	334	329
Y	25.7	22.5	19.0	20.7	22.3
Zr	203	123	83.3	103	135
Nb	12.1	7.20	7.00	7.10	8.73
Mo ^{†**}	1.12	0.520	1.90	1.21	1.18
Ru	0.240	-	-	-	-
Rh	-	-	-	-	-
Pd ^{**}	0.000800	0.000850	0.00554	0.00321	0.00214
Ag [†]	0.0329	0.0480	-	-	-
Cd	0.097	0.0600	-	-	-
In	0.058	0.0710	-	-	-
Sn	2.23	1.60	1.58	1.59	1.80
Sb	0.39	0.200	-	-	-

Te	-	0.0500	-	-	-
I	0.862	-	-	-	-
Cs	4.18	1.19	0.390	0.787	1.88
Ba	665	330	393	362	460
La	33.3	18.1	11.6	14.8	20.8
Ce	67.0	36.5	27.0	31.7	43.1
Pr	7.65	4.68	3.48	4.08	5.23
Nd	29.4	18.3	14.7	16.5	20.7
Sm	5.38	4.10	3.57	3.83	4.33
Eu	1.15	1.09	1.29	1.19	1.18
Gd	4.79	3.91	3.77	3.84	4.15
Tb	0.780	0.664	0.570	0.618	0.670
Dy	4.55	3.91	3.60	3.75	4.01
Ho	0.95	0.822	0.640	0.730	0.800
Er	2.67	2.29	1.89	2.09	2.28
Tm	0.380	0.350	0.254	0.302	0.327
Yb	2.44	2.19	1.70	1.94	2.10
Lu	0.370	0.330	0.245	0.287	0.315
Hf	5.73	3.42	2.05	2.73	3.70
Ta	0.900	0.540	0.597	0.569	0.677
W ^{†◇**}	1.59	0.440	0.310	0.374	0.773
Re [†]	0.000220	-	-	-	-
Os ^{†**}	0.0000418	-	0.0000210	-	-
Ir ^{†**}	0.0000274	0.0000154	0.0000290	0.0000222	0.0000160

Pt†	0.000618	0.000650	0.000249	0.000158	0.000127
Au	0.000938	0.000800	-	-	-
Hg**	0.0337	0.0300	-	-	-
Tl	0.805	0.500	-	-	-
Pb	16.5	11.7	4.50	8.08	10.8
Bi	0.190	0.0700	-	-	-
Th	10.6	3.68	0.767	2.21	4.90
U	2.64	1.00	0.248	0.621	1.27
Eu/Eu*	0.693	0.830	1.07	0.946	0.848
Heat prod. (nW/kg)	0.623	0.235	0.0640	0.149	0.303
Heat prod. (μ W/m ³)	1.81	0.682	0.186	0.432	0.878
Nb/Ta	13.4	13.3	11.7	12.5	12.9
Zr/Hf	35.5	36.0	40.6	37.7	36.6
Th/U	4.00	3.68	3.09	3.56	3.85
K/U	10500	12800	25100	15200	12300
La/Yb	13.7	8.27	6.82	7.63	9.89
Rb/Cs	22.2	36.8	27.2	34.4	25.7
K/Rb	278	272	547	325	301
La/Ta	36.9	33.5	19.4	26.1	0.71

Oxides reported in wt.%. All other abundances reported in ppm.

** denotes elements for which $N < 6$ for middle and/or lower crust

† denotes Upper crustal values from sources other than [Gaschnig et al. \(2016\)](#). Please see text for sources of † abundances.◊ W values from Archean granulite facies terrains ($n = 3$) excluded as outliers

Though amphibolite facies lithologies are held to represent the middle crust, the median SiO_2 for amphibolite lithologies is $\sim 10\%$ lower than existing estimates. This makes our bulk continental crustal model more mafic in the middle crust, leaving a potential mid-crustal deficit in many incompatible elements. A mafic bias in our amphibolite facies lithologies may exist and seems to stem from:

- 1) biasedly assigning medium grade mafic metamorphic samples to the “amphibolite” facies (due to the hallmark abundance of amphiboles) but not assigning medium grade felsic metamorphics to the amphibolite facies; and
- 2) the oversampling of mafic amphibolite facies locations.

Figure [3.9](#) orders elements left to right from the most to least abundant in the continental crust, relative to a BSE model. In doing so, we highlight the enrichments of Cs, Rb, Ba, Pb, U, Th, K, W, and La in the continental crust and identify the crust as an important host for these elements in the bulk silicate Earth (BSE), despite its insignificant mass contribution (0.55% of the BSE mass). Assuming a BSE composition ([McDonough and Sun, 1995](#); [Palme and O'Neill, 2014](#)), the estimated crustal contribution represents 15 to 50% of Earth's total budget of these elements sequestered in the continental crust. Importantly, $\sim 35\%$ of the heat producing elements are stored in the crust and not available for driving mantle convection.

The continental crust is often viewed as the complementary reservoir of the Depleted Mantle, particularly for the incompatible elements. In Figure 3.9 we compare the composition of the continental crust with that of the average MORB, a representation for the upper portion of the oceanic crust. A crude comparison of the composition of the Depleted Mantle (DM) can be taken as $\sim \frac{1}{10}$ the value of MORB. If the upper mantle (mantle above the 670 km seismic discontinuity; 26% of the mass of the BSE), often considered the Depleted Mantle, was uniformly depleted to create the continents, then it cannot be the complement to the continental crust. Elements whose crustal mass contribution exceeds 26% (i.e., Cs, Rb, Ba, Pb, U, Th, K, and W) require that the lower mantle has been accessed during the production of continental crust. Therefore, the part of the BSE referred to as the Depleted Mantle extends considerably into the lower mantle (nearly half a mantle mass is necessary to account for all of the Rb & Cs in the continental crust). The implications of this finding demand that all or most of the mantle has been involved, to some extent, in the production of the continents.

Apparent non-complementary relationships are found for K, Sr, and Li. The average MORB (Gale et al., 2013) pattern (Figure 3.9) shows marked depletion in the primitive mantle normalized abundance for K, Sr, and Li relative to adjacent elements. The relative incompatibility of these elements during mantle melting are $K \sim U$, $Sr \sim Pr-Nd$, and $Li \sim Dy$ (Sun and McDonough, 1989). The K/U values of continental crust and MORB are complementary relative to that of the BSE (Arevalo et al., 2009; Farcy et al., 2020), consistent with these patterns (Figure 3.9). The marked depletion in Sr seen in the MORB pattern is due to these basalts having experienced considerable plagioclase fractionation. However, as Tang et al. (2017) noted, MORBs with $\geq 10\text{wt.}\%$ MgO, primitive MORBs

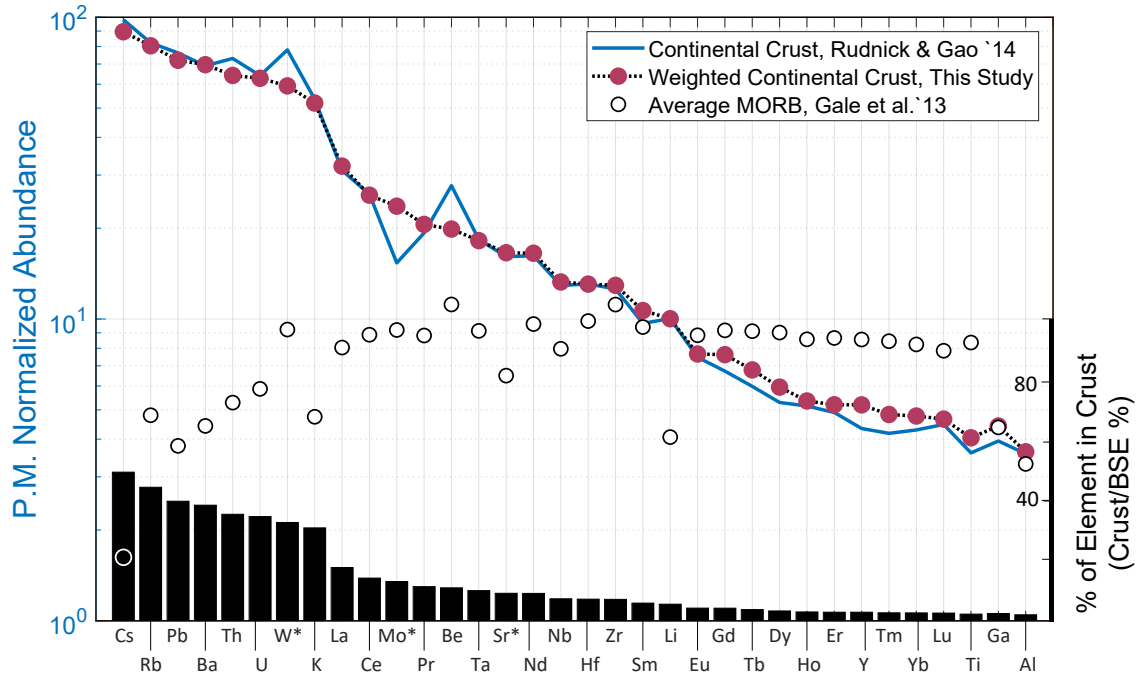


Figure 3.9: Weighted proportions of amphibolite and granulite facies lithologies were combined with upper crustal values (see text) to generate bulk silicate Earth incompatible element abundance estimates. Elements are ordered from left to right by relative enrichment in the continental crust. Alternatively, this ordering reflects element compatibility during mantle melting. Colored circles show our bulk continental crust model while empty circles plot MORB abundances. Black bars follow the right-hand y-axis and show the % of each element sequestered in the continental crust relative to BSE. Normalized to (McDonough and Sun, 1995) with W from (Arevalo and McDonough, 2008).

that have yet to experience plagioclase fractionation, do not show any depletion in Sr. Finally, the depletion in Li is somewhat more challenging to explain. Lithium's position (established by crustal abundances) suggests its relative incompatibility is enhanced due, most likely, to a combination of melting and weathering processes. That said, however, the continental crust, which hosts only $\sim 6\%$ of the BSE's Li budget, cannot account for the marked depletion seen in the MORB pattern. (Moving the MORB Li data point over to the heavy REE still does not account for its depletion.) We offer no explanation for this enigmatic observation.

Another method to determine deep crustal lithology is to test its composition with seismic ([Christensen and Mooney, 1995](#); [Rudnick and Fountain, 1995](#)) and gravity data. Seismic velocities and densities are controlled by mineral forming, major elements (e.g., SiO_2 and CaO), not the highly incompatibles. However, highly incompatibles correlate to a sample's degree of differentiation (i.e., SiO_2 content); therefore, incompatible element abundances can still be derived from major element concentrations. Table 3.4 compares the continent's expected velocities and densities, computed by Perple_X via Gibbs free energy minimization, for the middle (MC) and lower (LC) crust based on different geophysical and geochemical models. Our MC model has a middle crust of $V_p = 6.8$ km/s, $V_s = 4.0$ km/s, and density of 2980 kg/m^3 , values higher than other models. A middle crust calculated from [Rudnick and Gao \(2014\)](#)'s average composition also has a high density ([Brocher, 2005](#)) when compared to the V_p and surface wave predictions of CRUST 1.0 ([Laske et al., 2013](#)) and LITHO 1.0 ([Pasyanos et al., 2014](#)). These inconsistencies between geochemical and geophysical predictions extend into the deep crust, again with all but the lowest density and V_p geochemical estimates ([Hacker et al., 2015](#)) exceeding

Model # of Samples	Vp (km/s)	Vs(km/s)	Vp/Vs	Poiss.	Density (kg/m ³)
Our MC*	6.84	4.04	1.69	0.233	2980
RG MC*	6.74	3.97	1.70	0.235	2940
Low Vp MC†	6.57	3.80	1.73	0.235	2720
CRUST 1.0 MC	6.47	3.70	1.75	0.257	2830
LITHO 1.0 MC	6.51	3.75	1.74	0.252	2840
Our LC**	7.05	4.13	1.71	0.239	3090
RG LC**	7.00	4.01	1.74	0.255	3050
Low Vp LC†	6.80	3.92	1.74	0.235	2920
CRUST 1.0 LC	7.04	4.01	1.76	0.261	3010
LITHO 1.0 LC	7.05	4.00	1.76	0.263	2990
MORB (LC)**	7.40	4.23	1.75	0.258	3260
OIB (LC)**	7.46	4.28	1.74	0.255	3330

† [Hacker et al. \(2015\)](#) Table 4 Middle Crust Vp 6.5-6.6; Lower Crust 6.7-6.9

*MC Conditions - 300 C, 0.4 GPa

**LC Conditions - 500 C, 0.85 GPa

“RG” = [Rudnick and Gao \(2014\)](#)

Table 3.4: Comparison of Deep Crustal Physical Properties

seismic expectations. Though these global averaged seismic models are not infallible, a reconciliation is still required between the geochemical based and geophysical based models for Earth’s deep crust.

3.11 Conclusion

The deep continental crust will remain a topic of intense debate for years to come due to its inaccessibility. Amphibolite, granulite, and eclogite facies metamorphic terrains and xenoliths serve as our windows to middle and lower continental crust. Compositional variability even within these facies underscores the potential for deep crustal heterogeneity, though certain elements and patterns anchor our understanding of the chemistry of the crust. The Daly Gap, evident in amphibolite facies lithologies and granulite facies ter-

rains, dominates the relative abundance of SiO_2 in the deep crust, especially its lowermost portions. The more homogeneous mineralogy of eclogite facies lithologies distinguishes them from amphibolite or granulite facies lithologies, though all three could plausibly contribute to previous deep crustal compositional estimates. Constraining the proportion of mafic to felsic material in the deep crust will in turn constrain its trace element content and distribution, because its enrichment or depletion in these elements is most heavily influenced by differentiation processes.

Thankfully, all is not lost when using these samples to parse out the composition of the deep crust. The amount of Al_2O_3 is similar among all of the samples, as are the heavy rare earth elements. We find that the concentrations of Er, Tm, Yb, and Lu in particular show little variation among samples of different metamorphic grades. The controlling factor in incompatible element abundance among our deep crustal lithologies is how differentiated the material is, not metamorphic grade. This means that igneous processes and protolith composition rather than metamorphic processes dictate the chemical signatures of the deep crust.

Although the deep continental crust has been studied at length, many elements still lack sufficient concentration data (such as the highly siderophile elements). Future studies will be challenged to reduce the size of the uncertainty on element concentrations, and since instrumental precision is not the main source of uncertainty, inquiry into the processes that alter elemental abundances in different samples will have to be identified and explained. As it stands, there is also a density discrepancy between highly cited geochemical models (e.g. [Rudnick and Gao \(2014\)](#) versus [Hacker et al. \(2015\)](#)) and common geophysical crust models, with geochemical samples suggesting a deep crust that

has higher overall density than what is seismically observed. Ultimately, the future of deep crustal modeling will depend on the integration of multiple types of datasets, such as geochemical and seismological measurements and gravity analyses.

Chapter 4: Localized Lower Crustal Composition from Joint-Inversion of Geochemical and Geophysical Datasets

4.1 Overview

In this chapter, we develop a method for quantitatively determining the composition of the deep continental crust. We chose to test our method on an area that has been extensively studied, the southwestern United States, so that we could incorporate abundant geochemical, seismological, and geophysical data. We found that we could determine the major element composition of the crust in this area with $\sim 10\%$ uncertainty, and we used our compositional model to identify the most likely crustal deformation processes that were taking place.

[1] This chapter has been published as:

Sammon, L. G., Gao, C., & McDonough, W. F. (2020). Lower crustal composition in the southwestern United States. *Journal of Geophysical Research: Solid Earth*, 125(3), e2019JB019011.

[2] LGS wrote this text and modeled the lower crustal compositions with help from CG. CG provided ambient noise inversions. WFM contributed significantly to data analysis and text.

4.2 Abstract

The composition of the lower continental crust is well-studied, but poorly understood because of the difficulty of sampling large portions of it. Petrological and geochemical analyses of this deepest portion of the continental crust are limited to the study of high grade metamorphic lithologies, such as granulite. *In situ* lower crustal studies require geophysical experiments to determine regional-scale phenomena. Since geophysical properties, such as shear wave velocity (V_s), are nonunique among different compositions and temperatures, the most informative lower crustal models combine both geochemical and geophysical knowledge. We explored a combined modeling technique by analyzing the Basin and Range and Colorado Plateau of the United States, a region for which plentiful geochemical and geophysical data are available. By comparing seismic velocity predictions based on composition and thermodynamic principles to ambient noise inversions, we identified three compositional trends in the southwestern United States that reflect three different geologic settings. The Colorado Plateau (thick crust), Northern Basin and Range (medium crust), and Southern Basin and Range (thin crust) have intermediate, intermediate-mafic, and mafic deep crustal compositions. Identifying the composition of the lower crust depends heavily on its temperature because of the effect it has on rock mineralogy and physical properties. In this region, we see evidence for a lower crust that overall is intermediate-mafic in composition (53.7 ± 7.2 wt.% SiO_2), and notably displays a gradient of decreasing SiO_2 with depth.

4.3 Introduction

The composition of the lower continental crust, despite its influence over crust formation and geologic hazards, remains a mystery. Though as thin as 10 km in some regions (Rudnick and Gao, 2014), the lower crust contributes critically to the temperature, structure, and stress state of the continent (e.g. Hacker et al. (2015a); Jackson (2002); Shinevar et al. (2018)). Lower crustal deformation models are heavily informed by deep crust silica content, water, and mineralogy (Jackson, 2002). However, because of the relative scarcity (<1% of all samples listed on <http://www.EarthChem.org/>) and the compositional heterogeneity of deep crustal samples, it is difficult to constrain the bulk composition of the lower crust purely through geochemical or petrological measures.

Because the lower crust resides at depths >20 km, its composition can only be sampled indirectly. Granulite facies lithologies serve as metamorphic analogues for the lower crust due to their appearance in exposed crustal cross-sections (Rudnick and Gao, 2014). High grade metamorphic terrains, which have been tectonically emplaced in areas such as the Ivrea-Verbano Zone in Italy or the Fraser Range in western Australia (Fountain and Salisbury, 1981), and granulite facies xenoliths serve as two geochemical windows to the lower crust. As a metamorphic facies, characterized by the dehydration of hydrous minerals (Semprich and Simon, 2014), granulites span a confounding range of mafic (< 52 wt. % SiO₂) to felsic (> 68 wt. % SiO₂) compositions. Such wide variation leads to competing models for the lower crust's composition and density structure, as outlined recently by Dumond et al. (2018).

Combined modeling of high resolution geophysical and geochemical data can place

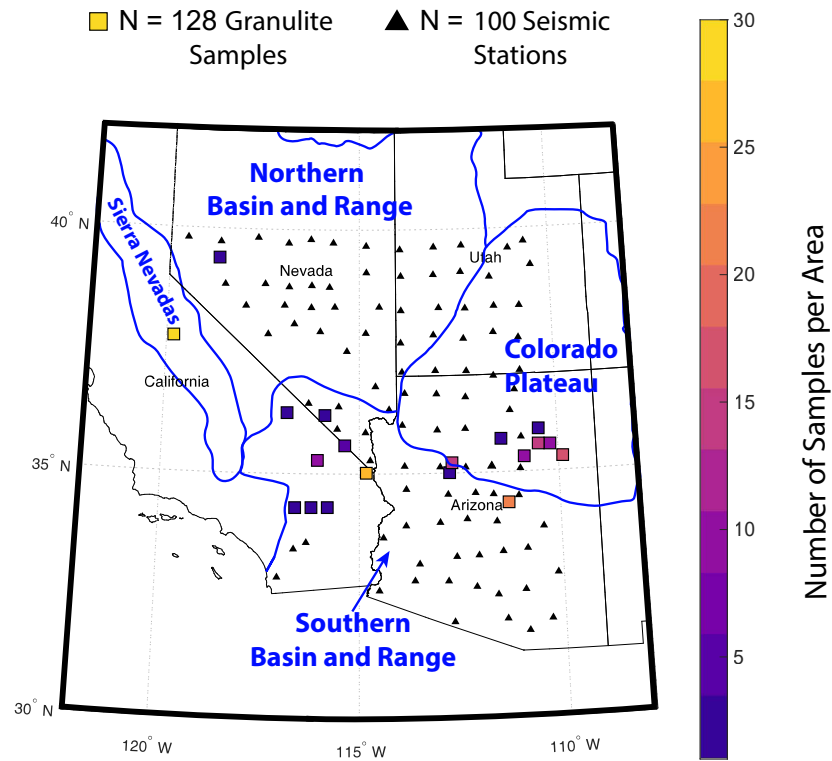


Figure 4.1: The southwestern United States has been sampled at high resolution through geochemical analyses and ambient noise seismology. The black triangles represent the placement of 100 Earthscope Transportable Array stations whose data were used in this study. Colored squares indicate the location of 128 granulite xenolith and terrain samples used as possible chemical compositions for the lower crust. The color of the squares indicates how many samples were collected from the area covered by the square. The overlaid blue lines demarcate three geologically distinct sub-regions within the study area.

tighter constraints on lower crustal composition. Seismic velocity measurements help differentiate among possible lower crustal compositions when compared to laboratory experiments (Holbrook et al., 1992). We use seismic inversions in conjunction with petrological data in an effort to form a less biased lower crustal composition model. In this study, we target the southwestern United States (Fig. 4.1) as a demonstration of such joint modeling efforts because of the variety of data available for the Basin and Range and Colorado Plateau physiographic provinces.

Global scale models (Bassin et al., 2000; Laske et al., 2013) predict seismic velocities in the Basin and Range that are 10% slower and densities 5% lower than those of adjacent tectonic regions. Slower seismic velocities could suggest that the Basin and Range has a more felsic lower crust than surrounding areas and stands in contrast to local velocity studies (Gao and Lekić, 2018; Olugboji et al., 2017; Plank and Forsyth, 2016; Shen et al., 2013). Schulte-Pelkum et al. (2017) provides an in depth reconciliation of various seismic data and lower crustal xenoliths, and proposes a mafic lower crust in the southwestern United States and throughout the country. Both mafic and felsic granulite facies terrains and xenoliths have been extensively characterized in the southwestern US, providing us with a geochemical dataset of 128 samples (<http://www.EarthChem.org/>). We incorporate high resolution, ambient noise, dispersion measurements (Ekström, 2014; Olugboji et al., 2017) from the Earthscope USArray (<http://www.usarray.org/>) project; Moho temperature models from Pn velocities (Schutt et al., 2018); and thermal gradient calculations to derive a distribution of compositions and compositional trends for the lower crust, addressing current model discrepancies.

4.4 Background

4.4.1 Compositional Modeling of the Lower Crust

The depth and thickness of the lower continental crust varies regionally and in the context of different studies. The Conrad discontinuity defines the lower crust seismically, but it is not ubiquitous. When the continental crust is split into thirds, the average lower crustal composition is typically ~ 53 wt.% SiO_2 (Rudnick and Gao, 2014). In some areas, however, the “lower crust” may refer to the bottom half of the continental crust, in which case the average SiO_2 becomes more felsic (Hacker et al., 2015b). The abundance of SiO_2 in the lower crust is not only a function of lower crustal composition, but also of one’s definition of the lower crust. For the purposes of this study, we define the lower crust as simply the bottom half of the crust between 11 km and the Moho after (Schmandt et al., 2015). For example, if the Moho depth were 31 km, the lower crust would have a thickness of $\frac{(31-11)}{2} = 10$ km, and range from 21 km - 31 km depth. We designate 11 km as the thickness of the upper crust because of changes seen in regional Rayleigh wave models from Lin et al. (2014). Though 11 km of upper crust and sediment throughout the entire southwestern United States is a sweeping generalization, it is similar to Roy et al. (1968) 7-11 km thick heat producing layer and Rudnick and Gao (2014)’s 12 km thick upper crust. Keep in mind that our compositional trends are more consequential than our somewhat arbitrary layer thicknesses.

Petrological and geochemical studies of the deep continental crust have sought to define composition through analysis of granulite facies xenoliths and terrains where avail-

able, usually analyzing in detail a small (5 - 20) set of samples. Similar practices have been used by many ([Al-Safarjalani et al., 2009](#); [Halliday et al., 1993](#); [Parsons et al., 1995](#); [Rudnick and Taylor, 1987](#); [Schaaf et al., 1994](#)) to determine the deep crustal structure in regions where samples are available, but it is hard to gauge if these isolated samples are representative of the whole lower crust. While studies of xenoliths provide insight into specific areas of the lower crust, limited sample sets and even smaller sample sizes prove to be recurring obstacles for geoscientists who seek to uncover the composition of the deep crust as it relates to global processes. Seismological crust models, on the other hand, are typically used to describe wide scale crustal phenomena. The use of seismic models for determining lower crust composition requires a conversion between seismic wave velocities and bulk rock compositions, typically achieved through laboratory experiments ([Christensen and Fountain, 1975](#); [Christensen and Mooney, 1995](#); [Holbrook et al., 1992](#)). Recent studies ([Hacker et al., 2015b](#)) give comprehensive assessments of shear and compressional waves velocities of granulite facies lithologies through thermodynamic modeling (calculations are based on empirical, composition-pressure-temperature relationships derived from rock mechanics and mineral physics experiments, and thermodynamic theory).

4.4.2 Geologic Setting

The southwestern United States has undergone multiple episodes of compression and extension since the Mesozoic ([Coney and Harms, 1984](#)). The elevated Colorado Plateau remains relatively undeformed ([Levander et al., 2011](#)) despite being sandwiched

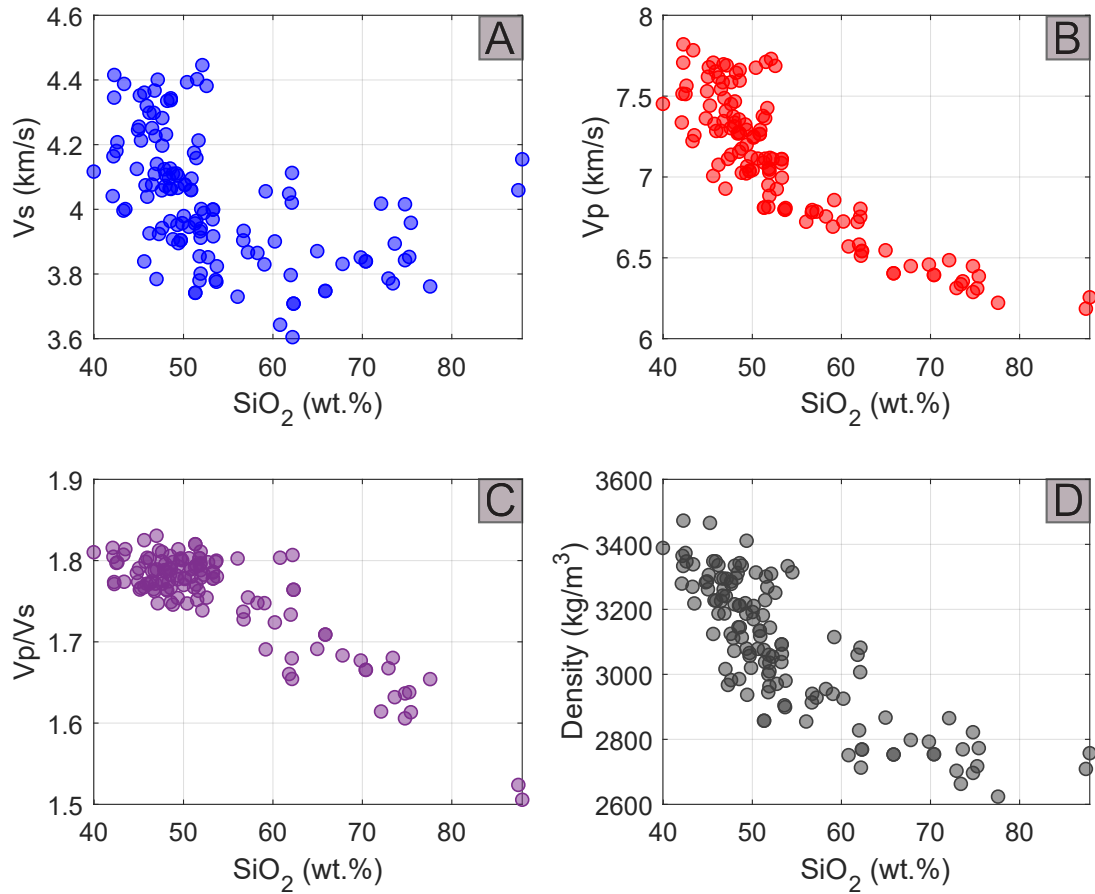


Figure 4.2: An example Perple_X velocity and density calculation shows a correlation with SiO_2 . The 128 samples from the southwestern United States span the range of mafic to felsic, showing especially strong correlations between V_p and SiO_2 and V_p/V_s and SiO_2 . These calculations were done at one temperature (650°C) and one pressure (0.85 GPa). Increasing temperature systematically shifts the distributions to lower V_s , V_p , and density, and, in general, higher V_p/V_s ratio.

between North American Cordillera and the Basin and Range. The Basin and Range province, on the other hand, is characterized by abruptly alternating basins and narrow mountain chains that arose from tensional stress and normal faulting in the Early Miocene (17 Ma) (Coney, 1980). The Basin and Range extended crust, in conjunction with the Colorado Plateau, houses Cenozoic volcanics that are thought to be linked to changes in plate interactions after the conclusion of the Laramide Orogeny (McKee, 1971). The deep crustal xenoliths delivered through Cenozoic volcanic eruptions provide one of our sources of geochemical data. A second data source are the Basin and Range's metamorphic core complexes - a belt of medium- to high-grade metamorphic terrains exhumed through crustal extension (Crittenden et al., 1980). A suite of crust deformation models have been proposed to produce these core complexes (Cooper et al., 2010), each model a different combination of brittle faulting and ductile extension.

4.5 Methods

We used a three-step joint geochemical-geophysical modeling process to constrain composition. Figure 4.3 provides a schematic walk-through of the inputs and outputs of each step.

First, we calculated physical properties over a range of pressures and temperatures for local granulite facies samples through the thermodynamic Gibbs free energy minimization software *Perple_X* (Connolly, 2005). Second, we determined pressure-temperature conditions at 1 km intervals within the lower crust of the southwestern United states, making the assumption that pressure uniformly increases 1 GPa per 35 km depth,

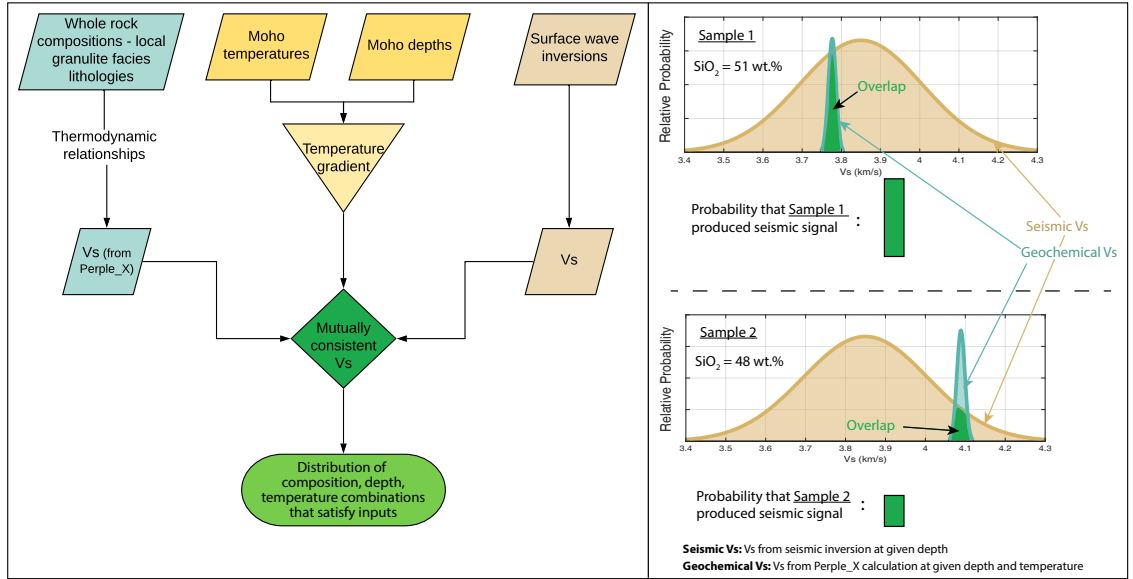


Figure 4.3: Crust modeling flowchart showing our procedure for finding consistent models based on Vs, temperature, depth, and composition. Seismic velocity map from [Olugboji et al. \(2017\)](#); Moho temperatures based on [Schutt et al. \(2018\)](#); Moho depths from [Shen and Ritzwoller \(2016\)](#). Colors in both panels match. The tan probability distribution reflects the spread of Vs solutions from seismic inversions at a given location and depth. The teal Vs is generated from the Perple_X calculations. The overlap between the two Vs distributions is shown in dark green, reflecting the mutually consistent Vs values.

or roughly 28.6 MPa (286 bars) per kilometer. Temperature inputs at the top and base of the crust allowed us to calculate a geothermal gradient and therefore a temperature for each kilometer within the crust. We assumed that the top of the crust resides at $5 \pm 5^\circ\text{C}$ and temperature at the Moho follows [Schutt et al. \(2018\)](#).

Third, we compared the Perple_X-calculated shear wave velocities (Vs) of each sample to seismic inversions for Vs. The seismic data were provided by [Olugboji et al. \(2017\)](#), and one-dimensional Vs profiles were calculated through Bayesian inversion of Rayleigh and Love wave data following [Gao and Lekić \(2018\)](#). We calculated the probability of each sample producing the observed seismic signal by convolving the two datasets. The overlap between the seismic Vs distribution and the geochemical Vs distri-

bution at a given pressure and temperature was totaled for each geochemical sample (Fig. 4.2).

In general, we favored what we considered to be the simplest parameter space that could explain the geochemical observations in our dataset, such as equilibrium conditions and common mineral assemblages. We used the `hpha11.ver` thermodynamic database, which contains the self-consistent (Holland and Powell, 2004) database but is augmented by Hacker et al. (2015a) to account for the α - β -quartz transition. A full explanation of the `Perple_X` parameters we used and our rationale is given in Appendix C. Olugboji et al. (2017) and Ekström (2014) explain the inversion techniques that produced our seismic profiles.

We evaluated the uncertainties associated with each step of our combined model, allowing for variations in lower crustal thickness, temperature, and seismic velocity (Table 4.1). Moho depths were assigned a 2 km uncertainty (Shen and Ritzwoller, 2016), and Moho temperature uncertainties range from 50 to 80C depending on location (Schutt et al., 2018). Combined variations in Moho temperature and depth, and a linear extrapolation of temperature through the crust gave us variable temperature gradients throughout the area of study, which we calculated via Monte Carlo simulation. The result was a distribution of possible lower crustal pressure-temperature conditions, which translated to a probability distribution of compositions. Convolving the distribution of `Perple_X` generated velocities with the seismic shear wave velocities produced our final distribution.

Systematic uncertainties may exist if our fundamental assumption of a dry, granulite facies lower crust is inaccurate. The accuracy of `Perple_X`'s velocity calculations depends largely on this assumption, as a lack of water restricts our compositions to anhydrous

Parameter	Uncertainty
Seismic velocity inversions	full distribution compared to geochemical results, uncertainties on seismic inversion methods given from Gao and Lekić (2018)
Perple_X calculations	subject to unknown systematic uncertainty, but < 1% uncertainty on Gibbs free energy calculations
Moho temperature	5%-10% (Schutt et al., 2018)
Lower crustal thickness	13% - 25%, assuming absolute uncertainty of 2 km (Shen and Ritzwoller, 2016)

Table 4.1: Uncertainties Associated with Methods

minerals. [Connolly \(2005\)](#) offers an overview of the software’s free energy minimization technique for calculating mineral assemblages.

4.6 Results

Overall, the hot lower crust of the southwestern United States trends towards intermediate and mafic compositions. When investigating sub-regional scale variations, however, three separate trends of composition emerge. Joint modeling of surface wave velocities and geochemical and petrological data yields a variety of compositions that depend on temperature. An iterative approach allows us to construct a distribution of probable compositions at each of 100 seismic stations, to account for uncertainties in temperature and composition. Any granulite compositions that were duplicates (i.e. samples whose Vs’s or compositions were indistinguishable from another sample’s) were removed to avoid

artificially weighting our results towards redundantly-sampled lithologies. Excluding all samples with similar silica content, though, would render the geochemical inputs to this model useless, since we would be ignoring geochemical trends in our samples.

Similar velocities and compositions are evident among three sub-provinces of the study area: the Colorado Plateau to the east, the beginnings of the Northern Basin and Range in the northwest, and the Southern Basin and Range in the southwest. As a whole, the shear wave velocities of all three regions range from 3.8 km/s to 4.2 km/s, with about half of the lower crust being faster than 4.0 km/s (Fig. 4.4). V_p , calculated from *Perple_X*, often exceeds 7.0 km/s in the Southern Basin and Range and in deeper portions of the Northern Basin and Range and Colorado Plateau. The Southern Basin and Range, which has experienced the most recent tectonic activity, is marked by the thinnest, hottest crust, while the Colorado Plateau has the thickest, coolest crust. Despite comparatively slow V_s in the Southern Basin and Range (3.9 ± 0.1 km/s), its high temperatures (often $> 800^\circ\text{C}$) require a V_p of 7.1 ± 0.1 km/s and a density of 3000 ± 190 kg/m³ at the base of the crust to satisfy the geophysical model (Fig. 4.4). The V_p/V_s ratio remains poorly constrained, with uncertainties upwards of 10% encompassing most lithologies (Brocher, 2005). Figure 4.4 illustrates a change in the median V_p/V_s from ~ 1.79 to ~ 1.72 separating the Colorado Plateau from the Basin and Range, a shift that reflects compositional variation.

Not surprisingly, compositional trends follow velocity trends, forming three distinct compositional provinces. Figure 4.5 shows representative distributions of SiO₂ content that result from our model. The Colorado Plateau, which has the coolest crust and the lowest V_p/V_s ratio, also has the widest distribution of possible compositions (Fig. 4.5A), which range from 45 to roughly 75 wt.% SiO₂. The Basin and Range favors narrower,

more mafic distributions (Fig. 4.5B and C). Regardless of location, though, mafic lithologies can explain the lower crust's seismic profile more frequently with increasing depth, as shown by the increasing blueness with depth of Figure 4.5. Figure 4.6 (and Figure S7) maps reveal clear compositional distinctions among the three sub-provinces. The differences in SiO_2 between the intermediate Colorado Plateau, intermediate-mafic Northern Basin and Range, and mafic Southern Basin and Range are most apparent in the shallow lower crust. Both the Colorado Plateau and the Northern Basin and Range increase in MgO and FeO content and decrease in SiO_2 content at greater depths, but the Colorado Plateau does not reach truly mafic compositions until 35 - 40 km depth. Caution is recommended when interpreting fine-scale variations in Vs, and therefore in composition. Since the seismic dispersion curves are sensitive to a range of depths (Ekström, 2014), values for the deepest regions of the crust might be elevated by averaging high mantle Vs into the crustal Vs estimates. However, the difference between the seismic velocity model and the combined velocity model (Fig. S11 and Fig. S12) suggests that the weight the geochemical inputs have on the combined model may override any averaging in of mantle velocities. The abundance of distinguishably different mafic granulite lithologies causes the combined model to favor higher velocities.

Six mineral groups dominate the modeled lower crustal mineralogy. Clinopyroxene and garnet grow at the expense of quartz and plagioclase and K-feldspars in deeper portions of the crust. Orthopyroxene abundances also decrease by a few weight % with depth. The high mode proportion (~ 3.5 wt.% to 16 wt.%) of K-feldspars (which primarily manifests sanidine under simulated pressure and temperature conditions) reflects the alkali-rich, latite-like compositions of crystalline rocks from the southwestern United

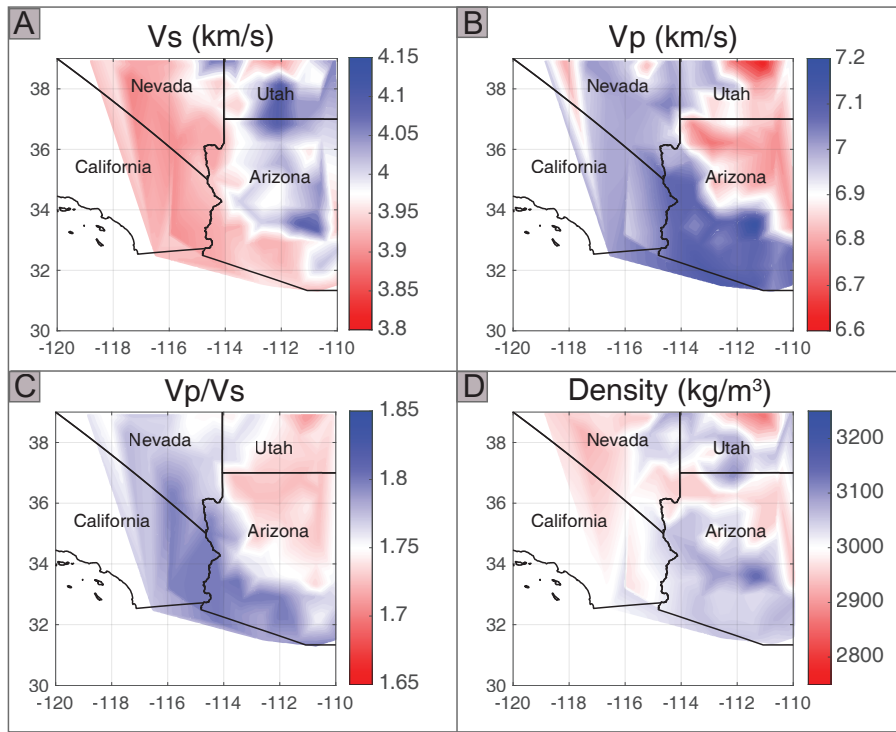


Figure 4.4: The Vs in (A) is a convolution of seismic and geochemical data. Panels (B) - (D) predict Vp, Vp/Vs, and density from the geochemical samples that were consistent with seismic Vs data. The Colorado Plateau is differentiated from the Basin and Range in Vp and Vp/Vs. Hotter temperatures in the south lead to slower Vs but faster Vp and higher densities in the Southern Basin and Range. Blue regions in B, C, and D correspond to more mafic compositions.

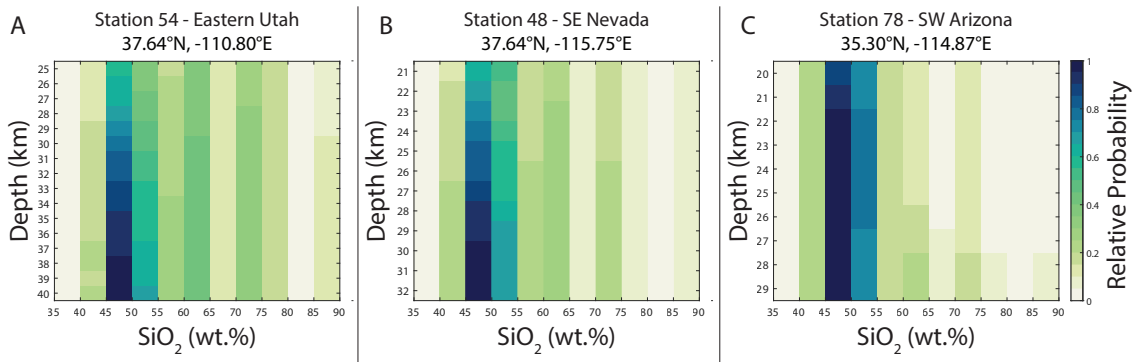


Figure 4.5: Representative histograms (A-C) of the three “sub-provinces” within the study area show increasing probability of a mafic crust with increasing depth. Color indicates the relative probability of a given SiO₂ abundance explaining the seismic signal at a given depth. The thicker, cooler Colorado Plateau (A) can, on average, accommodate a higher percentage of SiO₂ than the Northern (B) or Southern (C) Basin and Range.

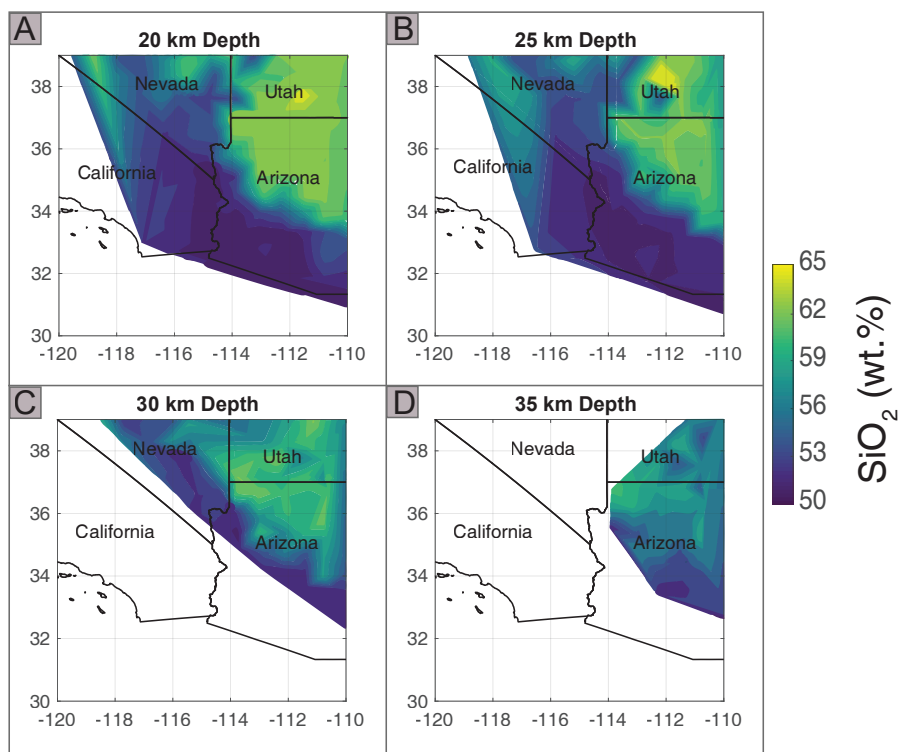


Figure 4.6: Variability in median SiO_2 abundance in the southwestern United States tracks the Colorado Plateau (high SiO_2), Northern Basin and Range (medium SiO_2), and Southern Basin and Range (low SiO_2). SiO_2 abundance overall decreases with increasing depth (A - D). Color scale indicates wt.% SiO_2 . Mantle compositions are not shown in this figure, and therefore deeper profiles (e.g. C - D) show only regions with greater crustal thickness.

States (Tyner and Smith, 1986). At shallower pressures and colder temperatures, minerals such as kyanite, sillimanite, or ilmenite can comprise anywhere from 5 - 15 wt.% of the “lower crust”. Mineral assemblages simplify at greater depths, with clinopyroxene, garnet, plagioclase, and quartz often controlling >80% of the mineralogy.

Though it is convenient to report one number and an uncertainty as representative for composition, we are mindful that the shapes of these major oxide and mineral distributions are non-normal and cannot be fully described by simple summary statistics. That being said, whether reporting mean or median value as representative of the lower crust, the depth trend of change in composition holds true for the Colorado Plateau and Northern Basin and Range (see Tables 4.2 - 4.3). The Southern Basin and Range mean composition shows this gradient to a lesser extent, while the median is homogeneously mafic. For the sake of convenience, our interpretations will reference the median $\pm \frac{1}{2}$ the inter-quartile range (IQR) compositions unless stated otherwise. We favor the median and IQR because they are more resistant to outliers than the mean.

4.7 Discussion

4.7.1 Lower Crust Composition

One value, one composition, is insufficient for describing the entirety of the continental lower crust. We can describe the lower crust more accurately by reporting changes in velocity, density, and composition as a function of depth and location. The lower continental crust, though less than 8 km thick in some sections of the southwestern United States (Buehler and Shearer, 2017), undoubtedly displays lateral and vertical heterogene-

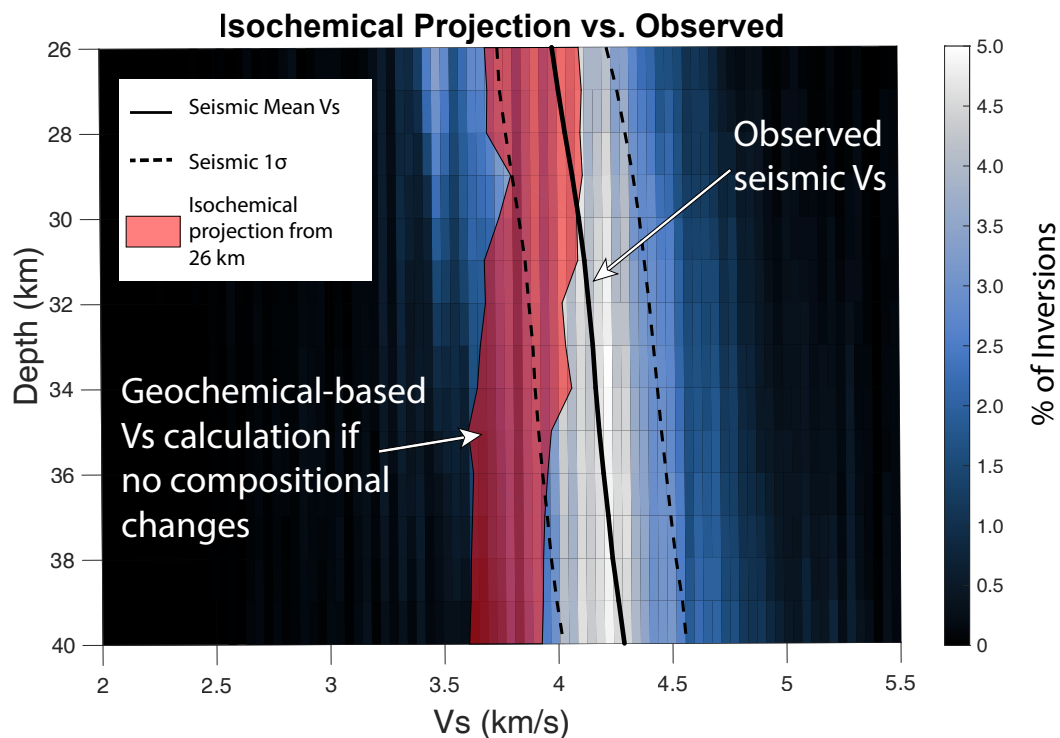


Figure 4.7: An example of granulite lithologies fitting the seismic signal at the top of the lower crust projected to higher temperature and pressure conditions (red field). This isochemical predicted Vs projection deviates substantially from the mean seismic Vs to the extent that by 38 km depth the distributions are distinguishable at 1σ . (For comparison, seismic Vs is typically reported as mean $\pm$ 1 standard error of the mean. Using this metric, the distributions become distinguishable at ~ 30 km.)

Table 4.2: Colorado Plateau SiO₂ Content

Depth	Mean	Median	Standard Devia- tion	1st Quartile	3rd tile	Quar- tile
26	60.0	58.8	11.6	50.0	69.7	
27	59.8	58.6	11.8	49.5	69.9	
28	59.6	58.3	11.7	49.5	69.5	
29	59.3	57.8	11.7	49.3	69.2	
30	59.0	57.4	11.7	49.1	68.7	
31	58.8	56.8	11.8	48.8	68.6	
32	58.6	56.2	11.8	48.6	68.1	
33	58.3	55.8	11.8	48.5	67.3	
34	58.1	55.2	11.7	48.4	66.8	
35	57.9	54.7	11.7	48.3	66.4	
36	57.7	54.3	11.7	48.3	65.9	
37	57.5	53.7	11.6	48.2	65.3	
38	57.3	53.4	11.6	48.1	64.8	
39	57.1	53.1	11.5	48.1	64.4	
40	56.9	52.8	11.5	48.0	64.0	

Oxide abundances reported in wt.%.

Table 4.3: Northern Basin and Range SiO₂ Content

Depth	Mean	Median	Standard Devia- tion	1st Quartile	3rd tile	Quar- tile
22	58.3	55.1	10.8	49.9	65.7	
23	58.3	55.1	10.9	50.0	65.4	
24	58.1	54.7	10.9	49.7	65.0	
25	58.0	54.4	11.0	49.7	64.9	
26	57.7	54.0	10.9	49.5	64.3	
27	57.7	53.9	11.1	49.2	64.2	
28	57.4	53.6	11.0	49.1	63.8	
29	57.4	53.6	11.1	48.9	64.0	
30	57.2	53.4	11.2	48.8	63.6	
31	57.0	52.9	11.2	48.6	63.4	
32	57.0	52.8	11.4	48.3	63.7	

Oxide abundances reported in wt.%.

Table 4.4: Southern Basin and Range SiO₂ Content

Depth	Mean	Median	Standard Devia- tion	1st Quartile	3rd tile	Quar- tile
19	54.6	51.2	10.0	48.0	58.1	
20	54.3	51.1	9.9	48.0	57.6	
21	54.1	51.1	9.8	47.9	57.0	
22	54.0	51.2	9.7	47.9	56.8	
23	54.0	51.2	9.7	48.0	56.6	
24	53.9	51.1	9.6	47.9	56.3	
25	53.9	51.2	9.6	47.8	56.4	
26	53.8	51.3	9.6	47.9	56.2	
27	54.0	51.4	9.7	47.9	56.7	

Oxide abundances reported in wt.%.

Table 4.5: Southwestern United States Lower Crust Major Oxide Content

	Mean	Median	Standard Devia- tion	1st Quartile	3rd tile	Quar- tile
SiO ₂	56.9	53.7	10.6	49.1	63.5	
Al ₂ O ₃	16.7	16.1	4.3	14.1	19.5	
MgO	5.4	3.8	4.2	2.5	7.3	
FeO	8.6	7.7	4.1	5.5	11.4	
CaO	7.0	5.6	4.9	2.2	10.6	
K ₂ O	1.6	1.3	1.5	0.4	2.2	
Na ₂ O	2.7	2.7	1.4	1.6	3.8	
TiO ₂	1.1	0.9	0.8	0.5	1.4	

Oxide abundances reported in wt.%.

Table 4.6: Summary of Lower Crust Seismic Properties and Mineralogy

	Mean	Median	Standard Devia- tion	1st Quartile	3rd Quar- tile
Vs*	3.62	3.61	0.36	3.30	3.64
Vs**	4.02	3.99	0.23	3.86	4.15
Vp	6.94	6.91	0.44	6.59	7.27
Vp/Vs	1.73	1.75	0.07	1.68	1.78
Density	3010	3000	220	2820	3190
Clinopyroxene	17.5	13.3	16.6	3.4	31.5
Garnet	13.0	11.8	8.7	6.5	18.1
K-feldspars	10.9	8.2	6.7	3.4	16.0
Kyanite	3.5	2.5	4.1	0.5	5.3
Olivine	1.3	0.03	5.8	0	0.3
Orthopyroxene	3.0	1.5	4.3	0.2	5.0
Plagioclase	30.3	27.6	14.0	15.5	43.5
Quartz	12.7	10.9	16.1	0.7	27.0

*Vs from surface wave inversions only

**Vs from combined surface wave and geochemical
model

ity (Fig. 4.6 and S7). Temperature plays a crucial role in determining lower crustal composition. Both cold intermediate and hot mafic granulites can produce the shear wave velocities of >3.9 km/s observed across the southwestern United States (Christensen and Mooney, 1995). The thicker, cooler crust of the Colorado Plateau (average Moho temperature 700°C , constant gradient of $17.5^{\circ}\text{C}/\text{km}$) and Northern Basin and Range (average Moho temperature 740°C , constant gradient of $22.4^{\circ}\text{C}/\text{km}$) can therefore accommodate 55.8 ± 9.4 and $53.9 \pm 7.5\%$ SiO_2 , respectively. The Southern Basin and Range, in contrast, must have a predominantly mafic composition of $51.2 \pm 4.3\%$ SiO_2 to reach similar Vs because of its thin crust and 800°C temperatures.

Our model does not account for contributions to the geotherm from heat production. The first order approximation of a linear, conductive geotherm could cause us to be underestimating temperatures in regions of high heat production, particularly at shallower depths in the crust. Correcting for heat production, however, is nontrivial. If we are underestimating temperature by excluding heat production, then it's possible that the deep crust is more mafic than predicted here. A more mafic crust, however, should have low heat production, causing a negative feedback loop: that is, by accounting for heat production, we are lowering the contribution from heat production. Calculating Moho temperature from surface heat flow measurements (Davies, 2013), literature heat production (Rudnick and Gao, 2014), and thermal conductivity (Jaupart et al., 2016) values yields results that vary by hundreds of degrees and are inconsistent at best. Future calculations regarding heat production and the geotherm will depend also on crustal heat producing element abundances, which are currently debated (Hacker et al., 2015b).

The temperature gradient in the lower crust necessitates a vertical gradient in miner-

alogy and composition. The crust becomes increasingly mafic with increasing depth. This trend is observed most prominently in areas of thicker crust. The increase in V_s cannot be explained by isochemical chemical changes in the lower crust - that is, we cannot explain the observed V_s by simply projecting mid-crustal compositions to higher pressures and temperatures (Figure 4.7). As noted by [Christensen and Mooney \(1995\)](#), we must invoke a compositional gradient within the lower crust to explain the increase in seismic velocity.

In the topmost portions of the Colorado Plateau's lower crust, our model can accommodate over 59 wt.% SiO_2 (Table 4.2). However, such intermediate-felsic material cannot reach high enough velocities to match the seismic signal deeper in the crust, where temperatures increase above 700°C ([Schutt et al., 2018](#)). Furthermore, our set of granulites can explain the seismic signal at the base of the crust more often than at the top, whereas we might expect equal probabilities at all depths if the lower crust were compositionally uniform (shown by the colors of Fig. 4.5). The Northern Basin and Range and Colorado Plateau (Fig. 4.6) show 3 wt.% to 6 wt.% decrease in SiO_2 and an increase in MgO , FeO , and CaO with increasing depth. The Southern Basin and Range, though, seems to lack this trend, the lower crust remaining consistently at 51 wt.% SiO_2 . This is possibly due to removal of more felsic material from the top of the crustal column, which we discuss in section 4.7.1.1.

The specific mineralogy of the lower crust is trickier to constrain than the bulk composition because of its strong dependence on our initial assumptions. Provided that our lower crust is dry and equilibrated in the granulite metamorphic facies, we expect to see mineral assemblages that are rich in clinopyroxenes, garnets, and plagioclase ([Rudnick and Fountain, 1995](#)). Few studies that characterize the whole rock compositions of gran-

ulite quantitatively report mineralogy. This makes comparison between our results and petrological studies of our samples difficult. Though *Perple_X* builds bulk rock velocities from mineral constituents, many mafic rock-forming minerals have similar V_s under lower crustal pressure and temperature conditions (e.g. at 650°C and 0.85 GPa diopside: 4.60 km/s; almandine: 4.57 km/s; spessartine: 4.65 km/s; anorthite: 3.65 km/s; sanidine: 3.49 km/s). A sample may therefore change mineralogy without drastically changing its bulk rock properties or composition. In addition, our model's mineralogy predictions are more sensitive to temperature than its seismic velocity predictions are, due to the abrupt and complete phase changes implemented by *Perple_X*. We do not have the seismic resolution to see such sharp changes in reality ([Olugboji et al., 2017](#)), if they exist at all.

Though we assume that the lower crust is predominantly granulite facies, projections of the geotherm through the Colorado Plateau leave the deep crust at temperatures cooler than what is required for granulite facies metamorphism. Disequilibrium likely exists here, but the assumption that the lower crust has, at some point, reached granulite facies metamorphism is based on observations of granulite grade metamorphic core complexes and xenoliths ([Coney and Harms, 1984](#)). Amphibolite facies lithologies could comprise shallower portions of the crust, or even portions of the lower crust where the crust is thin or cool. Ample amphibolite facies xenolith samples exist for the western United States. In the future, similar combined modeling assessments should consider the role amphibolite and other medium- to high-grade metamorphic lithologies play in the lower crust. However, if the lower crust reached granulite facies at some point in its history, retrograde metamorphism is unlikely to occur due to the thermodynamic barrier of rehydration ([Semprich and Simon, 2014](#)), and the base of the lower crust must be mafic

in our model no matter *which* mafic minerals specifically are present. Broadly speaking, the abundance of garnet and clinopyroxene increases with depth, driving the increase in Vs. Mineral assemblages simplify with increasing depth and temperature, leaving little room for accessory phases, such as ilmenite and kyanite, at the base of the crust.

Further seismic constraints could reduce the uncertainty on our compositions. The Vp/Vs ratio can often distinguish mafic from felsic compositions ([Holbrook et al., 1992](#)). A Vp/Vs of >1.65 would reduce the probability of lower crustal compositions >65 wt.% SiO₂ ([Holbrook et al., 1992](#)) (or, conversely, Vp/Vs of <1.65 would indicate that geochemical studies over-sample mafic compositions). Given that most crystalline rocks exhibit Vp/Vs between 1.6 - 1.9 at standard experimental conditions ([Brocher, 2005](#)), the ratio would have to be tightly constrained at $< \pm 7\%$ variation. Future quantitatively robust modeling efforts of the southwestern United States should also investigate the presence of hydrous minerals ([Dixon et al., 2004](#); [Valentine and Perry, 2007](#)) and melt ([Rey et al., 2009](#)) in the lower crust. The presence of fluids could lower the deep crust's Vs, requiring compositions that are even more mafic than those reported here. Alternatively, melt could cause the temperatures implemented in this study to be over-predicted ([Schutt et al., 2018](#)).

4.7.1.1 Implications for Crust Formation

Ductile spreading and uplift of the lower crust could explain the correlation between crustal thickness and composition. Brittle thinning of the upper and/or middle crust through normal faulting allows for isostatic uplift of ductile, deeper crust with little

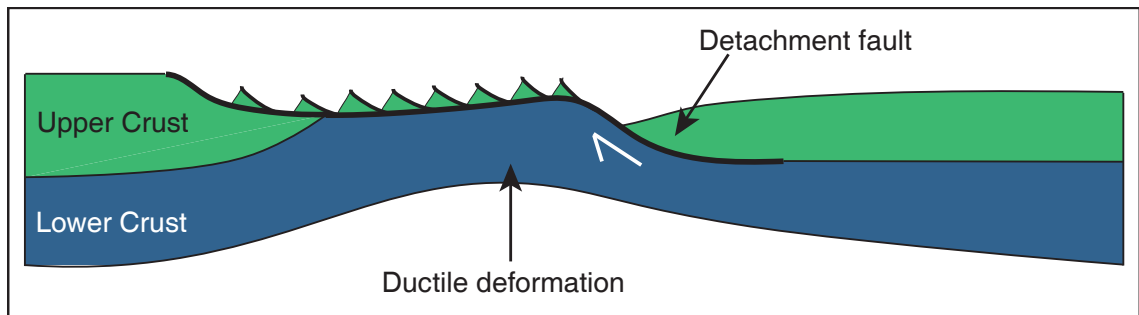


Figure 4.8: The west-east schematic cross-section of the southwestern United States deformation that our model supports. Crustal extension is accommodated primarily through brittle faulting in the upper crust, which is separated from the lower crust by a rolling hinge detachment [Cooper et al. \(2010\)](#). The ductilely deformed lower crust is uplifted but not substantially thinned.

change in lower crustal thickness ([Cooper et al., 2010](#)), illustrated by Figure 4.8. Because the thinnest regions of the southwestern United States are also the most mafic, what was once “lower crust” now likely comprises a greater volume of the 25-28 km thick crustal column. Crustal thickness was lost as intermediate and felsic material was removed from the top, rather than through lower crustal delamination ([Rudnick and Gao, 2014](#)). The composition of the deepest layers of Northern Basin and Range are similar to the deep Colorado Plateau 53 wt.% SiO₂, further suggesting that the extended crust has not lost a mafic root relative to the thicker crust. Had crustal thinning been caused by delamination, the thinnest Southern Basin and Range crust would be the most felsic rather than the most mafic. Based on the lower crust’s mafic composition, the eclogitization process required for delamination would likely only be triggered at high pressures ([Semprich and Simon, 2014](#)). While the Basin and Range is certainly hot enough to undergo delamination ([Jull and Kelemen, 2001](#)), it would have had to occur during a time of crustal thickening (for instance, during the Laramide Orogeny ([Bird, 1984](#); [Livaccari, 1991](#)), not during extension.

4.8 Conclusion

Joint modeling of geophysical and geochemical properties of the lower crust can help constrain lower crustal composition. As an example, in the southwestern United States, seismic velocities, when paired with Moho temperatures and thermodynamic calculations, indicates that the lower crust transitions from intermediate to mafic composition, SiO_2 content decreasing by up to 6% with increasing depth. Temperature gradients cause compositional distinctions to arise among the relatively cool Colorado Plateau, the warm Northern Basin and Range, and the hot Southern Basin and Range. The predominantly mafic composition of the lower crust reflects the tectonic history of this region and can help distinguish between different crustal deformation mechanisms.

Though global-scale models give a generalized view of the lower crust, nonunique solutions to composition can be better constrained in regional-scale studies by combining high resolution local datasets and compositional proxies. Combining seismological, petrological, and thermodynamic data opens avenues for future detailed investigation into deep crust composition and structure.

Chapter 5: The composition of the deep continental crust inferred from geochemical and geophysical data

5.1 Overview

Combining many different geophysical and geochemical techniques together helps us understand the composition of the bottom two-thirds of the continental crust. Building on the model developed in the previous chapter, we expand our joint geophysical-geochemical modeling techniques to a global scale. We account for various factors, such as crust temperature and tectonic setting, to create a large-scale model for the composition of the deep crust. Then, using joint probability analyses between major and trace elements, we predict heat production in the continental crust and heat flux across the Moho.

[1] This chapter has been accepted in the *Journal of Geophysical Research: Solid Earth* as: Sammon, L. G., McDonough, W. F., & Mooney, W. D. (2022). The composition of the deep continental crust inferred from geochemical and geophysical data.

[2] LGS, WFM and WDM contributed to the conceptualization and methodological development of this project. LGS did software development, modeling, visualization, and

writing. WFM and WDM contributed input and discussion throughout, as well as with the revising and editing. WDM provided the seismic surveys used to build this model.

5.2 Abstract

The combination of geochemical and seismological results places tighter constraints on the composition of the middle and lower continental crust than either field can achieve alone. The inaccessible depth of the deep crust (typically >15 km) forces reliance on analogue samples and modeling results to interpret its bulk composition, evolution, and physical properties. A common practice relates major oxide compositions of small- to medium-scale samples (e.g. medium to high metamorphic grade terrains and xenoliths) to large scale measurements of seismic velocities (V_p , V_s , V_p/V_s) to determine the composition of the deep crust. We provide a framework for building crustal models with multidisciplinary constraints on composition. We present a global deep crustal model that documents compositional changes with depth and accounts for uncertainties in Moho depth, temperature, and physical and chemical properties. Our 3-D deep crust global compositional model uses the USGS global seismic database ([Mooney, 2015](#)) and a compilation of geochemical analyses on amphibolite and granulite facies lithologies ([Sammon and McDonough, 2021](#)). We find that SiO_2 abundance decreases from 61.2 ± 7.3 to 53.8 ± 3.0 wt.% SiO_2 from the middle to the base of the crust. In addition, we calculate trace element abundances as a function of depth from their relationships to major oxides. From here, other lithospheric properties, such as Moho heat flux, are derived (18.8 ± 8.8 mW/m²). This study provides a global assessment of major element composition in the

deep continental crust.

5.3 Introduction

The deepest parts of Earth's crust are widely inaccessible to traditional geochemical sampling and so their composition is poorly understood. Only in areas where eruptions have brought xenoliths to the surface or where tectonic activity has exhumed medium and high grade metamorphic terrains are we able to *partially* determine the composition of the deep (middle and lower) continental crust. Even so, these *ex situ*, aged, weathered, and transported rocks may not adequately represent the overall, current composition of the deep crust. Such inaccessibility has challenged geochemists for decades, leading to competing models for continental crust and bulk silicate Earth (BSE) compositions, formation, and evolution. Dissonance in the geochemical community stems from known and unknown unknowns; that is, we are mostly certain of the uncertainties in our geochemical and petrological measurements, but we are uncertain if our samples are truly representative of large swathes of the deep crust or if they are merely point samples. Xenoliths and terrains are the sum of the processes that form them, which may cause them to differ from what is presently at 15-45 km and deeper. The deep crust is an enigma, and compositions of xenoliths and high grade metamorphic terrains provide only an incomplete cipher.

Seismological techniques, however, provide another piece of the cipher by directly measuring the physical state of large sections of the deep crust. Physical properties (e.g. density, Poisson's ratio, V_p , and V_s) determined from these *in situ* geophysical experiments can be compared to laboratory experiments on rocks of known compositions, par-

ticularly medium to high grade metamorphic lithologies (amphibolite and granulite facies rocks) to place constraints on estimates of deep crustal composition. Integrating geochemical and geophysical observations, related to each other by empirically (laboratory) derived thermodynamic properties, provides a reinforced, clearer, consistent picture of middle and lower crustal composition.

This study uses geophysical and geochemical datasets to build a global compositional model of the lower two-thirds of the continental crust. We generate a composition versus depth model for the middle and lower continental crust by applying thermodynamic modeling software to medium and high grade lithologies. We then compare the thermodynamically-generated seismic velocities to velocities obtained from seismological measurements to produce a jointly constrained geochemical-seismological compositional model.

5.4 Methods

Our model calculations are conducted in two steps: 1) assembling the data and performing thermodynamic calculations, and 2) adjusting model parameters to generate deep crustal compositional models with uncertainties. These calculations require seismic velocity depth profiles, Moho depths, and crustal temperature gradients for areas of interest. Using the thermodynamic modeling software *Perple_X* ([Connolly, 2005](#)), we calculate the probability that different deep crustal compositions could produce the observed seismic signal. These calculations are conducted using our modeling software, *CrustMaker*, which is provided as an electronic supplement. The calculation adopts a subdivision of the

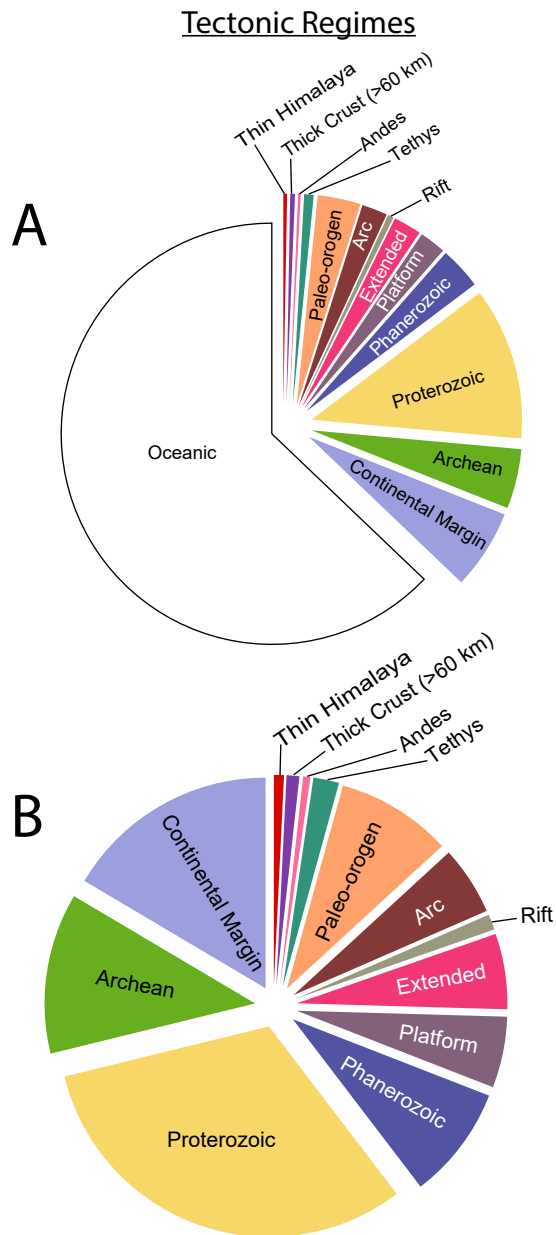


Figure 5.1: The weighted area proportion of crustal types, or “tectonic regimes”, used for our model as A) a fraction of total crust and B) a fraction of continental crust. Proterozoic crust is most abundant (32% of the continental crust), followed by continental margins (16%) and Archean crust (12%). Modern and paleo-orogens, including arcs, make up a combined 19% of the continental crust in our model.

global continental crust into 13 tectonic provinces (Figures 5.1 and 5.2) to speed calculations and extrapolate results to areas with lower data coverage. Individual tile resolution of this global model is set to 1°latitude x 1°longitude x 3 km depth as a default, but can be changed in the model to suit user needs. We chose this default resolution for our global model based on the resolution of our crustal categories (each 1°x 1°of crust was assigned a tectonic province based on models such as CRUST1.0, Litho1.0, and modifications discussed further in Section 5.4.1), and the resolution of our crustal thickness and temperature data, the ramifications of which are discussed further in the Results section. For considering higher resolution, regional scale data, the same methods can be used. Instead of simplifying the crust into tectonic provinces, calculations are run for individual seismic velocity profiles, so that if there are, for example, 34 seismic velocity profiles as inputs, there will be 34 locations for which compositional profiles are generated.

We calculated the overlapping probability between measured seismic velocities and the *Perple_X*-derived velocities for amphibolites and granulites equilibrated at middle and lower crustal pressures and temperatures. A reference average crustal density of $2900 \pm 200 \text{ kg/m}^3$ is used to calculate lithostatic pressure (Wipperfurth et al. (2020), cf., Christensen and Mooney (1995)). Integrating the area under both curves, the area shown as magenta in Figure 5.3, for a sample of composition *X* yields the total probability of sample *X* producing the observed seismic signal. Repeating this technique for a multitude of sample compositions at various depths and temperatures yields a final Monte Carlo model for a deep crustal composition. Probability distributions are generated for *V_p*, *V_s*, and *V_p/V_s* and then multiplied together to further constrain the final probability.

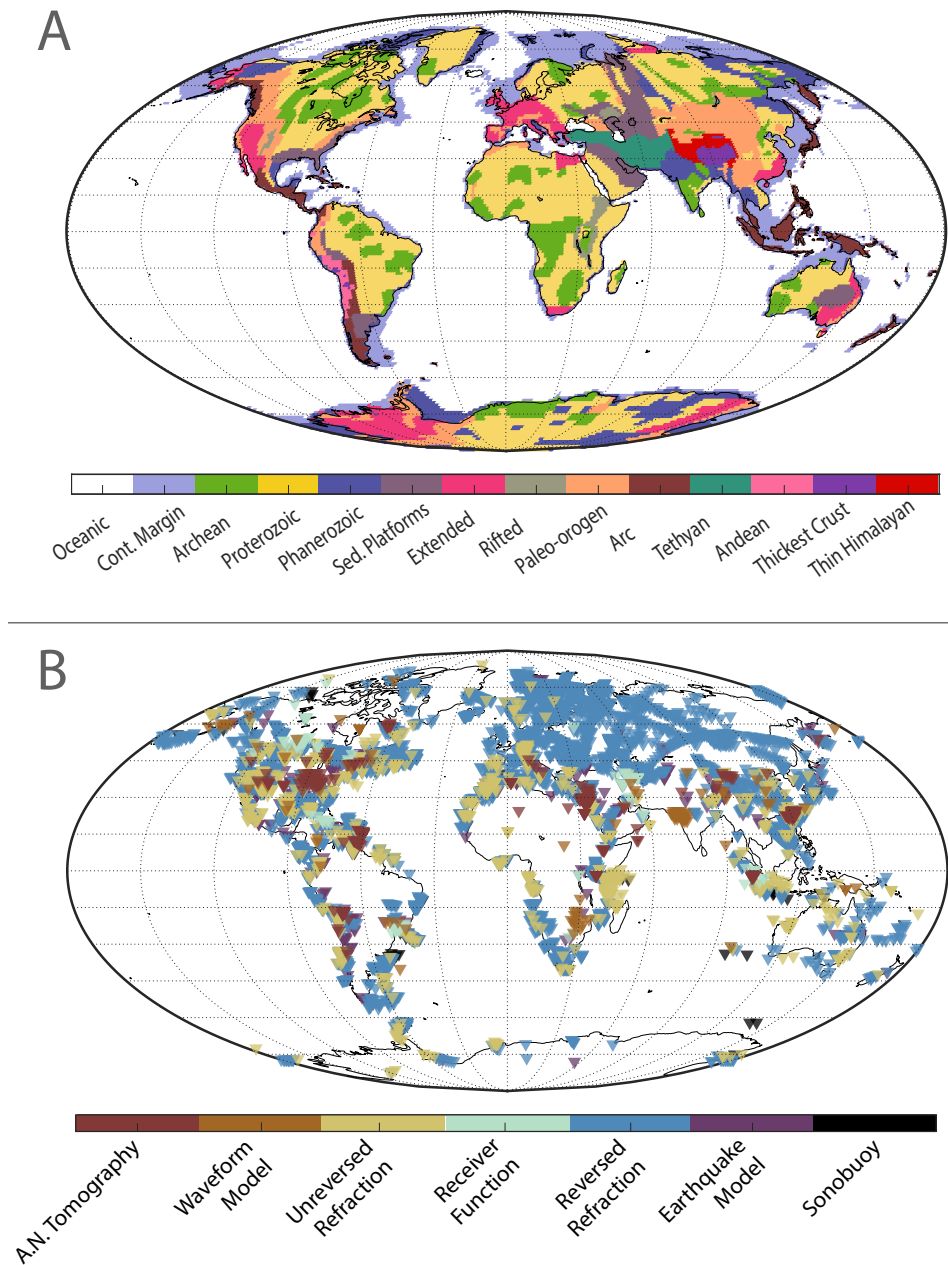


Figure 5.2: A) The mapped distribution of our 13 crust types and B) the seismic velocity profile data distribution from the USGS database. Data coverage is greatest in the northern hemisphere while places with less coverage, like Africa and Antarctica, rely more heavily on extrapolation of crust type.

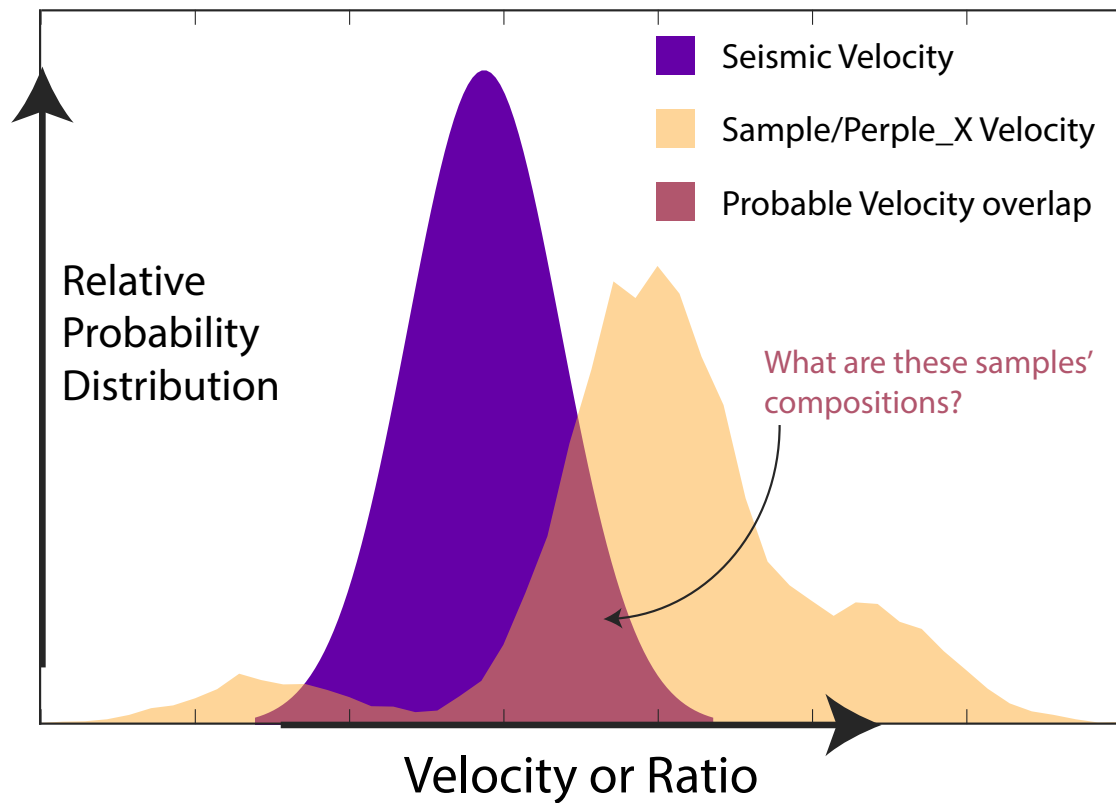


Figure 5.3: A conceptual illustration of overlapping velocity distributions used to identify probable crust compositions. The central pink region of the diagram, where the measured seismic velocity distribution (purple) overlaps the Perple_X-generated velocity distribution (tan), are the velocities that are considered the best-fit by the model. The model records the compositions of the samples that can produce the best-fit velocities.

5.4.1 Model Inputs

A global model of V_p , V_s , and V_p/V_s was generated from a compilation of over 5000 (V_p) and 1000 (V_s) 1-D seismic velocity profiles obtained from the Global Seismic Catalog (GSC) database (Mooney, 2015). Both controlled and passive source seismic velocity profiles were included to increase data coverage. We included profiles with V_p and V_s data that had been sampled at a minimum of 5 depth intervals within the crust. Figure 5.2 shows our tectonic province designations and the location of each seismic velocity profile used. We used global Moho depths from Litho1.0, except on the continental margins, where we reference Szwillus et al. (2019) Moho values. In comparison to Litho1.0, Szwillus et al. (2019) incorporated a larger dataset on the continental margins (~ 1600 profiles) and did not average depths across the continent-ocean transition. Global Moho temperatures were generated from the TC15 global temperature model of Artemieva (2006). We assumed a linear temperature gradient within the continental crust, though we address the contributions from crustal heat production in a later section of this paper. The Szwillus et al. (2019) and Artemieva (2006) models are based on the same USGS-maintained database as Mooney (2015). In most areas, the middle and lower crust are not in thermal equilibrium with the conditions of amphibolite and granulite facies, respectively. Temperature profiles suggest that the deep crust has cooled off since reaching peak metamorphic grade. To account for this in our composition calculations, we adopt the assumptions of Hacker et al. (2015), which uses an equilibration temperature of 500°C for amphibolite facies in the middle crust and 750°C for the granulite facies in the lower crust. We performed a temperature correction of $4 \times 10^{-4} \text{ km s}^{-1} \text{C}^{-1}$ (Christensen

and Mooney, 1995; Rudnick and Fountain, 1995) to account for the crust's cooling. We also perform a pressure correction of $-2 \times 10^{-4} \text{ km s}^{-1} \text{ MPa}^{-1}$ on both facies (Christensen and Mooney, 1995; Rudnick and Fountain, 1995).

The foundation of the tectonic provinces chosen for this global model are the classifications of crust provided by the Crust family of models (Mooney et al., 1998). To further identify tectonic provinces and group together geophysically similar crust, we incorporated crustal thickness, seismic velocity (V_p , V_s), gravity anomaly, sediment thickness, crust elevation, and surface heat flux observations in a tSNE test (t-distributed stochastic neighbor embedding). Results generally favored grouping the continental crust into 8 - 12 regimes, mostly matching the designations already given in Crust1.0. We augmented these regimes with additional groupings, such as “Thinner Himalyan” crust, when it became clear that the seismic velocity structure of the perimeter of the Himalayas differed from the thickest Himalayan crust. Areas with sparse seismic coverage such as central South America and northern Africa, rely heavily on extrapolation of measurements from similar tectonic provinces. Average V_p and V_s profiles for most tectonic provinces were created from a distribution of tens to hundreds of individual measurements. A notable exception is the “Continental Margins” province, which was represented by $> 1,600$ profiles. Highly localized regimes, such as Andean or Himalayan crust, tended to have < 100 profiles due to the uniqueness of their crustal profiles.

Figure 5.1 and Table 5.1 show the proportion of different crustal provinces by surface area coverage. These tectonic provinces consider only crust exposed at the surface, so that provinces such as “Platform” have underlying crystalline crust that may be Proterozoic or Archean in age. The Proterozoic crust covers the largest fraction (32%) of

Crustal Regime	% All Crust	% Continental Crust	Number of Profiles
Oceanic	63	-	-
Continental Margin	6	16	1693
Archean	5	12	416
Proterozoic	12	32	919
Phanerozoic	3	9	353
Platform	2	5	318
Extended	2	6	403
Rifted	<1	1	148
Arc	2	5	262
Paleo-orogenic	3	9	565
Tethyan	<1	2	59
Andean	<1	1	31
Thick Crust	<1	1	106
Thin Himalayan	<1	1	28

Table 5.1: Tectonic Regimes by Surface Area

the continental crust, followed by continental margins (17%). The thicknesses of the crust and lithospheric mantle in Table 5.1 were calculated from the Litho1.0 Moho and lithosphere-asthenosphere boundary depths. The masses and volumes of the lithospheric mantle are from [Wipperfurth et al. \(2020\)](#)’s physical properties associated with Litho1.0.

[Sammon and McDonough \(2021\)](#) serves as our geochemical constraint on the deep (middle and lower) continental crust. That study represents a compilation of major and trace element abundances for amphibolite and granulite facies rocks. We modeled amphibolite facies lithologies for the middle third of the crust and granulite facies lithologies for the bottom third, in agreement with the depth assignment of [Rudnick and Gao \(2014\)](#). We cannot confidently determine *a priori* which portions of the deep crust are more appropriately represented by amphibolite versus granulite facies data with our current model. In theory, one facies-classified lithology would have greater overall overlap with the seismic velocity profile(s), thus determining which is the more accurate rock type to use. In prac-

tice, however, amphibolite and granulite facies lithologies of the same SiO_2 abundance tend to have similar seismic velocities (see Section 5.5.1). As such, we have assumed a metamorphic facies switch from amphibolite to granulite at 2/3 the crustal depth. Future investigations should use seismic anisotropy of the deep crust to further constrain its mineralogy.

Though trace elements do not participate in thermodynamic calculations, we were able to estimate trace element abundances based on a joint probability analysis with the mineral-forming major oxides. Samples were placed into bins based on the abundance of the oxide and trace element of interest (e.g. SiO_2 and U). Bin width was selected using Sturge's rule ($N_{bins} = \log_2(N) + 1$). For each major oxide composition bin, there was then a correlated trace element abundance distribution.

5.4.2 Model Uncertainties

Errors in seismic and/or geochemical inputs will skew results. It is imperative to understand the uncertainties in the input datasets, if we want an accurate picture of the uncertainty of our crust compositional models.

The program also will not assess modeling errors stemming from foundational assumptions about what types of lithologies should be used as geochemical inputs and the tectonic provinces assigned to the global crust. These two assumptions are expected to control the systematic error of the model, which is why we made the program flexible and modular. Our approach facilitates testing different crustal models and highlights the projected differences in crust composition.

The primary sources of model error stem from uncertainty in the assumed crustal temperature gradient and Moho depth. Again, these are parameters that can be set by the user. For our preferred model, the uncertainty on Moho depth is on the order of 10% or less in most areas of the global model (Szwilius et al., 2019). The temperature uncertainty is much greater. Global Moho temperatures are taken from Artemieva (2006), which reports no uncertainties. Therefore, uncertainty is taken as the standard deviation of all temperatures found within a given crustal tectonic province (discussed below), and the model runs a Monte Carlo of at least 10,000 iterations to produce a distribution of Moho depths and temperatures. Future results could be improved with Moho temperature models that quantify uncertainty more directly.

We have also attempted to mitigate the bias introduced by oversampling of particular geochemical compositions. An over-sampled composition, such as 100 input compositions with nearly identical major oxide content artificially inflates the probability of that composition in our final combined model. However, we do consider the reporting of compositions to be at least somewhat reflective of the proportion of rock types present in the deep crust, i.e. if the distribution of reported compositions is bimodal, the rocks in the deep crust are likely bimodal in composition. Therefore, we only considered a sample redundant if its oxide content differed from another's by < 3 wt.% (9 major oxides, using the distance between vectors formula $d = \sqrt{x_1^2 + x_2^2 + \dots + x_n^2}$, where x_n is the difference in wt.% of an oxide between two samples), and its Perple_X generated values for Vp, Vs, and Vp/Vs were within uncertainty of each other.

The internal error contributed by calculational uncertainty is minimal. The overlap between of seismic velocity measurements and Perple_X-derived seismic velocities

is calculated via trapezoidal numerical integration at intervals determined by the uncertainty in the seismological data. When this interval is too large to use for the integration, the program reduces the interval by half. The precision errors of *Perple_X* are generally negligible compared to the uncertainty on our other inputs ([Connolly, 2005](#)). The systematic uncertainty of *Perple_X*, however, remains difficult to assess. As the software is built upon empirical measurements and thermodynamic equations, the experimental database for crustal minerals is incomplete and relies on ideal thermodynamic relationships to compute expected physical properties.

5.4.3 Quality, Expense, and Time: Global vs. Local Models

In numerical modeling, there is often a tradeoff between computation time and model resolution. For a global view of the continental crust, breadth and total model coverage may be more valuable than high data resolution, especially if results can be averaged over large areas. A large-scale, globe-encompassing model, however, comes with the choice of either short computation time and low resolution or longer computation time and higher resolution. Alternatively, those interested in a more in-depth analysis of a localized region may be able to accommodate higher resolution models. We suggest considering the following when determining whether to use a global or local scale model: data resolution (especially in seismic velocity profiles), data coverage, and model application. Those with data resolution on the scale of $> 0.5^\circ \times 0.5^\circ$ should consider using the global version of the script. Data sources with higher resolution, such as that provided by the Earthscope USArray, the AUSArray, or the J-ARRAY, should use the regional scale

model. For the remainder of this study, we will analyze global model results. [Sammon et al. \(2020\)](#) presents an example of a local-scale composition analysis using a nascent version of this method.

5.5 Results

5.5.1 Empirical Composition-Velocity Trends

For granulite and amphibolite facies lithologies seismic velocities correlate with SiO_2 content because of its high abundance, as compared to all other oxides. Perple_X-calculated V_p and V_s values at given pressure-temperature conditions show a quadratic relationship between SiO_2 and velocity (Figures [5.4](#) and [5.5](#)). The coefficients of the quadratic vary for different pressures and temperatures, and are ultimately correlated to the empirical mineral physics datasets used in the Perple_X Gibbs free energy minimization. Amphibolite and granulite facies lithologies span similar V_p and V_s values, though the shapes of their distributions are marginally different. This is because the velocities of many amphibolite and granulite facies minerals (e.g., garnet, pyroxene, plagioclase) have similar V_p (~ 7 km/s) and V_s (~ 3.6 km/s) values. Despite considerable scatter in the V_s data, when paired with V_p , a clear trend emerges: increasing SiO_2 abundance leads to decreasing velocities.

Higher V_p/V_s ratios have lower silica content (Figure [5.6](#)), though for a given SiO_2 , there is roughly a 10% spread in V_p/V_s . For both amphibolite and granulite lithologies, increasing V_s can lead to either an increase or a decrease in V_p/V_s ratio. The maximum V_p/V_s for amphibolite facies lithologies at typical middle crustal P-T conditions, is ex-

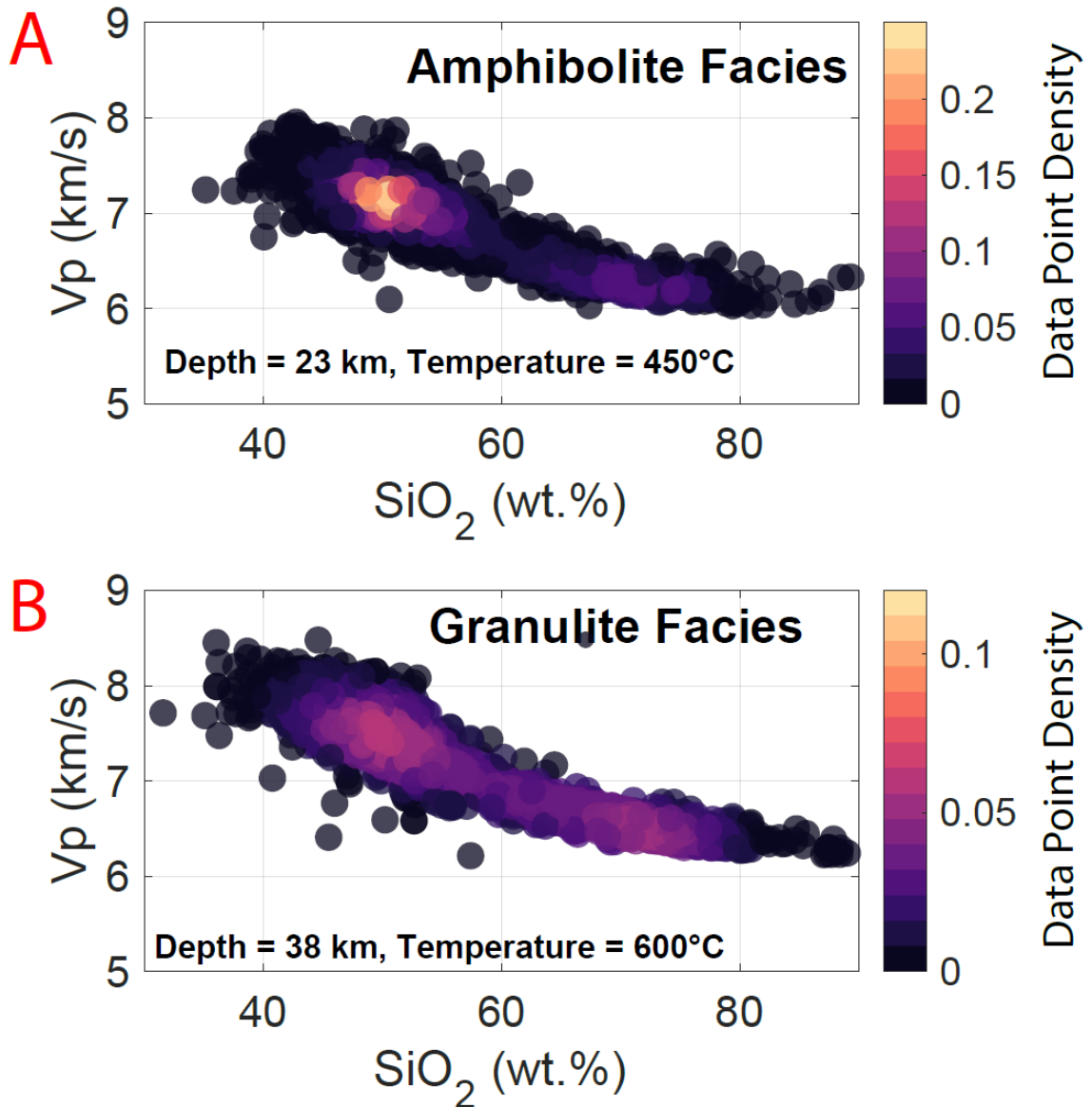


Figure 5.4: V_p as a function of SiO_2 wt.% for amphibolite (A) and granulite (B) facies lithologies at expected deep crustal pressures and temperatures. The color of the data points indicates percent data point density, with the brighter colors indicating more data points. The red line shows the best fit quadratic regression between V_p and SiO_2 and changes for different temperatures and pressures.

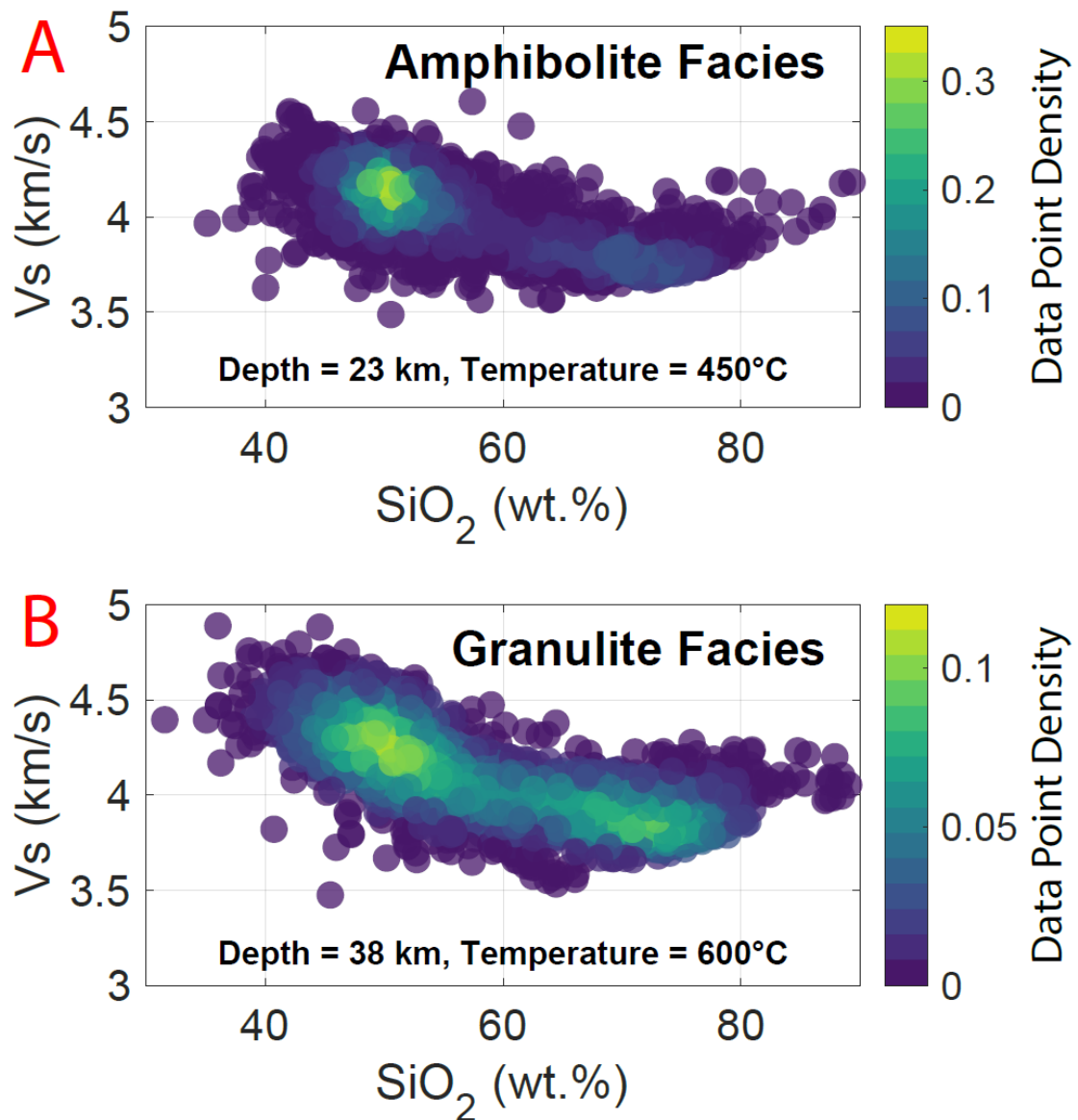


Figure 5.5: Vs as a function of SiO₂ wt.% for amphibolite (A) and granulite (B) facies lithologies at expected deep crustal pressures and temperatures, generated through *Perple_X*. The color of the data points indicates percent data point density, with the brighter colors indicating more data points. The red line shows the best fit quadratic regression between Vs and SiO₂ and changes for different temperatures and pressures. There is more scatter between SiO₂ and Vs than SiO₂ and Vp, but can be combined for a tighter constraint on composition than either compressional or shear velocity alone.

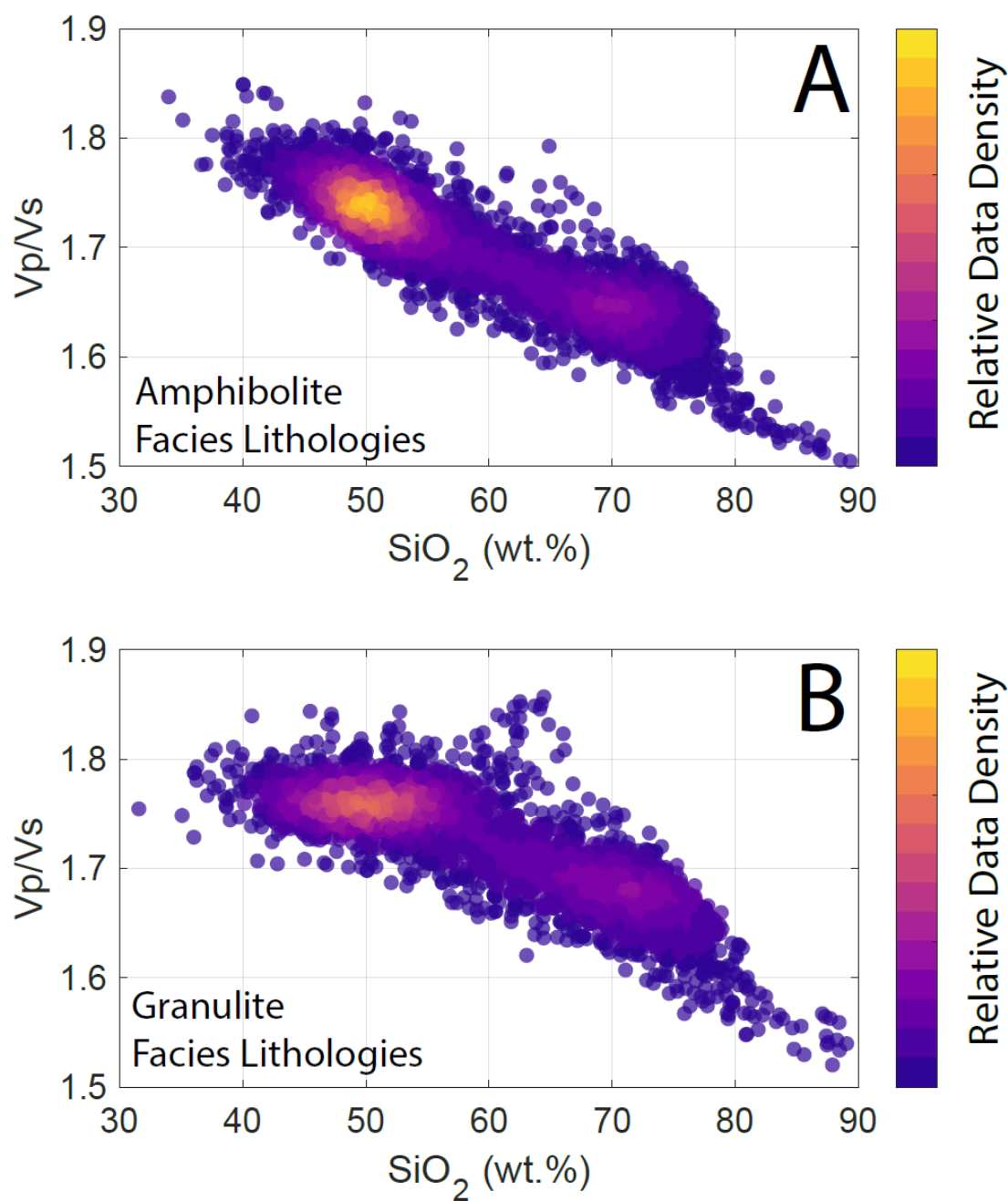


Figure 5.6: Perple_X-generated Vp/Vs plotted against SiO₂ for amphibolite (A) and granulite facies lithologies. Amphibolite results are shown for 450°C and 0.5 GPa. Granulite results are shown for 650°C and 0.8 GPa. Color indicates relative data point density.

pected at a V_s of about 3.5-3.8 km/s, a V_p of 6.5-7 km/s, and SiO_2 of 55 wt.%. For granulite, this maximum is expected at compositions closer to 60-63 wt.% SiO_2 . Interestingly, the maximum V_p/V_s in granulite lithologies corresponds to the lowest V_s rather than the highest V_p , suggesting that V_s variations exert a stronger control on V_p/V_s ratios.

5.5.2 Deep Crustal Density

We calculated deep crustal density by tracking the V_p and V_s values from Perple_X that overlapped with our seismological database back to their input samples. Then, we report the Perple_X-derived density of those input samples. We found that the depth-dependent variation deep crustal densities among the different tectonic provinces appear more similar when normalized to crustal thickness (Figure 5.7). Uncertainties in density are $\pm 3\%$ for each tectonic province, a result which is likely controlled by the $\sim 3\%$ uncertainty on velocities. Deep crustal density ranges from 2700-2780 kg/m^3 at 13 km depth to 3290-3340 kg/m^3 at the Moho.

We note that, in order to calculate deep crustal pressure, and thus mineralogy and composition, we *already assumed* a bulk crustal density of 2900 kg/m^3 to calculate lithostatic pressure. This initial assumption, though, does not significantly affect our composition results because there is, at most, a calculated pressure difference of $<15\%$ caused by using the 2900 kg/m^3 a-priori density vs. our model-generated density. This $<15\%$ pressure difference does not greatly change the stable mineral assemblages or velocities calculated by Perple_X for the deep crust.

The average densities and total masses of each tectonic province listed in Table 5.1

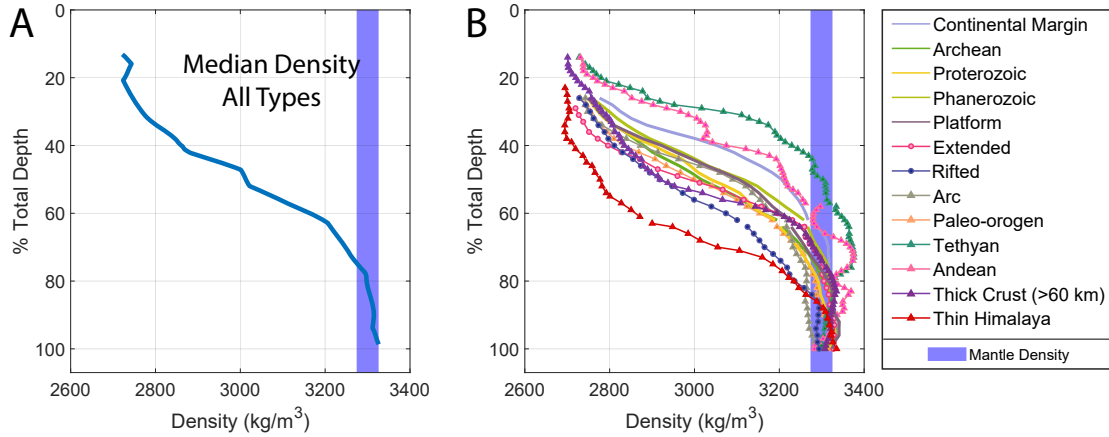


Figure 5.7: Calculated density normalized to depth for (A) average continental crust and (B) our different tectonic regimes with an imposed lithology transition (amphibolite to granulite facies) at 2/3 total crust depth. By this method, the bottom $\sim 20 - 30\%$ of the crust approaches or exceeds mantle density.

are calculated by this study. The average thicknesses, total surface area, and total volume of each tectonic province are calculated from Litho1.0's Moho and LAB depths, whose uncertainties are $\sim 12\%$ (c.f. [Dziewonski and Anderson \(1981\)](#); [Huang et al. \(2013\)](#); [Pasyanos et al. \(2014\)](#)). Continental Margin Moho depths are an exception and are from [Szwilius et al. \(2019\)](#) (Section 5.4.1). The densities of the oceanic crust and lithospheric mantle, not calculated by this study, were assumed to be $2900 \pm 116 \text{ kg/m}^3$ and $3400 \pm 136 \text{ kg/m}^3$, respectively ([Carlson and Herrick, 1990](#); [Huang et al., 2013](#)). The total volume of continental crust is $9.2 \times 10^9 \text{ km}^3$. The continental crust, including submerged continental margins, comprises 44% of Earth's surface area. Multiplying the density of each tectonic province by its volume yields the mass of crust predicted in each tectonic province. We assumed one-third of the crustal column was upper crust with a density of 2650 kg/m^3 ([Christensen and Mooney, 1995](#)). The total mass of the continental crust, $2.32 \times 10^{22} \text{ kg}$, is 4% higher than [Albarède \(1998\)](#); [Cogley \(1984\)](#); [Rudnick \(1995\)](#); [Wipperfurth et al. \(2020\)](#).

5.5.3 Composition

Our main analysis focuses on SiO_2 abundance and its uncertainties because of its strong correlation to seismic velocities, particularly V_p . The SiO_2 content at typical middle and lower crustal depth intervals (Figure 5.8) is given in Table 5.2. All 9 major oxide inputs (SiO_2 , TiO_2 , Al_2O_3 , CaO , MgO , FeO_T , MnO , K_2O , Na_2O) can be found in Table 5.3 and corresponding maps in Supplement Section 3. We use the notation “ $M_{x\%}$ ”, where $x\%$ is the percent distance to the Moho (M) from the surface. This notation normalizes comparisons among tectonic provinces of various thicknesses. We found that when we normalize depth in this manner, the composition vs. depth profiles for each province become more similar. The top of the deep crust starts at an intermediate composition, globally ranging from 58 - 68 wt.% SiO_2 , and the composition gradually decreases to 50-55 wt.% SiO_2 as it approaches the Moho. Global scale SiO_2 composition of the continental crust mostly decreases (or remains steadily mafic) with increasing depth for all tectonic provinces (Figure 5.9). Uncertainty in global SiO_2 also decreases with increasing depth due to fewer samples fitting the seismic signal at depth. However, the Andean and Himalayan tectonic provinces have larger uncertainties at depth because of the variation in geochemical data fitting the seismic signal and the sparsity of deep seismological profiles.

CaO content of the deep crust is also of interest due to its absolute abundance and significance as a contributor to sedimentary deposits, though only siliciclastic rocks and not carbonates were considered viable deep crustal components (Hartmann et al., 2012; Wilkinson et al., 2009). In our model, Ca is mostly contained in plagioclase, pyroxene,

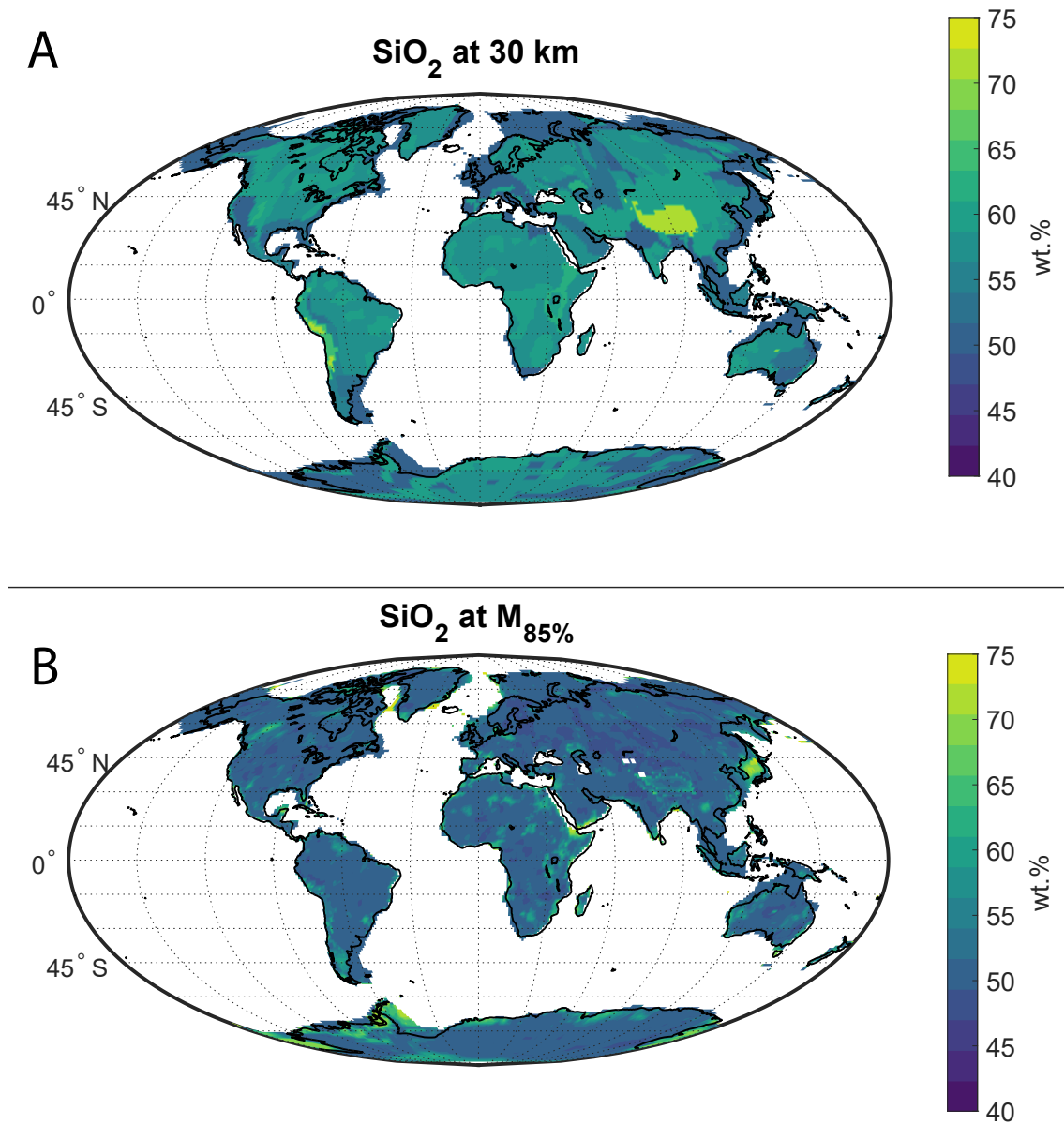


Figure 5.8: Global SiO_2 composition at a depth of 30 km shows regional distinctions whereas measuring composition at a crustal depth relative to the Moho ($M_{85\%}$ notation = 85% of the total crustal depth) produces a view of a deep crust that is contiguous and decreases in SiO_2 gradually with depth. Areas of high projected SiO_2 include the Himalayas, Andes, East African rift, and some continental margins. While the Himalayas and Andes may show compositional features, the high SiO_2 in some rifts and continental margins are likely from model input inaccuracies

	SiO ₂ at $M_{50\%}$ ($\sim$ <i>middle crust</i>)	Uncertainty $\pm$	SiO ₂ at $M_{85\%}$ ($\sim$ <i>lower crust</i>)	Uncertainty $\pm$
Continental Margin	58.8	10.2	50.8	2.7
Archean	59.1	7.8	50.3	2.9
Proterozoic	58.0	7.3	50.9	3.7
Phanerozoic	59.2	9.2	49.9	2.5
Platform	56.9	7.6	50.0	2.7
Extended	66.1	7.6	50.6	2.4
Rifted	63.9	7.3	55.3	6.5
Arc	65.9	9.8	55.2	6.9
Paleo-orogenic	61.6	9.2	50.8	3.4
Tethyan	61.7	9.4	50.5	3.1
Andean	57.7	9.1	54.0	6.9
Thick Crust	65.4	8.8	56.0	8.6
Thin Himalayan	68.9	5.9	50.0	2.2

Table 5.2: SiO₂ in wt.% for different tectonic regimes

and garnet. The CaO abundance tends to increase with depth because of the increasingly mafic nature of the deep crust, and therefore regions of low SiO₂ correlate with regions of high CaO. Globally, the median CaO at crustal depths of $M_{85\%}$ is 9.1 ± 3.1 wt.% (Figure 5.10).

We can also derive the global distribution of a trace element if that trace element has a quantifiable relationship to one of the thermodynamic components (major oxides) used in our model. We used a geochemical database of samples with both major and trace element concentrations (Sammon and McDonough, 2021) to generate trace element maps as a function of major oxide abundances. We used a bivariate probability analysis to generate trace element distributions from a major oxide abundance, such as SiO₂, at a specific depth or location. Although we suggest using regional analyses for high resolution interpretations of trace element abundances, we present here global predictions and uncertainties for Sr (Figure 5.11) and U (Figure 5.12) content based on their relationships

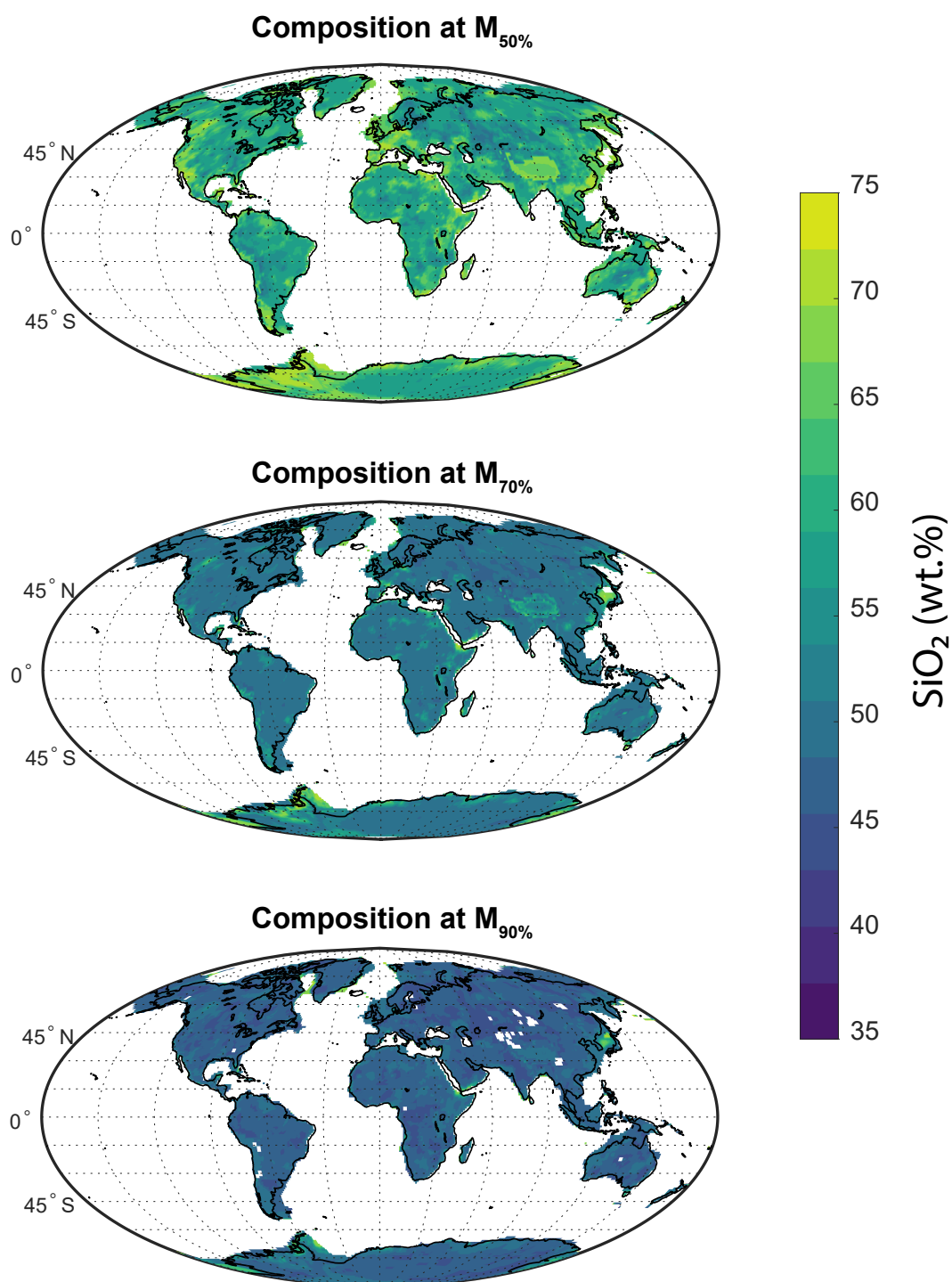


Figure 5.9: Global SiO_2 decreases with increasing depth from the middle to the bottom of the continental crust. The middle crust $M_{50\%}$ ranges from 60 to 65 wt.% SiO_2 in most areas and increases at a rate of about wt.% per km until reaching the base of the crust.

	Composition at $M_{50\%}$ ($\sim$ <i>middle crust</i>)	Uncertainty $\pm$	Composition at $M_{85\%}$ ($\sim$ <i>lower crust</i>)	Uncertainty $\pm$
SiO ₂	59.1	7.31	50.9	2.98
TiO ₂	0.74	0.38	0.82	0.40
Al ₂ O ₃	15.8	1.68	16.4	3.46
FeO	7.26	2.93	9.23	2.25
MnO	0.12	0.06	0.16	0.06
MgO	2.93	1.73	5.60	2.81
CaO	5.52	2.05	8.58	3.08
Na ₂ O	3.64	0.81	2.16	1.02
K ₂ O	1.41	0.97	0.77	0.96

Table 5.3: Middle and Lower Crust Bulk Composition in wt.%

with CaO and SiO₂, respectively, as examples. Global average Sr increases with increasing CaO until plagioclase is no longer the dominant Ca-bearing mineral. Uncertainties on global U concentration span an order of magnitude because the abundance of U in a given metamorphic sample set ranges from a few hundreds of ppb to a few ppm. U and SiO₂ abundances, however, are positively correlated, with median U increasing as median SiO₂ increases.

5.6 Discussion

5.6.1 SiO₂ and Overall Deep Crustal Composition

Figure 5.9 shows steady or decreasing SiO₂ with increasing depth in the deep crust. Figure 5.8 also makes it apparent, though, that the absolute SiO₂ at a given depth is not equal across different crustal types. For example, “Extended” crust appears mafic at 30 km depth while the “Thick Himalayan” crust is felsic at that depth, and “Proterozoic” crust falls in between. However, a more laterally consistent trend appears when compar-

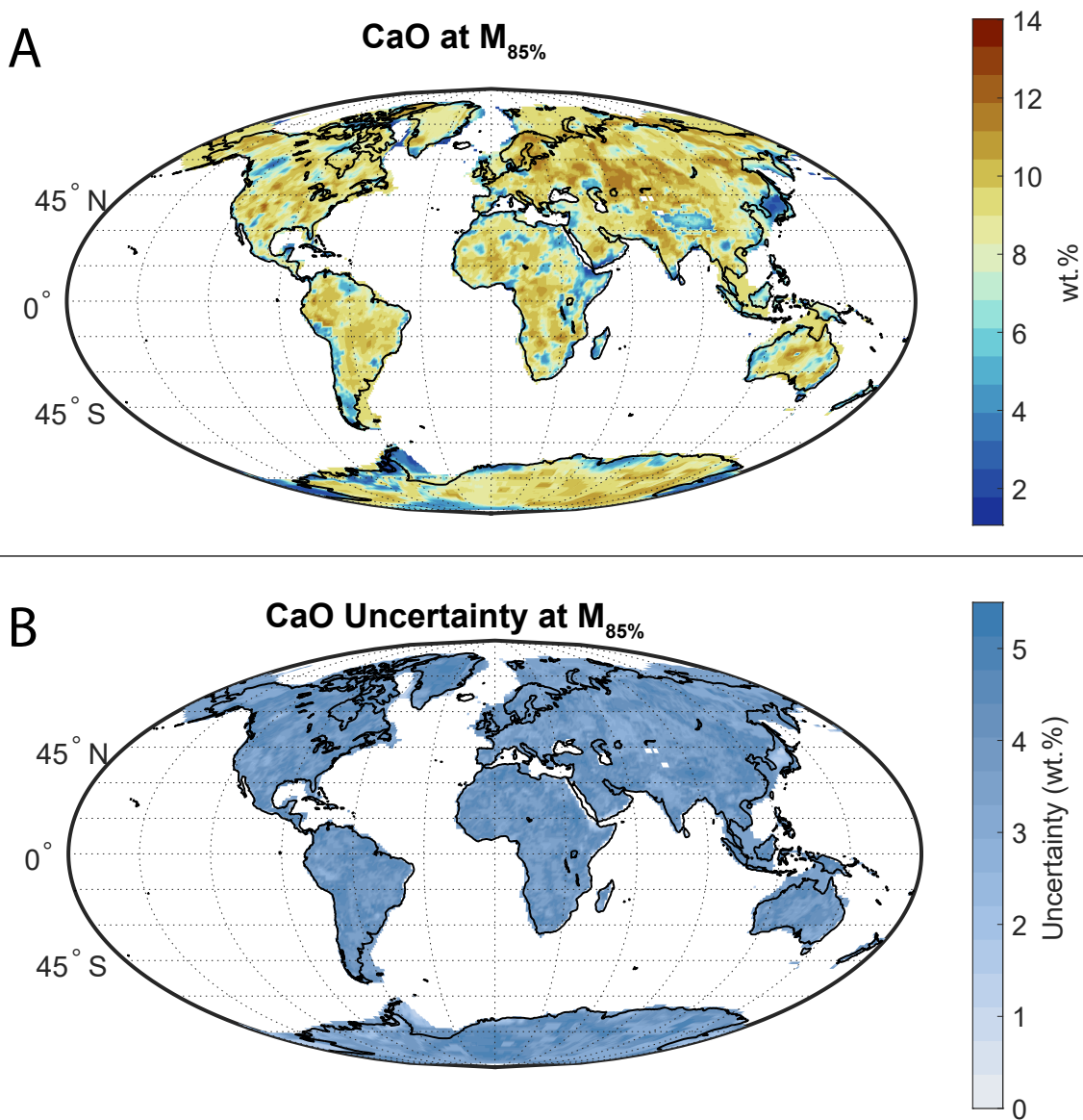


Figure 5.10: Global CaO abundance and uncertainty at 85% of the total crustal depth. Areas of low CaO correlate to areas of high SiO_2 . There does not appear to be any correlation between CaO content and uncertainty, with most regions having 3 to 4 wt.% uncertainty regardless of CaO abundance.

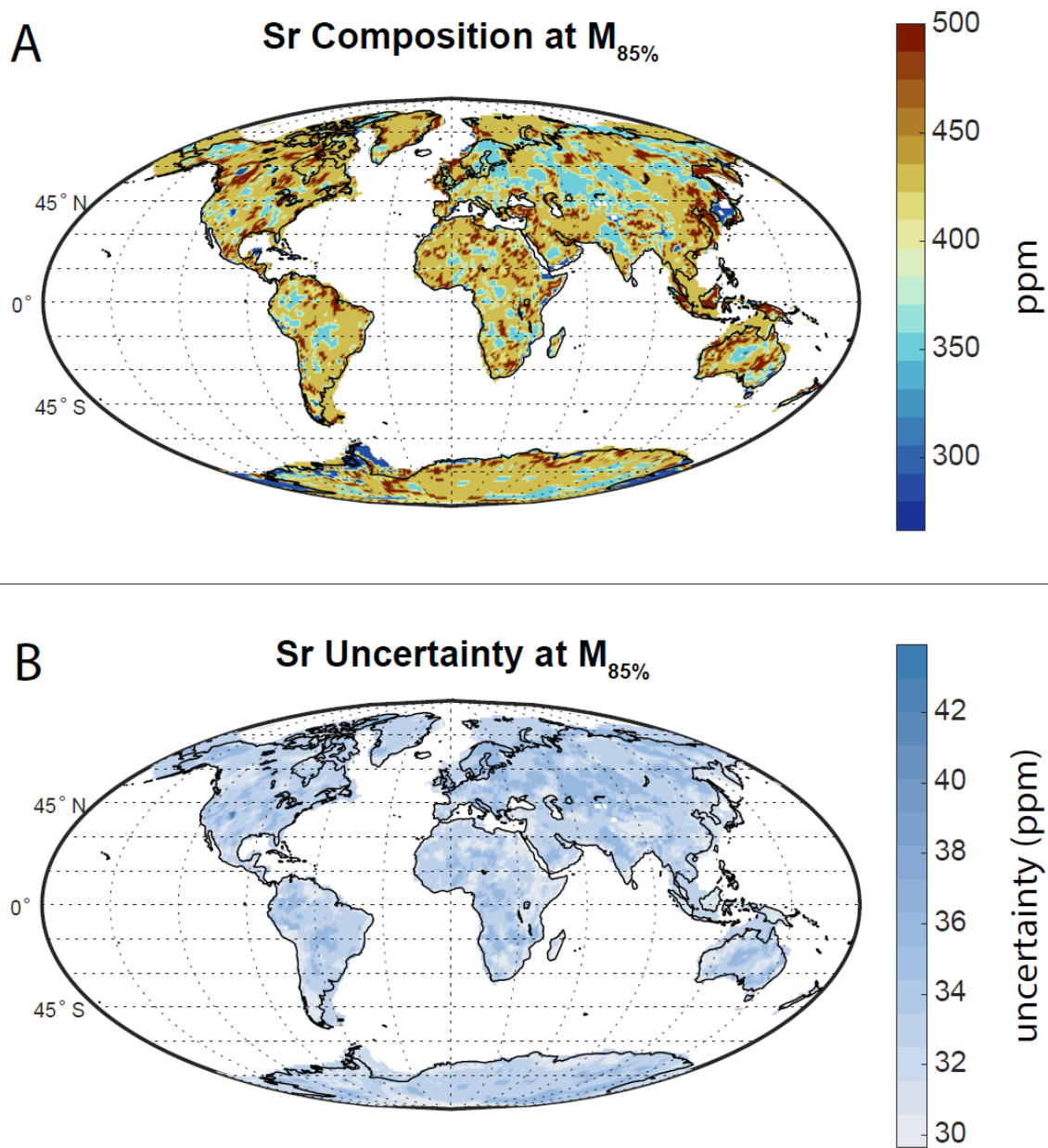


Figure 5.11: Global Sr abundance and uncertainty was derived from a joint probability analysis with CaO at 85% of the total crustal depth.

ing percent of the crustal column traversed rather than absolute depth (Figure 5.8). Most regions show a 5-10 wt.% decrease in median SiO₂ through the deep crust regardless of crustal thickness, so that SiO₂ decreases more rapidly in areas of thin crust than in areas of thick crust. We predict the global median SiO₂ at 50% above the Moho (or, alternatively, 50% crustal column thickness) to be 61.2 ± 7.3 wt.% SiO₂ with CIPW normative mineralogy of <10 wt.% alkali feldspar and <15 wt.% quartz. The average composition of middle continental crust is therefore expected to resemble that of a quartz monzonite; the lower crust average composition, with 50.5 ± 3.4 wt.% SiO₂ and 9.1 ± 3.1 wt.% CaO, is expected to resemble a gabbro-norite. It must be noted, however, that these compositions can be the result of seismic velocity profiles averaging felsic and mafic endmembers of the deep crust, and need not be produced by a monzonite or norite.

Density sorting provides a simple mechanism for producing the compositional structure of the continental crust. The process of crustal genesis leaves mafic, restitic material at the base of the crust regardless of crustal thickness except in the few cases discussed in the next paragraph. More buoyant, felsic material ascends to the top of the crust, producing a gradient of SiO₂ that scales with crustal thickness. In addition to density sorting, the deep crust could be more mafic because it is simply closer to the mantle and therefore has a greater number of mafic intrusions. Our results do not indicate any need for sharp compositional boundaries in the deep crust. The $M_X\%$ notation reinforces the importance of scaled, relative depth in the crust rather than absolute depth for making compositional comparisons.

Two regions that appear conspicuously more felsic than the global deep crustal median are the Andes and the Thin Himalayan crust (Figure 5.9). A low temperature

gradient could be the cause of the compositional difference between these two tectonic provinces and the global average, but we also must consider two other possibilities, particularly around the northern and northeastern Tibetan Plateau and Himalayan ramp. The first is that thick, convergent margins, especially in the Himalayas, might have layers of upper crustal material thrust deeper within the crust, incorporating more felsic material into the deep crust. In contrast, underthrust upper crustal material is less likely to appear in the Andes, which is a continent-ocean subduction zone. Alternatively, pockets of melt and partially melted material in the Andean and Himalayan middle and lower crust could reduce the shear wave velocity (Caldwell et al., 2009; Hacker et al., 2014; Nelson et al., 1996; Regis et al., 2016; Schilling and Partzsch, 2001; Schmitz et al., 1997; Searle et al., 2009). Because our current model does not factor in melt, slower V_s speeds would be attributed to a more felsic composition instead.

Other anomalous regions in Figure 5.8, particularly the continental margins of Antarctica, the East African rift zone, and the Sea of Japan, are likely caused by inaccurate temperature and Moho inputs. The East African Rift could appear felsic because the model's temperature gradient for that actively rifting region is too low; a cooler felsic composition can produce the same velocities as a warmer mafic composition. On the other hand, the highly localized, extremely felsic borders around Antarctica and between Japan and China likely indicate a misclassification of crust type and/or Moho depth. Oceanic-type crust has been documented in both regions (Cho et al., 2004; Gohl, 2008; Hirata et al., 1992; McCarthy et al., 2020). Better Moho and temperature resolution of the ocean-continent transition should increase the accuracy of compositional models in these regions.

Mafic granulite lithologies reach gravitational instability in the lower 10-20% of the average crustal column (Jagoutz et al., 2011), surpassing the upper mantle's density of 3300 kg/m^3 . Therefore, according to Figure 5.7, most of the granulite facies lower crust for continental margins, Andean crust, Tethyan crust, and Phanerozoic crust should be gravitationally unstable. On the other hand, most other tectonic provinces would just reach mantle-like densities around the Moho depths. Thinner Himalayan type crust has a middle crustal density $\sim 9\%$ lower than other provinces, correlating with negative seismic velocity anomalies. Arcs have the next lowest densities on average, suggesting that a potentially denser lower crust beneath some arcs has already foundered (Jagoutz et al., 2011). Figure 5.7B shows that accreted arcs of Andean type crust in particular display a stark decrease in density interpreted as being due to delamination of the lowermost crust (Ducea, 2011; Gao et al., 2021; Kay and Kay, 1993).

Forming continental crust via island arc processes, however, would then require the deep crust to become denser over time, since most of our tectonic provinces have a lower crust that is denser than arc lower crust. This can be achieved by cooling the crust, thickening it further, intra-crustal differentiation, or by mafic igneous injections into the lower crust. Moreover, if our Moho temperature model is too hot, though, it requires denser, more mafic lower crust than is actually present to explain the observed V_p and V_s values. As such, our compositional models are intrinsically tied to Moho temperature, and may be skewed towards mafic granulites. Reducing the assumed crustal Moho temperatures would bring the estimated average crustal density closer to arc crust density and increase the average weight percentage of SiO_2 .

There is a tradeoff between temperature and composition. V_p and V_s both carry

a temperature dependency through their bulk and shear moduli, so accurate temperature estimates are imperative for modeling the crust; decreased seismic velocities can be the result of either higher temperatures or greater SiO₂ content. As a reminder, this study implements a linear temperature gradient through the crust from the TC15 global temperature model ([Artemieva, 2006](#)).

Table [5.4](#) reports compositional models for the middle and lower continental crust, a practice that is required to make meaningful comparisons to previous crustal models. While we recognize the assumption of a three-layer crust as an oversimplification of the diversity of crustal compositions, it is useful for some calculations to have average composition numbers for the crust. A few examples include mantle tomography studies which require crustal correction, crustal corrections for geoneutrino studies; models of Earth's thermal history; and planetary scale compositional model for comparison with other rocky bodies. Compositional models in Table [5.4](#) have been normalized to 100 wt.%. Our middle crustal composition falls between two possible compositions given by [Hacker et al. \(2015\)](#): the fastest Vp endmember composition for the middle crust (62.7 wt.% SiO₂), and the middle crustal composition expected when the crust takes on a two compositional layer (upper and lower) structure, instead of three, (57.3 wt.% SiO₂). These SiO₂ estimates also overlap with the 62 wt.% SiO₂ reported by [Christensen and Mooney \(1995\)](#) and are more mafic than the 63.5 wt.% SiO₂ middle crust reported by [Rudnick and Gao \(2014\)](#). Our proposed lower crust composition agrees with [Rudnick and Gao \(2014\)](#) and other mafic models (e.g. [Hacker et al. \(2015\)](#)'s fast Vp lower crust; [Jagoutz and Schmidt \(2012\)](#)). Models which predict a more intermediate-felsic lower crust, such as the North China craton lower crustal model of [Liu et al. \(2001\)](#) or the higher SiO₂,

lower Vp options listed by [Hacker et al. \(2015\)](#), are not consistent with our global average, though isolated regions of more felsic lower crust may exist.

5.6.2 CaO and Sr

Bulk CaO concentration increases with depth (Figure [5.10](#)), but as a component of mafic, siliciclastic rocks, not carbonate. The lack of carbonate is due in part to our imposed amphibolite/granulite grade lithology restrictions on possible deep crust composition, but is reinforced by high density and Vp values observed in the deep crust. Carbonates, with densities of approximately 2750 kg/m³ and Vp's of 6.6 – 6.8 km/s at deep crustal conditions ([Christensen and Mooney, 1995](#)), cannot substantially contribute to the observed deep crustal velocities. Also, there are few carbonate-dominated granulite facies xenoliths and terrains as compared to silicate granulites. A comparison of Figures [5.8B](#) and [5.10](#) shows good correlation globally between regions of high SiO₂ and low CaO. Uncertainties in CaO estimates follow the same trends as SiO₂ uncertainties, though the former's uncertainty is roughly 10% higher. The relationship between Vp or Vs and CaO content has more scatter than with SiO₂, which increases the % uncertainty on CaO estimates.

The CaO content does, however, predictably track with Sr concentration (Figures [5.10](#) and [5.11](#)). Sr abundances cannot be directly derived from velocity calculations, but they can be predicted from its geochemical relationship with CaO. Patterns emerge when comparing the global distribution of Sr and CaO from two distinct sources: equilibrium mineralogy and data binning. First, Sr abundance increases for CaO contents between 2

	Christen & Mooney, 1995	Liu et al., 2001	Jagoutz & Schmidt, 2012	Rudnick & Gao, 2014	Hacker et al., 2015 [†]	Hacker et al., 2015 [‡]	This Study
	<i>Middle Crust</i>						
SiO ₂	62	-	-	63.5	62.7	57.3	61.2
TiO ₂	-	-	-	0.69	0.8	0.99	0.77
Al ₂ O ₃	-	-	-	15	15.7	16.8	16.4
FeO _T	-	-	-	6.02	6.76	8.15	7.52
MnO	-	-	-	0.10	0.13	0.16	0.12
MgO	-	-	-	3.59	3.51	4.46	3.04
CaO	-	-	-	5.25	5.27	6.63	5.72
Na ₂ O	-	-	-	3.39	3.42	3.89	3.77
K ₂ O	-	-	-	2.3	1.6	1.42	1.46
Mg#	-	-	-	51.5	48.1	43.4	41.9
	<i>Lower Crust</i>						
SiO ₂	47	58.3	52.16	53.4	50.7	57.3	53.8
TiO ₂	-	0.59	0.78	0.82	1.24	0.99	0.87
Al ₂ O ₃	-	13.6	18.68	16.9	16.5	16.8	16.3
FeO _T	-	5.32	8.41	8.57	10.39	8.15	9.75
MnO	-	0.08	0.17	0.10	0.19	0.16	0.17
MgO	-	9.58	5.86	7.24	7.03	4.46	5.92
CaO	-	4.54	10.79	9.59	10.1	6.63	9.07
Na ₂ O	-	2.54	2.56	2.65	2.8	3.89	2.28
K ₂ O	-	3.23	0.41	0.61	0.79	1.42	0.81
Mg#	-	76.2	55.4	60.1	54.7	49.4	52.0

[†] [Hacker et al. \(2015\)](#) fast Vp crustal model

[‡] [Hacker et al. \(2015\)](#) middle crust composition = lower crust composition model

Table 5.4: Continental crust composition estimates

to 6 wt.%, reaching a maximum at about 500 ppm Sr. However, Sr gradually decreases to 350 ppm as CaO increases to >6 wt.%. This shift in Sr abundance corresponds with the transition from plagioclase as the only Ca-bearing mineral phase to the addition of garnet and clinopyroxene as stable Ca-bearing phases.

Second, we see sharp jumps in Sr abundance in neighboring tectonic regions as a consequence of our data binning. The uncertainty on CaO content dictates that the compositional bin-widths are as wide as 2-3 wt.% for a total of six bins. Each bin has a central Sr value and distribution, leading to six possible median Sr abundances. The uncertainties on Sr are a combination of the systematic uncertainty (which CaO bin) and the statistical uncertainty (Sr variation within each bin).

5.6.3 Heat Production and Moho Heat Flux

Radioactive elements produce heat inside Earth, with K, Th, and U, the heat producing elements (HPEs), accounting for 99.5% of the total radiogenic energy output ([McDonough et al., 2020](#)). The few remaining radioactive elements in the planet contribute <0.5% to Earth's total heat production. We can model the abundance and distribution of HPEs in the crust, and Moho heat flux, with our model and supplementary geochemical relationships.

HPE abundance data are provided in the ([Sammon and McDonough, 2021](#)) geochemical dataset. Using a joint probability analysis, we derived U abundances with uncertainties from SiO₂. The middle continental crust has a median U concentration of 1.3 ± 0.2 ppm and the lower crust has 0.22 ± 0.03 ppm. The uncertainty on these median

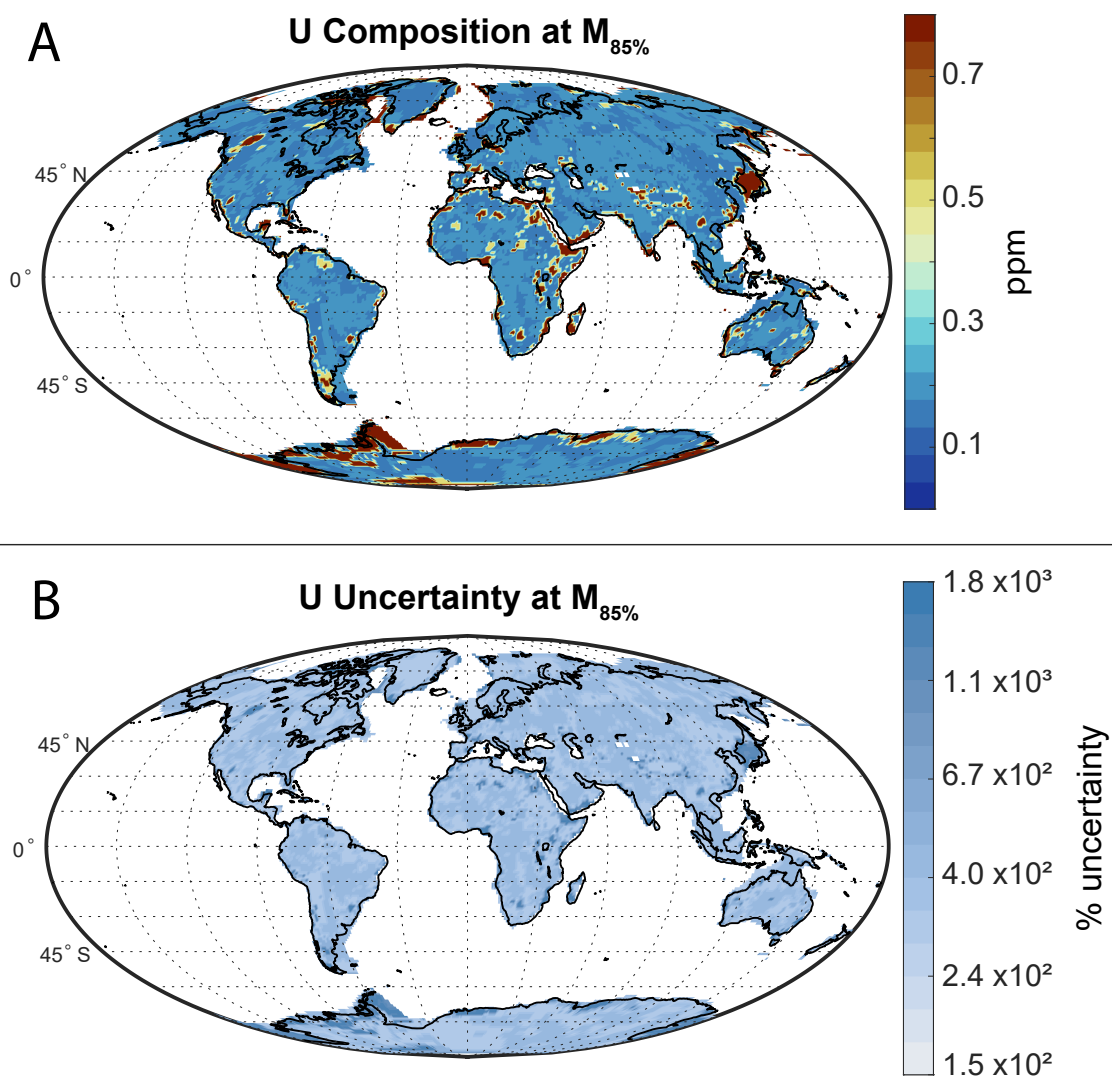


Figure 5.12: Global U abundance derived from a joint probability analysis with SiO_2 at 85% of the total crustal depth. Uncertainties span orders of magnitude because of the range of possible U values, but the global median at this depth is ~ 0.2 ppm U. Regions of high SiO_2 , especially the potentially inaccurate continental margin of Antarctica correlate with high U and the highest uncertainties.

values is reported as the standard error of the median. We do this to highlight that this is the confidence in the central value, the median. The full range of possible abundances spans five orders of magnitude (Fig. 5.12B) Th abundances could also be generated with a joint Th-SiO₂ probability analysis, but we use a constant Th/U_{mass} ratio of 3.77 ± 0.1 (Sammon and McDonough, 2021; Wipperfurth et al., 2018) which provides a narrower constraint on Th concentration. K₂O abundance is calculated directly as a thermodynamic component in our compositional model. We calculate heat production on the same 1°latitude by 1°longitude cell scale that we use for calculating major element abundances. The concentrations of K, Th, and U in each cell are multiplied by each cell's density (calculated in previous section) and volume. The resulting masses of K, Th, and U are then multiplied by the heat produced during their respective decays (McDonough et al., 2020), resulting in a heat production value for each cell in units of W/kg. Heat production values can be converted to W/m³ via multiplying by cell density once more. Uncertainties are calculated through standard propagation of errors on concentration and density. Figure 5.13 maps the total (depth-integrated) median contribution to surface heat flux from the upper, middle, and lower continental crust in mW/m².

We calculated a median heat production of 0.3 nW/kg (0.8 μW/m³) for the middle third of the crust and 0.05 nW/kg (0.1 μW/m³) for the lower third of the crust. Previous studies have also found low heat production in the deep continental crust (Fountain et al., 1987; Jaupart et al., 2016; Kukkonen et al., 1997). Our model is also consistent with local studies of HPE analyses of deep crustal xenoliths, such as Gruber et al. (2021); Pinet and Jaupart (1987); and Ashwal et al. (1987).

The uncertainties on this global model are dominated by uncertainties on U abun-

Contribution to Crustal Heat Production

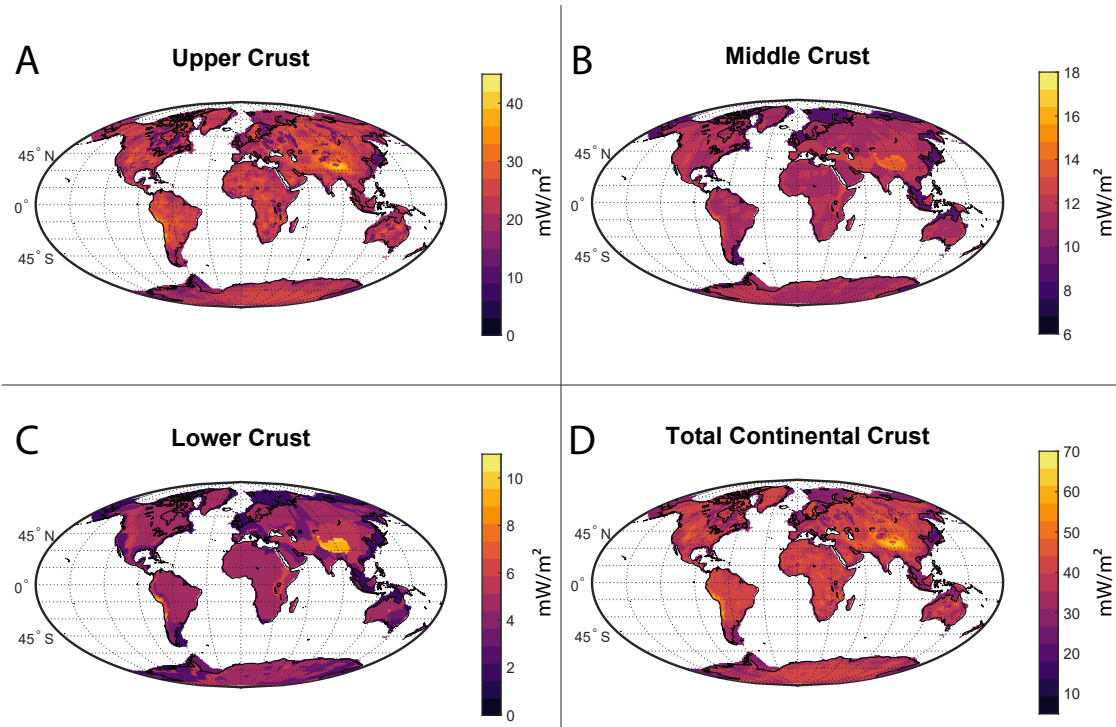


Figure 5.13: Panels A - C show the contribution from each crustal layer (upper, middle, lower) to the total continental surface heat flux (D). Continental crustal heat production is dominated by upper crustal heat production but with significant contributions from the middle crust.

dances, which span an order of magnitude. Even so, our uncertainty on the median or central value (standard error of the median) of HPEs or heat production is well constrained at pm 0.1%. While the range of possible heat production values span an order of magnitude, the median/average heat production value is tightly constrained. Provinces with high predicted SiO₂ content, such as the Andes, have estimated U content up to four times higher than the global lower crustal median because of the correlation between high SiO₂ samples and high U. We recommend using regional HPE data for understanding smaller scale variations and reserve this study's results for continent- or global-scale models.

We took our calculations a step further by estimating Moho heat flux. Once we determined deep crustal heat production, we calculated a subcontinental Moho heat flux by subtracting continental heat production from the global surface heat flux of [Lucazeau \(2019\)](#), supplemented with [Shen et al. \(2020\)](#)'s Antarctica surface heat flux. We assumed a [Gaschnig et al. \(2016\)](#) upper crustal heat production of 1.75 $\mu\text{W}/\text{m}^3$ to complete the initial calculation. Pairing an upper crustal composition of [Gaschnig et al. \(2016\)](#) with our deep crustal composition yielded Moho heat fluxes for tectonically stable regions that agree with [Jaupart et al. \(2007\)](#) and marginally agree with ([Hacker et al., 2015](#))'s models of crust with slow V_p. However, Moho heat flux calculations depend substantially on the assumed HPE abundance model for the upper crust, which contributes $\sim 70\%$ of the total crustal heat production in most regions. Using [Gaschnig et al. \(2016\)](#)'s upper crustal U and Th abundances in low heat flux, cratonic regions, results in roughly 6% (by area) of the continents having a negative heat flux across the Moho (an unreasonable condition) – or more likely, other factors, such as heat dissipation through fluid circulation in the near surface, would be needed to explain these low surface heat flux (e.g., 20-40 mW/m²)

Parameter	Value
Global Surface Heat Flux	inputs from Lucazeau (2019)
Antarctica Surface Heat Flux	inputs from Shen et al. (2020)
Upper Crust Heat Production	$1.75 \mu\text{W}/\text{m}^3$ (Gaschnig et al., 2016)
Upper Crust Heat Production (cratonic)	$0.8 \mu\text{W}/\text{m}^3$ (see discussion for sources)
Deep Crustal Densities	inputs from this study

Table 5.5: Heat production calculation parameters

regions. Alternatively, the assumed upper crustal heat production values may need to be lowered. Most low heat flux areas coincide with stable cratonic lithosphere, where low heat flux and low heat production is not a new observation (e.g., [Cammarano and Guerri \(2017\)](#); [Jaupart et al. \(2007\)](#); [Kukkonen et al. \(1997\)](#); [Nyblade and Pollack \(1993\)](#)). Estimates of upper crustal heat production in cratonic regions range from 0.6 to $1 \mu\text{W}/\text{m}^3$ ([Gruber et al., 2021](#); [Jaupart et al., 2014, 2016](#); [Mareschal and Jaupart, 2013](#); [Phaneuf and Mareschal, 2014](#)), so we approximated Archean upper crustal heat production at $0.8 \mu\text{W}/\text{m}^3$, which is the maximum permissible heat production value found by [Rudnick and Nyblade \(1999\)](#) for the Kalahari craton and also the maximum average crustal heat production expected for crust ≥ 2.5 Ga ([Jaupart et al., 2016](#)). Our final model for Moho heat flux uses the data and resources listed in Table 5.5. The median global continental Moho heat flux, shown in Figure 5.15, is $21.5 \pm 11.6 \text{ mW}/\text{m}^2$. Figure 5.14 summarizes the distribution of Moho heat flux values for each tectonic province.

Table 5.6 compares our average crustal heat production and Moho heat flux values to the average values calculated from other studies. We calculated the Moho heat fluxes from the previous studies by taking the average surface heat flux and subtracting the crust’s average radiogenic contribution. For consistency, we recalculated our study’s Moho heat flux in this simplified manner as well. Total crustal HPE concentrations av-

	Models of the continental crust					
	RG13	HKB11	HKB15 low	HKB15 high	WSM20	This Study
K (ppm)	15000	15800	11900	18800	11700	15400
Th (ppm)	5.60	7.30	4.20	5.31	5.64	5.40
U (ppm)	1.30	1.40	1.09	1.33	1.27	1.40
Heat production K (nW/kg)	0.0508	0.0535	0.0403	0.0637	0.0396	0.0522
Heat production Th (nW/kg)	0.147	0.191	0.110	0.139	0.148	0.141
Heat production U (nW/kg)	0.128	0.138	0.107	0.131	0.125	0.138
Total crustal heat production (nW/kg)	0.326	0.383	0.258	0.334	0.312	0.331
Total crustal heat production (mW/m ³)	0.945	1.111	0.748	0.969	0.905	0.960
Crustal radiogenic heat production (TW)	7.56	8.89	5.99	7.75	7.24	7.68
Upper crust contribution to total heat production	59.4%	50.5%	75.0%	57.9%	62.0%	58.4%
Radiogenic contribution to surface heat flux (mW/m ²)	33.7	39.6	26.7	34.6	32.3	34.2
Average crustal Moho heat flux (mW/m ²)	29.3	23.4	36.3	28.4	30.7	28.9

Data sources: RG13 = (Rudnick and Gao, 2014), HKB11 = (Hacker et al., 2011), HKB15 = (Hacker et al., 2015), WSM20 = (Wipperfurth et al., 2020), Šrámek et al. 13 = (Šrámek et al., 2013). [‡]Moho heat flux is calculated by taking the average surface heat flux for the continents 63 mW/m² (and the oceans 65 mW/m²) (Lucazeau, 2019) and subtracting the continental crust's (and ocean) radiogenic contribution. [†]Upper crust contribution is calculated using the upper crust composition from Rudnick and Gao (2014) relative to the total crust production; note all models assume this same upper crust composition.

Table 5.6: Heat production in the continental crust

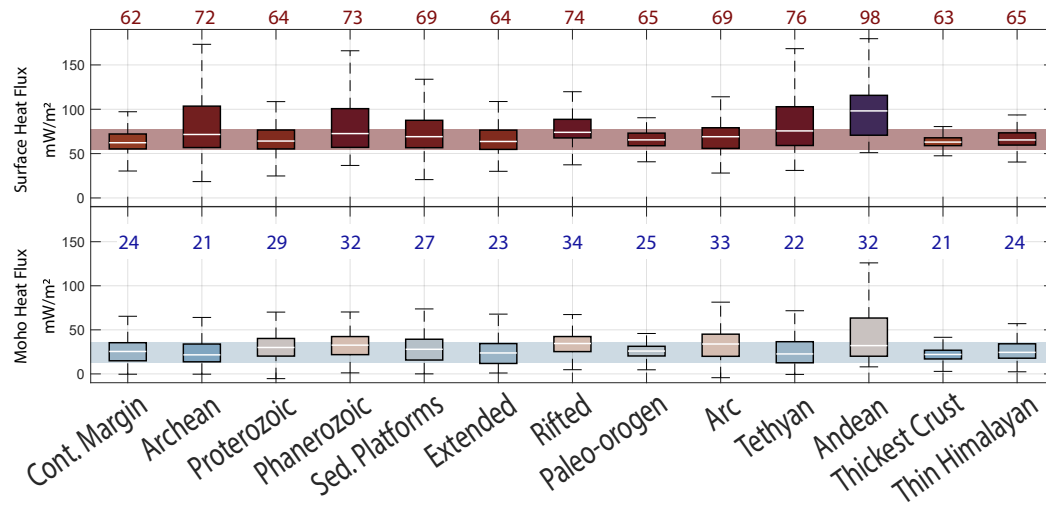


Figure 5.14: The median global Moho heat flux is $21.6 \pm 11.6 \text{ mW/m}^2$, shown by the blue shaded area at the bottom of the graph. Global surface heat flux is $63.8 \pm 12.3 \text{ mW/m}^2$, indicated by the red shaded band towards the top. Although variations exist among the tectonic provinces, they show remarkably consistent central heat flux values. These statistics were produced by calculating the median heat flux for each latitude x longitude cell. Outliers are not shown on the graphs. The white line in the middle of each box and whisker plot is the median value. This value is written above the box and whisker plot in blue for the Moho heat flux and in red for the surface heat flux for each province.

erage upper, middle, and lower crustal abundances. Figure 5.16 compares the different models of U abundance in the bulk continental crust from Table 5.6 to their corresponding Moho heat flux values.

Notably, all models reported in Table 5.6 assume a similar compositional model for the upper crust (Gaschnig et al., 2016; Rudnick and Gao, 2014). A 1% change in upper crustal U has eight times the effect on the Moho heat flux as a 1% change in lower crustal U content. Thus, having an accurate model for the upper crust is of paramount importance to determining Moho heat flux. Only after upper crustal heat production is constrained will the uncertainties on deep crustal heat production impact Moho heat flux calculations.

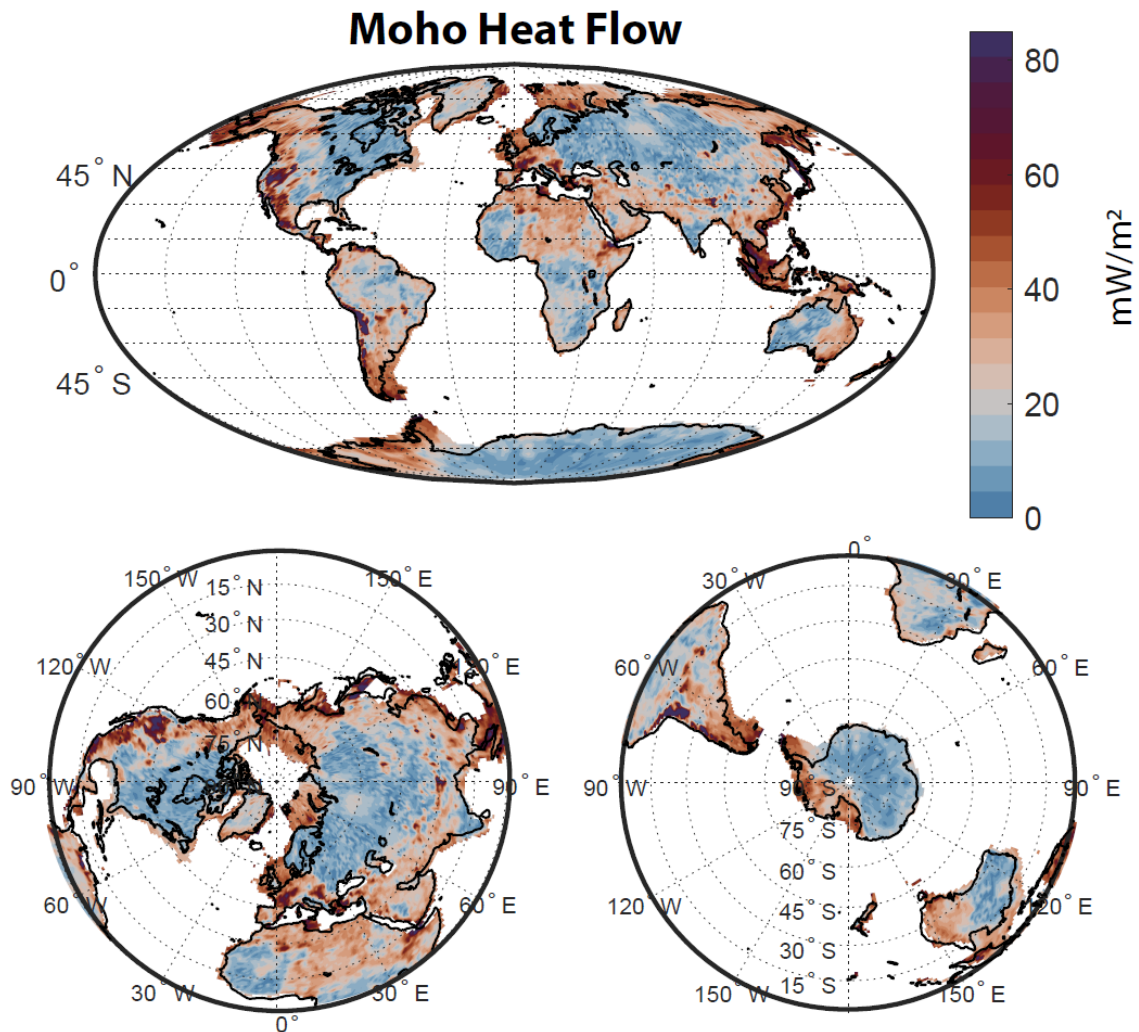


Figure 5.15: Global heat flux across the Moho calculated by subtracting crustal heat production from measurements of surface heat flux. The median subcontinental Moho heat flux is $21.5 \pm 11.6 \text{ mW/m}^2$ globally. This result is heavily sensitive to the assumption of a uniform upper crustal heat production of $0.8 \text{ } \mu\text{W/m}^3$ for cratonic and $1.65 \text{ } \mu\text{W/m}^3$ for non-cratonic regions.

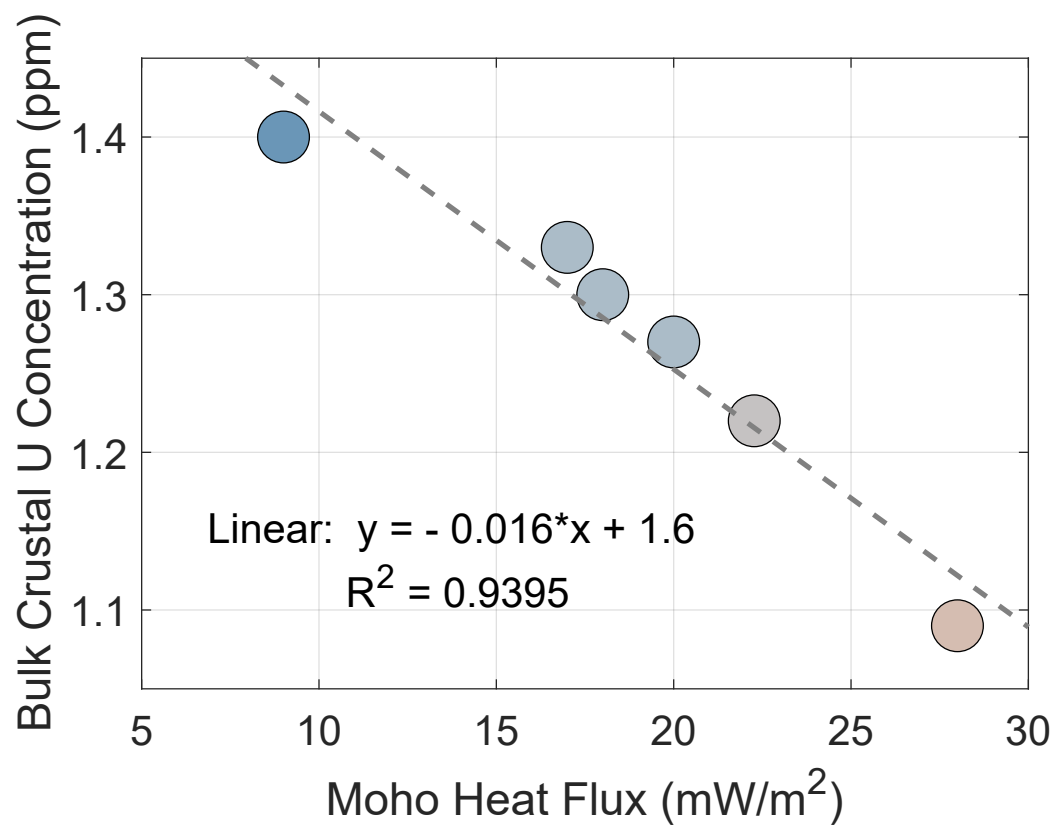


Figure 5.16: There is a linear dependency between crustal U concentration and calculated Moho heat flux. When the different models of Table 5.6 are compared, differences in each models' predicted Moho heat flux are mostly the result of differences in upper crustal HPE concentrations.

5.7 Conclusions

We have constructed a global compositional model of the deep continental crust by synthesizing seismic, temperature, heat flux, and geochemical data. We predict deep crustal compositions on the global scale using major and trace element compositions from amphibolite and granulite facies lithologies, and seismic velocity profiles. Our proposed global compositional model reconciles the USGS Global Seismic Catalog of crustal velocities, compositions for thousands of medium and high grade metamorphic rocks, and constraints on Moho depth ([Pasyanos et al., 2014](#); [Szwillus et al., 2019](#)), crust temperature ([Artemieva, 2006](#)), and surface heat flux ([Lucazeau, 2019](#); [Shen et al., 2020](#)).

V_p values, and to a lesser extent V_s and V_p/V_s values, correlate with bulk rock SiO₂ content, and SiO₂ can be used as a predictor of seismic velocities if temperature can be estimated accurately. Globally, SiO₂ concentrations tend to decrease with increasing depth, leading to a predominantly mafic to intermediate-mafic base of the crust. The decreased density and less mafic nature of the lower crust in younger and tectonically active crust, such as arcs and active mountain ranges, suggests that those tectonic provinces are hotter than our temperature model predicts, that they have undergone lower crustal delamination, or both. Global median SiO₂ concentrations for the middle and lower crust are 61.2 ± 7.31 and 50.5 ± 3.4 wt.%, respectively, though steady composition and velocity gradients in the deep crust urge us to embrace a less distinctly layered view of the crust. This mid-to-deep crustal gradient in wt.% SiO₂ is the equivalent of a lithological gradient ranging from quartz monzonite to gabbro-norite. We predict the abundances of multiple major oxides, many of which are correlated to trace element abundances. This correlation

allows us to derive expected heat production in the deep crust. We also predict a Moho heat flux of $21.5 \pm 11.6 \text{ mW/m}^2$.

Chapter 6: Conclusions and Future Work

6.1 Key conclusions

The chemical composition of Earth's continental crust is the sum of its evolutionary processes throughout time. Radiogenic heat production in Earth's interior is a particularly important feature to understand if we want to decipher Earth's starting materials and for how long our mantle will maintain convection. Once the planet runs out of heat energy in its interior, its core will fully solidify, its protective magnetic field will dissipate followed by a significant loss of atmosphere, and the surface-renewing forces of plate tectonics will cease.

The concentrations of heat producing elements (HPEs: K, U, and Th) and their distribution throughout the crust can be inferred from many different geological, geochemical, and physical methods. Ultimately, using all of these techniques in tandem provides the most clarity, constraint, and quantifiability on the chemical attributes of the crust and mantle.

Geoneutrino detection remains the only method to directly sample the HPE content of Earth's inaccessible mantle, but without accurate knowledge of the geology surrounding geoneutrino detectors, geoneutrino signal measurements alone constrain nothing. Chapter 2 shows the importance of applying an accurate, precise lithospheric HPE

correction to the geoneutrino signal. Changes in U and Th content surrounding a detector balloon into vastly different results for Earth's total radiogenic heat production.

While up to 70% of the geoneutrino signal originates from the uppermost portions of the continental crust, the 30% from the deep crust proves more challenging to constrain. Moreover, the composition of the deeper portion of the continental crust holds crucial information about how Earth's outermost layer forms, deforms, and evolves through time. The chemical makeup of the deep crust controls not only its contribution to the geoneutrino signal (through its HPE content), but its stability and strength overall.

The deep continental crust is unreachable by traditional geological sampling methods. However, we can study medium- and high-grade metamorphic lithologies, which serve as analogues or windows into the higher pressure and temperature regions of the crust (Chapter 3). Analyzing the composition of thousands of amphibolite, granulite, and eclogite facies xenoliths and terrains gives insight into the chemical and petrological behaviors of major element oxides and trace elements. In Chapter 3, we learned that the "Daly Gap" phenomenon is strongly reflected in the compositions of deep crustal analogue samples, leading to an abundance of mafic (<52 wt.% SiO_2) and felsic (>68 wt.% SiO_2) rocks, and few rocks of intermediate composition. Though metamorphic grade does play a role in the composition of these medium- and high-grade metamorphic rocks, the degree of melt evolution/differentiation has an even stronger correlation with trace element abundances. In short, it is just as important to constrain major element composition as it is metamorphic grade.

Because deep crustal lithologies range from mafic to felsic in composition, studying the geochemistry of these rocks alone leaves multiple possibilities for deep crustal

major and trace element content. Similarly, geophysical applications, such as seismology, provides multiple nonunique solutions to crustal composition, since different rock types can have the same seismic velocities. The key to determining composition, both on the regional and global scale, lies in reconciling geochemical trends with geophysical observations. Creating joint geochemical-geophysical models sheds more light on deep crustal composition.

We built a probabilistic model of deep crustal composition by calculating the expected physical properties of thousands of amphibolite and granulite samples at specific pressures and temperatures. We then used a Monte Carlo numerical simulation to find which compositions best reproduced the seismic properties (i.e., V_p , V_s , and V_p/V_s ratios) observed by regional velocity surveys. In Chapter 4, we applied a geochemical-geophysical data and models to a specific region, where pairing our model with observations of surface geology allowed us to infer crustal deformation schemes. We adapted and expanded our joint model by incorporating global seismic datasets and splitting the crust into tectonic provinces in Chapter 5. The resulting code gave us a lower resolution but larger scale map of the middle and lower continental crust (Chapter 5).

In both the local and global models, we observed that the deep crust displays a gradient of decreasing SiO_2 with increasing depth, though the severity of this gradient depends on specific location. In most regions, the deep crust reaches an intermediate composition of 61.2 ± 7.3 wt.% SiO_2 at about half its total thickness. At the base of the continental crust, bulk composition is as mafic as 53.8 ± 3 wt.% SiO_2 .

An increasingly mafic composition in the crust with depth has three significant implications. First, there are no universal, sharp compositional changes within the crust, and

we should therefore move away from thinking of the continental crust as a three-layer ensemble. We have enough technology and data to in most cases to take a more gradual composition change within the crust into consideration. Second, The lowermost parts of the continental crust are, by and large, mafic in nature. This means that relamination of sediments onto the base of the crust is likely not a dominant crustal evolutionary process, though it is currently impossible to differentiate between mafic sediments and mafic restites. Third, the mafic composition of the deep continental crust correlates to lower abundances of HPEs in that portion of the crust. This means that for geoneutrino studies, the uppermost continental crust remains the major focus of chemical modeling, as changes in the HPE-depleted deep crust do not affect the lithospheric correction to the geoneutrino signal as much as the abundance of HPEs in the uppermost crust.

6.2 Future geoneutrino project locations

The under-development Jiangmen Underground Neutrino Observatory (JUNO, 20 kton) experiment in Jiangmen, Guangdong province, China ([An et al., 2016](#)) is currently being built. This detector will be 20 times bigger than the any existing detector. It will begin detecting geoneutrinos in 2023 or 2024. Because of its large size, one year of JUNO data will present us with more detected geoneutrino events than what has been collected by all the previous experiments. The current $S_{lithosphere}$ predictions for the for the JUNO detector differ significantly ([Gao et al., 2020](#); [Strati et al., 2015](#)). A more accurate and precise estimate of the distribution of K, Th, and U in the continental lithosphere surrounding the JUNO detector is needed. Implications for these findings go far beyond

this immediate applications of neutrino geosciences for southeastern China in particular. An understanding of the continental lithosphere in this region will constrain the crustal growth history and architecture of southeastern China, and the nature of its granitic magmatism, geothermal energy, and ore potential.

An under development detector adjacent to the Himalayas, at the China Jinping Underground Laboratory (CJPL), poses unique challenges for crustal modeling. The principle goals of this instrument will be detecting geoneutrinos and solar neutrinos, simultaneously revealing the secrets of the fuel powering both the Earth and the Sun. At 2.4 kilometers beneath Mount Jinping and 1000 km away from the closest commercial nuclear reactor, CJPL is the deepest of the world's underground laboratories, shielding it against unwelcome background signals from reactor antineutrinos and atmospheric muons. Sited on the eastern slope of the Tibetan plateau, Jinping will sense geoneutrinos from the planet's thickest crust. The Sichuan region of southern China is geologically one of the most active regions of the world with deadly earthquakes, some of the fastest moving crust (5 - 20 mm/yr), and nearly unrivaled topographic gradients (10%) between the Sichuan basin and the Tibetan plateau. The composition of the Himalayan crust and Tibetan Plateau remains highly debated, and global velocity models conflict with local assessments, leaving geoscientists with competing hypotheses for the strength and deformation mechanisms of the region's crust ([Jackson, 2002](#)).

Last but not least, teams have proposed to remove the issue of the continental crust altogether by building an ocean bottom detector (OBD) ([Sakai et al., 2021](#)). Because the oceanic crust is thin and depleted in HPEs, 70% of the OBD signal would come directly from mantle geoneutrinos, with minimal crustal corrections needed. A team of scientists

and engineers have proposed working with JAMSTEC to build a portable prototype detector that would be moved to different ocean bottom locales every few years. This project is in its infancy but represents a massive feat in engineering.

Appendix A: Supporting Information for Chapter 2

A.1 Full electron antineutrino flux equation

Table S1 explains the meaning of each symbol and its units.

$$\frac{dN(E_{\bar{\nu}_e}, \vec{r})}{d(E_{\bar{\nu}_e})} = \varepsilon \frac{N_A \lambda}{\mu} \sigma_P(E_{\nu_e}) \frac{dn(E_{\bar{\nu}_e})}{d(E_{\bar{\nu}_e})} \int_{\oplus} P_{ee}(E_{\bar{\nu}_e}, |\vec{r} - \vec{r}'|) d\vec{r}' \frac{a(\vec{r}) \rho(\vec{r}')}{4\pi |\vec{r} - \vec{r}'|^2} \quad (\text{A.1})$$

Symbol	Description	Units
$\frac{dN(E_{\bar{\nu}_e}, \vec{r})}{d(E_{\bar{\nu}_e})}$	$\bar{\nu}_e$ detection spectrum	$\bar{\nu}_e$
ε	10^{32} protons x 3.154×10^7 s x 100%*	$proton \times s$
N_A	Avogadro's number	$\frac{atom}{mol}$
λ	Decay constant	$\frac{decay}{s \times atom}$
μ	Atomic mass	$\frac{kg}{mol}$
$\sigma_P(E_{\nu_e})$	$\bar{\nu}_e$ cross-section (function of $E_{\bar{\nu}_e}$)	$\frac{m^2}{proton}$
$\frac{dn(E_{\bar{\nu}_e})}{d(E_{\bar{\nu}_e})}$	$\bar{\nu}_e$ emission spectrum	$\frac{\bar{\nu}_e}{decay}$
$P_{ee}(E_{\bar{\nu}_e}, \vec{r} - \vec{r}')$	Oscillation probability (function of $E_{\bar{\nu}_e}$)	unitless
$a(\vec{r})$	Concentration of radionuclide in cell	$\frac{kg}{kg}$
$\rho(\vec{r}')$	Density of rock in cell	$\frac{kg}{m^3}$
$ \vec{r} - \vec{r}' ^2$	Distance from cell to detector	m
size and efficiency normalization factor		

*detector

Table A.1: Heat production and geoneutrino flux results

A.2 Heat production from K, Th, and U decay

Radionuclide	Mole Fraction (%)	λ (a ⁻¹)	Q(MeV)	Q(pJ)	
²³² Th	100	4.916 x 10 ⁻¹¹	42.646	6.8326	*Total
²³⁵ U	0.72049	9.8531 x 10 ⁻¹⁰	46.397	7.4336	
²³⁸ U	99.2740	1.5513 x 10 ⁻¹⁰	51.694	8.2823	
⁴⁰ K	0.01167	5.491 x 10 ⁻¹⁰ *	1.331*	2.132*	
from all ⁴⁰ K decay modes					

Table A.2: Radionuclide heat production

Appendix B: Supporting Information for Chapter 3

B.1 Eu Anomalies

All amphibolite, granulite, and eclogite facies lithologies have similar median Eu anomalies ($\frac{Eu}{\sqrt{Sm_N * Gd_N}}$). While granulite facies lithologies tend to show more variation in Eu anomalies (Figure B.1), the median Eu/Eu* values for all of our groups fall between 0.89 and 1.2. The distributions of Eu/Eu*, shown in Figure B.1, are skewed towards the right, moreso for granulite facies than for other facies, favoring Eu/Eu* values around 1 but reaching as high as 2-3. These values of Eu/Eu* < 1 (which translate to negative median Eu anomalies) are in line with previous estimates for the middle and lower crust and could indicate voluminous crustal recycling (Tang et al., 2017). The lack of strong, positive Eu anomalies in the deep crust counters the argument that the deep (or lower) crust is complementary to the upper crust, which also has a negative Eu anomaly (Taylor and McLennan, 1985).

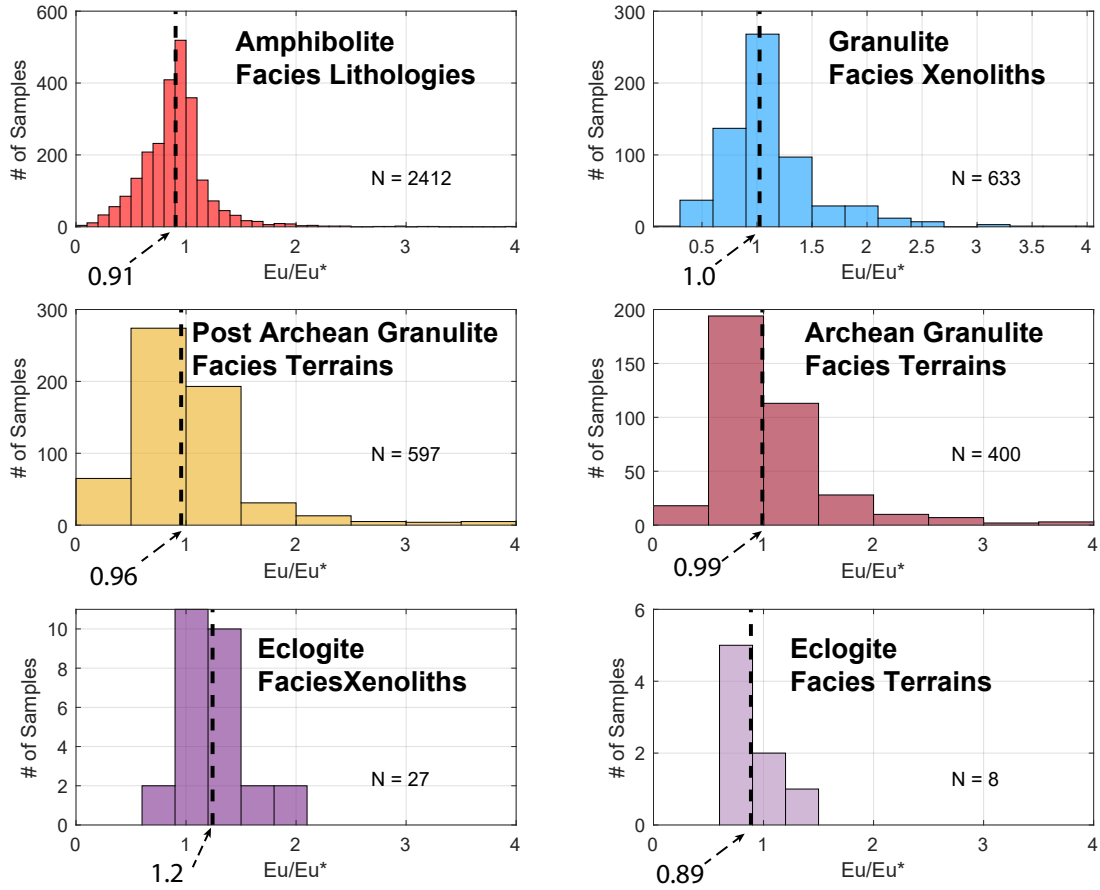


Figure B.1: The right-skewed Eu/Eu^* distributions of our datasets all find a median value close to 1, which is 15 - 30% less than previous estimates. Because of the asymmetry in the distributions, both granulite facies xenoliths and terrains have similar values, though the more felsic Archean granulite facies lithologies have a higher percentage of Eu/Eu^* values >1 than their more mafic Post Archean counterparts.

B.2 Fluid Mobile Elements

B.2.1 K, Rb, and Cs

The transition from amphibolite to granulite facies can be considered, for the major elements, an isochemical dehydration reaction, but this is not the case for the most fluid mobile elements: K, Rb, and Cs. Consistent with the observations of ([Rudnick and Presper, 1990](#)), K_2O/Rb ratios in amphibolite and granulite facies lithologies are negatively correlated with K_2O content, especially at $K_2O < 1.2$ wt.%, indicating Rb depletion relative to K_2O . We restrict our analysis of K_2O/Rb to compositions to rock with > 55 wt.% SiO_2 because the K_2O/Rb ratios of unmetamorphosed basalts are highly variable, making it difficult to evaluate Rb depletion in mafic lithologies ([Rudnick and Presper, 1990](#)). K_2O/Rb ratios reach a maximum at about 1 in most of the metamorphic datasets. Although we omit them, mafic lithologies follow the same trend, though lower K_2O values are reached. As suggested by various studies, this trend likely reflects a mineralogical control on the partitioning of Rb and K between minerals and a fluid phase ([Fowler, 1986](#); [Rudnick et al., 1985](#)). Low concentrations of K_2O and Rb, and high K_2O/Rb ratios, may reflect igneous processes rather than metamorphic processes ([Van Calsteren et al., 1986](#)). However, several mafic granulite facies and amphibolite facies lithologies have high K/Rb ratios compared to basalts, which suggests that they have experienced some Rb depletion during metamorphism ([Stosch et al., 1986](#)). The similarities between the amphibolite and granulite facies K/Rb are striking, because high K_2O/Rb was thought to be the providence of dehydration metamorphism. Partial dehydration of amphibolite facies litholo-

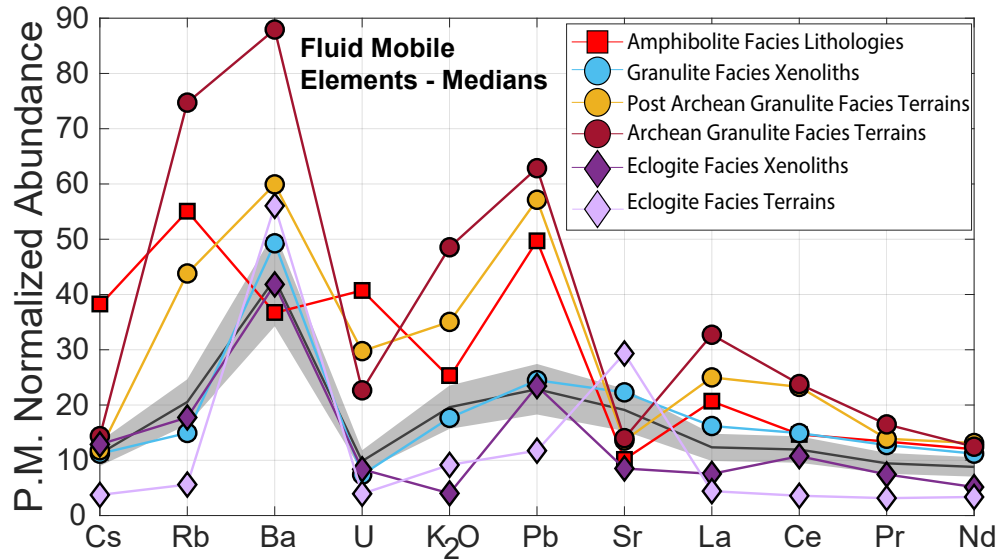


Figure B.2: Fluid mobile element concentrations from amphibolite and granulite facies lithologies are comparable with previous estimates for the lower crust. Eclogite facies lithologies are up a half an order of magnitude more depleted. Variable labile element depletions cause the amphibolite facies lithologies to overlap with the granulite facies terrains, suggesting that either the amphibolite facies samples have experienced some fluid depletion or that the behavior of the labile elements is caused by an igneous process and not metamorphic dehydration. The black line and gray shaded region surrounding it is the [Rudnick and Gao \(2014\)](#) lower crustal composition $\pm 15\%$.

gies might lead to high K_2O/Rb ratios while still leaving behind enough hydrous mineral assemblages to classify the rock as amphibolite. Future studies should explore the quantification of dehydration as it relates to K_2O/Rb ratios. Alternatively, metamorphosing gabbro to the amphibolite facies can produce uneven increases in K and Rb ([Field and Elliott, 1974](#)), leading to lower K_2O/Rb ratios. Low K_2O/Rb ratios in amphibolites could also be inherited from the retrograde metamorphism of granulite with low K_2O/Rb ([Field and Clough, 1976](#)).

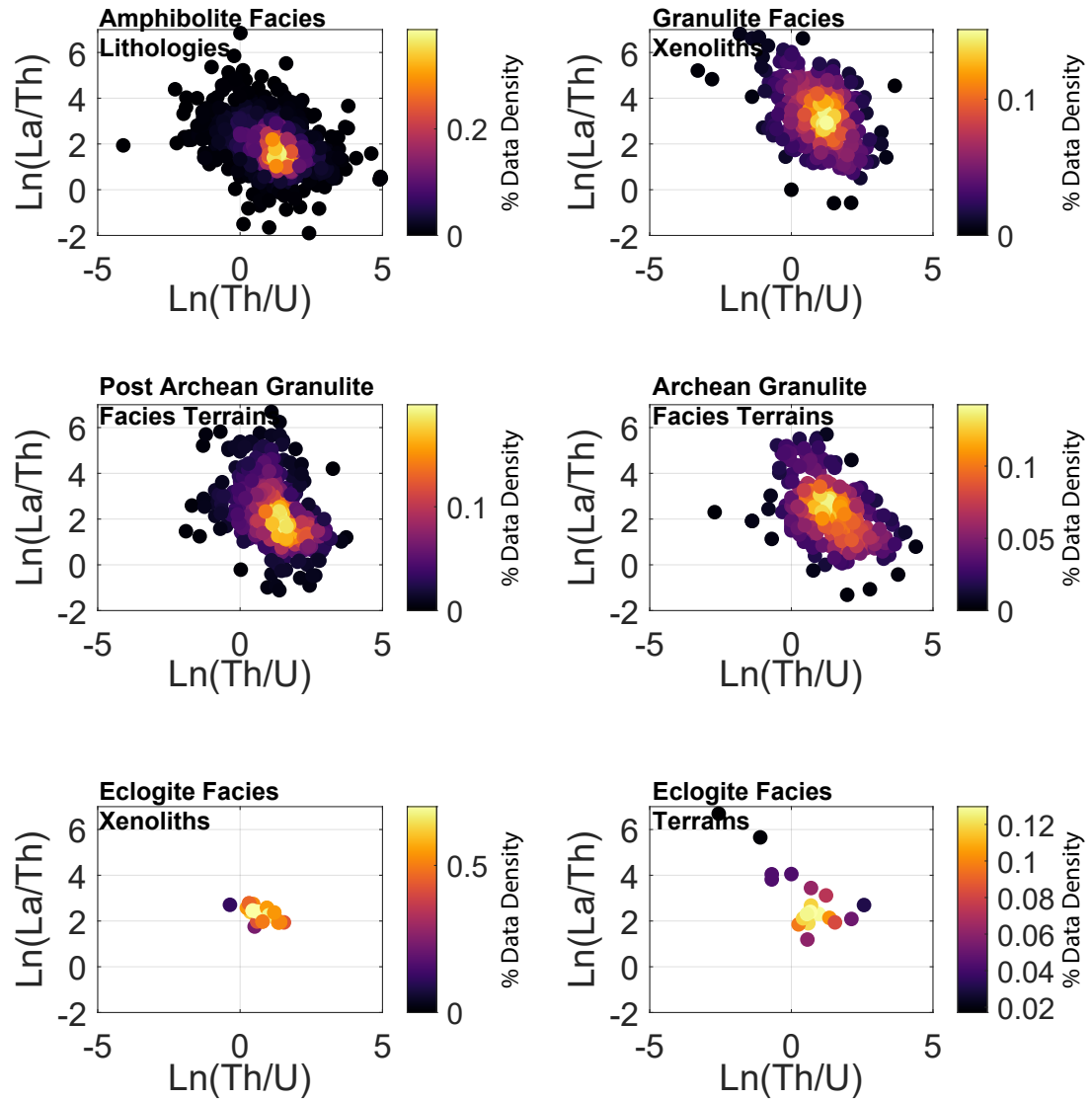


Figure B.3: La/Th vs. Th/U in natural log space. Color indicates the data point density. High La/Th values correlate with low Th/U concentrations, indicating that low La/Th is due to depletions in Th instead of enrichments in La.

B.2.2 La/Th and Th/U

Granulite facies lithologies and eclogite facies terrains show greater Th and U depletion than amphibolite facies lithologies or eclogite facies xenoliths. While the granulite and amphibolite facies lithologies show comparable amounts of U loss, indicated by elevated Th/U and low La/Th values, the granulite facies lithologies can reach almost an order of magnitude greater La/Th values with low Th/U, indicating a loss of both U and Th. U can be depleted through loss of water due to its redox sensitivity (U^{6+} is fluid mobile; U^{4+} is not), though both U and Th depletions typically indicate a lack of U and Th-bearing accessory phases (Fowler, 1986; Rudnick et al., 1985). Archean granulite facies terrains reach higher levels of U depletion than Post-Archean terrains, a curious observation since many Archean terrains contain monazite, a Th and U-bearing accessory phase (Rudnick and Presper, 1990). Eclogite facies lithologies on the whole have Th/U ratios around 2 and a wider range (0.1 to 20) in La/Th.

B.2.3 Alkalis and Alkaline Earth Metals, in General

The alkali and alkaline earth metals are, generally, large ion lithophiles due to their low charges and high ionic radii. The abundance of these fluid mobile elements becomes more variable as ionic radius increases, with Rb and Cs having the greatest variability and greatest fluid mobility. Samples that lack K-feldspar typically display depletions in K, Rb, and Cs while Na and Ca remain rather remarkably invariable outside of their felsic and mafic modes. Ba is also quite variable with relative uncertainties ranging from 100 to 200% (similar to Rb, Cs, and La, though Ba is more normally distributed than those

three). Many of the amphibolite and granulite facies lithologies are also metaluminous, though our samples span the range of peraluminous to peralkaline.

As alluded to in Section [B.2.3](#), the alkali metals show considerable variability. K, Rb, and Cs increase in skewness as their ionic radius increases, with Cs being the largest and most labile. Rb is also incompatible and fluid mobile. We suspect much of the variation in K stems from the presence or absence of K-feldspar. Lithium also shows a moderate to high level of variability in amphibolite and granulite facies, which could be the result of Li's fluid mobility. Ba and Sr, also incompatible and fluid mobile elements, vary more than their smaller radii counterparts, Mg and Ca, but not nearly as much as the alkalis.

B.3 High Field-Strength Elements

Eclogite facies lithologies display the most fractionation of the generally insoluble high field strength elements (HFSEs). Of the HFSEs, Ta and Mo display the largest spread and highest uncertainties for most of our lithologies. There is not enough data to plot Mo or Ta for eclogite facies lithologies. Once more Mo data are reported, paired spikes in both Mo and Nb content in eclogites could indicate that Mo and Nb, which have similar ionic radii, are both being retained in rutile ([Rudnick et al., 2000](#)). With 64 data points, granulite facies xenoliths' elevated Mo is more of a conundrum. High abundances of rutile are not expected for granulite facies xenoliths ([Rudnick and Presper, 1990](#)), and the xenoliths do not have a complementary Nb spike. Since titanite (sphene), ilmenite, and magnetite typically host the greatest abundance of Mo in crustal rocks ([Greaney et al.,](#)

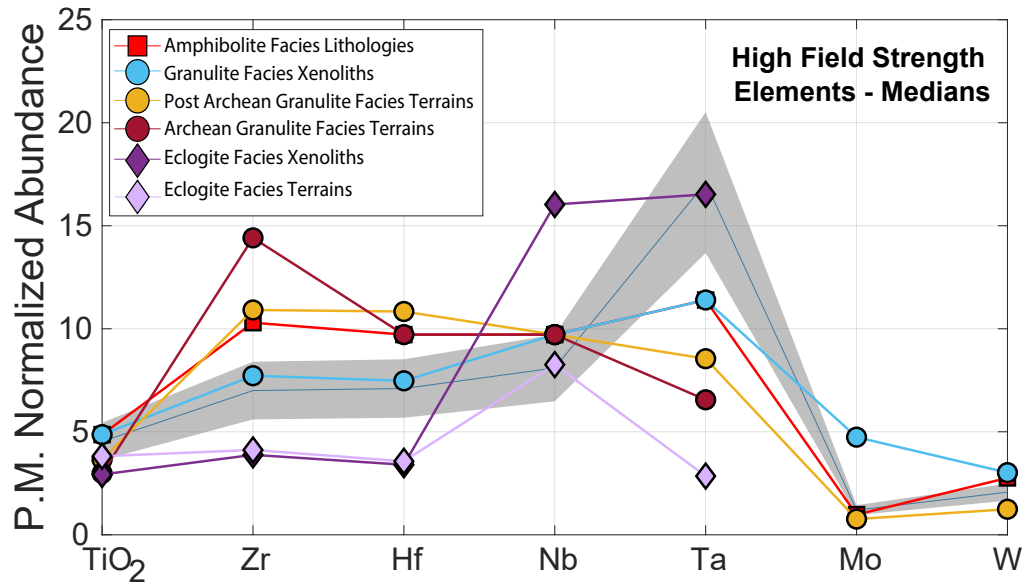


Figure B.4: As with the fluid mobile elements, the high field strength element concentrations are broadly in agreement with previous estimates. Granulite facies xenoliths and eclogite facies lithologies show elevated concentrations of Mo, partly due to lack of data but also potentially because of the abundance of rutile, titanite, ilmenite, and magnetite in these predominantly mafic samples. Eclogite facies lithologies have negative Ta spikes, which are also potentially caused by the presence of rutile. The black line and gray shaded region surrounding it is the [Rudnick and Gao \(2014\)](#) lower crustal composition $\pm 15\%$.

[2018](#)), along with rutile, it is possible that these minerals exist in higher abundances in granulite facies xenoliths and/or the deep crust than previously thought.

The data also show variable depletions in Ta compared to ([Rudnick and Gao, 2014](#))’s value. Nb/Ta values also exceed previous lower crustal estimates ([Rudnick and Gao, 2014](#)), mimicking more the middle crust’s ratio, suggesting that the deep crust might have a super-chondritic Nb/Ta ratio. Plotting Nb vs. Ta (Figure [B.7](#)) yields a linear log-log plot for amphibolite, granulite, and eclogite facies lithologies. In many cases, we do not have high resolution measurements for Ta concentrations at the sub-ppm level, evidenced by the diagonal lines of data on the Nb vs. Nb/Ta plots (ratios dominated solely by changing Nb) and horizontal array of dots on the Ta vs. Nb/Ta plots. Concentrations which have

been rounded to the nearest 100 ppb or ppm when reported. Eclogite facies xenoliths and terrains have notably different Nb/Ta ratios: 15.6 ± 1.9 and 22.4 ± 10.2 , respectively. Studies have hypothesized that Nb can be sequestered by rutile in eclogite formation, but Ta is not expected to increase ([Rudnick et al., 2000](#)). Eclogitic material in down-going crustal slabs could be a complementary reservoir to the apparent Nb depletion in the rest of the observable bulk silicate Earth.

B.4 Transition Metals

Transition metal data exists in more abundance for amphibolite facies lithologies than for granulite or eclogite. Trends in the first row transition elements are discussed in depth in Section 6 of the main text, but we wanted to briefly mention them in context with the other elements here. The dearth of data on moderately and highly siderophile elements is because they exist in such low abundances in the silicate portion of Earth that only recently have we been able to obtain high precision measurements.

Ti, Zr, and Hf have similar variability, with more mafic lithologies having lower absolute concentrations of Zr. Eclogite facies lithologies have the lowest concentrations of all three elements, a curious finding considering that eclogite facies lithologies tend to have an abundance of rutile (FeTiO_2). Though they track similarly, what controls the abundance of Ti family elements is unclear.

The Cr family, on the other hand, shows much greater variability, perhaps due to Cr concentrating in garnets, spinels, chromite, or even olivine. Olivine fractionation in the source melt could also explain the variations also seen in Ni ([Arndt, 1994](#)) among

amphibolite and granulite facies lithologies. However, Mo's slightly chalcophile behavior ([Barnes, 2016](#)) could be lending the Cr-family a more discontinuous and variable look than, for example, the Sc, Ti, or Mn families. The Ni and Cu display more variable abundances as well, especially in amphibolite facies lithologies. Their behavior could be attributed to S-affinity and/or the presence of Ni- or Cu-bearing accessory mineral phases (e.g., olivine, diopside). Weathering has been shown to ([Arndt, 1994](#); [Polat et al., 2002](#)), making it more likely that the spread of Cr and Ni abundances is controlled by igneous factors.

A stepped increase in variability is visible from the top to the bottom of the V family for both amphibolite and granulite facies lithologies. This trend, notable in other parts of the periodic table is discussed in Section 6 of the main text. Of course, limited sample numbers may be causing the fluctuation in third-row transition metal abundances. More data is always welcomed to better constrain the abundances and behaviors of the transition metals.

Not much can be said about the elements towards the bottom and right of the periodic table for granulite facies lithologies due to a lack of data, though we can devise a few hypotheses to explain the increased spread in abundances of heavier elements compared to their lighter counterparts. Especially in heavy elements that do not lack for data, such as Ta, Pb, Th, and U, the continuous but wide distribution of data could be caused by the vertical fractionation of these elements within the continental crust. Vertical fractionation of elements within the crust has been discussed in depth for decades ([Rudnick and Gao, 2014](#); [Sighinolfi, 1971](#); [Taylor and McLennan, 1985](#)). One means for testing this hypothesis is comparing trace element abundance to inferred maximum pressure conditions -

the higher the pressure, the deeper from within the crust the sample originated. Sadly, most of our data does not have precise pressures and/or temperatures associated with it, though some studies have explored these composition-pressure relationships with quantitative methods ([Barnhart et al., 2012](#); [Bohlen and Mezger, 1989](#); [Rudnick and Jackson, 1995](#)).

Cr, Ni, and Cu break the homogeneity of the first row transition elements. Cr and the two other elements it shares a family with, Mo and W, are more variable than any other transition metals, though we must consider that Mo and W have fewer analyses than many of the others (89 and 42 data points, respectively, compared to the 1000's of analyses done on first row transition elements and Y, Zr, Nb, Hf, and Ta). The Cr family takes on an abnormal valence electron structure ($4s^1 3d^5$). This causes Cr to have a range of valence states, though it commonly exists in a 3+ valence state as a substitution in garnet (uvarovite, $\text{Ca}_3\text{Cr}_2(\text{SiO}_4)_3$). Mo and W are high field strength elements with high valences and large radii, which make it harder for them to substitute into common minerals and may contribute to their variability. Mo is typically hosted in Ti-oxides though displays much more variation than Ti.

The Ni and Cu families might also be special cases, not because of their valence structure, but because of their metal and sulfide affinity. The variability in S content may cause fluctuations chalcophile-siderophile elements (CSEs), including Ni and Cu. Of the other CSE's, Zn, a moderately chalcophile element ([Jenner, 2017](#)), is much more normally distributed than either Ni or Cu. There are too few measurements on Au and Ag to determine whether their apparent variation is caused by CSE ([Jenner, 2017](#)) behavior or if they are under-sampled. The Cu and Ni distributions have far more outliers than any

other first row transition metals. Cross-referencing the original data sources show that > 1100 amphibolite and > 900 granulite measurements come from samples with 0.01 - 0.1 wt.% Cu or Ni. The spread in Cu and Ni compositions could be due to cumulate formation and recycling processes, whereby CSE's can be partitioned by magnetite-induced sulfide fractionation in lower crustal settings. While we expect enrichment in CSE's in these lower crustal cumulates, delamination of such material could also cause depletion and contribute to the bulk crust's low Cu concentrations.

B.5 Halogens

Limited data is available for the halogens in these medium and high grade metamorphic lithology datasets. F and Cl are the only elements measured for amphibolite and granulite facies lithologies. No halide data is available for eclogite facies lithologies in our dataset. Halides can be transported in fluids or incorporated into select mineral phases, such as Cl-rich biotite and amphibole, apatite, and fluorite ([Kusebauch et al., 2015](#)). High solubility in water means these elements are often mobilized by fluid fluxing and can often be found as chemical weathering and alteration byproducts in granulites, despite granulites being nominally anhydrous ([Banks et al., 2000](#); [Kusebauch et al., 2015](#); [Markl et al., 1997](#)). Stable Cl isotopes have been studied in granulites to track mantle participation in silicate Earth's Cl cycle and fluid infiltration processes ([Markl et al., 1997](#)). However, the volatility of halogens and their incompatibility in many melt systems has resulted in relatively low halide abundances in the deep crust outside of fluid processes.

B.6 Supplementary Tables

Table B.1: Amphibolite Facies Lithologies

	Mean	Median	Geometric γ Mean		STD	IQR	Geo STD	γ STD	N (filtered)	N (original)
			Mean							
Li	17.9	15.0	14.8	17.9	10.7	14.5	1.89	10.6	455	727
Be	1.57	1.44	1.21	1.57	1.02	1.10	2.22	1.08	212	374
B	14.0	9.00	8.44	14.0	14.5	15.6	2.81	13.1	184	362
N	-	-	-	-	-	-	-	-	0	0
F	435	399	60	435	351	512	48.2	746	180	269
S	41.2	22.0	25.2	41.2	46.1	44.0	2.6	38.2	93	292
Cl	51.0	29.3	1.2	51.0	62.2	86.7	83.3	114	81	196
Sc	23.1	21.0	18.3	23.1	13.6	24.4	2.11	15.2	1360	3160
V	152	134	104	152	108	192	2.71	125	1840	3690
Cr	119	81.0	63.5	119	121	151	3.57	124	1820	4200
Co	30.1	29.9	22.2	30.1	19.1	33.1	2.42	22.4	1420	3480
Ni	51.8	39.7	31.2	51.8	47.1	62.0	3.08	49.0	1810	4290
Cu	41.3	30.0	25.8	41.3	36.4	49.5	2.89	37.6	1290	2970
Zn	80.7	78.0	73.0	80.7	34.6	49.5	1.60	35.7	1550	3310
Ga	18.1	18.0	17.7	18.1	3.66	4.32	1.23	3.69	1120	2320
Ge	-	-	-	-	-	-	-	-	0	0.0
As	2.9	1.3	1.4	2.9	4.2	2.4	3.1	3.2	200	368
Se	77.5	53.0	57.1	77.5	63.9	69.0	2.2	57.8	136	158

Br	-	-	-	-	-	-	-	-	0	0
Rb	59.1	43.8	35.1	59.1	51.8	80.0	3.16	56.5	1910	4030
Sr	245	201	201	245	151	204	1.91	149	2100	4730
Y	24.3	22.5	21.9	24.3	11.0	15.0	1.60	11.0	1970	4460
Zr	133	123	115	133	69.5	97.0	1.77	70.9	2070	4650
Nb	8.61	7.20	6.90	8.61	5.47	7.93	2.02	5.55	1790	3790
Mo	1.29	0.520	0.670	1.29	1.56	1.48	3.10	1.37	208	354
Ru	-	-	-	-	-	-	-	-	0	0
Rh	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0	0
Pd	1.71	0.85	0.89	1.71	2.20	1.70	3.09	1.81	118	218
Ag	47.7	48.0	34.2	47.7	22.1	25.0	3.4	37.1	150	200
Cd	66.2	60.0	49.1	66.2	36.4	58.0	3.0	49.0	146	192
In	0.0712	0.0712	0.0712	-	0.0	0.0	1.0	-	1	28
Sn	3.08	1.60	1.85	3.08	3.58	2.75	2.67	2.90	215	377
Sb	0.30	0.20	0.21	0.30	0.27	0.32	2.38	0.25	253	497
Te	0.08	0.05	0.05	0.08	0.11	0.05	2.76	0.08	8	14
I	-	-	-	-	-	-	-	-	0	0
Cs	1.8	1.2	1.1	1.8	1.8	2.0	2.9	1.7	752	1680
Ba	399	330	268	399	306	486	2.7	337	2030	4400
La	22.3	18.1	15.8	22.3	16.7	25.5	2.5	17.5	1820	3920
Ce	44.4	36.5	33.1	44.4	31.8	46.4	2.3	32.7	1790	3870
Pr	5.6	4.7	4.3	5.6	3.8	5.4	2.1	3.9	1110	2120
Nd	21.6	18.3	17.7	21.6	13.5	18.0	1.9	13.3	1710	3500
Sm	4.6	4.1	4.0	4.6	2.4	3.0	1.7	2.3	1630	3470

Eu	1.1	1.1	1.0	1.1	0.5	0.6	1.5	0.5	1640	3470
Gd	4.4	3.9	3.9	4.4	2.1	2.7	1.6	2.1	1390	2700
Tb	0.7	0.7	0.7	0.7	0.3	0.4	1.6	0.3	1250	2860
Dy	4.3	3.9	3.8	4.3	2.0	2.8	1.6	2.0	1150	2330
Ho	0.9	0.8	0.8	0.9	0.4	0.6	1.6	0.4	1050	2100
Er	2.5	2.3	2.2	2.5	1.2	1.7	1.7	1.2	1150	2250
Tm	0.4	0.4	0.3	0.4	0.2	0.2	1.6	0.2	821	1960
Yb	2.4	2.2	2.1	2.4	1.1	1.5	1.7	1.1	1740	3740
Lu	0.4	0.3	0.3	0.4	0.2	0.2	1.7	0.2	1550	3240
Hf	3.9	3.4	3.3	3.9	2.3	3.1	1.9	2.3	1320	2830
Ta	0.7	0.5	0.5	0.7	0.5	0.7	2.3	0.5	1120	2540
W	1.0	0.4	0.5	1.0	2.0	0.8	2.9	1.1	182	454
Re	-	-	-	-	-	-	-	-	0	6
Os	-	-	-	-	-	-	-	-	0	6
Ir	0.2	0.2	0.2	0.2	0.1	0.1	1.8	0.1	6	19
Pt	1.7	0.7	0.9	1.7	2.0	1.9	3.1	1.8	120	217
Au	2.0	0.8	1.0	2.0	2.7	1.9	3.0	2.2	160	271
Hg	0.0	0.0	0.0	0.0	0.0	0.0	2.2	0.0	4	12
Tl	0.6	0.5	0.3	0.6	0.5	0.8	3.9	0.6	20	60
Pb	13.2	11.7	10.5	13.2	8.3	11.5	2.1	8.6	1260	2490
Bi	0.1	0.1	0.1	0.1	0.1	0.1	2.2	0.1	143	200
Th	5.6	3.7	3.0	5.6	5.4	8.2	3.6	5.9	1600	3330
U	1.4	1.0	0.9	1.4	1.1	1.5	2.6	1.1	1390	2700

Table B.2: Granulite Facies Xenoliths

	Mean	Median	Geometric γ Mean		STD	IQR	Geo STD	γ STD	N (filtered)	N (original)
			Mean							
Li	11.8	6.9	9.2	11.8	10.5	6.5	1.9	8.0	12	154
Be	0.5	0.5	0.5	0.5	0.0	0.0	1.1	0.0	2	20
B	-	-	-	-	-	-	-	-	0	0
N	-	-	-	-	-	-	-	-	0	0
F	-	-	-	-	-	-	-	-	0	35
S	143	140	89.8	143.0	110.8	220.6	2.85	129.8	4	67
Cl	151.0	151.0	151.0	-	0.0	0.0	1.0	-	1	28
Sc	28.0	28.0	26.5	28.0	9.0	11.9	1.4	9.1	97	863
V	186.0	186.0	166.0	186.0	79.4	100.0	1.7	88.1	92	1030
Cr	210.0	168.0	165.0	210.0	145.0	162.0	2.1	140.0	120	1140
Co	48.9	46.8	46.5	48.9	15.1	23.5	1.4	15.5	87	759
Ni	103.0	100.0	91.6	103.0	46.7	65.0	1.7	48.3	100	1080
Cu	43.4	37.8	36.0	43.4	27.4	33.1	1.9	25.9	72	774
Zn	77.6	81.1	73.0	77.6	25.2	43.3	1.4	26.9	72	751
Ga	17.1	17.3	16.8	17.1	3.1	4.6	1.2	3.2	48	391
Ge	-	-	-	-	-	-	-	-	0	0
As	-	-	-	-	-	-	-	-	0	0
Se	-	-	-	-	-	-	-	-	0	0
Br	-	-	-	-	-	-	-	-	0	0
Rb	15.5	10.6	10.0	15.5	13.0	20.2	2.8	13.6	112	1180

Sr	482.0	465.0	437.0	482.0	207.0	315.0	1.6	210.0	121	1280
Y	19.9	19.0	17.3	19.9	9.9	16.6	1.8	10.5	102	1030
Zr	91.3	83.3	70.5	91.3	58.9	88.7	2.2	63.1	105	1060
Nb	8.8	7.0	7.0	8.8	5.8	7.7	2.0	5.8	99	888
Mo	2.3	1.9	2.1	2.3	1.2	0.8	1.6	1.1	6	62
Ru	-	-	-	-	-	-	-	-	0	0
Rh	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0	0
Pd	5.54	5.54	5.54	-	0.00	0.00	1.00	-	1	5
Ag	-	-	-	-	-	-	-	-	0	0
Cd	-	-	-	-	-	-	-	-	0	0
In	-	-	-	-	-	-	-	-	0	0
Sn	1.70	1.58	1.62	1.70	0.52	0.96	1.39	0.54	9	77
Sb	-	-	-	-	-	-	-	-	0	39
Te	-	-	-	-	-	-	-	-	0	0
I	-	-	-	-	-	-	-	-	0	0
Cs	0.752	0.390	0.316	0.752	0.960	0.871	4.280	0.901	44	340
Ba	470	393	358	470	330	428	2	334	107	1120
La	14.1	11.6	11.5	14.1	8.6	12.9	2.0	8.7	118	1080
Ce	30.0	27.0	25.0	30.0	17.2	27.4	1.9	17.6	119	1090
Pr	3.6	3.5	3.0	3.6	2.1	3.3	2.0	2.2	63	535
Nd	15.8	14.7	13.7	15.8	8.0	12.3	1.8	8.4	115	1080
Sm	3.8	3.6	3.4	3.8	1.7	2.3	1.6	1.7	114	1070
Eu	1.2	1.3	1.2	1.2	0.4	0.6	1.4	0.4	108	983
Gd	3.8	3.8	3.4	3.8	1.7	2.4	1.7	1.8	79	705

Tb	0.6	0.6	0.5	0.6	0.3	0.3	1.6	0.3	93	778
Dy	3.7	3.6	3.2	3.7	1.8	3.0	1.8	2.0	73	656
Ho	0.7	0.6	0.6	0.7	0.3	0.6	1.7	0.3	63	538
Er	2.0	1.9	1.7	2.0	1.1	1.9	1.9	1.2	71	645
Tm	0.3	0.3	0.2	0.3	0.2	0.3	2.0	0.2	41	344
Yb	1.9	1.7	1.6	1.9	1.0	1.6	1.7	1.0	109	973
Lu	0.3	0.2	0.2	0.3	0.2	0.3	1.7	0.1	99	815
Hf	2.5	2.1	2.0	2.5	1.6	1.3	1.8	1.4	82	680
Ta	0.8	0.6	0.6	0.8	0.5	0.7	2.1	0.5	62	444
W	2.4	1.9	1.8	2.4	1.7	2.6	2.2	1.7	4	30
Re	-	-	-	-	-	-	-	-	0	1
Os	0.2	0.2	0.2	-	0.0	0.0	1.0	-	1	4
Ir	0.3	0.3	0.3	-	0.0	0.0	1.0	-	1	5
Pt	2.5	2.5	2.5	-	0.0	0.0	1.0	-	1	6
Au	-	-	-	-	-	-	-	-	0	3
Hg	-	-	-	-	-	-	-	-	0	0
Tl	-	-	-	-	-	-	-	-	0	0
Pb	5.6	4.5	4.6	5.6	3.6	4.9	2.0	3.5	77	690
Bi	-	-	-	-	-	-	-	-	0	0
Th	1.8	0.8	0.9	1.8	2.4	1.5	3.0	1.9	94	750
U	0.4	0.2	0.3	0.4	0.5	0.3	2.5	0.4	77	604

Table B.3: Post-Archean Granulite Terrains

	Mean	Median	Geometric γ Mean		STD	IQR	Geo STD	γ STD	N (filtered)	N (original)
			Mean							
Li	7.9	7.2	7.3	7.9	3.2	4.7	1.5	3.1	6	74
Be	2.0	1.6	1.9	2.0	0.7	1.1	1.4	0.6	3	36
B	-	-	-	-	-	-	-	-	0	0
N	-	-	-	-	-	-	-	-	0	0
F	-	-	-	-	-	-	-	-	3	9
S	300.0	300.0	300.0	-	-	-	-	-	1	10
Cl	100.0	100.0	100.0	-	0.0	0.0	1.0	-	2	6
Sc	25.4	25.6	22.7	25.4	10.5	14.3	1.7	11.9	50	635
V	163.0	157.0	145.0	163.0	68.9	90.9	1.7	78.9	54	1030
Cr	198.0	132.0	126.0	198.0	209.0	183.0	2.6	176.0	81	1130
Co	32.3	31.8	27.8	32.3	16.5	21.6	1.8	17.2	41	677
Ni	80.4	59.0	55.7	80.4	78.4	64.8	2.3	65.5	82	1110
Cu	34.2	29.0	28.2	34.2	20.4	27.7	1.9	20.6	45	748
Zn	81.1	77.7	76.4	81.1	29.4	37.6	1.4	27.7	67	949
Ga	18.6	18.0	18.4	18.6	2.8	3.9	1.2	2.8	38	774
Ge	-	-	-	-	-	-	-	-	0	0
As	-	-	-	-	-	-	-	-	0	0
Se	-	-	-	-	-	-	-	-	0	0
Br	-	-	-	-	-	-	-	-	0	0
Rb	65.2	59.0	42.6	65.2	48.1	79.0	3.0	56.9	127	1340

Sr	280.0	267.0	240.0	280.0	144.0	190.0	1.8	152.0	134	1420
Y	28.7	24.2	24.9	28.7	15.8	17.3	1.7	14.9	67	1170
Zr	166.0	136.0	127.0	166.0	114.0	147.0	2.2	118.0	90	1330
Nb	9.9	9.5	7.7	9.9	6.2	7.9	2.3	6.9	63	948
Mo	2.4	3.2	1.8	2.4	1.3	2.1	2.3	1.7	3	37
Ru	-	-	-	-	-	-	-	-	0	0
Rh	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0	0
Pd	-	-	-	-	-	-	-	-	0	0
Ag	-	-	-	-	-	-	-	-	0	0
Cd	-	-	-	-	-	-	-	-	0	0
In	-	-	-	-	-	-	-	-	0	0
Sn	3.6	2.9	3.0	3.6	2.6	2.7	1.8	2.3	5	223
Sb	0.1	0.1	0.0	0.1	0.1	0.1	2.9	0.1	6	60
Te	-	-	-	-	-	-	-	-	0	0
I	-	-	-	-	-	-	-	-	0	0
Cs	1.7	0.7	0.6	1.7	2.0	2.5	6.0	2.3	24	530
Ba	470.0	461.0	360.0	470.0	280.0	447.0	2.4	330.0	83	1260
La	26.8	24.5	20.9	26.8	16.7	24.4	2.2	18.3	63	1050
Ce	55.6	50.5	45.0	55.6	33.3	42.5	2.0	35.0	62	1060
Pr	6.0	6.1	4.3	6.0	3.9	5.4	2.7	4.7	27	588
Nd	25.9	24.8	21.6	25.9	15.3	17.1	1.9	15.1	57	878
Sm	5.2	4.9	4.5	5.2	2.8	3.0	1.8	2.8	55	842
Eu	1.3	1.4	1.2	1.3	0.4	0.7	1.4	0.4	53	836
Gd	4.9	4.4	4.3	4.9	2.6	2.3	1.7	2.4	42	683

Tb	0.9	0.7	0.8	0.9	0.5	0.5	1.8	0.5	44	759
Dy	4.3	3.7	3.9	4.3	2.0	2.4	1.6	1.9	32	634
Ho	0.9	0.8	0.8	0.9	0.4	0.6	1.6	0.4	31	607
Er	2.5	2.1	2.3	2.5	1.2	1.3	1.6	1.2	33	642
Tm	0.5	0.4	0.4	0.5	0.2	0.4	1.6	0.2	13	67
Yb	2.5	2.0	2.2	2.5	1.3	1.5	1.7	1.3	61	862
Lu	0.4	0.4	0.3	0.4	0.2	0.2	1.7	0.2	48	802
Hf	4.1	4.1	3.3	4.1	2.3	3.0	2.0	2.5	41	666
Ta	0.7	0.5	0.5	0.7	0.5	0.7	2.4	0.5	29	578
W	-	-	-	-	-	-	-	-	0	29
Re	-	-	-	-	-	-	-	-	0	0
Os	-	-	-	-	-	-	-	-	0	0
Ir	-	-	-	-	-	-	-	-	0	0
Pt	-	-	-	-	-	-	-	-	0	0
Au	-	-	-	-	-	-	-	-	0	0
Hg	-	-	-	-	-	-	-	-	0	0
Tl	-	-	-	-	-	-	-	-	0	18
Pb	13.2	11.7	10.4	13.2	8.2	12.1	2.1	8.7	35	728
Bi	-	-	-	-	-	-	-	-	0	14
Th	9.9	5.4	4.2	9.9	10.2	16.2	4.9	11.7	90	816
U	1.0	0.8	0.7	1.0	0.9	1.0	2.8	0.9	83	732

Units: Pd,W, Re, Os, Ir, Pt, and Au are in ppb. All other elements are in ppm.

Table B.4: Archean Granulite Terrains

	Mean	Median	Geometric Mean	γ Mean	STD	IQR	Geo STD	γ STD	N (filtered)	N (original)
Li	15.3	16.3	14.4	15.3	4.6	7.6	1.4	5.0	4	84
Be	-	-	-	-	-	-	-	-	0	0
B	-	-	-	-	-	-	-	-	0	0
N	-	-	-	-	-	-	-	-	0	0
F	-	-	-	-	-	-	-	-	0	1
S	1772	1772.0	698	1772	2303	3260	4.9	2190	2	5
Cl	90.0	90.0	90.0	-	0.0	0.0	1.0	-	1	6
Sc	16.6	16.1	13.1	16.6	9.8	13.9	2.2	11.0	33	468
V	119.0	115.0	102.0	119.0	62.0	81.0	1.8	64.0	48	1000
Cr	146.0	117.0	104.0	146.0	108.0	170.0	2.4	114.0	67	1060
Co	41.3	35.7	33.6	41.3	25.4	38.1	1.9	25.7	35	604
Ni	72.4	55.0	54.9	72.4	52.1	76.7	2.2	51.7	69	1090
Cu	28.5	24.7	22.4	28.5	17.5	30.3	2.1	18.9	32	736
Zn	66.0	65.9	61.9	66.0	22.0	32.6	1.5	23.6	40	875
Ga	20.2	19.5	19.9	20.2	3.7	4.9	1.2	3.6	27	703
Ge	-	-	-	-	-	-	-	-	0	0
As	30.0	30.0	30.0	-	0.0	0.0	1.0	-	1	11
Se	-	-	-	-	-	-	-	-	0	6
Br	-	-	-	-	-	-	-	-	0	0
Rb	44.2	37.0	23.9	44.2	39.9	63.2	3.7	45.5	89	1340

Sr	307.0	295.0	237.0	307.0	185.0	333.0	2.2	212.0	87	1360
Y	18.4	17.8	15.8	18.4	9.4	13.0	1.8	9.9	70	1130
Zr	155.0	137.0	135.0	155.0	86.4	75.7	1.7	79.8	84	1270
Nb	9.2	7.3	7.5	9.2	6.6	5.7	1.9	5.7	66	973
Mo	-	-	-	-	-	-	-	-	0	2
Ru	-	-	-	-	-	-	-	-	0	0
Rh	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0	0
Pd	-	-	-	-	-	-	-	-	0	3
Ag	-	-	-	-	-	-	-	-	0	0
Cd	-	-	-	-	-	-	-	-	0	0
In	-	-	-	-	-	-	-	-	0	0
Sn	2.6	2.5	2.5	2.6	0.7	1.2	1.4	0.8	23	95
Sb	-	-	-	-	-	-	-	-	0	6
Te	-	-	-	-	-	-	-	-	0	0
I	-	-	-	-	-	-	-	-	0	0
Cs	1.8	0.9	0.8	1.8	2.2	2.4	4.2	2.1	38	229
Ba	561.0	559.0	459.0	561.0	316.0	425.0	2.0	346.0	69	1220
La	27.3	20.0	21.8	27.3	20.6	24.5	1.9	17.6	66	993
Ce	49.8	40.0	41.6	49.8	33.8	34.5	1.8	28.9	65	987
Pr	5.0	4.4	4.4	5.0	2.6	2.7	1.7	2.4	39	376
Nd	18.1	16.4	15.8	18.1	9.2	8.6	1.7	9.3	64	594
Sm	3.6	3.3	3.3	3.6	1.5	2.3	1.5	1.5	55	661
Eu	1.1	1.1	1.1	1.1	0.3	0.5	1.4	0.33	53	641
Gd	3.5	3.3	3.2	3.5	1.5	2.2	1.6	1.56	45	455

Tb	0.5	0.5	0.5	0.5	0.2	0.3	1.6	0.22	50	571
Dy	3.3	3.0	3.0	3.3	1.2	1.9	1.5	1.28	39	481
Ho	0.6	0.6	0.6	0.6	0.3	0.4	1.7	0.30	40	417
Er	1.8	1.7	1.5	1.8	0.8	0.9	1.8	0.93	42	484
Tm	0.3	0.3	0.3	0.3	0.1	0.1	1.2	0.05	9	86
Yb	1.6	1.6	1.4	1.6	0.8	1.3	1.9	0.91	53	630
Lu	0.3	0.3	0.2	0.3	0.1	0.1	1.8	0.13	35	524
Hf	3.2	3.1	3.0	3.2	0.9	1.4	1.4	0.95	44	377
Ta	0.7	0.4	0.4	0.7	0.9	0.3	2.5	0.61	26	258
W	6.7	6.7	6.7	-	0.0	0.0	1.0	-	1	7
Re	-	-	-	-	-	-	-	-	0	0
Os	-	-	-	-	-	-	-	-	0	0
Ir	-	-	-	-	-	-	-	-	0	7
Pt	-	-	-	-	-	-	-	-	0	3
Au	-	-	-	-	-	-	-	-	0	6
Hg	-	-	-	-	-	-	-	-	0	0
Tl	-	-	-	-	-	-	-	-	0	0
Pb	10.0	8.0	8.8	10.0	5.4	5.4	1.6	4.77	61	758
Bi	-	-	-	-	-	-	-	-	0	0
Th	7.4	2.7	2.9	7.4	14.6	5.2	3.7	9.21	73	772
U	1.2	0.5	0.6	1.2	1.7	1.4	3.3	1.28	66	562

Units: Pd,W, Re, Os, Ir, Pt, and Au are in ppb. All other elements are in ppm.

Table B.5: Eclogite Facies Xenoliths

	Mean	Median	Geometric Mean	γ Mean	STD	IQR	Geo STD	γ STD	N (filtered)	N (original)
Li	28.7	28.7	28.7	-	0.0	0.0	1.0	-	1	8
Be	-	-	-	-	-	-	-	-	0	0
B	-	-	-	-	-	-	-	-	0	0
N	-	-	-	-	-	-	-	-	0	0
F	-	-	-	-	-	-	-	-	0	4
S	-	-	-	-	-	-	-	-	0	0
Cl	-	-	-	-	-	-	-	-	0	4
Sc	56.2	57.0	55.6	56.2	7.5	13.8	1.15	7.6	3	33
V	320	274	314	320	62.4	120	1.21	61.0	5	60
Cr	626	764	552	626	278	554	1.70	308	5	60
Co	62.7	65.2	61.6	62.7	11.4	20.5	1.21	11.9	4	35
Ni	182	157	174	182	53.5	110	1.34	53.2	5	60
Cu	55.2	55.2	55.1	55.2	4.13	8.26	1.08	4.1	2	31
Zn	72.3	68.9	70.9	72.3	14.3	25.5	1.21	14.0	4	58
Ga	11.4	11.4	11.4	-	0.0	0.0	1.00	-	1	46
Ge	-	-	-	-	-	-	-	-	0	0
As	-	-	-	-	-	-	-	-	0	0
Se	-	-	-	-	-	-	-	-	0	0
Br	-	-	-	-	-	-	-	-	0	0
Rb	7.4	7.1	5.4	7.4	4.9	11.0	2.4	5.6	5	41

Sr	165.0	178.0	157.0	165.0	53.5	72.0	1.4	52.0	7	67
Y	19.6	20.7	19.1	19.6	4.1	7.2	1.2	4.2	7	66
Zr	40.1	41.1	34.4	40.1	20.6	29.3	1.8	21.7	6	65
Nb	10.4	6.3	5.1	10.4	9.9	19.2	3.9	11.3	6	58
Mo	-	-	-	-	-	-	-	-	0	2
Ru	-	-	-	-	-	-	-	-	0	0
Rh	-	-	-	-	-	-	-	-	0	0
Pd	-	-	-	-	-	-	-	-	0	18
Ag	-	-	-	-	-	-	-	-	0	0
Cd	-	-	-	-	-	-	-	-	0	0
In	-	-	-	-	-	-	-	-	0	0
Sn	-	-	-	-	-	-	-	-	0	2
Sb	-	-	-	-	-	-	-	-	0	2
Te	-	-	-	-	-	-	-	-	0	0
I	-	-	-	-	-	-	-	-	0	0
Cs	-	-	-	-	-	-	-	-	0	6
Ba	370.0	239.0	242.0	370.0	350.0	382.0	2.6	323.0	7	67
La	5.6	6.0	4.3	5.6	3.5	6.0	2.2	3.9	7	62
Ce	16.5	13.9	12.3	16.5	10.9	20.7	2.3	12.3	7	62
Pr	2.2	1.9	1.9	2.2	1.3	1.9	1.9	1.3	4	30
Nd	9.7	7.9	7.9	9.7	6.2	9.5	1.9	6.0	6	37
Sm	2.6	2.3	2.5	2.6	1.1	1.0	1.4	1.0	5	34
Eu	1.0	0.8	1.0	1.0	0.2	0.4	1.3	0.2	5	34
Gd	2.8	2.4	2.7	2.8	0.8	1.1	1.3	0.8	4	30

Tb	0.4	0.4	0.4	0.4	0.1	0.1	1.2	0.1	5	34
Dy	3.0	3.0	2.9	3.0	0.5	0.9	1.2	0.5	4	30
Ho	0.6	0.6	0.6	0.6	0.1	0.2	1.2	0.1	4	30
Er	1.6	1.7	1.6	1.6	0.2	0.3	1.2	0.2	4	30
Tm	0.3	0.3	0.3	0.3	0.0	0.0	1.1	0.0	2	27
Yb	2.1	1.9	2.0	2.1	0.8	1.3	1.5	0.8	5	34
Lu	0.3	0.3	0.3	0.3	0.1	0.1	1.3	0.1	4	32
Hf	1.5	1.4	1.3	1.5	0.7	1.2	1.6	0.7	4	31
Ta	0.6	0.7	0.6	0.6	0.1	0.1	1.1	0.1	3	29
W	-	-	-	-	-	-	-	-	0	0
Re	-	-	-	-	-	-	-	-	0	0
Os	-	-	-	-	-	-	-	-	0	0
Ir	-	-	-	-	-	-	-	-	0	18
Pt	-	-	-	-	-	-	-	-	0	18
Au	-	-	-	-	-	-	-	-	0	18
Hg	-	-	-	-	-	-	-	-	0	0
Tl	-	-	-	-	-	-	-	-	0	0
Pb	2.9	2.9	2.3	2.9	1.7	3.3	2.0	1.8	4	38
Bi	-	-	-	-	-	-	-	-	0	0
Th	0.4	0.3	0.4	0.4	0.2	0.3	1.7	0.2	5	55
U	0.2	0.2	0.2	0.2	0.2	0.3	1.9	0.1	4	54

Units: Pd,W, Re, Os, Ir, Pt, and Au are in ppb. All other elements are in ppm.

Table B.6: Eclogite Facies Terrains

	Mean	Median	Geometric Mean	γ Mean	STD	IQR	Geo STD	γ STD	N (filtered)	N (original)
Li	-	-	-	-	-	-	-	-	0	2
Be	-	-	-	-	-	-	-	-	0	0
B	-	-	-	-	-	-	-	-	0	0
N	-	-	-	-	-	-	-	-	0	0
F	-	-	-	-	-	-	-	-	0	2
S	-	-	-	-	-	-	-	-	0	0
Cl	-	-	-	-	-	-	-	-	0	1
Sc	37.9	30.3	31.0	37.9	23.2	41.4	1.9	23.4	3	33
V	239.0	244.0	224.0	239.0	75.4	86.5	1.5	84.3	5	14
Cr	167.0	187.0	140.0	167.0	76.6	127.0	2.0	98.2	8	39
Co	38.0	38.1	36.2	38.0	11.5	13.2	1.4	11.7	7	38
Ni	83.7	86.0	77.9	83.7	27.0	35.8	1.5	31.1	8	37
Cu	26.8	17.8	21.7	26.8	18.6	26.5	1.9	17.0	6	13
Zn	108.0	105.0	106.0	108.0	19.8	36.5	1.2	19.7	4	30
Ga	20.3	19.0	19.7	20.3	5.0	9.0	1.3	4.9	3	9
Ge	-	-	-	-	-	-	-	-	0	0
As	-	-	-	-	-	-	-	-	0	23
Se	-	-	-	-	-	-	-	-	0	0
Br	-	-	-	-	-	-	-	-	0	0
Rb	7.6	4.4	5.5	7.6	6.0	10.2	2.2	5.8	3	35

Sr	207.0	183.0	165.0	207.0	118.0	208.0	2.1	134.0	9	45
Y	27.6	26.2	24.8	27.6	12.0	22.0	1.6	12.3	6	13
Zr	89.2	107.0	81.4	89.2	33.1	60.3	1.6	37.6	7	36
Nb	6.5	5.4	6.1	6.5	2.5	4.3	1.4	2.4	3	7
Mo	-	-	-	-	-	-	-	-	0	2
Ru	-	-	-	-	-	-	-	-	0	0
Rh	-	-	-	-	-	-	-	-	0	0
Pd	-	-	-	-	-	-	-	-	0	0
Ag	-	-	-	-	-	-	-	-	0	0
Cd	-	-	-	-	-	-	-	-	0	0
In	-	-	-	-	-	-	-	-	0	0
Sn	-	-	-	-	-	-	-	-	0	2
Sb	-	-	-	-	-	-	-	-	0	25
Te	-	-	-	-	-	-	-	-	0	0
I	-	-	-	-	-	-	-	-	0	0
Cs	0.1	0.1	0.1	-	0.0	0.0	1.0	-	1	26
Ba	212.0	135.0	136.0	212.0	188.0	319.0	2.7	189.0	9	39
La	7.9	6.4	6.3	7.9	5.3	9.4	2.0	5.2	4	35
Ce	15.6	14.8	12.9	15.6	8.8	16.9	1.9	9.3	4	35
Pr	0.8	0.8	0.8	-	0.0	0.0	1.0	-	1	4
Nd	6.7	5.8	6.3	6.7	2.5	3.5	1.4	2.3	4	36
Sm	2.5	2.0	2.4	2.5	1.0	1.4	1.4	0.9	4	36
Eu	1.0	1.0	1.0	1.0	0.1	0.1	1.1	0.1	3	34
Gd	5.0	5.0	5.0	5.0	0.2	0.5	1.1	0.2	2	8

Tb	0.7	0.8	0.7	0.7	0.2	0.4	1.4	0.2	3	34
Dy	6.3	6.3	6.3	-	0.0	0.0	1.0	-	1	4
Ho	1.3	1.3	1.3	1.3	0.1	0.1	1.0	0.1	2	8
Er	3.8	3.8	3.8	3.8	0.2	0.3	1.0	0.2	2	8
Tm	0.5	0.5	0.5	-	0.0	0.0	1.0	-	1	6
Yb	3.4	3.8	3.0	3.4	1.5	2.5	1.7	1.6	6	40
Lu	0.2	0.2	0.2	-	0.0	0.0	1.0	-	1	31
Hf	1.8	1.8	1.8	-	0.0	0.0	1.0	-	1	32
Ta	0.1	0.1	0.1	-	0.0	0.0	1.0	-	1	29
W	-	-	-	-	-	-	-	-	0	0
Re	-	-	-	-	-	-	-	-	0	0
Os	-	-	-	-	-	-	-	-	0	0
Ir	-	-	-	-	-	-	-	-	0	0
Pt	-	-	-	-	-	-	-	-	0	0
Au	-	-	-	-	-	-	-	-	0	0
Hg	-	-	-	-	-	-	-	-	0	0
Tl	-	-	-	-	-	-	-	-	0	0
Pb	2.3	2.3	2.2	2.3	0.7	1.4	1.4	0.7	2	8
Bi	-	-	-	-	-	-	-	-	0	0
Th	0.2	0.2	0.2	0.2	0.1	0.3	2.1	0.1	2	32
U	0.1	0.1	0.1	-	0.0	0.0	1.0	-	1	32

Units: Pd,W, Re, Os, Ir, Pt, and Au are in ppb. All other elements are in ppm.

	Mean	Median	Geo Mean	γ Mean	STD	IQR	Geo STD	γ STD	N(filtered)	N(original)
La/Yb	11.6	8.16	7.15	11.6	10.9	13.40	2.88	10.7	1740	3740
Nb/Ta	14.4	14.1	12.6	14.4	5.99	6.33	1.99	7.37	719	1360
Zr/Hf	35.5	35.6	35.0	35.5	6.08	7.70	1.20	6.27	1320	2820
Rb/Cs	48.5	34.0	31.4	48.5	45.4	41.90	2.80	42.7	753	1580
K/Rb	331	294	299	331	160	183.00	1.56	147	1900	3960
La/Th	6.62	5.02	5.30	6.62	4.75	5.42	1.93	4.27	1590	3200
Th/U	4.20	3.65	3.35	4.20	2.67	3.10	2.08	2.73	1380	2600
K/U	15900	12100	11300	15900	13000	14200.00	2.42	12600	1380	2600
Eu/Eu*	0.859	0.884	0.826	0.859	0.229	0.32	1.34	0.241	1372	2674
κPb	3.89	3.86	3.84	3.89	0.646	0.48	1.17	0.621	-	165
Heat Prod. (nW/kg)	0.470	0.278	0.240	0.470	1.99	0.46	3.16	0.503	490	1467
Heat Prod. (mW/m ³)	1.36	0.81	0.70	1.36						

Table B.7: Amphibolite Facies Lithologies

	Mean	Median	Geo Mean	γ Mean	STD	IQR	Geo STD	γ STD	N(filtered)	N(original)
La/Yb	8.71	7.52	7.29	8.71	5.33	6.37	1.83	5.05	103	957
Nb/Ta	18.0	18.6	15.3	18	7.54	9.1	2.14	9.98	41	424
Zr/Hf	39.9	36.3	36.4	39.9	17.8	11.5	1.56	16.9	62	669
Rb/Cs	73.4	40.6	37.2	73.4	84.1	52	3.83	79	39	317
K/Rb	788	643	653	788	491	685	1.86	470	104	1140
La/Th	27.0	17.1	18.0	27.0	25.0	25.8	2.51	23.0	89	683
Th/U	3.79	3.39	3.05	3.79	2.47	2.67	2.03	2.42	67	571
K/U	46300	36900	32300	46300	36100	43000	2.66	37400	72	571
Eu/Eu*	1.28	1.09	1.21	1.28	0.479	0.38	1.38	0.425	79	703
κ_{Pb}	4.17	4.04	4.14	4.17	0.520	0.24	1.11	0.464	-	357
Heat Prod.	0.147	0.0716	0.0861	0.147	0.187	0.11	2.68	0.142	62	79
Heat Prod. (mW/m ³)	0.43	0.21	0.25	0.43						

Table B.8: Granulite Facies Xenoliths

	Mean	Median	Geo Mean	γ Mean	STD	IQR	Geo STD	γ STD	N(filtered)	N(original)
La/Yb	11.7	10.1	8.54	11.7	7.75	8.96	2.77	8.84	41	658
Nb/Ta	14.7	13.9	9.46	14.7	10.4	8.57	4.00	13.0	22	416
Zr/Hf	35.9	35.7	35.1	35.9	7.2	11	1.24	7.46	29	479
Rb/Cs	100	44.6	44.8	100	159	81.4	3.50	116	18	386
K/Rb	423	291	346	423	335	265	1.78	259	109	1090
La/Th	23.7	11.2	12.3	23.7	30.2	18	3.05	25.0	36	777
Th/U	10.1	6.58	6.50	10.1	9.91	10	2.61	8.90	73	564
K/U	37500	31000	27500	37500	32900	32200	2.22	28300	73	564
Eu/Eu*	1.02	1.03	0.94	1.02	0.411	0.46	1.48	0.395	40	673
κ_{Pb}	3.99	4.00	3.99	3.99	0.159	0.19	1.04	0.158	-	33
Heat Prod.	0.514	0.407	0.267	0.514	0.490	0.62	4.41	0.544	70	91
Heat Prod. (mW/m ³)	1.49	1.18	0.77	1.49						

Table B.9: Post Archean Granulite Facies Terrains

	Mean	Median	Geo Mean	γ Mean	STD	IQR	Geo STD	γ STD	N(filtered)	N(original)
La/Yb	25.4	16	18.2	25.4	20.3	31.1	2.33	19.8	45	572
Nb/Ta	17.5	16.3	16.4	17.5	6.24	6.92	1.45	6.26	16	218
Zr/Hf	35.4	35.8	34.7	35.4	7.21	9.54	1.22	7.00	39	341
Rb/Cs	86.0	61.0	44.7	86.0	94.0	93.7	3.61	90.9	30	196
K/Rb	640	433	506	640	477	515	1.95	424	81	1250
La/Th	13.4	7.32	7.8	13.4	16.1	14.2	2.88	13.1	47	728
Th/U	9.17	7.26	6.58	9.17	7.00	9.37	2.39	7.13	60	527
K/U	38100	29300	22900	38100	31900	44900	3.30	36100	60	527
Eu/Eu*	1.16	1.09	1.11	1.16	0.331	0.46	1.34	0.330	44	451
κ_{Pb}	4.16	4.11	4.16	4.16	0.164	0.23	1.03	0.141	-	4
Heat Prod.	0.558	0.179	0.207	0.558	1.04	0.36	3.73	0.708	54	70
Heat Prod. (mW/m^3)	1.62	0.52	0.60	1.62						

Table B.10: Archean Granulite Facies Terrains

	Mean	Median	Geo Mean	γ Mean	STD	IQR	Geo STD	γ STD	N(filtered)	N(original)
La/Yb	6.53	4.69	4.87	6.53	5.16	7.37	2.14	4.8	7	34
Nb/Ta	16.70	17.5	7.89	16.7	10.1	16	6.43	18.8	5	29
Zr/Hf	30.20	31.7	28.4	30.2	9.24	13.5	1.47	10.6	5	31
Rb/Cs	26.60	26.6	18.5	26.6	19.1	38.3	2.48	21.6	2	6
K/Rb	542	503	458	542	287	476	1.85	306	7	41
La/Th	9.22	9.56	8.77	9.22	2.66	4.61	1.4	2.91	6	31
Th/U	2.18	2.17	2.03	2.18	0.798	1.47	1.45	0.798	5	29
K/U	16100	5190	7040	16100	17400	25600	4.29	18800	5	29
Eu/Eu*	1.01	0.968	1.01	1.01	0.095	0.14	1.09	0.0926	4	30
κ_{Pb}	5.48	5.75	5.44	5.48	0.668	1.01	1.13	0.668	-	21
Heat Prod.	0.0792	0.0433	0.0585	0.0792	0.0771	0.07	2.08	0.0590	3	5
Heat Prod. (mW/m ³)	0.23	0.13	0.17	0.23						

Table B.11: Eclogite Facies Xenoliths

	Mean	Median	Geo Mean	γ Mean	STD	IQR	Geo STD	γ STD	N(filtered)	N(original)
La/Yb	7.46	5.48	3.96	7.46	6.9	10.4	3.69	7.77	5	40
Nb/Ta	29.40	29.4	29.4	-	0	0	1	-	1	25
Zr/Hf	27.20	27.2	20.6	27.2	17.7	35.4	2.18	19.4	2	32
Rb/Cs	53.30	53.3	52.7	53.3	7.52	15	1.15	7.55	2	23
K/Rb	693	526	564	693	498	435	1.84	431	5	35
La/Th	60.50	60.5	31.7	60.5	51.5	103	3.53	63.6	2	19
Th/U	2.73	2.56	2.7	2.73	0.424	0.746	1.16	0.415	3	26
K/U	68000	44300	24700	68000	65600	117000	6.04	87100	3	26
Eu/Eu*	0.875	0.875	0.875	0.875	0.0182	0.036	1.02	0.0182	2	8
K^P_b	-	-	-	-	-	-	-	-	-	0
Heat Prod.	0.0647	0.0437	0.0539	0.0647	0.0489	0.0681	1.81	0.0380	1	3
Heat Prod. (mW/m ³)	0.19	0.13	0.16	0.19						

Table B.12: Eclogite Facies Terrains

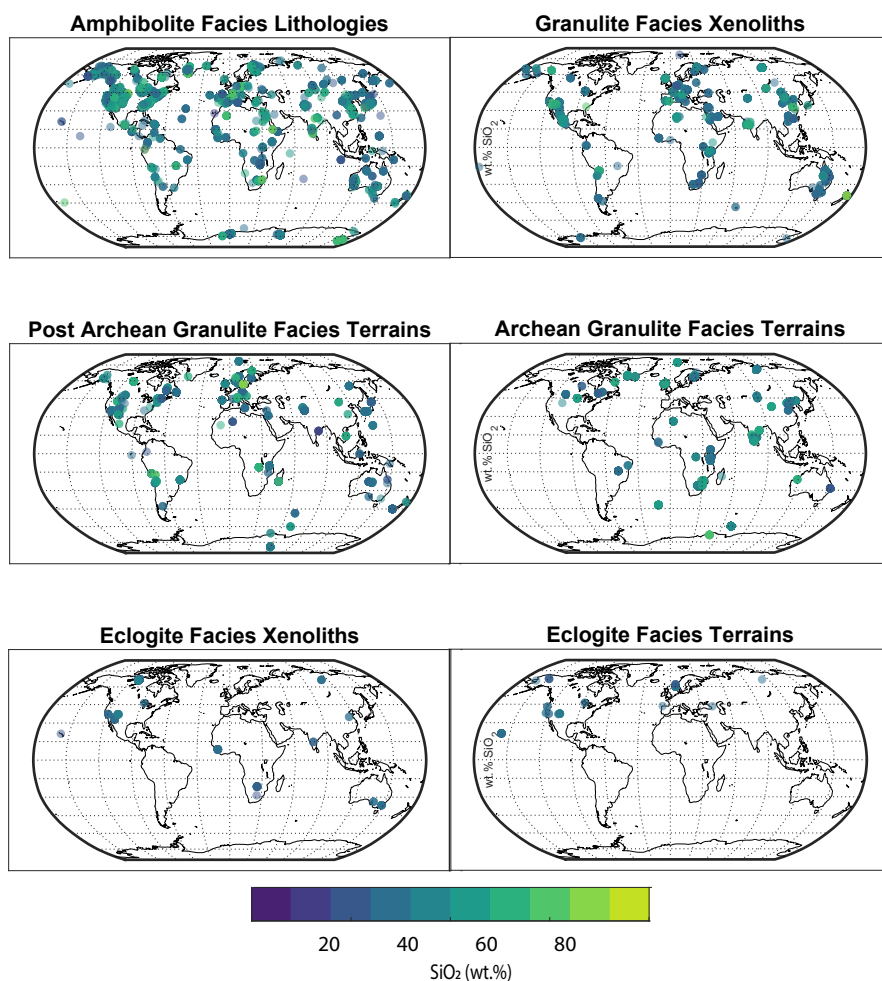


Figure B.5: Maps of the metamorphic sample datasets that we have extensive sample coverage for amphibolite and granulite facies lithologies with more limited exposures of eclogite facies lithologies. Each circle is a single sample whose color is determined by weight percent SiO₂. Many circles overlap. There is little correlation between SiO₂ and location.

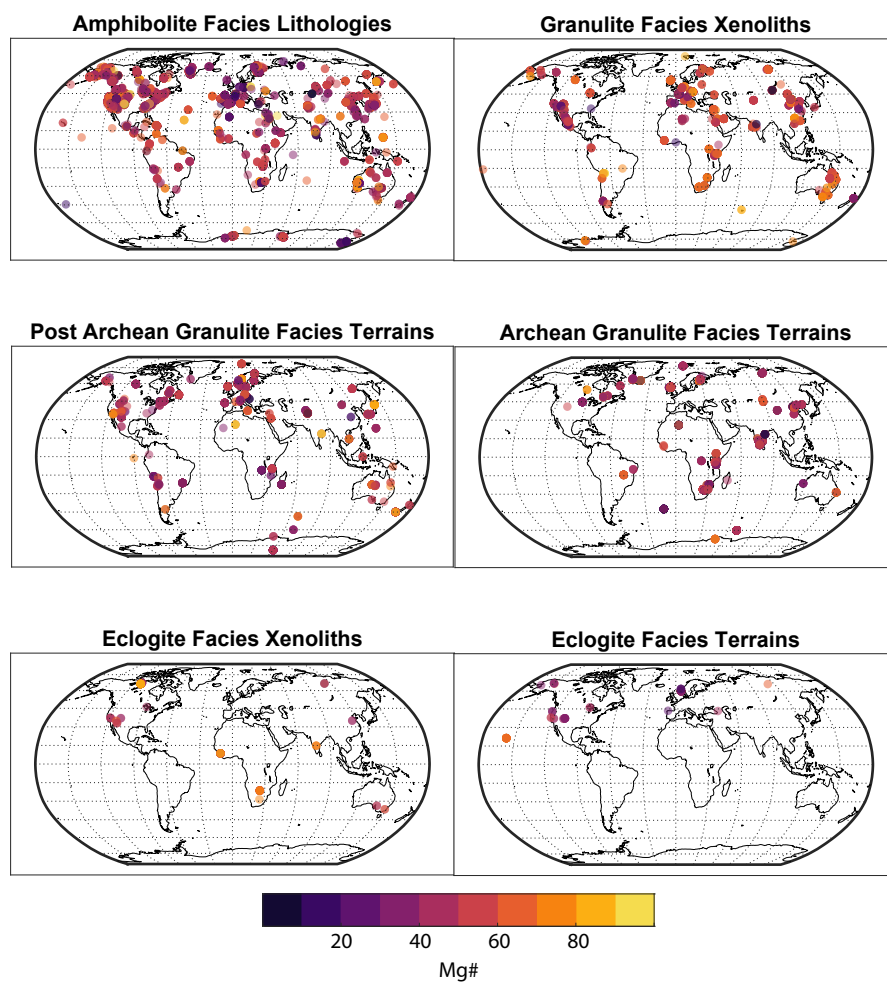


Figure B.6: Maps of the metamorphic sample datasets that we have extensive sample coverage for amphibolite and granulite facies lithologies with more limited exposures of eclogite facies lithologies. Each circle is a single sample whose color is determined by weight percent SiO_2 . Many circles overlap. There is no correlation between Mg# and location.

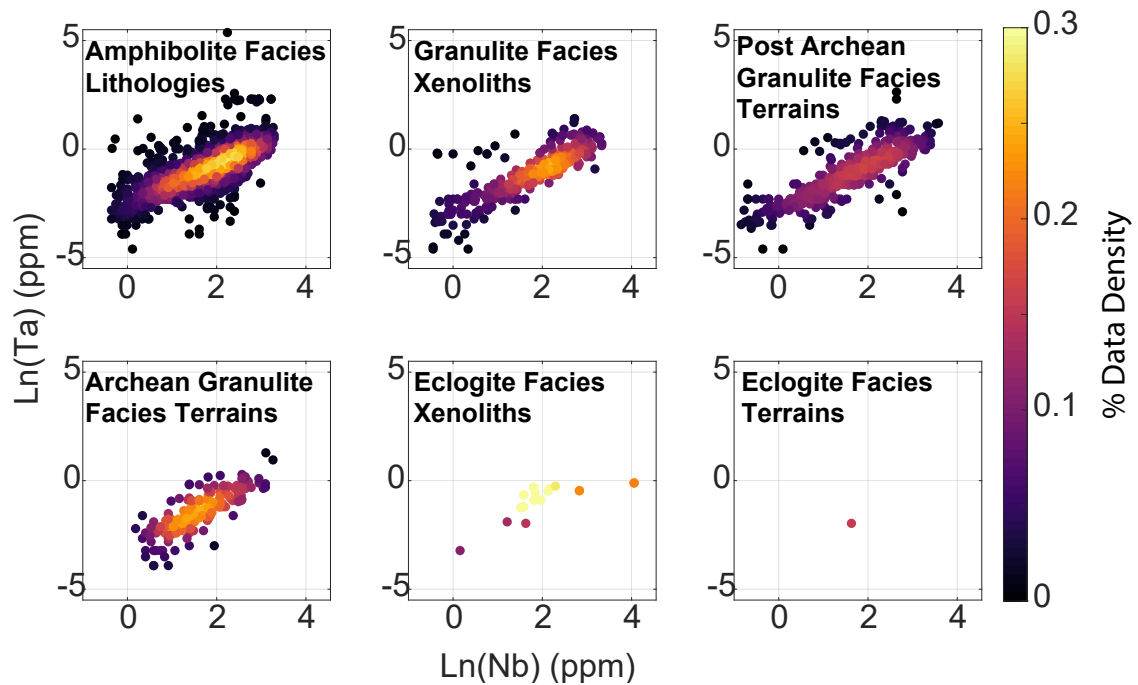


Figure B.7: Nb vs. Ta forms a log-linear relationship with consistent Nb/Ta values for amphibolite and granulite facies lithologies. Color is relative data point density.

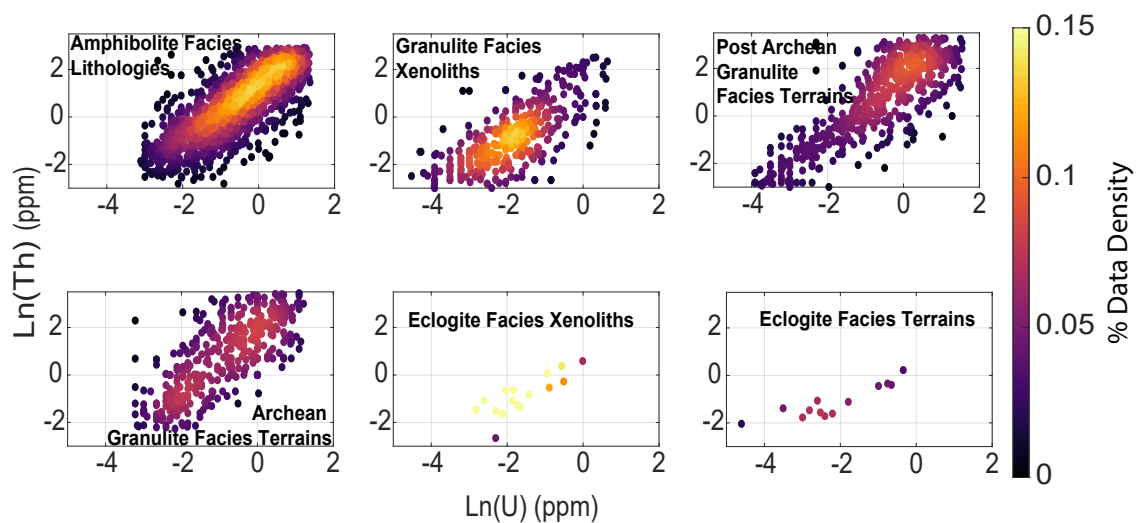


Figure B.8: $\text{Ln}[\text{Th}]$ vs. $\text{Ln}[\text{U}]$ for amphibolite and granulite facies lithologies forms a linear trend. Color indicates relative data point density. The Th/U ratio for all lithologies forms a linear trend in log-log space. Scatter in U concentration could be caused by U's mobility in certain oxidation states since amphibolites can contain hydrous minerals.

Appendix C: Supporting Information for Chapter 4

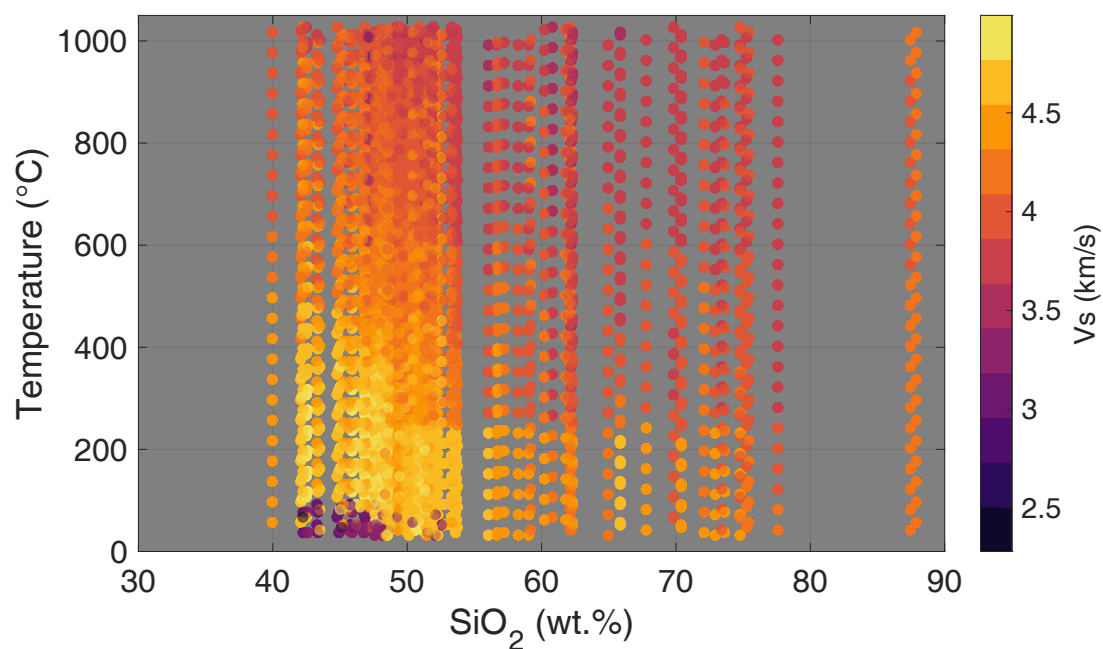


Figure C.1: V_s varies primarily as a function of temperature and composition. The figure shows the calculated V_s for each sample from the southwestern US at different temperature intervals. For this example, pressure is kept constant at 0.858 GPa (~ 30 km depth). Lowering temperature or SiO₂ content generally decreases V_s .

C.1 Compositional Maps and Uncertainties

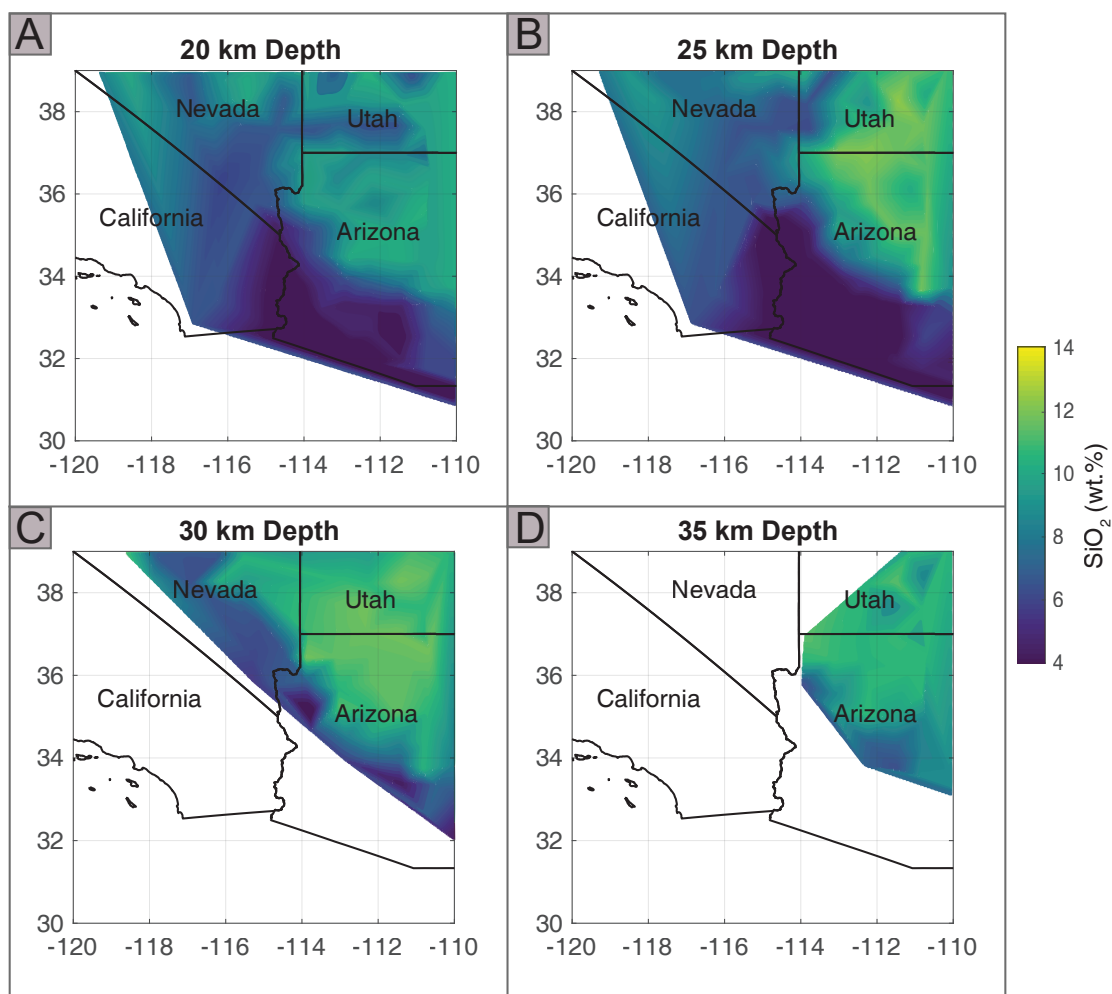


Figure C.2: Uncertainty in wt.% associated with our SiO_2 calculations. Uncertainty is calculated as $\frac{1}{2}$ the inter-quartile range at various depths. Uncertainty is lowest in the Southern Basin and Range and remains relatively consistent with depth.

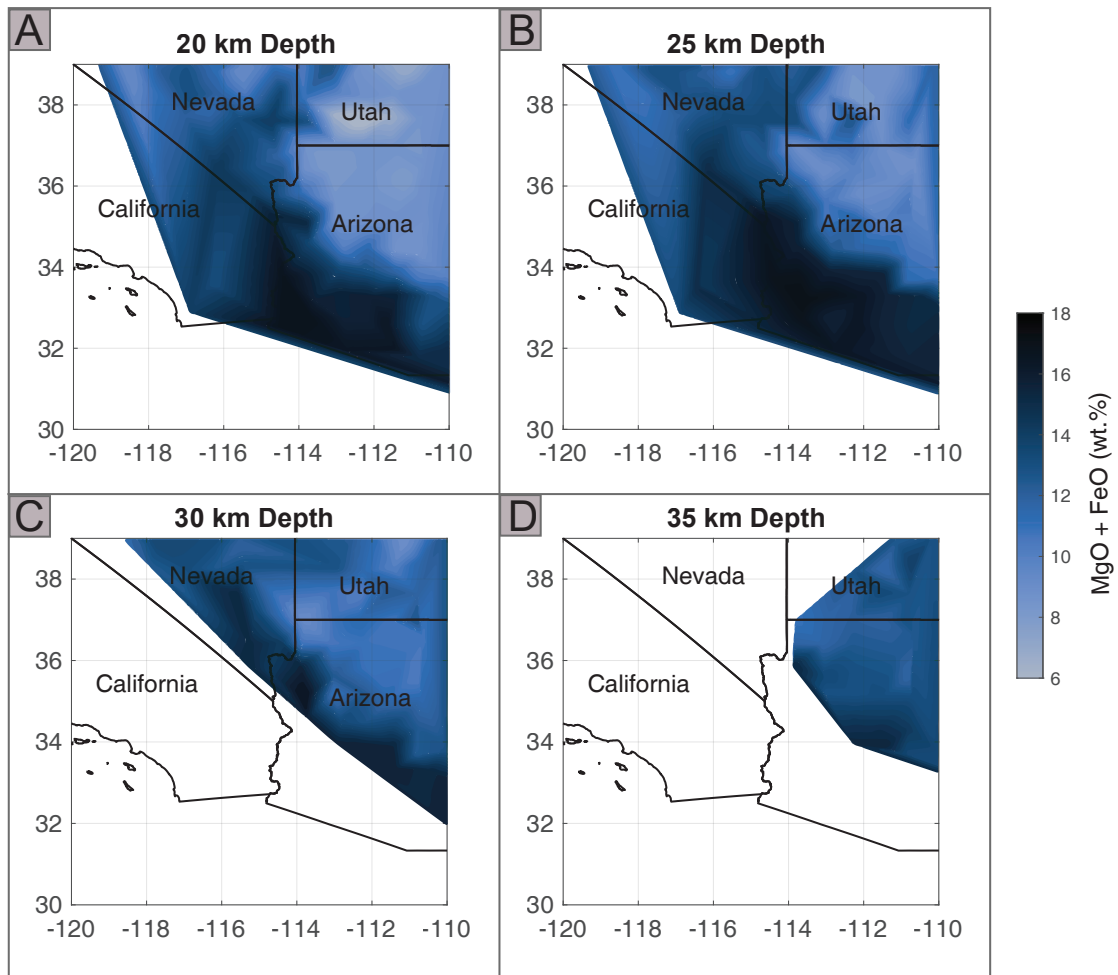


Figure C.3: Variations in median MgO + FeO abundance across the Basin and Range and Colorado Plateau. Where SiO_2 abundance decreases, amount of mafics increases with depth. The Southern Basin and Range has a consistently high mafic content while the Colorado Plateau is more intermediate.

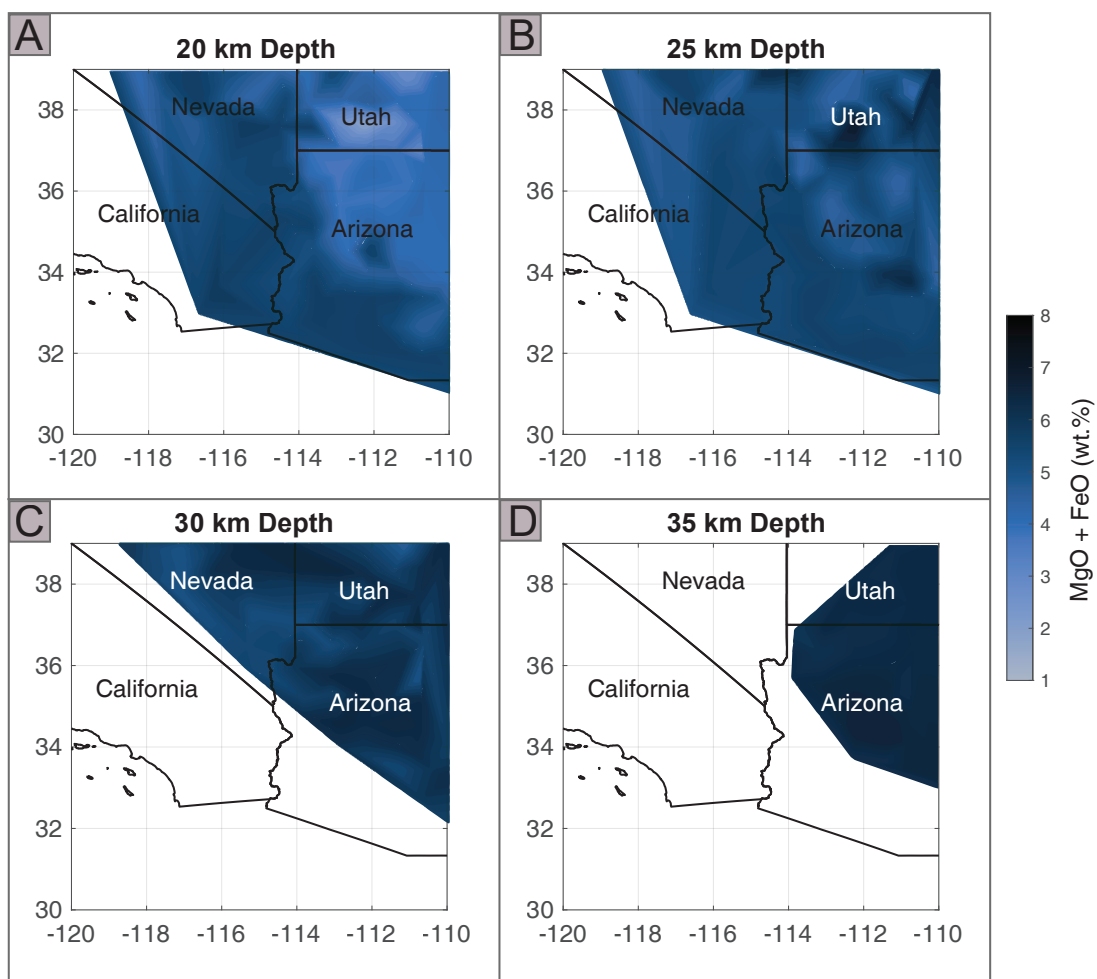


Figure C.4: Uncertainty in wt.% associated with our MgO + FeO calculations. Uncertainty is calculated as $\frac{1}{2}$ the inter-quartile range at various depths, and increases by a few wt.% with increasing depth.

Change in Composition Relative to Uncertainty

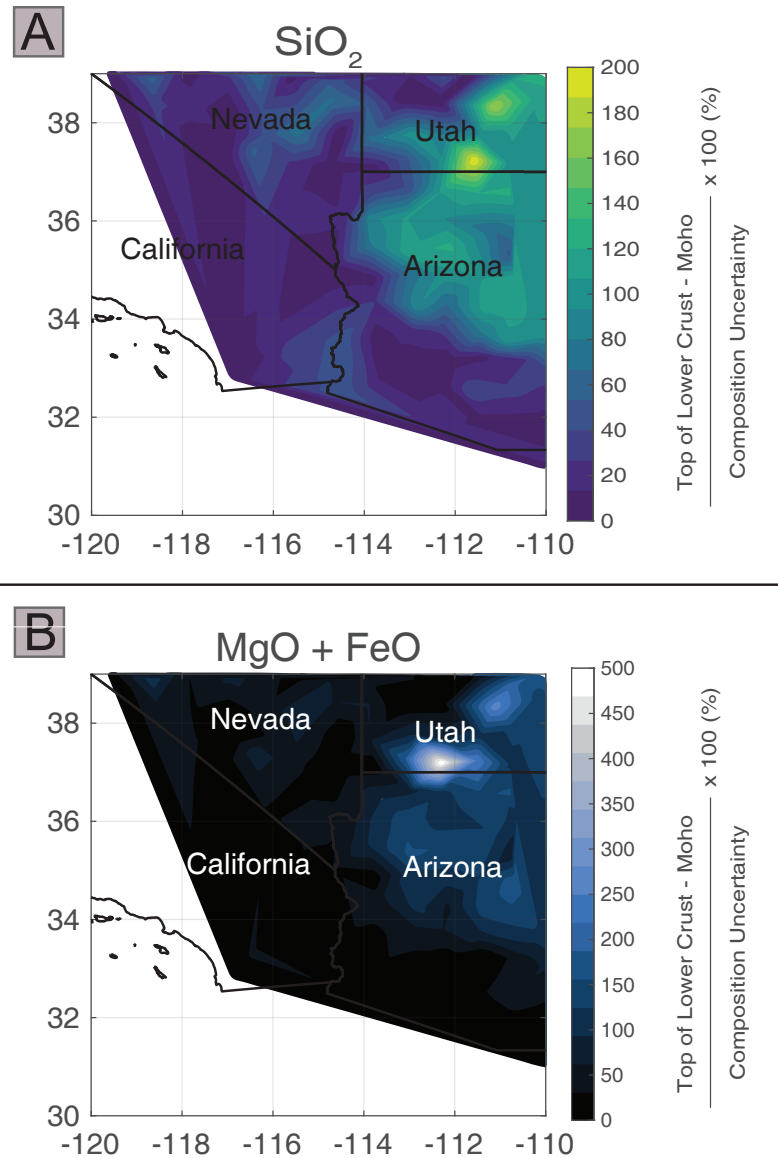


Figure C.5: The Colorado Plateau shows the greatest % change in composition when compared to the uncertainty in our measurements. There is almost no composition change from the top to the bottom of the lower crust in the Southern Basin and Range. The color scale indicates the difference in (A) SiO_2 or (B) mafics between 20 km and at Moho depth, divided by the uncertainty in the composition at Moho depth.

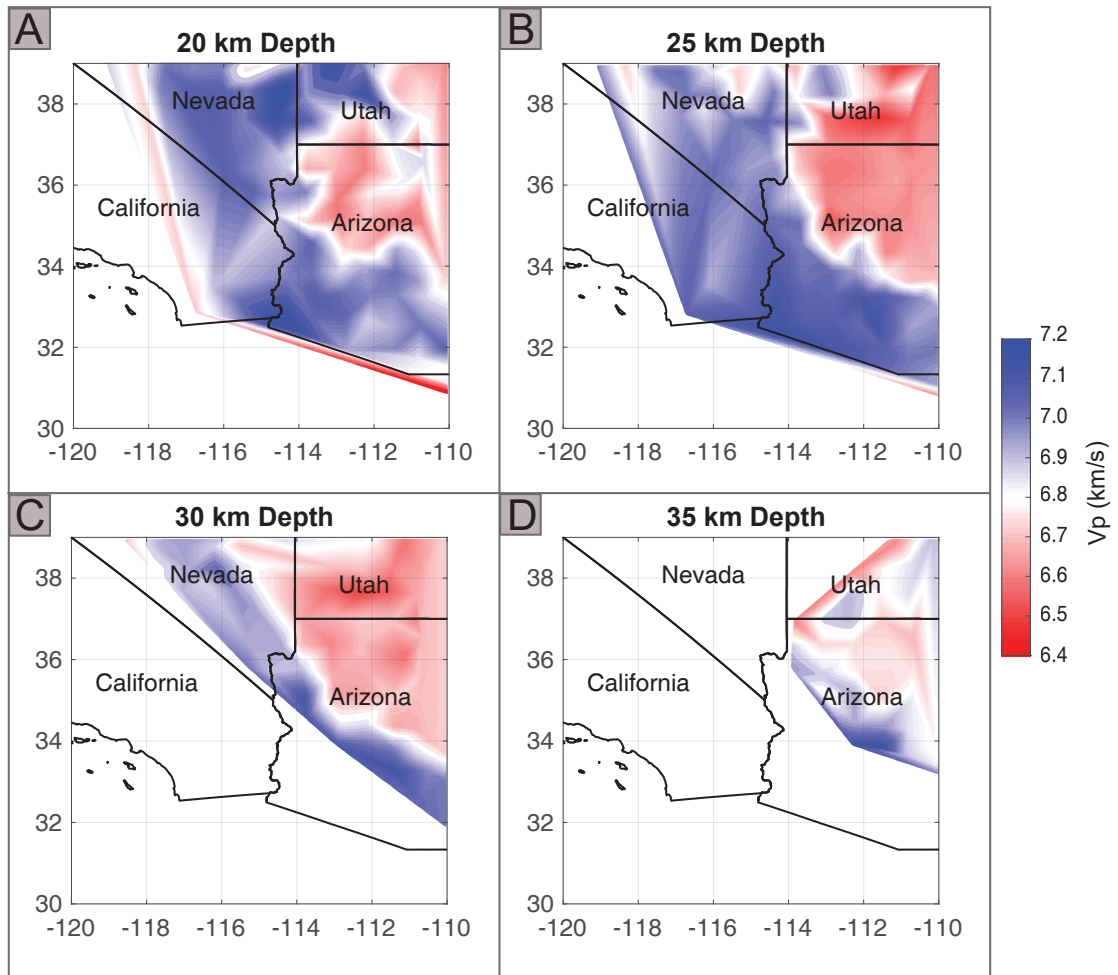


Figure C.6: Median V_p calculated from the joint geochemical-geophysical model. Original V_p 's were calculated in `Perple_X`. The Colorado Plateau is clearly visible as a slower V_p region to the east.

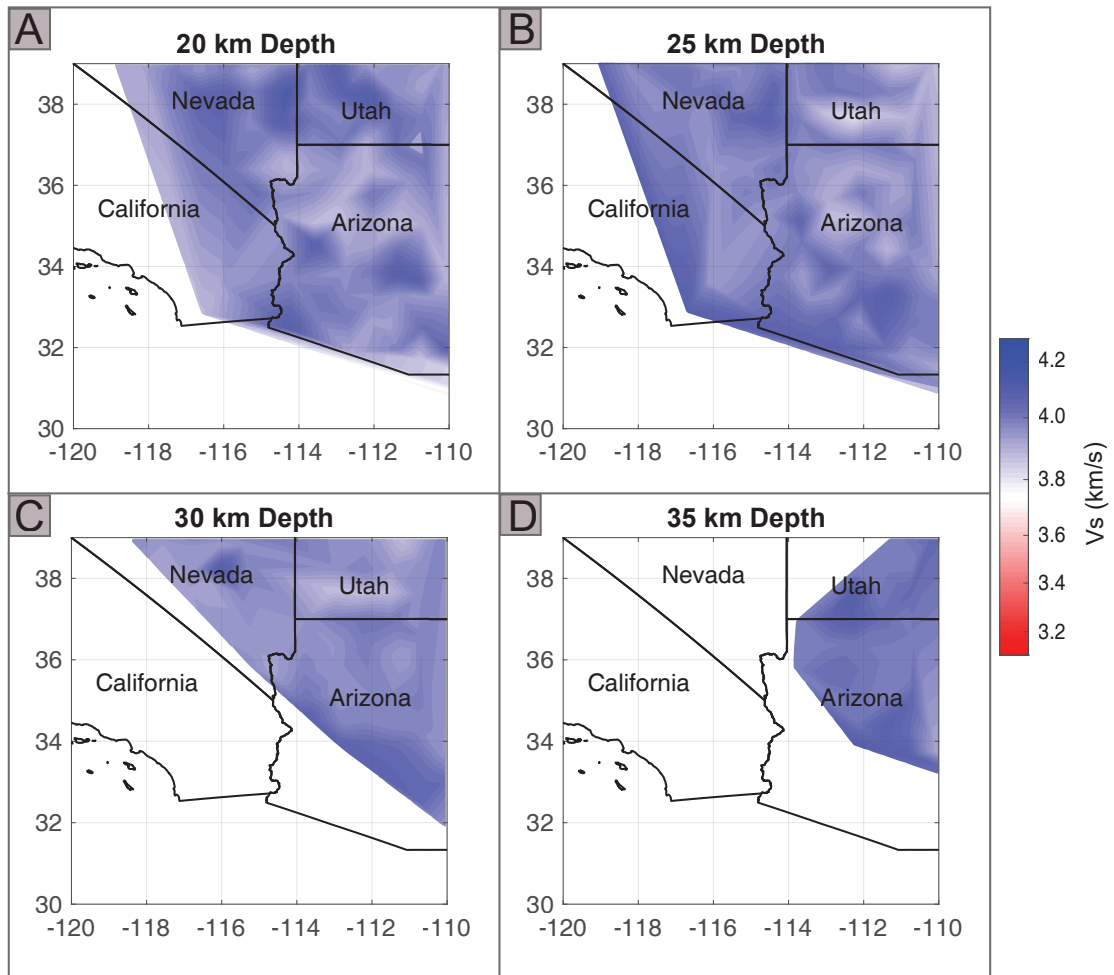


Figure C.7: Joint model selections of median V_s . The geochemical data favor faster V_s solutions and therefore weight the model.

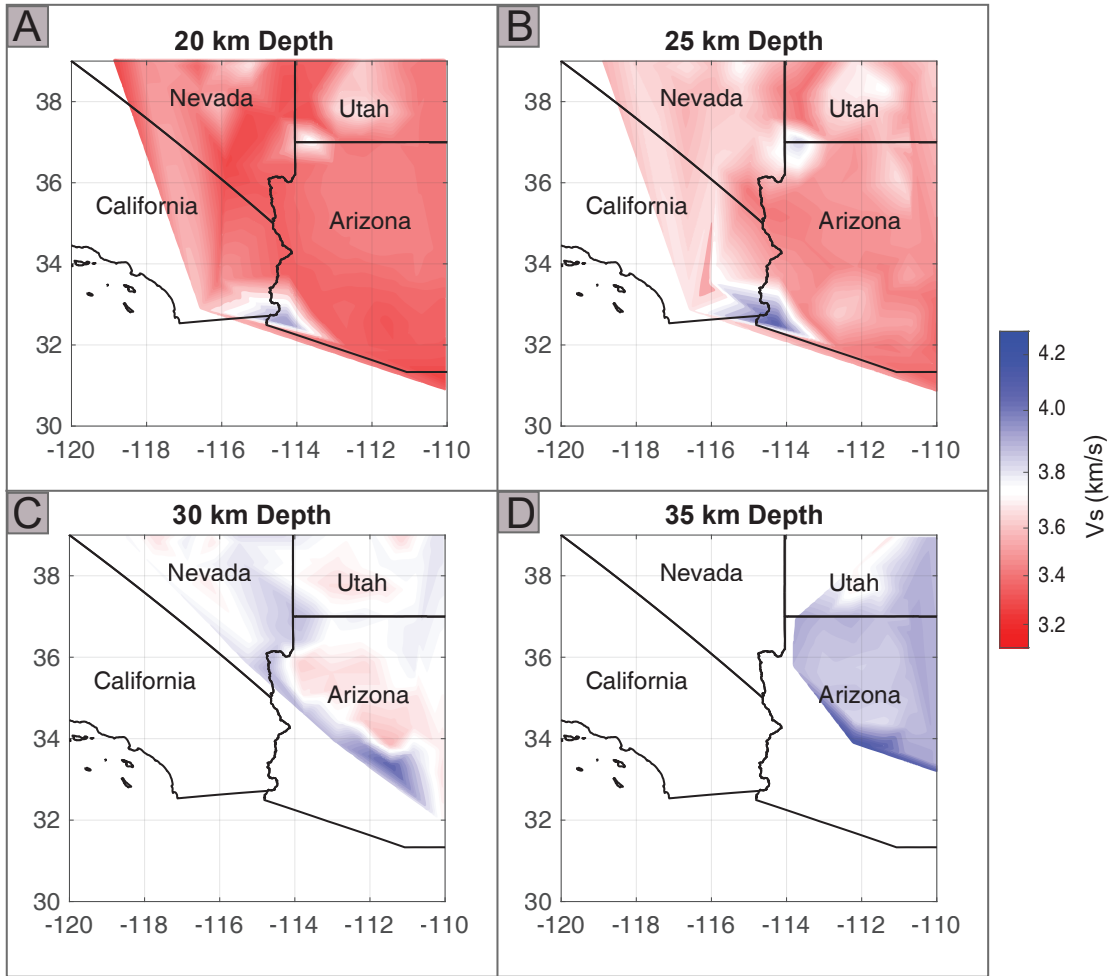


Figure C.8: Median V_s from inversion of Earthscope USArray seismic data at various depths. V_s increases noticeably with increasing depth.

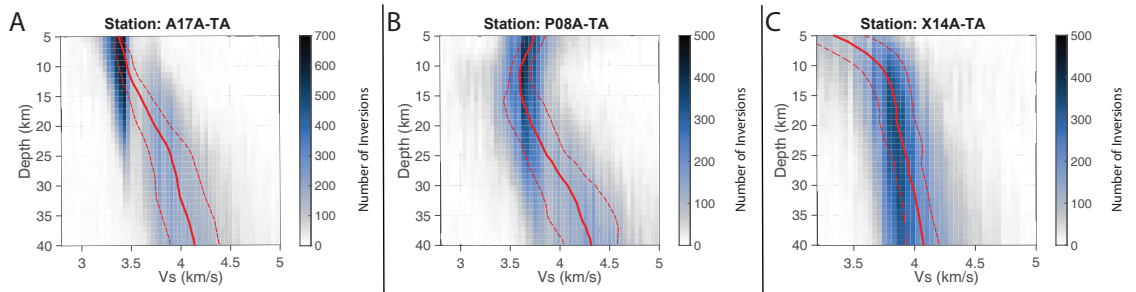


Figure C.9: Examples of the results of the dispersion wave inversions from three different TA stations used to produce the "seismic V_s " in the model. At each depth (y-axis) an ensemble of results shows the distribution of possible V_s values from the seismic inversions alone.

Bibliography

- RN Abbott and JP Greenwood. Retrograde metamorphism of eclogite in the southern appalachian mountains, usa—a case involving seamount subduction? *Journal of Metamorphic Geology*, 19(4):433–443, 2001.
- M. Agostini, S. Appel, G. Bellini, J. Benziger, D. Bick, G. Bonfini, D. Bravo, B. Caccianiga, F. Calaprice, A. Caminata, P. Cavalcante, A. Chepurnov, K. Choi, D. D’Angelo, S. Davini, A. Derbin, L. Di Noto, I. Drachnev, A. Empl, A. Etenko, G. Fiorentini, K. Fomenko, D. Franco, F. Gabriele, C. Galbiati, C. Ghiano, M. Giammarchi, M. Goeger-Neff, A. Goretti, M. Gromov, C. Hagner, T. Houdy, E. Hungerford, Aldo Ianni, Andrea Ianni, K. Jedrzejczak, M. Kaiser, V. Kobychiev, D. Korablev, G. Korga, D. Kryn, M. Laubenstein, B. Lehnert, E. Litvinovich, F. Lombardi, P. Lombardi, L. Ludhova, G. Lukyanchenko, I. Machulin, S. Manecki, W. Maneschg, F. Mantovani, S. Marcocci, E. Meroni, M. Meyer, L. Miramonti, M. Misiaszek, M. Montuschi, P. Mosteiro, V. Muratova, B. Neumair, L. Oberauer, M. Obolensky, F. Ortica, K. Otis, L. Pagani, M. Pallavicini, L. Papp, L. Perasso, A. Pocar, G. Ranucci, A. Razeto, A. Re, B. Ricci, A. Romani, R. Roncin, N. Rossi, S. Schönert, D. Semenov, H. Simgen, M. Skorokhvatov, O. Smirnov, A. Sotnikov, S. Sukhotin, Y. Suvorov, R. Tartaglia, G. Testera, J. Thurn, M. Toropova, E. Unzhakov, R. B. Vogelaar, F. von Feilitzsch, H. Wang, S. Weinz, J. Winter, M. Wojcik, M. Wurm, Z. Yokley,

O. Zaimidoroga, S. Zavatarelli, K. Zuber, and G. Zuzel. Spectroscopy of geoneutrinos from 2056 days of borexino data. *Phys. Rev. D*, 92:031101, Aug 2015. doi: 10.1103/PhysRevD.92.031101.

M. Agostini, K. Altenmüller, S. Appel, V. Atroshchenko, Z. Bagdasarian, D. Basilico, G. Bellini, J. Benziger, D. Bick, G. Bonfini, D. Bravo, B. Caccianiga, F. Calaprice, A. Caminata, L. Cappelli, P. Cavalcante, F. Cavanna, A. Chepurnov, K. Choi, D. D’Angelo, S. Davini, A. Derbin, A. Di Giacinto, V. Di Marcello, X. F. Ding, A. Di Ludovico, L. Di Noto, I. Drachnev, G. Fiorentini, A. Formozov, D. Franco, F. Gabriele, C. Galbiati, M. Gschwender, C. Ghiano, M. Giammarchi, A. Goretti, M. Gromov, D. Guffanti, C. Hagner, E. Hungerford, Aldo Ianni, Andrea Ianni, A. Jany, D. Jeschke, S. Kumaran, V. Kobychiev, G. Korga, T. Lachenmaier, T. Lasserre, M. Laubenstein, E. Litvinovich, P. Lombardi, I. Lomsкая, L. Ludhova, G. Lukyanchenko, L. Lukyanchenko, I. Machulin, F. Mantovani, G. Manuzio, S. Marocchi, J. Maricic, J. Martyn, E. Meroni, M. Meyer, L. Miramonti, M. Misiasek, M. Montuschi, V. Muratova, B. Neumair, M. Nieslony, L. Oberauer, A. Onillon, V. Orekhov, F. Ortica, M. Pallavicini, L. Papp, Ö. Penek, L. Pietrofaccia, N. Pilipenko, A. Pocar, G. Raikov, M. T. Ranalli, G. Ranucci, A. Razeto, A. Re, M. Redchuk, B. Ricci, A. Romani, N. Rossi, S. Rottenanger, S. Schönert, D. Semenov, M. Skorohvatov, O. Smirnov, A. Sotnikov, V. Strati, Y. Suvorov, R. Tartaglia, G. Testera, J. Thurn, E. Unzhakov, A. Vishneva, M. Vivier, R. B. Vogelaar, F. von Feilitzsch, M. Wojcik, M. Wurm, O. Zaimidoroga, S. Zavatarelli, K. Zuber, and G. Zuzel. Comprehensive geoneutrino analysis with Borexino. *Phys. Rev. D*, 101(1):012009, 2020.

doi: 10.1103/PhysRevD.101.012009.

Abdulrahman Al-Safarjalani, Sobhi Nasir, Thomas Fockenberg, and Hans-Joachim Massonne. Chemical composition of crustal xenoliths from southwestern Syria: Characterization of the upper part of the lower crust beneath the Arabian plate. *Geochemistry*, 69(4):359–375, 2009. ISSN 0009-2819. doi: 10.1016/j.chemer.2009.05.005.

Francis Albarède. The growth of continental crust. *Tectonophysics*, 296(1-2):1–14, 1998.

Kiara L Alessio, Martin Hand, David E Kelsey, Megan A Williams, Laura J Morrissey, and Karin Barovich. Conservation of deep crustal heat production. *Geology*, 46(4):335–338, 2018.

Fengpeng An, Guangpeng An, Qi An, Vito Antonelli, Eric Baussan, John Beacom, Leonid Bezrukov, Simon Blyth, Riccardo Brugnera, Margherita Buizza Avanzini, et al. Neutrino physics with junos. *Journal of Physics G: Nuclear and Particle Physics*, 43(3):030401, 2016.

T Araki, S Enomoto, K Furuno, Y Gando, K Ichimura, H Ikeda, K Inoue, Y Kishimoto, M Koga, Y Koseki, et al. Experimental investigation of geologically produced antineutrinos with kamland. *Nature*, 436(7050):499–503, 2005.

Ricardo Arevalo and William F McDonough. Tungsten geochemistry and implications for understanding the earth’s interior. *Earth and Planetary Science Letters*, 272(3-4):656–665, 2008.

Ricardo Arevalo and William F McDonough. Chemical variations and regional diversity

- observed in MORB. *Chemical Geology*, 271(1):70–85, 2010. ISSN 0009-2541. doi: 10.1016/j.chemgeo.2009.12.013.
- Ricardo Arevalo, William F McDonough, and Mario Luong. The K/U ratio of the silicate earth: Insights into mantle composition, structure and thermal evolution. *Earth and Planetary Science Letters*, 278(3-4):361–369, 2009.
- NT Arndt. Archean komatiites. *Archean crustal evolution*, 10:11–44, 1994.
- Irina M Artemieva. Global 1×1 thermal model TC1 for the continental lithosphere: implications for lithosphere secular evolution. *Tectonophysics*, 416(1-4):245–277, 2006.
- LD Ashwal, P Morgan, SA Kelley, and JA Percival. Heat production in an Archean crustal profile and implications for heat flow and mobilization of heat-producing elements. *Earth and Planetary Science Letters*, 85(4):439–450, 1987.
- Lewis D Ashwal, Robert D Tucker, and Ernst K Zinner. Slow cooling of deep crustal granulites and pb-loss in zircon. *Geochimica et Cosmochimica Acta*, 63(18):2839–2851, 1999.
- D A Banks, R Green, RA Cliff, and B W D Yardley. Chlorine isotopes in fluid inclusions: determination of the origins of salinity in magmatic fluids. *Geochimica et Cosmochimica Acta*, 64(10):1785–1789, 2000.
- Sarah-Jane Barnes. *Chalcophile elements*. Springer International Publishing, 2016.
- Katherine R Barnhart, Kevin H Mahan, Terrence J Blackburn, Samuel A Bowring, and Francis O Dudas. Deep crustal xenoliths from central montana, usa: Implications for

- the timing and mechanisms of high-velocity lower crust formation. *Geosphere*, 8(6): 1408–1428, 2012.
- C Bassin, G Laske, and G Masters. The current limits of resolution for surface wave tomography in North America. *Eos*, 81, 2000.
- John F. Beacom, Shaomin Chen, Jianping Cheng, Sayed N. Doustimotlagh, Yuanning Gao, Guanghua Gong, Hui Gong, Lei Guo, Ran Han, Hong-Jian He, Xingtao Huang, Jianmin Li, Jin Li, Mohan Li, Xueqian Li, Wei Liao, Guey-Lin Lin, Zuowei Liu, William McDonough, Ondřej Šrámek, Jian Tang, Linyan Wan, Yuanqing Wang, Zhe Wang, Zongyi Wang, Hanyu Wei, Yufei Xi, Ye Xu, Xun-Jie Xu, Zhenwei Yang, Chunfa Yao, Minfang Yeh, Qian Yue, Liming Zhang, Yang Zhang, Zhihong Zhao, Yangheng Zheng, Xiang Zhou, Xianglei Zhu, and Kai Zuber. Physics prospects of the Jinping neutrino experiment. *Chinese Physics C*, 41(2):023002, 2017. doi: 10.1088/1674-1137/41/2/023002.
- G Bellini, A Ianni, L Ludhova, Fabio Mantovani, and William F McDonough. Geoneutrinos. *Progress in Particle and Nuclear Physics*, 73:1–34, 2013.
- Gianpaolo Bellini, Kunio Inoue, Fabio Mantovani, Andrea Serafini, Virginia Strati, and Hiroko Watanabe. Geoneutrinos and geoscience: an intriguing joint-venture. *L'Espresso Rivista del Nuovo Cimento*, pages 1–105, 2021. doi: <https://doi.org/10.1007/s40766-021-00026-7>.
- Peter Bird. Laramide crustal thickening event in the Rocky Mountain Foreland and

- Great Plains. *Tectonics*, 3(7):741–758, 1984. ISSN 1944-9194. doi: 10.1029/TC003i007p00741.
- S R Bohlen and Klaus Mezger. Origin of granulite terranes and the formation of the lowermost continental crust. *Science*, 244(4902):326–329, 1989.
- Thomas M Brocher. Empirical Relations between Elastic Wavespeeds and Density in the Earth’s Crust. *Bulletin of the Seismological Society of America*, 95(6):2081–2092, 2005. ISSN 0037-1106. doi: 10.1785/0120050077.
- Michael Brown and Tim Johnson. Time’s arrow, time’s cycle: Granulite metamorphism and geodynamics. *Mineralogical Magazine*, 83(3):323–338, 2019.
- JS Buehler and PM Shearer. Uppermost mantle seismic velocity structure beneath usarray. *Journal of Geophysical Research: Solid Earth*, 122(1):436–448, 2017.
- Roland Bürgmann and Georg Dresen. Rheology of the lower crust and upper mantle: Evidence from rock mechanics, geodesy, and field observations. *Annu. Rev. Earth Planet. Sci.*, 36:531–567, 2008.
- Warren B Caldwell, Simon L Klemperer, Shyam S Rai, and Jesse F Lawrence. Partial melt in the upper-middle crust of the northwest himalaya revealed by rayleigh wave dispersion. *Tectonophysics*, 477(1-2):58–65, 2009.
- Fabio Cammarano and Mattia Guerri. Global thermal models of the lithosphere. *Geophysical Journal International*, 210(1):56–72, 2017.
- R L Carlson and C N Herrick. Densities and porosities in the oceanic crust and their

- variations with depth and age. *Journal of Geophysical Research: Solid Earth*, 95(B6): 9153–9170, 1990.
- DA Carswell and SJ Cuthbert. Eclogite facies metamorphism in the lower continental crust. *Geological Society, London, Special Publications*, 24(1):193–209, 1986.
- Bruce W Chappell, Colleen J Bryant, and Doone Wyborn. Peraluminous i-type granites. *Lithos*, 153:142–153, 2012.
- Bernard Charlier, Olivier Namur, Michael J Toplis, Pierre Schiano, Nicolas Cluzel, Michael D Higgins, and Jacqueline Vander Auwera. Large-scale silicate liquid immiscibility during differentiation of tholeiitic basalt to granite and the origin of the daly gap. *Geology*, 39(10):907–910, 2011.
- Kang Chen, Richard J Walker, Roberta L Rudnick, Shan Gao, Richard M Gaschnig, Igor S Puchtel, Ming Tang, and Zhao-Chu Hu. Platinum-group element abundances and re–os isotopic systematics of the upper continental crust through time: Evidence from glacial diamictites. *Geochimica et Cosmochimica Acta*, 191:1–16, 2016.
- Kang Chen, Roberta L Rudnick, Zaicong Wang, Ming Tang, Richard M Gaschnig, Zongqi Zou, Tao He, Zhaochu Hu, and Yongsheng Liu. How mafic was the archean upper continental crust? insights from cu and ag in ancient glacial diamictites. *Geochimica et Cosmochimica Acta*, 278:16–29, 2020.
- Hyun-Moo Cho, Han-Joon Kim, Hyeong-Tae Jou, Jong-Kuk Hong, and Chang-Eob Baag. Transition from rifted continental to oceanic crust at the southeastern korean margin in the east sea (japan sea). *Geophysical Research Letters*, 31(7), 2004.

Nikolas I. Christensen and David M. Fountain. Constitution of the Lower Continental Crust Based on Experimental Studies of Seismic Velocities in Granulite. *Geological Society of America Bulletin*, 86(2):227, 1975. ISSN 0016-7606. doi: 10.1130/0016-7606(1975)86(227:COTLCC)2.0.CO;2.

Nikolas I Christensen and Walter D Mooney. Seismic velocity structure and composition of the continental crust: A global view. *Journal of Geophysical Research: Solid Earth*, 100(B6):9761–9788, 1995a.

Nikolas I Christensen and Walter D Mooney. Seismic velocity structure and composition of the continental crust: A global view. *Journal of Geophysical Research: Solid Earth*, 100(B6):9761–9788, 1995b. ISSN 2156-2202. doi: 10.1029/95JB00259.

J Graham Cogley. Continental margins and the extent and number of the continents. *Reviews of Geophysics*, 22(2):101–122, 1984.

Massimo Coltorti, Riccardo Boraso, Fabio Mantovani, Michele Morsilli, Giovanni Fiorentini, Alberto Riva, G Rusciadelli, Renzo Tassinari, C Tomei, G Di Carlo, et al. U and Th content in the Central Apennines continental crust: A contribution to the determination of the geo-neutrinos flux at LNGS. *Geochimica et Cosmochimica Acta*, 75(9):2271–2294, 2011. doi: 10.1016/j.gca.2011.01.024.

Peter J Coney. Cordilleran metamorphic core complexes: An overview. *Cordilleran metamorphic core complexes: Geological Society of America Memoir*, 153:7–31, 1980.

Peter J Coney and Tekla A Harms. Cordilleran metamorphic core complexes: Cenozoic extensional relics of mesozoic compression. *Geology*, 12(9):550–554, 1984.

J A D Connolly. Computation of phase equilibria by linear programming: A tool for geodynamic modeling and its application to subduction zone decarbonation. *Earth and Planetary Science Letters*, 236(1):524–541, 2005. ISSN 0012-821X. doi: 10.1016/j.epsl.2005.04.033.

Sandro Conticelli, Richard W Carlson, Elisabeth Widom, and Giancarlo Serri. Chemical and isotopic composition (Os, Pb, Nd, and Sr) of Neogene to Quaternary calc-alkalic, shoshonitic, and ultrapotassic mafic rocks from the Italian peninsula: Inferences on the nature of their mantle sources. *Cenozoic volcanism in the Mediterranean area*, 418: 171, 2007.

F. J. Cooper, J. P. Platt, R. Anczkiewicz, and M. J. Whitehouse. Footwall dip of a core complex detachment fault: Thermobarometric constraints from the northern Snake Range (Basin and Range, USA). *Journal of Metamorphic Geology*, 28(9):997–1020, 2010. ISSN 1525-1314. doi: 10.1111/j.1525-1314.2010.00907.x.

Domenico Cosentino, Paola Cipollari, Pietro Marsili, and Davide Scrocca. Geology of the central apennines: a regional review. *Journal of the virtual explorer*, 36(11):1–37, 2010.

Max D Crittenden, Peter J Coney, George Herbert Davis, and George Hamilton Davis. *Cordilleran metamorphic core complexes*, volume 153. Geological Society of America, 1980.

RA Daly. Igneous rocks and their origin. *Preliminary Note on Alkaline Rhyolites from Tokati, Hokkaido*, 319:20, 1914.

- J Huw Davies. Global map of solid earth surface heat flow. *Geochemistry, Geophysics, Geosystems*, 14(10):4608–4622, 2013.
- R De Argollo and J-G Schilling. Ge/si and ga/al fractionation in hawaiian volcanic rocks. *Geochimica et Cosmochimica Acta*, 42(6):623–630, 1978.
- de H De La Roche, J t Leterrier, Philippe Grandclaude, and M Marchal. A classification of volcanic and plutonic rocks using r1r2-diagram and major-element analyses—its relationships with current nomenclature. *Chemical geology*, 29(1-4):183–210, 1980.
- E Di Luzio, G Mele, MM Tiberti, GP Cavinato, and M Parotto. Moho deepening and shallow upper crustal delamination beneath the central apennines. *Earth and Planetary Science Letters*, 280(1-4):1–12, 2009. doi: <https://doi.org/10.1016/j.epsl.2008.09.018>.
- Jacqueline E Dixon, T. H Dixon, D. R Bell, and R Malservisi. Lateral variation in upper mantle viscosity: Role of water. *Earth and Planetary Science Letters*, 222(2):451–467, 2004. ISSN 0012-821X. doi: 10.1016/j.epsl.2004.03.022.
- Ksenia Dokukina and Michael Mints. Subduction of the mesoarchaeon spreading ridge and related metamorphism, magmatism and deformation by the example of the gridino eclogitized mafic dyke swarm, the belomorian eclogite province, eastern fennoscandian shield. *Journal of Geodynamics*, 123:1–37, 2019.
- Mihai N Ducea. Fingerprinting orogenic delamination. *Geology*, 39(2):191–192, 2011.
- Josef Dufek and Olivier Bachmann. Quantum magmatism: Magmatic compositional gaps generated by melt-crystal dynamics. *Geology*, 38(8):687–690, 2010.

- Gregory Dumond, Michael L Williams, and Sean P Regan. The athabasca granulite terrane and evidence for dynamic behavior of lower continental crust. *Annual Review of Earth and Planetary Sciences*, 46:353–386, 2018.
- Adam M Dziewonski and Don L Anderson. Preliminary reference earth model. *Physics of the earth and planetary interiors*, 25(4):297–356, 1981.
- Göran Ekström. Love and rayleigh phase-velocity maps, 5–40 s, of the western and central usa from usarray data. *Earth and Planetary Science Letters*, 402:42–49, 2014.
- Sanshiro Enomoto. Experimental study of geoneutrinos with kamland. In *Neutrino Geophysics: Proceedings of Neutrino Sciences 2005*, pages 131–146. Springer, 2006.
- Sanshiro Enomoto, Eiji Ohtani, K Inoue, and A Suzuki. Neutrino geophysics with KamLAND and future prospects. *Earth and Planetary Science Letters*, 258(1-2):147–159, 2007. doi: 10.1016/j.epsl.2007.03.038.
- B Farcy, R Arevalo, and William F McDonough. K/U of the morb source and silicate earth. *Journal of Geophysical Research: Solid Earth*, 125(12):e2020JB020245, 2020.
- D Field and R B Elliott. The chemistry of gabbro/amphibolite transitions in south norway. *Contributions to Mineralogy and Petrology*, 47(1):63–76, 1974.
- Dennis Field and Peter W L Clough. K/Rb ratios and metasomatism in metabasites from a precambrian amphibolite–granulite transition zone. *Journal of the Geological Society*, 132(3):277–288, 1976.

David M Fountain and Matthew H Salisbury. Exposed cross-sections through the continental crust: implications for crustal structure, petrology, and evolution. *Earth and Planetary Science Letters*, 56:263–277, 1981.

David M Fountain, Kevin P Furlong, and Matthew H Salisbury. A heat production model of a shield area and its implications for the heat flow-heat production relationship. *Geophysical Research Letters*, 14(3):283–286, 1987.

M B Fowler. Large-ion lithophile element characteristics of an amphibolite facies to granulite facies transition at gruinard bay, north-west scotland. *Journal of Metamorphic Geology*, 4(3):345–359, 1986.

Allison Gale, Muriel Laubier, Stephane Escrig, and Charles H Langmuir. Constraints on melting processes and plume-ridge interaction from comprehensive study of the famous and north famous segments, mid-atlantic ridge. *Earth and Planetary Science Letters*, 365:209–220, 2013.

Azusa Gando, Y Gando, H Hanakago, H Ikeda, K Inoue, K Ishidoshiro, H Ishikawa, M Koga, R Matsuda, S Matsuda, et al. Reactor on-off antineutrino measurement with kamland. *Physical Review D*, 88(3):033001, 2013.

Chao Gao and Vedran Lekić. Consequences of parametrization choices in surface wave inversion: Insights from transdimensional Bayesian methods. *Geophysical Journal International*, 215(2):1037–1063, 2018. ISSN 0956-540X. doi: 10.1093/gji/ggy310.

Ruohan Gao, Zhiwei Li, Ran Han, Andong Wang, Yufeng Li, Yufei Xi, Jingao Liu, Xin Mao, Yao Sun, and Ya Xu. Juloc: A local 3-d high-resolution crustal model in south

- china for forecasting geoneutrino measurements at junco. *Physics of the Earth and Planetary Interiors*, 299:106409, 2020.
- Shan Gao, Ben-Ren Zhang, Zhen-Ming Jin, Hartmut Kern, Ting-Chuan Luo, and Zi-Dan Zhao. How mafic is the lower continental crust? *Earth and Planetary Science Letters*, 161(1-4):101–117, 1998.
- Yajian Gao, Frederik Tilmann, Dirk-Philip van Herwaarden, Solvi Thrastarson, Andreas Fichtner, Benjamin Heit, Xiaohui Yuan, and Bernd Dieter Schurr. Full waveform inversion beneath the central andes: Insight into the dehydration of the nazca slab and delamination of the back-arc lithosphere. *Earth and Space Science Open Archive ES-SOAr*, 2021.
- M Gard, D Hasterok, M Hand, and G Cox. Variations in continental heat production from 4 ga to the present: Evidence from geochemical data. *Lithos*, 342:391–406, 2019.
- Richard M Gaschnig, Roberta L Rudnick, William F McDonough, Alan J Kaufman, John W Valley, Zhaochu Hu, Shan Gao, and Michelle L Beck. Compositional evolution of the upper continental crust through time, as constrained by ancient glacial diamictites. *Geochimica et Cosmochimica Acta*, 186:316–343, 2016. doi: 10.1016/j.gca.2016.03.020.
- T V Gerya, L L Perchuk, and J-P Burg. Transient hot channels: perpetrating and regurgitating ultrahigh-pressure, high-temperature crust–mantle associations in collision belts. *Lithos*, 103(1-2):236–256, 2008.
- Taras V Gerya, Leonid L Perchuk, Walter V Maresch, Arne P Willner, Dirk D Van

- Reenen, and C Andre Smit. Thermal regime and gravitational instability of multi-layered continental crust: implications for the buoyant exhumation of high-grade metamorphic rocks. *European Journal of Mineralogy*, 14(4):687–699, 2002.
- Karsten Gohl. Antarctica’s continent-ocean transitions: consequences for tectonic reconstructions. In *Antarctica: A Keystone in a Changing World: Proceedings of the 10th International Symposium on Antarctic Earth Sciences/Alan K. Cooper, Peter Barrett, Howard Stagg, Bryan Storey, Edmund Stump, Woody Wise, and the 10th ISAES editorial team, Polar Research*, pages 29–38, 2008.
- Allison T Greaney, Roberta L Rudnick, Richard M Gaschnig, Joseph B Whalen, Béatrice Luais, and John D Clemens. Geochemistry of molybdenum in the continental crust. *Geochimica et Cosmochimica Acta*, 238:36–54, 2018.
- Benjamin Gruber, Thomas Chacko, D Graham Pearson, Claire Currie, and Andrew Menzies. Heat production and maximum temperatures in cratonic crust: evidence from lower crustal xenoliths from the slave craton. *Lithos*, 380:105889, 2021.
- BR Hacker, MH Ritzwoller, and J Xie. Partially melted, mica-bearing crust in central tibet. *Tectonics*, 33(7):1408–1424, 2014.
- Bradley R Hacker, Torgeir B Andersen, Scott Johnston, Andrew RC Kylander-Clark, Emily M Peterman, Emily O Walsh, and David Young. High-temperature deformation during continental-margin subduction & exhumation: The ultrahigh-pressure western gneiss region of norway. *Tectonophysics*, 480(1-4):149–171, 2010.

- Bradley R Hacker, Peter B Kelemen, and Mark D Behn. Differentiation of the continental crust by relamination. *Earth and Planetary Science Letters*, 307(3-4):501–516, 2011.
- Bradley R Hacker, Peter B Kelemen, and Mark D Behn. Continental lower crust. *Annual Review of Earth and Planetary Sciences*, 43:167–205, 2015a.
- Bradley R Hacker, Peter B Kelemen, and Mark D Behn. Continental Lower Crust. *Annual Review of Earth and Planetary Sciences*, 43(1):167–205, 2015b. doi: 10.1146/annurev-earth-050212-124117.
- Bradley R. Hacker, Peter B. Kelemen, and Mark D. Behn. Continental Lower Crust. *Annual Review of Earth and Planetary Sciences*, 43(1):167–205, 2015c. doi: 10.1146/annurev-earth-050212-124117.
- Alex N Halliday, Alan P Dickin, Robert N Hunter, Gareth R Davies, Tim J Dempster, P Joseph Hamilton, and Brian G J Upton. Formation and composition of the lower continental crust: Evidence from Scottish xenolith suites. *Journal of Geophysical Research: Solid Earth*, 98(B1):581–607, 1993. ISSN 2156-2202. doi: 10.1029/92JB02276.
- Jens Hartmann, Hans H Dürr, Nils Moosdorf, Michel Meybeck, and Stephan Kempe. The geochemical composition of the terrestrial surface (without soils) and comparison with the upper continental crust. *International Journal of Earth Sciences*, 101(1):365–376, 2012.
- D Hasterok, M Gard, G Cox, and M Hand. A 4 ga record of granitic heat production:

- Implications for geodynamic evolution and crustal composition of the early earth. *Pre-cambrian Research*, 331:105375, 2019a.
- D Hasterok, Matthew Gard, C M B Bishop, and David Kelsey. Chemical identification of metamorphic protoliths using machine learning methods. *Computers & Geosciences*, 132:56–68, 2019b.
- Christopher J Hawkesworth and AIS Kemp. Evolution of the continental crust. *Nature*, 443(7113):811–817, 2006.
- Naoshi Hirata, Boris Ya Karp, Takashi Yamaguchi, Toshihiko Kanazawa, Kiyoshi Suehiro, Junzo Kasahara, Hajime Shiobara, Masanao Shinohara, and Hajimu Kinoshita. Oceanic crust in the japan basin of the japan sea by the 1990 japan-ussr expedition. *Geophysical Research Letters*, 19(20):2027–2030, 1992.
- W Steven Holbrook, Walter D Mooney, and Nikolas I Christensen. The seismic velocity structure of the deep continental crust. *Continental lower crust*, 23:1–43, 1992.
- T J B Holland and R Powell. An internally consistent thermodynamic data set for phases of petrological interest: An Internally Consistent Thermodynamic Dataset. *Journal of Metamorphic Geology*, 16(3):309–343, 2004. ISSN 02634929, 15251314. doi: 10.1111/j.1525-1314.1998.00140.x.
- Yu Huang, Viacheslav Chubakov, Fabio Mantovani, Roberta L Rudnick, and William F McDonough. A reference Earth model for the heat-producing elements and associated geoneutrino flux. *Geochemistry, Geophysics, Geosystems*, 14(6):2003–2029, 2013. ISSN 1525-2027. doi: 10.1002/ggge.20129.

JA Jackson. Strength of the continental lithosphere: time to abandon the jelly sandwich?

GSA today, 12:4–10, 2002.

M D Jackson, J Blundy, and R S J Sparks. Chemical differentiation, cold storage and remobilization of magma in the earth's crust. *Nature*, 564(7736):405–409, 2018.

Oliver Jagoutz and Max W Schmidt. The formation and bulk composition of modern juvenile continental crust: The kohistan arc. *Chemical Geology*, 298:79–96, 2012.

Oliver Jagoutz, Othmar Müntener, Max W Schmidt, and Jean-Pierre Burg. The roles of flux-and decompression melting and their respective fractionation lines for continental crust formation: Evidence from the kohistan arc. *Earth and Planetary Science Letters*, 303(1-2):25–36, 2011.

C Jaupart and J C Mareschal. Constraints on heat production from heat flow data, treatise on geochemistry, vol. 3, the crust, edited by r. rudnick, 2003.

C Jaupart, S Labrosse, F Lucazeau, and JC Mareschal. 7.06-temperatures, heat and energy in the mantle of the earth. *Treatise on geophysics*, 7:223–270, 2007a.

C Jaupart, JC Mareschal, and AB Watts. Heat flow and thermal structure of the lithosphere. *Treatise on geophysics*, 6:217–252, 2007b.

C Jaupart, J-C Mareschal, H Bouquerel, and C Phaneuf. The building and stabilization of an archean craton in the superior province, canada, from a heat flow perspective. *Journal of Geophysical Research: Solid Earth*, 119(12):9130–9155, 2014.

C. Jaupart, S. Labrosse, F. Lucazeau, and J.-C. Mareschal. Temperatures, heat, and energy

- in the mantle of the earth. In Gerald Schubert, editor, *Treatise on Geophysics (Second Edition)*, volume 7, pages 223–270. Elsevier, Oxford, 2015. ISBN 978-0-444-53803-1. doi: 10.1016/B978-0-444-53802-4.00126-3.
- Claude Jaupart, Jean-Claude Mareschal, and Lidia Iarotsky. Radiogenic heat production in the continental crust. *Lithos*, 262:398–427, 2016.
- M Javoy. The integral enstatite chondrite model of the earth. *Geophysical Research Letters*, 22(16):2219–2222, 1995.
- Marc Javoy, Edouard Kaminski, François Guyot, Denis Andrault, Chrystèle Sanloup, Manuel Moreira, Stéphane Labrosse, Albert Jambon, Pierre Agrinier, Anne Davaille, and Claude Jaupart. The chemical composition of the Earth: Enstatite chondrite models. *Earth and Planetary Science Letters*, 293(3):259–268, 2010. doi: 10.1016/j.epsl.2010.02.033.
- Frances Elaine Jenner. Cumulate causes for the low contents of sulfide-loving elements in the continental crust. *Nature Geoscience*, 10(7):524–529, 2017.
- M. Jull and P. B. Kelemen. On the conditions for lower crustal convective instability. *Journal of Geophysical Research: Solid Earth*, 106(B4):6423–6446, 2001. ISSN 2156-2202. doi: 10.1029/2000JB900357.
- Robert Woodbury Kay and S Mahlburg Kay. Delamination and delamination magmatism. *Tectonophysics*, 219(1-3):177–189, 1993.
- Peter B Kelemen and Mark D Behn. Formation of lower continental crust by relamination of buoyant arc lavas and plutons. *Nature Geoscience*, 9(3):197–205, 2016.

- C Brenhin Keller, Blair Schoene, Melanie Barboni, Kyle M Samperton, and Jon M Husson. Volcanic–plutonic parity and the differentiation of the continental crust. *Nature*, 523(7560):301–307, 2015.
- Jun Korenaga. Plate tectonics, flood basalts and the evolution of earth’s oceans. *Terra Nova*, 20(6):419–439, 2008.
- EJ Krogh, CW Oh, and JC Liou. Polyphase and anticlockwise p-t evolution for franciscan eclogites and blueschists from jenner, california, usa. *Journal of Metamorphic Geology*, 12(2):121–134, 1994.
- IT Kukkonen, IV Golovanova, Yu V Khachay, VS Druzhinin, AM Kosarev, and VA Schapov. Low geothermal heat flow of the urals fold belt—implication of low heat production, fluid circulation or palaeoclimate? *Tectonophysics*, 276(1-4):63–85, 1997.
- Christof Kusebauch, Timm John, Jaime D Barnes, Andreas Klügel, and Håkon O Austrheim. Halogen element and stable chlorine isotope fractionation caused by fluid–rock interaction (bamble sector, se norway). *Journal of Petrology*, 56(2):299–324, 2015.
- N J Kusznir and R G Park. The extensional strength of the continental lithosphere: its dependence on geothermal gradient, and crustal composition and thickness. *Geological Society, London, Special Publications*, 28(1):35–52, 1987.
- s Labrosse. Core-mantle boundary, heat flow across. In D Gubbins and E Herrero-Bervera, editors, *Encyclopedia of Geomagnetism and Paleomagnetism*. Springer, Dordrecht, 2007. doi: https://doi.org/10.1007/978-1-4020-4423-6_51.

- Stéphane Labrosse. Hotspots, mantle plumes and core heat loss. *Earth and Planetary Science Letters*, 199(1-2):147–156, 2002.
- Stephane Labrosse, Jean-Paul Poirier, and Jean-Louis Le Mouél. The age of the inner core. *Earth and Planetary Science Letters*, 190(3-4):111–123, 2001.
- Gabi Laske, Guy Masters, Zhitu Ma, and Mike Pasyanos. Update on crust1.0 — a 1-degree global model of earth’s crust. In *Geophys. res. abstr*, volume 15, page 2658, 2013.
- Mary L Leech. Arrested orogenic development: Eclogitization, delamination, and tectonic collapse. *Earth and Planetary Science Letters*, 185(1):149–159, 2001. ISSN 0012-821X. doi: 10.1016/S0012-821X(00)00374-5.
- A. Levander, B. Schmandt, M. S. Miller, K. Liu, K. E. Karlstrom, R. S. Crow, C.-T. A. Lee, and E. D. Humphreys. Continuing Colorado plateau uplift by delamination-style convective lithospheric downwelling. *Nature*, 472(7344):461–465, 2011. ISSN 1476-4687. doi: 10.1038/nature10001.
- A Lima, S Albanese, and D Cicchella. Geochemical baselines for the radioelements k, u, and th in the campania region, italy: a comparison of stream-sediment geochemistry and gamma-ray surveys. *Applied Geochemistry*, 20(3):611–625, 2005.
- Fan-Chi Lin, Victor C Tsai, and Brandon Schmandt. 3-d crustal structure of the western united states: application of rayleigh-wave ellipticity extracted from noise cross-correlations. *Geophysical Journal International*, 198(2):656–670, 2014.

- Yong-Sheng Liu, Shan Gao, Shu-Yan Jin, Sheng-Hong Hu, Min Sun, Zu-Bin Zhao, and Jia-Lin Feng. Geochemistry of lower crustal xenoliths from neogene hannuoba basalt, north china craton: Implications for petrogenesis and lower crustal composition. *Geochimica et Cosmochimica Acta*, 65(15):2589–2604, 2001.
- Richard F. Livaccari. Role of crustal thickening and extensional collapse in the tectonic evolution of the Sevier-Laramide orogeny, western United States. *Geology*, 19(11):1104–1107, 1991. ISSN 0091-7613. doi: 10.1130/0091-7613(1991)019<1104:ROCTAE>2.3.CO;2.
- Bruno Lombardo and Franco Rolfo. Two contrasting eclogite types in the Himalayas: Implications for the Himalayan orogeny. *Journal of Geodynamics*, 30(1):37–60, 2000. ISSN 0264-3707. doi: 10.1016/S0264-3707(99)00026-5.
- F Lucazeau. Analysis and mapping of an updated terrestrial heat flow data set. *Geochemistry, Geophysics, Geosystems*, 20(8):4001–4024, 2019.
- Gordon A MacDonald and Takashi Katsura. Chemical composition of hawaiian lavas. *Journal of petrology*, 5(1):82–133, 1964.
- Jean-Claude Mareschal and Claude Jaupart. Radiogenic heat production, thermal regime and evolution of continental crust. *Tectonophysics*, 609:524–534, 2013.
- Gregor Markl, Masaaki Musashi, and Kurt Bucher. Chlorine stable isotope composition of granulites from lofoten, norway: Implications for the cl isotopic composition and for the source of cl enrichment in the lower crust. *Earth and Planetary Science Letters*, 150(1-2):95–102, 1997.

- H Martin. Effect of steeper archean geothermal gradient on geochemistry of subduction-zone magmas. *Geology*, 14(9):753–756, 1986.
- A McCarthy, TJ Falloon, I Sauermilch, JM Whittaker, K Niida, and DH Green. Revisiting the australian-antarctic ocean-continent transition zone using petrological and geophysical characterization of exhumed subcontinental mantle. *Geochemistry, Geophysics, Geosystems*, 21(7):e2020GC009040, 2020.
- W F McDonough and S-s Sun. The composition of the Earth. *Chemical Geology*, 120(3):223–253, 1995a. ISSN 0009-2541. doi: 10.1016/0009-2541(94)00140-4.
- William F McDonough. Constraints on the composition of the continental lithospheric mantle. *Earth and Planetary Science Letters*, 101(1):1–18, 1990.
- William F McDonough. Partial melting of subducted oceanic crust and isolation of its residual eclogitic lithology. *Philosophical Transactions of the Royal Society of London. Series A: Physical and Engineering Sciences*, 335(1638):407–418, 1991.
- William F McDonough and S-S Sun. The composition of the earth. *Chemical geology*, 120(3-4):223–253, 1995b.
- William F McDonough, Ondřej Šrámek, and Scott A Wipperfurth. Radiogenic power and geoneutrino luminosity of the earth and other terrestrial bodies through time. *Geochemistry, Geophysics, Geosystems*, 21(7):e2019GC008865, 2020.
- EH McKee. Tertiary igneous chronology of the great basin of western united states—implications for tectonic models. *Geological Society of America Bulletin*, 82(12):3497–3502, 1971.

- W D Mooney. *Crust and Lithospheric Structure—Global Crustal Structure*, volume 1. Oxford: Elsevier, 2015.
- Walter D Mooney, Gabi Laske, and T Guy Masters. CRUST 5.1: A global crustal model at $5^{\circ} \times 5^{\circ}$. *Journal of Geophysical Research: Solid Earth*, 103(B1):727–747, 1998a.
- Walter D Mooney, Gabi Laske, and T Guy Masters. CRUST 5.1: A global crustal model at $5^{\circ} \times 5^{\circ}$. *Journal of Geophysical Research: Solid Earth*, 103(B1):727–747, 1998b. ISSN 01480227. doi: 10.1029/97JB02122.
- Takashi Nakagawa and Paul J Tackley. Influence of initial cmb temperature and other parameters on the thermal evolution of earth’s core resulting from thermochemical spherical mantle convection. *Geochemistry, Geophysics, Geosystems*, 11(6), 2010.
- K Douglas Nelson, Wenjin Zhao, LD Brown, J Kuo, Jinkai Che, Xianwen Liu, SL Klemperer, Y Makovsky, RJJM Meissner, James Mechie, et al. Partially molten middle crust beneath southern tibet: synthesis of project indepth results. *Science*, 274(5293): 1684–1688, 1996.
- Andrew A Nyblade and Henry N Pollack. A global analysis of heat flow from precambrian terrains: implications for the thermal structure of archean and proterozoic lithosphere. *Journal of Geophysical Research: Solid Earth*, 98(B7):12207–12218, 1993.
- Tohru Ohta and Hiroyoshi Arai. Statistical empirical index of chemical weathering in igneous rocks: A new tool for evaluating the degree of weathering. *Chemical Geology*, 240(3-4):280–297, 2007.

TM Olugboji, V Lekic, and W McDonough. A statistical assessment of seismic models of the us continental crust using bayesian inversion of ambient noise surface wave dispersion data. *Tectonics*, 36(7):1232–1253, 2017.

Herbert Palme and H St C O’Neill. Cosmochemical estimates of mantle composition. In Heinrich D Holland and Karl K Turekian, editors, *Treatise on Geochemistry (Second Edition)*, volume 3, pages 1–39. Elsevier, Oxford, 2014. ISBN 978-0-08-098300-4. doi: 10.1016/B978-0-08-095975-7.00201-1.

Tom Parsons, Nikolas I. Christensen, and Howard G. Wilshire. Velocities of southern Basin and Range xenoliths: Insights on the nature of lower crustal reflectivity and composition. *Geology*, 23(2):129–132, 1995. ISSN 0091-7613. doi: 10.1130/0091-7613(1995)023(0129:VOSBAR)2.3.CO;2.

Michael E Pasyanos, T Guy Masters, Gabi Laske, and Zhitu Ma. LITHO1.0: An updated crust and lithospheric model of the Earth. *Journal of Geophysical Research: Solid Earth*, 119(3):2153–2173, 2014. ISSN 2169-9356. doi: 10.1002/2013JB010626.

Etta Patacca, Paolo Scandone, Emiliano Di Luzio, Gian Paolo Cavinato, and Maurizio Parotto. Structural architecture of the central apennines: Interpretation of the CROP 11 seismic profile from the adriatic coast to the orographic divide. *Tectonics*, 27(3), 2008. doi: <https://doi.org/10.1029/2005TC001917>.

C Pauselli, G Gola, P Mancinelli, E Trumpy, M Saccone, A Manzella, and Giorgio Ranalli. A new surface heat flow map of the northern apennines between latitudes 42.5

- and 44.5°N. *Geothermics*, 81:39–52, 2019. doi: <https://doi.org/10.1016/j.geothermics.2019.04.002>.
- Catherine Phaneuf and Jean-Claude Mareschal. Estimating concentrations of heat producing elements in the crust near the sudbury neutrino observatory, ontario, canada. *Tectonophysics*, 622:135–144, 2014.
- Christophe Pinet and Claude Jaupart. The vertical distribution of radiogenic heat production in the precambrian crust of norway and sweden: geothermal implications. *Geophysical Research Letters*, 14(3):260–263, 1987.
- Terry Plank and Donald W Forsyth. Thermal structure and melting conditions in the mantle beneath the basin and range province from seismology and petrology. *Geochemistry, Geophysics, Geosystems*, 17(4):1312–1338, 2016.
- A Polat, AW Hofmann, and Minik Thorleif Rosing. Boninite-like volcanic rocks in the 3.7–3.8 ga isua greenstone belt, west greenland: geochemical evidence for intra-oceanic subduction zone processes in the early earth. *Chemical geology*, 184(3-4): 231–254, 2002.
- Jason R Price and Michael A Velbel. Chemical weathering indices applied to weathering profiles developed on heterogeneous felsic metamorphic parent rocks. *Chemical geology*, 202(3-4):397–416, 2003.
- Daniele Regis, CJ Warren, Catherine Mary Mottram, and Nick MW Roberts. Using monazite and zircon petrochronology to constrain the P–T–t evolution of the middle crust in the bhutan himalaya. *Journal of Metamorphic Geology*, 34(6):617–639, 2016.

- Olivier Reubi and Jon Blundy. A dearth of intermediate melts at subduction zone volcanoes and the petrogenesis of arc andesites. *Nature*, 461(7268):1269–1273, 2009. ISSN 1476-4687. doi: 10.1038/nature08510.
- P. F. Rey, C. Teyssier, and D. L. Whitney. Extension rates, crustal melting, and core complex dynamics. *Geology*, 37(5):391–394, 2009. ISSN 0091-7613. doi: 10.1130/G25460A.1.
- Paul H Roberts, Christopher A Jones, and Arthur R Calderwood. Energy fluxes and ohmic dissipation in the earth’s core. *Earth’s core and lower mantle*, pages 100–129, 2003.
- Kaylynn M Rockow, Larry A Haskin, * BL Jollif, and David M Fountain. Constraints on element mobility associated with the conversion of granulite to eclogite along fractures in an anorthositic complex on holsnøy, norway. *Journal of Metamorphic Geology*, 15(3):401–418, 1997.
- NW Rogers and C J Hawkesworth. Proterozoic age and cumulate origin for granulite xenoliths, lesotho. *Nature*, 299(5882):409–413, 1982.
- Robert F Roy, David D Blackwell, and Francis Birch. Heat generation of plutonic rocks and continental heat flow provinces. *Earth and Planetary Science Letters*, 5:1–12, 1968.
- R. L. Rudnick and S. R. Taylor. The composition and petrogenesis of the lower crust: A xenolith study. *Journal of Geophysical Research: Solid Earth*, 92(B13):13981–14005, 1987a. ISSN 01480227. doi: 10.1029/JB092iB13p13981.
- Roberta L Rudnick. Making continental crust. *Nature*, 378(6557):571–578, 1995.

Roberta L Rudnick and David M Fountain. Nature and composition of the continental crust: A lower crustal perspective. *Reviews of Geophysics*, 33(3):267–309, 1995. ISSN 1944-9208. doi: 10.1029/95RG01302.

Roberta L Rudnick and S Gao. Composition of the Continental Crust. In *Treatise on Geochemistry*, pages 1–51. Elsevier, 2014. ISBN 978-0-08-098300-4. doi: 10.1016/B978-0-08-095975-7.00301-6.

Roberta L Rudnick and Ian Jackson. Measured and calculated elastic wave speeds in partially equilibrated mafic granulite xenoliths: Implications for the properties of an underplated lower continental crust. *Journal of Geophysical Research: Solid Earth*, 100(B6):10211–10218, 1995.

Roberta L Rudnick and Andrew A Nyblade. The thickness and heat production of archean lithosphere: constraints from xenolith thermobarometry and surface heat flow. *Mantle petrology: field observations and high pressure experimentation: a tribute to Francis R.(Joe) Boyd*, 6:3–12, 1999.

Roberta L Rudnick and T Presper. Geochemistry of intermediate/-to high-pressure granulites. In *Granulites and crustal evolution*, pages 523–550. Springer, 1990.

Roberta L Rudnick and S R Taylor. The composition and petrogenesis of the lower crust: a xenolith study. *Journal of Geophysical Research: Solid Earth*, 92(B13):13981–14005, 1987b.

Roberta L Rudnick, S M McLennan, and S R Taylor. Large ion lithophile elements in

- rocks from high-pressure granulite facies terrains. *Geochimica et Cosmochimica Acta*, 49(7):1645–1655, 1985.
- Roberta L Rudnick, Matthias Barth, Ingo Horn, and William F McDonough. Rutile-bearing refractory eclogites: missing link between continents and depleted mantle. *Science*, 287(5451):278–281, 2000.
- James K Russell, Lucy A Porritt, Yan Lavallée, and Donald B Dingwell. Kimberlite ascent by assimilation-fuelled buoyancy. *Nature*, 481(7381):352–356, 2012.
- T Sakai, K Inoue, H Watanabejg, WF McDonough, N Abe, E Araki, T Kasaya, M Kyo, N Sakurai, K Uek, et al. Study of ocean bottom detector for observation of geo-neutrino from the mantle. In *Journal of Physics: Conference Series*, volume 2156, page 012144. IOP Publishing, 2021.
- T Sakai, K Inoue, H Watanabe, W F McDonough, N Abe, E Araki, T Kasaya, M Kyo, N Sakurai, K Ueki, and H Yoshida. Study of Ocean Bottom Detector for observation of geo-neutrino from the mantle. *Proceedings for TAUP*, 2022.
- L G Sammon, C Gao, and W F McDonough. Lower crustal composition in the southwestern united states. *Journal of Geophysical Research: Solid Earth*, 125(3): e2019JB019011, 2020.
- Laura G Sammon and William F McDonough. A geochemical review of amphibolite, granulite, and eclogite facies lithologies: Perspectives on the deep continental crust. *Journal of Geophysical Research: Solid Earth*, page e2021JB022791, 2021a.

Laura G Sammon and William F McDonough. A geochemical review of amphibolite, granulite, and eclogite facies lithologies: Perspectives on the deep continental crust. *Journal of Geophysical Research: Solid Earth*, 2021b.

Laura G Sammon, William F McDonough, and Walter D Mooney. The composition of the deep continental crust inferred from geochemical and geophysical data. *Journal of Geophysical Research: Solid Earth*, page *in review*, 2022. doi: <https://doi.org/10.1002/essoar.10507582.1>.

P. Schaaf, W. Heinrich, and T. Besch. Composition and Sm-Nd isotopic data of the lower crust beneath San Luis Potosí, central Mexico: Evidence from a granulite-facies xenolith suite. *Chemical Geology*, 118(1):63–84, 1994. ISSN 0009-2541. doi: 10.1016/0009-2541(94)90170-8.

FR Schilling and GM Partzsch. Quantifying partial melt fraction in the crust beneath the central andes and the tibetan plateau. *Physics and Chemistry of the Earth, Part A: Solid Earth and Geodesy*, 26(4-5):239–246, 2001.

Brandon Schmandt, Fan-Chi Lin, and Karl E Karlstrom. Distinct crustal isostasy trends east and west of the rocky mountain front. *Geophysical Research Letters*, 42(23):10–290, 2015.

M Schmitz, W-D Heinsohn, and FR Schilling. Seismic, gravity and petrological evidence for partial melt beneath the thickened central andean crust (21–23 s). *Tectonophysics*, 270(3-4):313–326, 1997.

Vera Schulte-Pelkum, Kevin H Mahan, Weisen Shen, and Josh C Stachnik. The distri-

- bution and composition of high-velocity lower crust across the continental us: Comparison of seismic and xenolith data and implications for lithospheric dynamics and history. *Tectonics*, 36(8):1455–1496, 2017.
- Derek L. Schutt, Anthony R. Lowry, and Janine S. Buehler. Moho temperature and mobility of lower crust in the western United States. *Geology*, 46(3):219–222, 2018. ISSN 0091-7613. doi: 10.1130/G39507.1.
- MP Searle, JM Cottle, MJ Streule, and DJ Waters. Crustal melt granites and migmatites along the himalaya: melt source, segregation, transport and granite emplacement mechanisms. *Earth and Environmental Science Transactions of the Royal Society of Edinburgh*, 100(1-2):219–233, 2009.
- J Semprich and N S C Simon. Inhibited eclogitization and consequences for geophysical rock properties and delamination models: Constraints from cratonic lower crustal xenoliths. *Gondwana Research*, 25(2):668–684, 2014. ISSN 1342-937X. doi: 10.1016/j.gr.2012.08.018.
- Weisen Shen and Michael H. Ritzwoller. Crustal and uppermost mantle structure beneath the United States. *Journal of Geophysical Research: Solid Earth*, 121(6):4306–4342, 2016. ISSN 2169-9356. doi: 10.1002/2016JB012887.
- Weisen Shen, Michael H. Ritzwoller, and Vera Schulte-Pelkum. A 3-D model of the crust and uppermost mantle beneath the Central and Western US by joint inversion of receiver functions and surface wave dispersion. *Journal of Geophysical Research: Solid Earth*, 118(1):262–276, 2013. ISSN 2169-9356. doi: 10.1029/2012JB009602.

Weisen Shen, Douglas A Wiens, Andrew J Lloyd, and Andrew A Nyblade. A geothermal heat flux map of antarctica empirically constrained by seismic structure. *Geophysical Research Letters*, 47(14):e2020GL086955, 2020.

William J Shinevar, Mark D Behn, Greg Hirth, and Oliver Jagoutz. Inferring crustal viscosity from seismic velocity: Application to the lower crust of Southern California. *Earth and Planetary Science Letters*, 494:83–91, 2018. ISSN 0012-821X. doi: 10.1016/j.epsl.2018.04.055.

G Paolo Sighinolfi. Investigations into deep crustal levels: fractionating effects and geochemical trends related to high-grade metamorphism. *Geochimica et Cosmochimica Acta*, 35(10):1005–1021, 1971.

Ondřej Šrámek, William F. McDonough, Edwin S. Kite, Vedran Lekić, Stephen T. Dye, and Shijie Zhong. Geophysical and geochemical constraints on geoneutrino fluxes from Earth’s mantle. *Earth and Planetary Science Letters*, 361:356–366, 2013. doi: 10.1016/j.epsl.2012.11.001.

H-G Stosch, G W Lugmair, and H A Seck. Geochemistry of granulite-facies lower crustal xenoliths: implications for the geological history of the lower continental crust below the eifel, west germany. *Geological Society, London, Special Publications*, 24(1):309–317, 1986.

Virginia Strati, Marica Baldoncini, Ivan Callegari, Fabio Mantovani, William F McDonough, Barbara Ricci, and Gerti Xhixha. Expected geoneutrino signal at junco. *Progress in Earth and Planetary Science*, 2(1):1–7, 2015.

- S-s Sun and William F McDonough. Chemical and isotopic systematics of oceanic basalts: Implications for mantle composition and processes. *Geological Society, London, Special Publications*, 42(1):313–345, 1989. ISSN 0305-8719, 2041-4927. doi: 10.1144/GSL.SP.1989.042.01.19.
- H Svensen, Bjorn Jamtveit, D Austrheim Banks, and Håkon Austrheim. Halogen contents of eclogite facies fluid inclusions and minerals: Caledonides, western norway. *Journal of Metamorphic Geology*, 19(2):165–178, 2001.
- Wolfgang Szwillus, Juan Carlos Afonso, Jörg Ebbing, and Walter D Mooney. Global crustal thickness and velocity structure from geostatistical analysis of seismic data. *Journal of Geophysical Research: Solid Earth*, 124(2):1626–1652, 2019.
- Ming Tang, William F McDonough, and Richard D Ash. Europium and strontium anomalies in the morb source mantle. *Geochimica et Cosmochimica Acta*, 197:132–141, 2017.
- Stuart Ross Taylor and Scott M McLennan. *The continental crust: its composition and evolution*. Blackwell Scientific Pub., Palo Alto, CA, 1985.
- Tatsuki Tsujimori, Virginia B Sisson, Juhn G Liou, George E Harlow, and Sorena S Sorensen. Very-low-temperature record of the subduction process: A review of world-wide lawsonite eclogites. *Lithos*, 92(3-4):609–624, 2006.
- Donald Turcotte and Gerald Schubert. *Geodynamics*. Cambridge University Press, 2014.
- G. Nell Tyner and Douglas Smith. Peridotite xenoliths in silica-rich, potassic latite from the transition zone of the Colorado Plateau in north-central Arizona. *Contributions*

to Mineralogy and Petrology, 94(1):63–71, 1986. ISSN 1432-0967. doi: 10.1007/BF00371227.

Greg A. Valentine and Frank V. Perry. Tectonically controlled, time-predictable basaltic volcanism from a lithospheric mantle source (central Basin and Range Province, USA). *Earth and Planetary Science Letters*, 261(1):201–216, 2007. ISSN 0012-821X. doi: 10.1016/j.epsl.2007.06.029.

PWC Van Calsteren, NBW Harris, CJ Hawkesworth, MA Menzies, and NW Rogers. Xenoliths from southern africa: a perspective on the lower crust. *Geological Society, London, Special Publications*, 24(1):351–362, 1986.

Massimo Verdoya, Paolo Chiozzi, and Gianluca Gola. Unravelling the terrestrial heat flow of a young orogen: The example of the northern Apennines. *Geothermics*, 90: 101993, 2021. doi: <https://doi.org/10.1016/j.geothermics.2020.101993>.

G Vignaroli, M Mancini, F Bucci, M Cardinali, GP Cavinato, M Moscatelli, ML Putignano, P Sirianni, M Santangelo, F Ardizzone, et al. Geology of the central part of the amatrice basin (central apennines, italy). *Journal of Maps*, 15(2):193–202, 2019.

K Hans Wedepohl. The composition of the continental crust. *Geochimica et cosmochimica Acta*, 59(7):1217–1232, 1995.

Bruce H Wilkinson, Brandon J McElroy, Stephen E Kesler, Shanan E Peters, and Edward D Rothman. Global geologic maps are tectonic speedometers—rates of rock cycling from area-age frequencies. *Geological Society of America Bulletin*, 121(5-6): 760–779, 2009.

Scott A Wipperfurth, Meng Guo, Ondřej Šrámek, and William F McDonough. Earth's chondritic Th/U: Negligible fractionation during accretion, core formation, and crust–mantle differentiation. *Earth and Planetary Science Letters*, 498:196–202, 2018. ISSN 0012-821X. doi: 10.1016/j.epsl.2018.06.029.

Scott A Wipperfurth, Ondřej Šrámek, and William F McDonough. Reference models for lithospheric geoneutrino signal. *Journal of Geophysical Research: Solid Earth*, 125(2):e2019JB018433, 2020.

M Xhixha, Marica Baldoncini, GP Bezzon, GP Buso, L Carmignani, L Casini, I Callegari, T Colonna, S Cuccuru, E Guastaldi, et al. A detailed gamma-ray survey for estimating the radiogenic power of sardinian variscan crust. 2014.

Toru Yamasaki. The role of bimodal magmatism in seafloor massive sulfide (sms) ore-forming systems at the middle okinawa trough, japan. *Ocean Science Journal*, 53(2): 413–436, 2018.

Takashi Yoshizaki and William F McDonough. Earth and mars–distinct inner solar system products. *Geochemistry*, 81(2):125746, 2021.

E-An Zen. Phase relations of peraluminous granitic rocks and their petrogenetic implications. *Annual Review of Earth and Planetary Sciences*, 16(1):21–51, 1988.