

The Dynamic Crustal and Mantle Structure Around the Afar Triple Junction



Miriam Gauntlett
Linacre College
University of Oxford

A thesis submitted for the degree of
Doctor of Philosophy
Trinity 2024

*For my grandmothers:
Judith Gauntlett and
Pwu Yih Dowker*

Abstract

The Afar Triple Junction marks the confluence of the Red Sea rift, the Gulf of Aden and the East African Rift System. The Afro-Arabian Rift System (AARS) encompasses these three rifts, each in a different stage of continental breakup. I investigate the crust and mantle beneath this complex tectonic region using three different seismological techniques.

I explore mantle flow using measurements of seismic anisotropy, and show that mantle flow is coherent across the three branches of the AARS. Shear-wave splitting of the SKS phase reveals patterns of density-driven flow in the upper mantle beneath the Afar Triple Junction. Fast shear-wave polarisation directions in northeastern Afar are directed NE-SW, coherent with observations further south. Rotation of the fast direction in southwestern Yemen suggests that the flow then bifurcates around the thick, cold lithosphere of Arabia and is channeled along both the Gulf of Aden and Red Sea spreading centres. The magnitude of splitting is smaller in the Gulf of Aden, indicating that mantle flow is less persistent beneath this branch of the AARS.

I interrogate crustal thickness and structure by calculating teleseismic receiver functions across northeastern Afar and the Danakil microplate. I use the H- κ stacking method to estimate the Moho depth and bulk crustal V_P/V_S ratio. My results reveal a heterogeneous crust with variable crustal thickness, and regions of partial melt indicated by anomalously high V_P/V_S ratios. These variations are on a shorter length scale than the mantle flow patterns, which suggests the influence of pre-existing lithospheric structures on crustal properties. By investigating the relationship between crustal thickness and topographic elevation, I estimate that the regional crustal bulk density is elevated compared to the global average. I show that topography is higher than would be expected due to crustal isostasy alone, implying that dynamic support from the mantle contributes to the topography in the region.

Extensional processes in the region have facilitated magmatism throughout the crust, producing volcanic episodes such as the 2011 eruption of Nabro volcano in Eritrea. I use local earthquake traveltime tomography to invert for the seismic velocity structure (V_P , V_S , and V_P/V_S) beneath Nabro. Key observations are an aseismic region of low V_P , low V_S and high V_P/V_S ratio at depths of 6–10 km b.s.l., interpreted as the primary melt storage region that fed the eruption; a zone of high V_S , low V_P and low V_P/V_S ratio, representing an intrusive complex of fractured rocks partially-saturated with over-pressurised gases.

By conducting my investigation on three different scales (upper crust, bulk crust, and upper mantle), I provide critical insights into the dynamic extensional processes driving the evolution of the Afar Triple Junction. Whilst rifting is driven by the mantle, pre-existing structures in the crust control crustal properties and the distribution of magmatism. Lithospheric thickness has dictated the pattern of continental breakup in the region. The thick, cold, dense lithosphere of the Arabian Platform acts as a backstop to density-driven flow in the mantle as it moves north of the Afar Triple Junction. Rifting progresses along thin lithospheric corridors beneath the Gulf of Aden and the Red Sea rift. The impingement of mantle flow on the Arabian Platform helps drive the plate motion of Arabia, and thus the mantle exerts a strong control on regional kinematics across the AARS.

Statement of Authorship

Chapter 4 has been published by the Journal of Geophysical Research: Solid Earth under the title:

Gauntlett, M., Hudson, T., Kendall, J.-M., Rawlinson, N., Blundy, J., Lapins, S., Goitom, B., Hammond, J., Oppenheimer, C., and Ogubazghi, G. (2023). Seismic tomography of Nabro caldera, Eritrea: Insights into the magmatic and hydrothermal systems of a recently erupted volcano. *Journal of Geophysical Research: Solid Earth* (p. e2022JB025742). <http://doi.org/10.1029/2022JB025742>

Miriam Gauntlett conducted the seismic tomography analysis, came up with interpretations for the results and authored the paper and figures. Nick Rawlinson provided the FMTOMO code for the seismic tomography inversion. Sacha Lapins produced the machine learning code that produced the seismic catalogue which was subsequently modified for seismic tomography by Miriam Gauntlett. Clive Oppenheimer was the PI for the UKRI grant “Mechanisms and implications of the 2011 eruption of Nabro volcano, Eritrea” and, along with James Hammond, Ghebrebrhan Ogubazghi and Berhe Goitom, conducted the fieldwork at Nabro. Mike Kendall and Tom Hudson supervised the work. Jon Blundy provided suggestions on interpretation from a volcanological perspective. All authors read and provided feedback on the finalised manuscript.

Chapter 3 has been published by Geochemistry, Geophysics, Geosystems under the title:

Gauntlett, M., Stephenson, S. N., Kendall, J.-M., Ogden, C., Hammond, J. O. S., Hudson, T., Goitom, B., Ogubazghi, G. (2024). The dynamic crust of northern Afar and adjacent rift margins: new evidence from receiver function analysis in Eritrea and Ethiopia. *Geochemistry, Geophysics, Geosystems*, 25(6), e2023GC011314. <https://doi.org/10.1029/2023GC011314>

Miriam Gauntlett conducted the receiver function analysis, came up with interpretations for the results, modified the original code for receiver function calculation

(provided by Chris Ogden) and authored the paper and figures. Simon Stephenson carried out the dynamic topography analysis and produced Figure 3.3. James Hammond, Ghebrebrhan Ogubazghi and Berhe Goitom conducted the fieldwork in Eritrea. Mike Kendall and Thomas Hudson supervised the work. All authors read and provided feedback on the finalised manuscript.

Chapter 2 has been submitted to Geology under the title:

Gauntlett, M., Kendall, J-M., Hammond, J. O. S., Hudson, T. (2024). Channelized mantle flow through the Afar Triple Junction and around the Arabian plate: evidence from seismic anisotropy. *Geology*.

Miriam Gauntlett conducted the shear-wave splitting analysis, came up with interpretations for the results and authored the paper and figures. Mike Kendall and Thomas Hudson supervised the work. All authors assisted with the genesis of the project and gave interpretations and feedback on the finalised manuscript.

Acknowledgements

“Marx was right: The smallest indivisible human unit is two people, not one; one is a fiction. From such nets of souls societies, the social world, human life springs. And also plays,” wrote Tony Kushner in his afterword to *Angels in America*. I would add: and also PhDs. At times when I was particularly daunted by the difficulty of writing my thesis, I would start to draft these acknowledgements in my head. It was a reminder that no matter how isolating the task might seem, I have been surrounded by so much love, care and support over the last four years, without which I would never have been able to finish my PhD.

First thanks go to my supervisors, Mike Kendall and Tom Hudson, for their unwavering support and guidance. I am very grateful for the care and time you have both taken to mentor me, and for all that I have learned from you over the past four years. Particular thank you to Mike for taking me on as a student right after your arrival at Oxford and for constantly coming up with new and exciting ideas for me to work on. Thank you also to James Hammond; without your hard work putting out seismometers across Eritrea and wealth of knowledge about the Afar region, this thesis would not have been possible.

Thank you to my collaborators: Nick Rawlison, who was eternally patient when teaching me the ins and outs of FMTOMO; Chris Ogden, who painstakingly guided me through the receiver function calculation; Simon Stephenson, who shared his expertise on dynamic topography and spent a lot of time helping me with that chapter; Sacha Lapins, for his pivotal work on Nabro and machine learning techniques, and for always taking the time to meet; Jon Blundy, for helpful, enthusiastic conversations on volcanology and interpretations of the Nabro seismic tomography. Thank you to Tim Greenfield for mentoring me during the summer of 2019, and particularly for teaching me the ins and outs of GMT. I am grateful to co-authors Clive Oppenheimer and Ghebrebrhan Ogubazghi, for reading through my manuscripts, and to Lara Wagner and Helen Janiszewski for guidance and help. This work was funded by a UKRI Natural Environment Research Council studentship, which I am grateful to have received.

All my gratitude to Matthew Kemp, whose companionship throughout the past four years has been a godsend. I don’t know how I would have made it through without our regular lunches to commiserate and, more importantly, discuss

everything we've read, watched and listened to recently. Thank you to Alice Turner, Rita Kounoudis and Rebecca Colquhoun for support, help and inspiration.

Thank you to my friends for sharing space with me, in person and online, for making me laugh, for keeping me company, for loving me, for teaching me how to live. Thank you to ATEEZ and BTS for reminding me that life is better when shared and for joyfully soundtracking the last four years. Particular thank you to:

My second family of sisters: Saskia Borchardt-Hume, Lydia Day, Dorothy Hoskins, Alex Rowe, and Holly Hamilton; to be known and cared for by you is a blessing. Over the past nine years you have shown me that friendship can surpass all my expectations. Arenike Adebajo, for your wisdom, infinite care and compassion, which has made me braver than I ever thought possible. Thank you for growing with me; I'm proud of the people we've become.

Iman Alshahib, for the corona-oke sessions, for making me laugh until I cry, for the four hour long FaceTimes, for always having the right answer, for making me feel so safe and loved over the last 12 years. Huda Abdirahman, for all the years of being absolutely unhinged together but particularly the last one, for your insightful, considerate care, and for being the funniest person I know. Lizzie Bell, for being such a steadfast presence in my life.

Zoë Bourque, whose friendship came along at a time I desperately needed it. Writing with you has been one of the most joyful and fulfilling experiences imaginable. Thank you for your curiosity, your enthusiasm in all things and for driving me to and from the airport while blasting ATEEZ. Lee Byrne, for your quiet kindness, warm honesty and enthusiasm for the outdoors. Agnes Cameron, thank you for poems, for teaching me about putting problems in the fridge, for holding my hand over the table, for seeing me and understanding.

Marisa Clements, for the unconditional love, support and care, for never letting me feel alone, for making me feel safe. Eddie Conway, for our years-long back and forth messages, for all our exchanged book thoughts and for always asking what I'm reading. Alfie Coates, your thoughtfulness, openness and bravery have been such a balm over the years. I am so grateful to have you in my life. Lucy Foreman, for being the first person to show me the joys of a best friendship.

Surina Fordington, I'm so lucky we met on the first day of university. There aren't words for what your friendship has meant to me over the years, but I know for certain it's made me a kinder, more compassionate person. Thank you for all the miles we've walked together. Micha Frazer-Carroll, for your warm intelligence and for always making me feel safe and understood. Kallie Gregg, I'll be with you til the end of the line. Alice Hawkins, for your warmth and openness and for a friendship that has always felt easy; I can't wait to see you down under soon! Akua Henaku, for always making me laugh, keeping me updated on our boys, and for

understanding exactly how I feel about Song Mingi. Julian, for the Nona proof and for all your support. Em Kaiser, for your generous heart and beautiful art.

Abeera Khan, who was there throughout with whatever I needed: reassurance, mutual rage, a shoulder to cry on, a delicious meal. I will always remember the home we built together at 18 Moreton Close. Jocelyne Laflamme, for being the older sister I never had, for our beautiful walks in the mountains, for taking me under your wing. Sarah Lasoye, for being there for me during the hardest part, for all the crazy trips abroad to see concerts. Rox Middleton, for being so smart and thoughtful, for your wise emails and for the words “we have the rest of our lives”.

Martha Mitchell, for laughter and weird noises and long voicenotes about books we hate and being so full of love. Nandini Mitra, for the biggest tightest hugs, for the warmth and ease of your friendship, for how well you listen. Elete Nelson-Fearon, my not-twin twin, co-president of the 7th May 1997 club, what would I do without you? Life with you and your big laugh and your ceaseless support and the way you wordlessly understand me: all of it is a joy and full of love.

Jun Pang, for every time you planned something for me, for travelling across the city to be there, for always picking up the phone. Your fierce capacity for love has kept me going. Hana Pereira, meeting you turned my luck around. Thank you for making me hopeful again, for holding me tight, for running down the road to meet me. I love you and I am so lucky to love you! Lola Olufemi, for expansive horizons, for believing in better, for refusing to let go, for telling me I can demand things. Shahla Omar, for your companionship in very dark times, for your gentle compassion, for trusting me.

Audrey Sebatindira, whose friendship has sharpened my political convictions and shaped my world in so many ways. Long may our Marxist theory/gay trash TV club of two continue. Ygraine Siemens, thank you for weekly updates, for the dyke-to-dyke communication and understanding, for being so wise and gentle. Sakina Shahid, for your steadfast loyalty, for the best advice (and for loving me even when I don’t take it), for the podcast-length voicenotes, for your big warm heart.

Claire Sosienski Smith, my right arm, my rock, my companion in the void, my life partner. Emailing you every day (give or take) for the last 4.5 years has taught me so much about vulnerability and trust and unconditional love. Thank you for making life feel possible and beautiful. Stella Swain, for your indefatigable commitment to making the world a better place and for doing it with such warmth and care. Siyang Wei, for the personal and political solidarity, for all the Chinese food we’ve eaten and cooked, for understanding the deepest recesses of my depraved mind. Chloe Wong, thank you for all the climbs and battles!

Hayley Wu, for whom I will always know what time it is in Hong Kong. Thank you for being my safety zone, for being so curious and brave and wise. And to Iman

Alshahib, Kallie Gregg, Holly Ganderton, Nicola Jackson, Chelsi Holt, and Asyikin Yusoff, who have been here since the beginning. I am so lucky I met you all.

Finally, thank you to my family near and far, particularly my parents and my brother, who raised me with a sharp curiosity about the world and always encouraged me to pursue it. It means the world to me that I always have your support behind me.

Contents

List of Figures	xv
List of Tables	xviii
List of Abbreviations	xix
List of Symbols	xx
1 Introduction	1
1.1 Plate tectonics and continental rifting	2
1.2 Afro-Arabian Rift System	6
1.2.1 Arabian Plate	8
1.2.2 Gulf of Aden	8
1.2.3 Red Sea	9
1.2.4 Afar	11
1.2.5 Main Ethiopian Rift	15
1.3 The African Large Low Shear Velocity Province	16
1.4 Magmatism in the Afro-Arabian Rift System	19
1.4.1 Recent volcanic activity in Afar	23
1.4.2 The Nabro Volcanic Range and Nabro volcano	24
1.5 Seismic methods	27
1.5.1 Shear-wave splitting	28
1.5.2 Receiver functions	33
1.5.3 Seismic body-wave velocity tomography	38
1.6 Thesis overview	43
2 Channelized mantle flow through the Afar Triple Junction and around the Arabian plate	46
2.1 Introduction	47
2.2 Data and methodology	48
2.3 Results	50
2.4 Discussion	51
2.4.1 Asthenospheric flow in northeastern Afar	51

2.4.2	Bifurcation of asthenospheric flow around the Arabian Plate	51
2.5	Conclusion	55
3	The dynamic crust of northern Afar and adjacent rift margins: evidence from receiver functions	57
3.1	Introduction	59
3.2	Data and Methodology	63
3.2.1	Seismic Network	63
3.2.2	Teleseismic Dataset	64
3.2.3	Methods	66
3.3	Results	69
3.4	Discussion	73
3.4.1	Stretching of the Danakil block	73
3.4.2	Crustal sills beneath Nabro volcano	74
3.4.3	Implications for rifting through the Gulf of Zula	75
3.4.4	Topographic elevation and crustal structure	76
3.5	Conclusion	83
4	Seismic tomography of Nabro caldera	87
4.1	Introduction	89
4.2	Data	92
4.2.1	Network and Data Collection	92
4.2.2	Data Selection	94
4.3	Methodology	96
4.3.1	Tomographic Method	96
4.3.2	1D Model Selection	100
4.3.3	Defining a Grid for 3D Inversions	102
4.3.4	Optimising Data-Model Fit	102
4.3.5	Earthquake Relocations	103
4.4	Results	104
4.4.1	Checkerboard Resolution Tests	104
4.4.2	Inverting for 3D Velocity Structure	106
4.5	Discussion	110
4.5.1	Defining a Local 1D Velocity Model at Nabro	110
4.5.2	Earthquake Detection Using U-GPD	113
4.5.3	Interpretation of V_P/V_S Variations	113
4.5.4	The Crustal Substructure Beneath Nabro Caldera	119
4.5.5	Comparisons with Other Volcanoes	120
4.5.6	Limitations and Future Work	121
4.6	Conclusion	122

5	Conclusion	125
5.1	Summary	126
5.1.1	Dynamic mantle flow constrained by seismic anisotropy . . .	126
5.1.2	Heterogeneous crustal structure in northern Afar and dynamic support from the mantle	127
5.1.3	Magma storage and fluid transport through an intrusive complex beneath Nabro volcano	128
5.2	Future work	129
5.2.1	Further investigation of Nabro volcano and the Nabro Volcanic Range	129
5.2.2	Other seismic tomography studies in the AARS	131
5.2.3	Improving understanding of the crustal structure beneath northern Afar	132
5.2.4	Regional mantle dynamics	133
5.3	Final remarks	134

Appendices

A	Supplementary information for “Channelized mantle flow through the Afar Triple Junction and around the Arabian plate”	137
A.1	Introduction	137
B	Supplementary information for “The dynamic crust of northern Afar and adjacent rift margins: evidence from receiver functions”	143
B.1	Introduction	143
B.2	Text S1: H- κ Stacking Method	144
B.3	Text S2: H- κ Stacking Result Quality Control and Uncertainty Estimates	145
B.4	Text S3: Result Quality Control–Eritrea Seismic Project Stations .	147
B.5	Text S4: Result Quality Control–AFAR0911 Stations	148
B.6	Text S5: Isostatic calculations	149
C	Supplementary information for “Seismic tomography of Nabro caldera”	173
C.1	Introduction	173
C.2	Text S1: Nonlinear Earthquake Relocation	174
C.3	Text S2: Formulation of the V_P/V_S Inversion	175
C.4	Text S3: Perturbation of Derived 1D Starting Model	176
C.5	Text S4: Finding Regularisation Parameters	176
C.6	Text S5: Synthetic Resolution Tests	178

D Differential Ps times from receiver functions	189
D.1 Introduction	189
D.2 Plotting differential Ps times at Nabro stations	190
Bibliography	193

List of Figures

1.1	Tectonic versus magmatic extension	5
1.2	Topographic map of the Afro-Arabian Rift System	7
1.3	Bathymetric and topographic map of the Gulf of Aden area	9
1.4	Dated oceanic magnetic anomalies	10
1.5	Topographic map of Afar and the Danakil microplate	12
1.6	Whole mantle cross-section centred on the AARS	17
1.7	Cenozoic magmatic activity in the AARS	21
1.8	Shear-wave splitting occurs for a shear wave traveling through an anisotropic medium	28
1.9	Suggested anisotropy mechanisms of the Ethiopian rift mantle	31
1.10	P-to-S converted phases and reverberations	34
1.11	The impact of crustal structure on receiver functions	36
1.12	Moho depth and V_P/V_S ratios across Afar and Ethiopia	37
1.13	Earthquake raypaths used for a hypothetical tomographic inversion	39
1.14	The Aluto seismic tomography model for V_P/V_S ratio.	43
2.1	Topographic map and seismic networks	49
2.2	Map of the SKS splitting results	52
2.3	Map of the SKS splitting parameters plotted over a tomography model	54
3.1	Topographic map of Afar	65
3.2	Regional map showing the crustal thickness and V_P/V_S ratio results	72
3.3	Crustal and mantle structure	84
4.1	The final event catalog after relocation in a joint inversion for velocity structure and earthquake location	95
4.2	The V_P and V_S velocity model output from a quasi-1D inversion	101
4.3	Input V_P/V_S checkerboard models of differing scale lengths	107
4.4	Output V_P/V_S checkerboard models of differing scale lengths	108
4.5	Latitude and longitude cross-sections through the final solution models	111
4.6	Cross sections at depths of 1km, 3 km, 5 km and 7 km below sea level (b.s.l.) through the final solution models	112

4.7 The V_P/V_S , V_P and V_S solution models plotted along the 41.7°E cross-section beneath Nabro	119
A.1 Shear-wave splitting parameters as a function of source backazimuths	138
A.2 SKS splitting analysis for a good event	139
A.3 SKS splitting analysis for a poor result	140
A.4 SKS splitting analysis for a null event	141
B.1 H- κ result for station ASSE	151
B.2 H- κ result for station CAYE	152
B.3 H- κ result for station DOLE	153
B.4 H- κ result for station EITE	154
B.5 H- κ result for station EITE (E)	155
B.6 H- κ result for station EITE (W)	156
B.7 H- κ result for station FAME	157
B.8 H- κ result for composite station NAB	158
B.9 H- κ result for station LIDE	159
B.10 H- κ result for station TIOE	160
B.11 H- κ result for station ABAE	161
B.12 H- κ result for station BTIE	162
B.13 H- κ result for station FINE	163
B.14 H- κ result for station HITE	164
B.15 H- κ result for station IGRE	165
B.16 H- κ result for station KOZE	166
B.17 H- κ result for station LULE	167
B.18 H- κ result for station MAYE	168
B.19 H- κ result for station SAHE	169
B.20 Estimates of V_P/V_S ratio (κ) plotted against crustal thickness (H) .	171
B.21 Parameter sweeps for different asthenospheric channel and initial lithospheric thickness values	172
C.1 Cross-sections through solution models using perturbed starting models	178
C.2 Trade-off curves used to find the optimum values of the regularisation parameters	179
C.3 The event relocations following six iterations of the FMTOMO inversion procedure plotted over a topographic map	180
C.4 Input V_P checkerboard models of differing scale lengths	181
C.5 Output V_P checkerboard models of differing scale lengths	182
C.6 Input V_S checkerboard models of differing scale lengths	183
C.7 Output V_S checkerboard models of differing scale lengths	184

C.8	Cross-sections through the final V_P , V_S and V_P/V_S solution models	185
C.9	Cross-sections through the V_P/V_S model obtained through direct division	185
C.10	The workflow followed in this paper	186
C.11	Source mislocations during the synthetic checkerboard tests	187
D.1	Composite plot of receiver functions at each Nabro station	191
D.2	Plot of differential P_s times around Nabro volcano	192

List of Tables

1.1	The Wilson Cycle	3
3.1	H- κ and V_P Results	71
4.1	Inversion Statistics for the Final V_P , V_S and V_P/V_S Inversions . . .	106
A.1	Shear-wave Splitting Stacked Results.	142
B.1	H- κ Results	170
C.1	Inversion statistics for the V_P/V_S inversions using the perturbed 1D models as starting models, as compared to the original model. . . .	187
C.2	Inversion Statistics for the Synthetic V_P , V_S and V_P/V_S Checker- board Tests	188

List of Abbreviations

AARS		Afro-Arabian Rift System
a.s.l.		Above sea level
b.s.l.		Below sea level
EARS		East African Rift System
FMTOMO	. .	Fast-Marching TOMOgraphy
GOA		Gulf of Aden
LLSVP		Large Low Shear Velocity Province
MER		Main Ethiopian Rift
Ma		Million years ago
NVR		Nabro Volcanic Range
RSR		Red Sea Rift
USGS		United States Geological Survey

List of Symbols

P	Primary/pressure/compressional seismic phase
V_P	P-wave velocity
S	Secondary/shear seismic phase
V_S	S-wave velocity
V_P/V_S	Ratio of P- to S-wave velocity
SKS	Seismic phase travelling as an S-wave from source to core, a P-wave through the outer core, and an S-wave from core to receiver
ϕ	Orientation of the fast shear-wave component after splitting in an anisotropic medium
δt	Time delay between the fast and slow shear-wave components after splitting in an anisotropic medium
M_b	Body-wave magnitude
H	Crustal thickness (distance from surface to the base of the Mohorovičić discontinuity)
κ	Ratio of P- to S-wave velocity
ϵ	Topographic elevation
ρ_c	Crustal bulk density
ρ_a	Asthenospheric mantle density
ρ_{m_o}	Mantle density at surface pressure and temperature
$\bar{\rho}$	Overburden density
T_P	Mantle potential temperature
P_1	Pressure at the base of the lithosphere
α	Thermal expansivity
K	Bulk modulus
g	Acceleration due to gravity
C_p	Heat capacity

1

Introduction

Contents

1.1	Plate tectonics and continental rifting	2
1.2	Afro-Arabian Rift System	6
1.2.1	Arabian Plate	8
1.2.2	Gulf of Aden	8
1.2.3	Red Sea	9
1.2.4	Afar	11
1.2.5	Main Ethiopian Rift	15
1.3	The African Large Low Shear Velocity Province	16
1.4	Magmatism in the Afro-Arabian Rift System	19
1.4.1	Recent volcanic activity in Afar	23
1.4.2	The Nabro Volcanic Range and Nabro volcano	24
1.5	Seismic methods	27
1.5.1	Shear-wave splitting	28
1.5.2	Receiver functions	33
1.5.3	Seismic body-wave velocity tomography	38
1.6	Thesis overview	43

The process of continental breakup, commonly referred to as ‘continental rifting’, has shaped our planet as we know it today. Not only has continental rifting determined the configuration of continents and oceans across the globe, but it is also responsible for the distribution of natural resources along continental margins and within rift basins, and has produced dramatic geological features. Notably, the East African Rift System was the home of the first known hominids, our

earliest ancestors. The oldest fossils representing *Homo sapiens* were discovered in Ethiopia and date back to 230,000 years ago. Rifting processes have been ongoing for millennia in this region, creating a natural habitat for hominid migration—a valley extending over thousands of kilometers, containing numerous lakes (both food and water sources), as well as volcanic products for making tools. The East African Rift System forms one branch of the Afro-Arabian Rift System, one of the best examples of active continental rifting on Earth.

This introductory chapter reviews the considerable literature that has been previously published on the Afro-Arabian Rift System (AARS) and its three rift branches. I specifically focus on the region encompassing the Afar Triple Junction, the meeting point of these rift branches—the East African Rift System, the Red Sea rift, and the Gulf of Aden. I detail the main debates and ongoing questions regarding the nature of the crust and mantle beneath the Afar Triple Junction and its associated rifts, the driving forces responsible for rifting processes, and surface manifestations of magmatism. I begin by summarising the theory of plate tectonics and continental rifting mechanisms in Section 1.1. Section 1.2 gives an in-depth overview of the rift zones within the AARS and the Afar region. In Section 1.3, I review the debates on the nature of hot mantle flow from the African Large Low Shear Velocity Province (LLSVP) and its interaction with the lithosphere under the AARS. Section 1.4 looks at magmatism and volcanism throughout the region, with a particular focus on Nabro volcano and its 2011 eruption. In Section 1.5, I describe the seismic methods I use to investigate these questions: shear-wave splitting, receiver function analysis, and local velocity tomography. I focus on giving an overview of each method, with further methodological details described later in each relevant chapter. Finally, Section 1.6 summarises the contents of this thesis.

1.1 Plate tectonics and continental rifting

Plate tectonics is the term given to the evolution of Earth’s lithosphere, defined as the rigid part of the crust and upper mantle. The idea that the continents may not be fixed through time was first proposed by Wegener (1912). In the 1960s, Wilson

Table 1.1: The Wilson Cycle of the opening and the closing of the ocean basins. Table modified from Burke (2011).

Stage	Examples	Dominant motions	Characteristic features
1. Embryonic	East African Rift Valleys	Uplifts	Rift valleys
2. Young	Red Sea, Gulf of Aden	Spreading	Narrow seas with central depression
3. Mature	Atlantic Ocean	Spreading	Ocean basin with active mid-ocean ridge
4. Declining	Pacific Ocean	Shrinking	Island arcs, adjacent trenches
5. Terminal	Mediterranean Sea	Shrinking, uplifts	Young mountains
6. Relic scar (geosuture)	Indus Line in Himalayas	Shrinking, uplifts	Young mountains

further developed the theory, observing that the lithosphere could be divided into several large rigid plates (e.g., Wilson, 1965; Wilson et al., 1966). These plates move relative to one another due to convective heat flow from the Earth’s deep interior. The Earth’s lithosphere can be classified as either continental or oceanic lithosphere. Oceanic lithosphere is thinner, denser, younger, and of more mafic composition than continental lithosphere. Crucially, continental lithosphere is weaker and deforms more under a given stress (e.g., Molnar, 2015), allowing continents to rupture and separate. The Wilson Cycle (Table 1.1) describes this cycle of continental breakup and reassembly, and concurrent ocean opening and closing.

The formation and progression of continental rifts have been the focus of study for geophysicists, geochemists and geodynamicists alike, aiming to quantify the causes of continental breakup and how/whether it might develop into an ocean basin. Some rifts fail before getting to the oceanic spreading stage and remain preserved as fossil features within continents, such as the Midcontinental Rift in North America (e.g., Stein et al., 2018).

Continental rifting involves structural and rheological changes to the lithosphere. There are two end-member mechanisms proposed to explain the initiation of continental rupture: passive or active rifting (e.g., Sengör and Burke, 1978).

The former is mechanical, with thinning resulting from horizontal extension of the continental lithosphere. Far-field stresses generated within the lithosphere may arise from plate margins due to ridge push or slab pull (e.g., Turcotte and Emeryman, 1983; McKenzie, 1978). Thinning of the crust and lithosphere through this passive mechanism will lead to crustal uplift and volcanism as secondary processes, as hot asthenospheric mantle ascends to replace the thinned lithosphere (White and McKenzie, 1989).

Active rifting, by contrast, invokes crustal uplift and volcanism as primary processes that directly contribute to the tensional failure of the lithosphere (e.g., Neugebauer, 1978). Active rifting begins when a heat source directly acts on the base of the lithosphere; for example, after the impingement of a mantle plume (Burke and Dewey, 1973). The ascending hot material is responsible for crustal uplift and generating large volumes of magma. The elevated temperatures associated with the magmatic intrusion act to lower the yield stress of the lithosphere (Figure 1.1; Buck, 2004). Continental breakup can therefore begin without marked thinning of the crust.

These two basic modes of rifting are complicated by various local physical parameters (e.g., rate and duration of extension, geothermal gradient, rheological profile of the lithosphere, initial lithospheric thickness, magmatic intrusions). A great diversity of continental rifting style is thus observed globally: from wide areas of extension such as the Basin and Range province (e.g., Brun, 1999) to narrow rift basins observed in the Western branch of the East African Rift System in Tanzania (e.g., Ebinger et al., 1989b); from magma-poor rifted margins (e.g., the Iberian-Newfoundland margins; Péron-Pinvidic and Manatschal, 2009) to volcanic rifted margins (e.g., the Africa-Madagascar margins; Menzies et al., 2002).

Eventually, if a continent rifts to completion, an oceanic spreading centre is produced between two conjugate continental margins. Seafloor spreading is associated with a narrow zone of magma, produced by decompression melting of the mantle, which intrudes the plate to create new oceanic crust (e.g., Crisp, 1984; Schouten et al., 1985). However, the evolution from initiation of breakup to

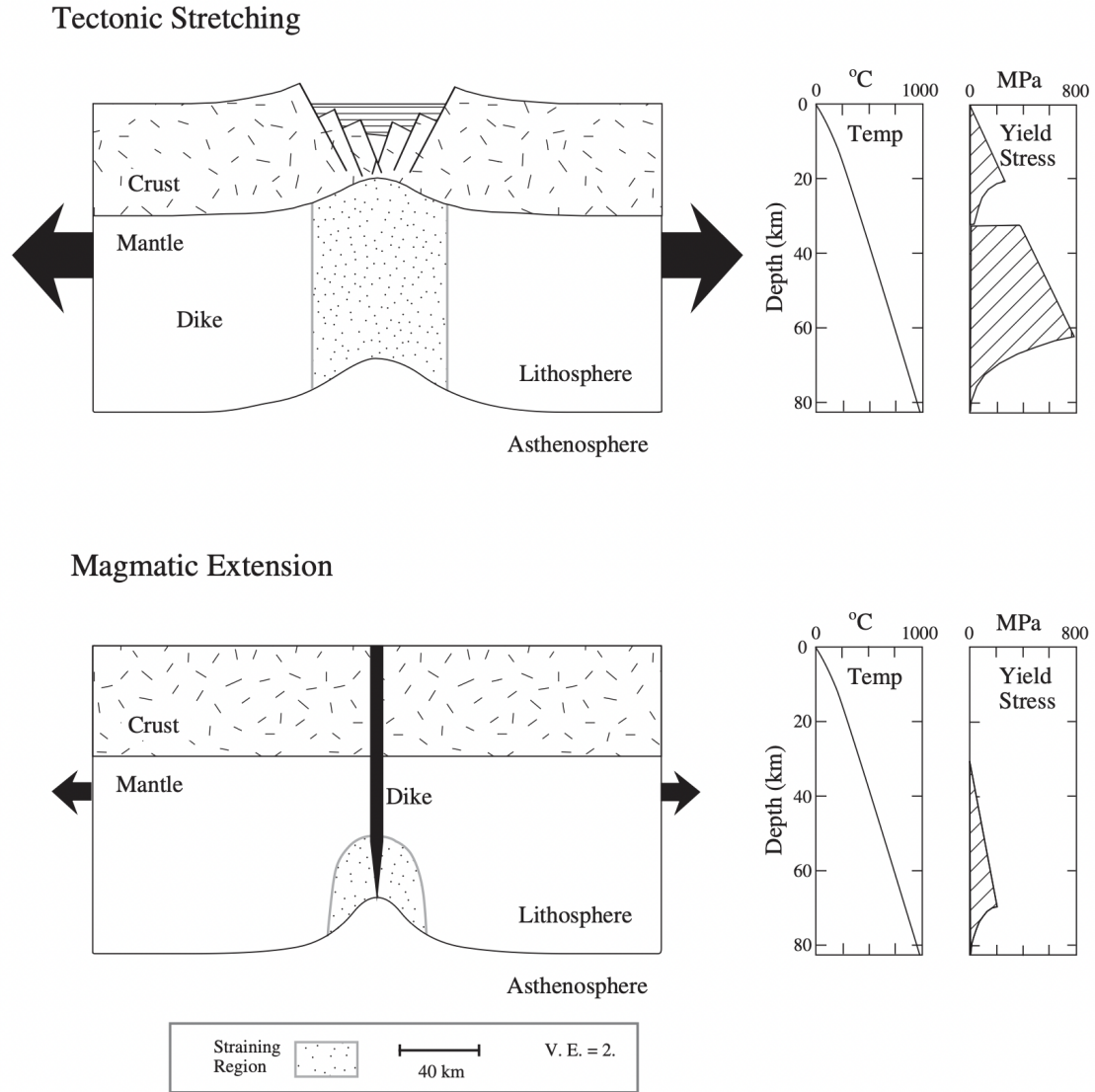


Figure 1.1: Schematic of the difference between extension of thick lithosphere with and without magmatic intrusion. Note the large difference in the yield stress, the stress difference needed to get extensional separation of two lithospheric blocks. Figure and caption from Buck (2004).

complete rupture is complex. The transition is characterised by extensional faulting governed by large-scale border faults and half-grabens (e.g., Hayward and Ebinger, 1996), voluminous magmatism (e.g., White and McKenzie, 1989), and subsidence and ductile deformation (e.g., Bastow and Keir, 2011; Keir et al., 2013; Daniels et al., 2014), but the relative timing of each of these mechanisms can vary.

To comprehensively understand rifting processes, studies of both crust and mantle are required: the mantle, to map mantle flow and anomalous hot upwellings

that influence rifting processes; the crust, to understand surface manifestations of strain distribution caused by rifting and and quantify magmatic intrusion throughout the rifting cycle.

1.2 Afro-Arabian Rift System

The Cenozoic Afro-Arabian Rift System (AARS) is one of Earth's most dramatic tectonic features (Figure 1.2). The contemporaneous existence of different active rifting stages, from initial continental breakup to full seafloor spreading, within the AARS means it is an ideal locale to study the evolution and development of continental rifting. The AARS encompasses the following elements: the East African Rift System, which extends from Mozambique through the Turkana Depression to the Main Ethiopian Rift and marks the separation of the Nubian and Somalian plates; the Red Sea rift, an incipient ocean basin separating the Arabian and Nubian plates; and the Gulf of Aden rift, where full seafloor spreading has been established between the Arabian and Somalian plates (e.g., McKenzie et al., 1970; Ghebreab, 1998; Tesfaye et al., 2003; Leroy et al., 2004; Bosworth et al., 2005). The meeting point of these three rifts is the Afar Triple Junction, currently located near Lake Abhe on the north-eastern Ethiopia-Djibouti border (Tespaye et al., 2003). The existence of rift-rift-rift triple junctions is required by the geometrics of plate evolution, and is the most stable of the plate boundary combinations (i.e., transform, rift or thrust) where three plates meet (McKenzie and Morgan, 1969).

The Arabian plate is currently moving away from Africa at $\sim 15 \text{ mm yr}^{-1}$, which has been fairly constant since 56 Ma (Chu and Gordon, 1999; McQuarrie et al., 2003; DeMets et al., 2010). The absolute motion of the Nubian plate in a Eurasian frame of reference decreased from $\sim 30 \text{ mm yr}^{-1}$ to $\sim 10 \text{ mm yr}^{-1}$, co-eval with the divergence of Arabia from Africa (Jolivet and Faccenna, 2000; Reilinger and McClusky, 2011). The Somalian plate is moving apart from Nubia at $\sim 6 \text{ mm yr}^{-1}$ (Chu and Gordon, 1999; Stamps et al., 2008). The variation in extension rates across the three rift segments means that the location of the Afar Triple Junction has migrated over time (McKenzie and Morgan, 1969). Tesfaye et al. (2003) find

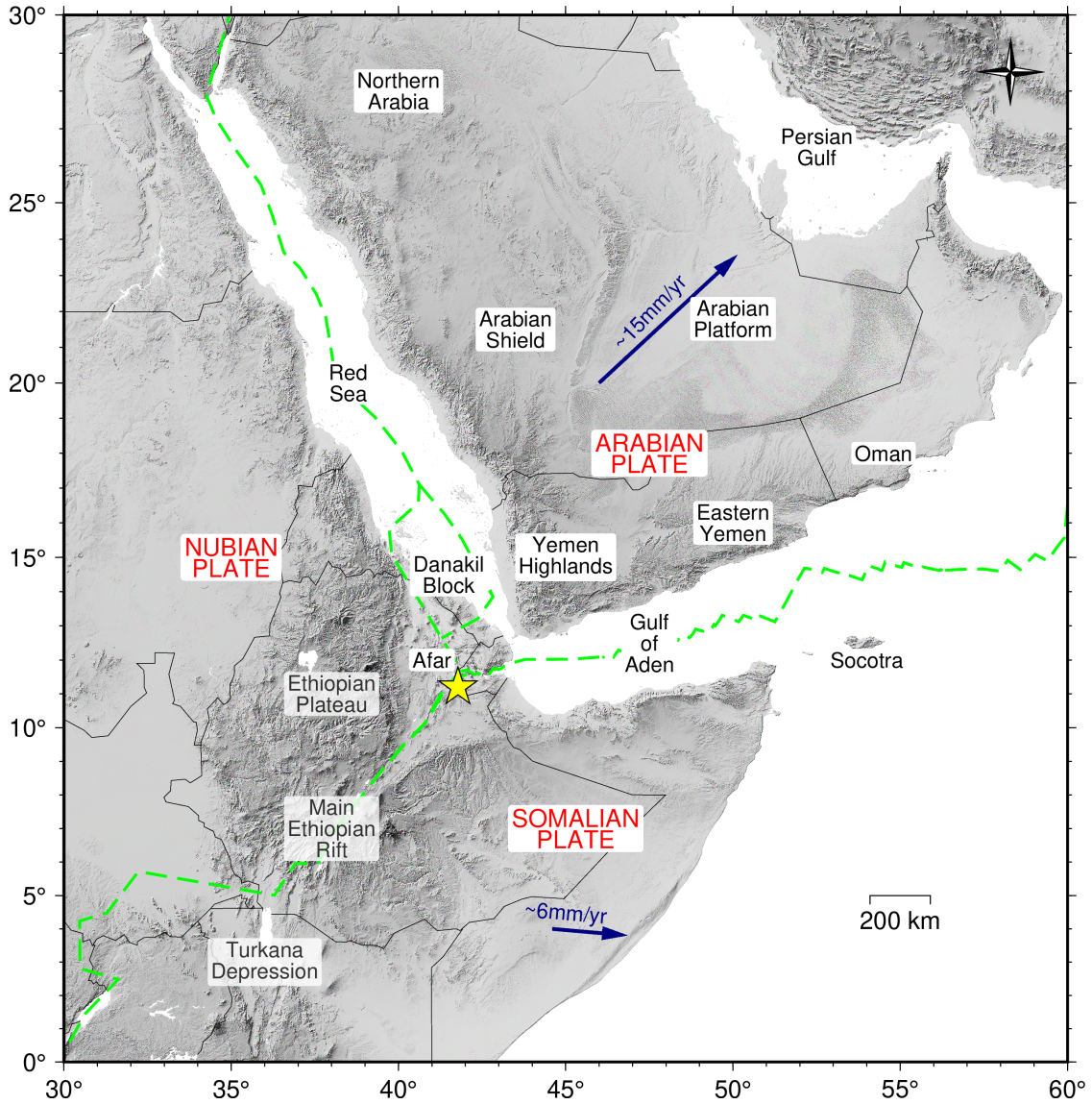


Figure 1.2: Topographic map of the Afro-Arabian Rift System. The location of the Afar Triple Junction is marked with a yellow star (Tesfaye et al., 2003) and its surrounding plate boundaries (the Red Sea rift, the Gulf of Aden, and the Main Ethiopian Rift), are outlined with dashed green lines (Viltres et al., 2020; Bird, 2003). The three tectonic plates are labelled with red text, and other geographic domains are labelled in black text. Navy arrows mark the absolute plate motions of Arabia and Somalia in a Nubia fixed frame of reference (Chu and Gordon, 1999; Stamps et al., 2008; DeMets et al., 2010). National borders are marked with thin black lines.

that the triple junction is currently located at 11.20°N, 41.78°E, having migrated ~ 160 km in a NNE direction over the past 25 Ma.

1.2.1 Arabian Plate

Rifting processes initiated on the African plate in the mid-Oligocene with the opening of the Gulf of Aden, followed by the Red Sea rift (Laughton, 1966; McKenzie et al., 1970). Eventually, the Arabian plate separated from the African plate. The fit of the southern coast of Arabia into the Horn of Africa was one of the first case studies used as evidence for plate tectonic theory (Keir et al., 2013). The two broad geological terranes that make up the Arabian plate are the Arabian Shield in the west (close to the Red Sea margin) and the Arabian Platform in the east (Stern and Johnson, 2010). The boundary between these two terranes marks a fundamental crustal and mantle heterogeneity in the plate, which is expressed magnetically by the Central Arabian Magnetic Anomaly (Stern and Johnson, 2010). The Arabian Shield has elevations of up to 3 km in the west, while the Arabian Platform descends to sea level, leading to an asymmetric tilt of the plate towards the Persian Gulf (Daradich et al., 2003; Bosworth et al., 2005). Seismic tomography studies have shown that the Arabian Platform is underlain by thick (200–250 km), cold, seismically-fast lithosphere (Sandvol et al., 1998; Hansen et al., 2007; Koulakov et al., 2016).

1.2.2 Gulf of Aden

Rifting in the Gulf of Aden began in approximately the mid-Oligocene (~ 30 Ma); there is no firm evidence that extension pre-dated the onset of magmatism in the region (Bosworth et al., 2005). The dating of magnetic anomalies indicates that the earliest inception of seafloor spreading in the Aden Ridge is ~ 17.5 Ma (Leroy et al., 2004; Fournier et al., 2010). The young ocean basin trends N75°E, oblique to the extension direction (d’Acremont et al., 2010). The trend is also oblique to the relative plate motion of Arabia with respect to Africa (Bosworth et al., 2005). The Gulf of Aden is divided by two major discontinuities: the Shukra-El Sheikh discontinuity in the west, near the Gulf of Tadjoura, and the Alula-Fartak fracture zone, separating

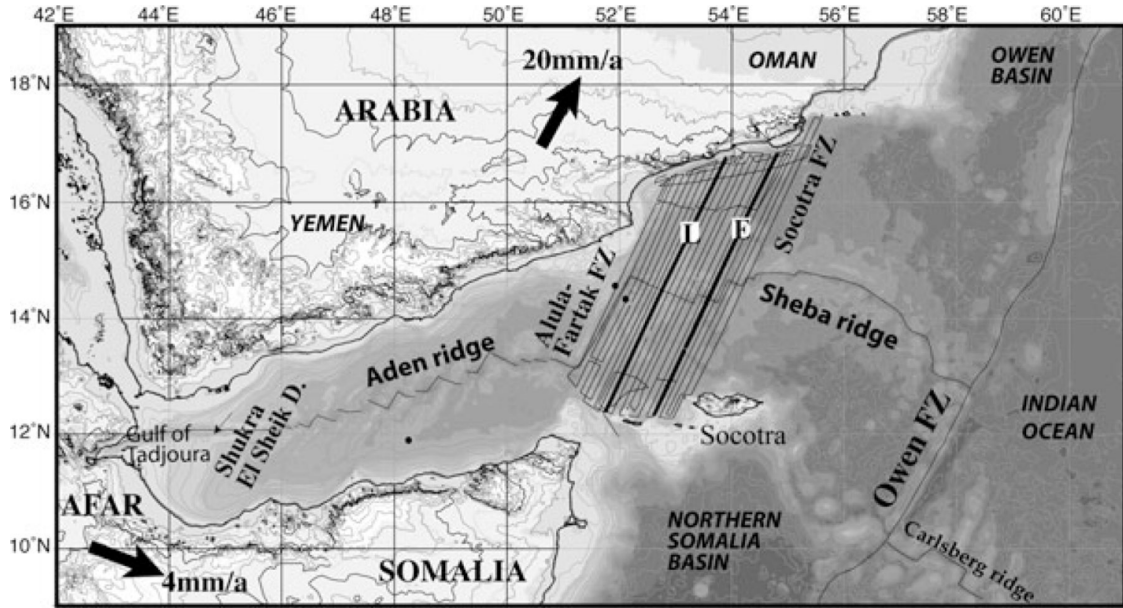


Figure 1.3: Bathymetric and topographic map of the Gulf of Aden area. Figure and caption from d'Acremont et al. (2010).

the Aden Ridge from the Sheba Ridge (Figure 1.3; Manighetti et al., 1997). The Sheba Ridge lies to the east of the Alula-Fartak fracture zone and connects to the Owen fracture zone, with seafloor spreading dating to 19.7 Ma (Stein and Cochran, 1985; Fournier et al., 2010). Magnetic dating youngs to the west, showing the direction of propagation (along-strike migration) of the rift was from east to west (Figure 1.4; Leroy et al., 2004; d'Acremont et al., 2005; Fournier et al., 2010), with major stalling in places, particularly at the Shukra-El Sheikh discontinuity (Manighetti et al., 1997). The Gulf of Aden propagated westward into Afar through the Gulf of Tadjoura at ~ 2 Ma, coherent with the anticlockwise rotation of the Arabian plate (Courtillot et al., 1980; Manighetti et al., 1998; Audin et al., 2001).

1.2.3 Red Sea

The opening of the Red Sea rift post-dated the Gulf of Aden by some millions of years, in the Late Oligocene (~ 24 Ma; Ghebreab, 1998). The tectonic evolution of the rift is debated amongst authors. There is general agreement that oceanic crust exists along parts of the rift, and the onset of seafloor spreading has been dated to ~ 5 Ma (Figure 1.4; Izzeldin, 1987; Schettino et al., 2016). ~ 85 km of oceanic

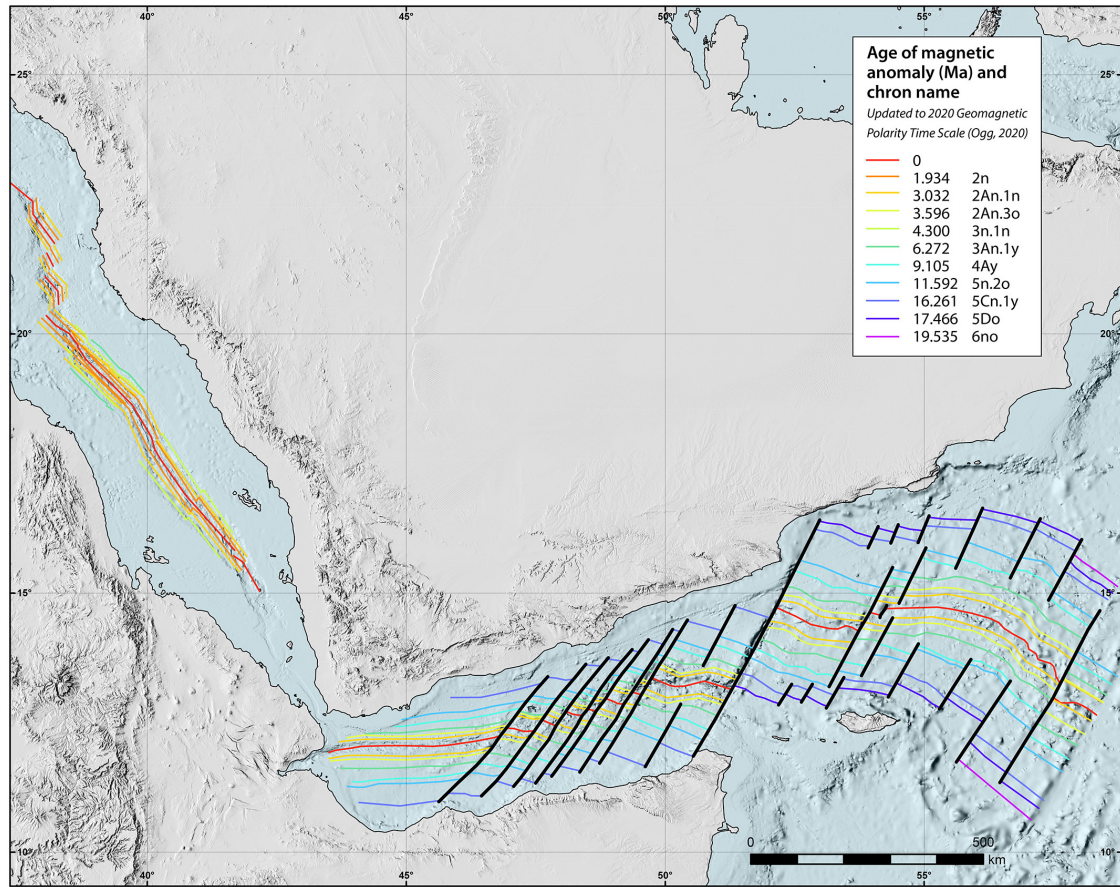


Figure 1.4: Dated oceanic magnetic anomalies in the Red Sea and Gulf of Aden rifts. Figure and caption from Rime et al. (2023) and references therein.

crust is thought to have formed in the south-central Red Sea (e.g., Cochran, 1983; Bosworth et al., 2005). However, there have been disagreements as to whether the oceanic crust is widespread within the Red Sea, as proposed by McKenzie et al. (1970), or whether part of the extension between Arabia and Nubia is still occurring through thinning of continental lithosphere, particularly in the northern Red Sea (e.g., Cochran, 1983; Martinez and Cochran, 1988).

Another open question is when and how rifting developed, with the Red Sea featuring widely in the debate described in Section 1.1 of passive versus active rifting mechanisms. Authors have claimed passive origins for the rift, with magmatism occurring only after the crust had been sufficiently thinned by faulting (e.g., McGuire and Bohannon, 1989; Bohannon, 1989), while others have argued for an intermediary between passive and active (e.g., Davison et al., 1994; Drury et al., 1994). See

Ghebreab (1998) for a comprehensive discussion of the different models proposed for the Red Sea tectonic development.

A notable feature of the rift is that it does not directly join with the Gulf of Aden via the Bab el Mandeb Strait, despite their marine connection there. Instead, the Red Sea bifurcates south of $\sim 16^\circ\text{N}$ (Barberi and Varet, 1977; Eagles et al., 2002; McClusky et al., 2010). It has been suggested that this transfer of strain is a complete ‘rift jump’ into Afar (e.g., Wood, 1983), but kinematic models show that strain is partly accommodated in one arm continuing south to at least the Hanish-Zukur Islands (Viltres et al., 2020). The rest of the extension is accommodated on land in Afar, reaching $\sim 20\text{mmyr}^{-1}$ at 13°N (McClusky et al., 2010).

1.2.4 Afar

Afar is a diffuse extensional province, where plate boundaries are difficult to identify (Courtillet et al., 1980). Topographically, Afar is extremely low-lying, with parts lying beneath sea level, and is sometimes termed the ‘Afar depression’ in the literature (e.g., Tazieff et al., 1972; CNR-CNRS, 1973). Border faults mark the transition from the rift valley floor to the 2–3 km high Ethiopian and Southeastern plateaus that flank the Afar depression (Keir et al., 2013). Rift-related processes began $\sim 29\text{--}30\text{Ma}$ on the western Afar margin (Wolfenden et al., 2005) and in northern Afar at $\sim 23\text{--}25\text{ Ma}$ (Barberi et al., 1972).

The northern part of Afar is known as the Danakil Depression (e.g., Bastow et al., 2018). To the west of the Danakil Depression are the Danakil Alps, part of the Danakil block, alternatively called the Danakil horst or the Danakil microplate, separating the southern Red Sea rift basin from the Afar depression (Mohr, 1970; Tazieff et al., 1972). The differential extensional motion between the two rift branches is fully accommodated by rigid rotation of the Danakil block, as shown by various kinematic models from GPS motions (e.g., Eagles et al., 2002; McClusky et al., 2010; Viltres et al., 2020).

Tectonics and volcanism in Afar are characterised by complex faulting over wide areas, with as many as 8 linear Quaternary volcanic axial ranges in Afar

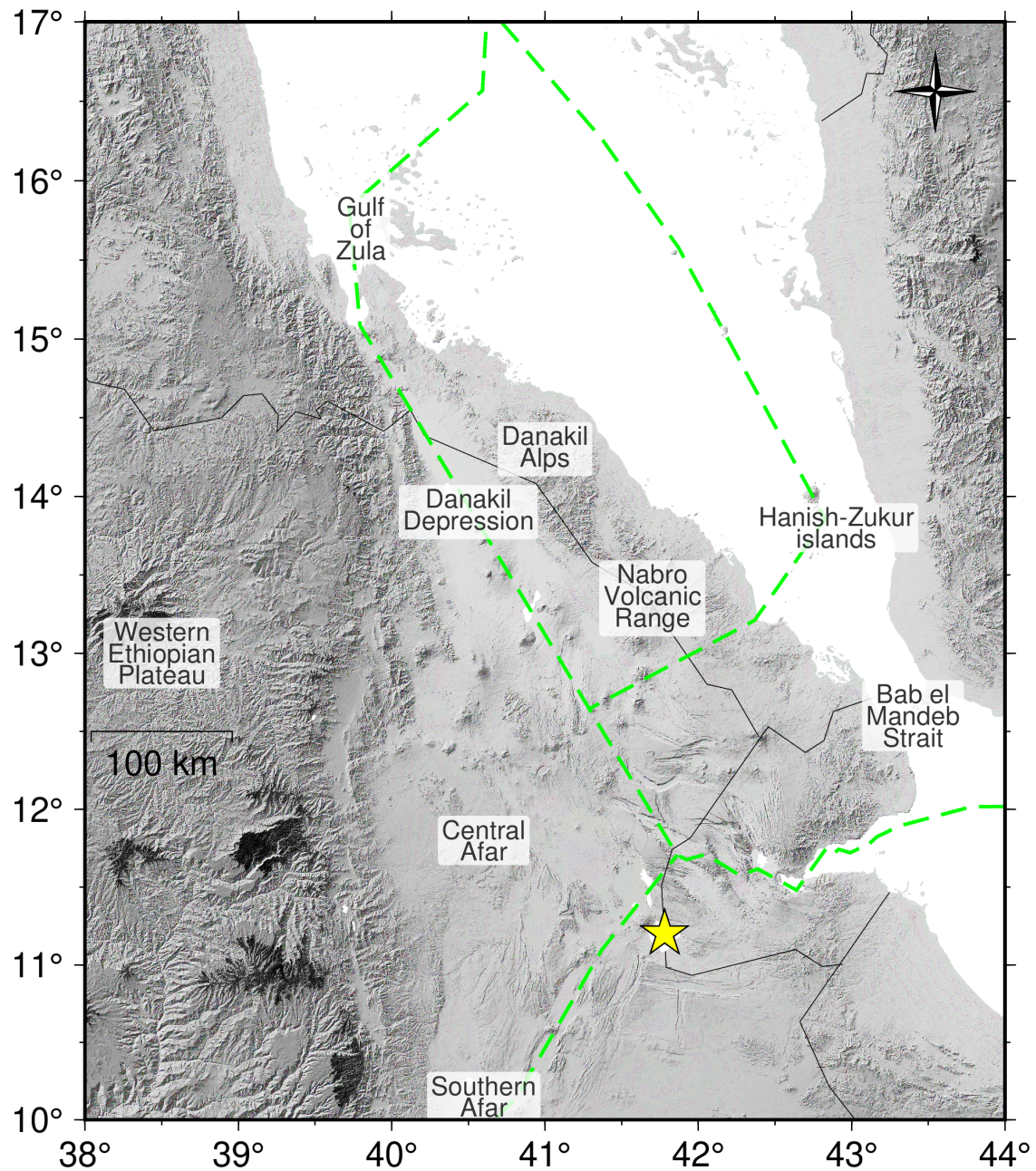


Figure 1.5: Topographic map of Afar and the Danakil microplate. The location of the Afar Triple Junction is marked with a yellow star (Tesfaye et al., 2003). The boundaries of the Danakil microplate and other plate boundaries are outlined by a dashed green line (Viltres et al., 2020). Solid black lines demarcate national borders.

(e.g., Barberi and Varet, 1977; CNR-CNRS, 1973). These axial ranges alternately parallel the rifting trends of the three rifts that meet in the Afar Triple Junction: the Gulf of Aden, the Main Ethiopian Rift and the Red Sea Rift (Manighetti et al., 2001; Keir et al., 2013). Extension across these mismatched rift segments may be accommodated by combinations of transform faults, strike-slip faults, basins, folds and penetrative strain (Tesfaye et al., 2003). Rift segments are linked through “accommodation zones” marked by intense zones of deformation (Tesfaye et al., 2003). Courtillot et al. (1980) proposes a model where the intense deformation in Afar is that which takes place before a throughgoing lithospheric crack is established. The motion of the Arabian plate has had a significant impact on Afar’s tectonic evolution (e.g., Keir et al., 2011b).

While the propagation of the Gulf of Aden into Afar is well-established (e.g., Manighetti et al., 1997), the development and progression of the Red Sea strain migration into Afar is debated. The literature overwhelmingly describes the rifting within Afar as a southward propagation of the Red Sea rift, or even ‘the subaerial Southern Red Sea’ (e.g., Manighetti et al., 2001; Wolfenden et al., 2004; Reed et al., 2014). However, it has also been proposed that rifting propagates northward through Afar, facilitated by the rotation of the Danakil block (e.g., Eagles et al., 2002; Sani et al., 2017; Molnar et al., 2018). See Rime et al. (2023) for further discussion of the propagation direction of rifting in Afar.

Wide-angle seismic refraction studies and teleseismic receiver functions have investigated crustal thicknesses in Afar. The results show significant variation across the region: ~ 15 km thick beneath the Danakil Depression; ~ 40 – 45 km thick beneath the western Ethiopian plateau; ~ 25 km beneath central and southern Afar (Makris and Ginzburg, 1987; Maguire et al., 2006; Stuart et al., 2006; Dugda et al., 2005; Hammond et al., 2011c). The nature of this crust is controversial. The Afar crust has relatively fast shear-wave velocity, suggesting that most of the Afar crust is made up of solidified mafic rock (Chambers et al., 2019). The origin of the crust in Afar is debated. Some authors argue that new oceanic crust is being created in the region (e.g., Mohr, 1989), whilst others conclude that the area is entirely underlain

by intruded, stretched continental crust (e.g., Makris and Ginzburg, 1987). An alternative proposal is that the overall crustal composition is transitional between continental and oceanic (e.g., Redfield et al., 2003; Hammond et al., 2011c). These studies suggest that the region west of the current rift axis is predominantly made up of mafic material with oceanic affinities, whereas continental material exists to the east, potentially connected to the Danakil block.

Seismic moment release from earthquakes in Afar (from 1960–2000) contributes less than half of the total geodetic moment expected from current plate separation rates (Hofstetter and Beyth, 2003). Thus, magmatic extensional mechanisms such as dyke intrusion are proposed to be the principal strain accommodation mechanisms in Afar at present (e.g., Bastow and Keir, 2011; Keir et al., 2011a). Crustal anisotropy observations of the strain field around the Afar Triple Junction are explained by an increase in dyke-induced fracturing and faulting into zones of magma intrusion, combined with oriented melt pockets near active magmatic centres, corroborating this conclusion (Keir et al., 2011a). By comparing variations in volcanism, elevation and crustal thickness with latitude, Bastow and Keir (2011) show that magmatism during continental breakup is controlled principally by the strength and thermal structure of the extending plate. They propose that extension by magma intrusion initially occurs without marked crustal thinning, supported by the lack of crustal thinning in Afar between 9.5°N–13°N (Makris and Ginzburg, 1987; Maguire et al., 2006). Eventually, the plate strength is reduced sufficiently for ductile extension to commence, with associated adiabatic decompression melting (Bastow and Keir, 2011). The seismic reflection study of Hurman et al. (2023) finds that upper crustal axial faulting in a magma-poor area of the depression accommodates 30% of the total extension. Thus, it has been suggested that the final stages of continental breakup in northern Afar may be characterised by a return to ductile plate stretching and upper-crustal brittle faulting (Bastow and Keir, 2011).

1.2.5 Main Ethiopian Rift

In the Miocene, rifting initiated between the Nubian and Somalian plates to form the Main Ethiopian Rift (MER), the third and final branch of rifting into the Afar Triple Junction (Wolfenden et al., 2004; Kazmin et al., 1978). The onset of rifting in the MER post-dated extensional processes in southern Afar, as shown by NE-NNE striking normal faults of the MER cross-cutting NW-NNW striking faults in southern Afar (Tesfaye et al., 2003; Keir et al., 2011b). The northernmost boundary of the MER is the Tendaho–Goba’ad discontinuity in southern Afar, and rifting continues southwards through the Turkana Depression to terminate in Mozambique; the entire system is referred to as the East African Rift System (Wolfenden et al., 2004; Chorowicz, 2005). For a comprehensive synthesis of the tectonic evolution of the EARS, see Chorowicz (2005).

Extension began in the MER on large-scale border faults (Wolfenden et al., 2004), and has since shifted to ~ 30 km-wide, ~ 60 km-long en-echelon axial magmatic segments (Ebinger and Casey, 2001; Keir et al., 2006). The existence of partial melt distributed throughout the lower crust and upper mantle beneath the MER is well-established by seismicity and other geophysical studies, both within the rift axis and beneath the bordering NW plateau (e.g., Kendall et al., 2005; Keir et al., 2009a; Chambers et al., 2019). Seismic images of low velocity beneath the NW plateau indicate that localised melt emplacement and related hydrothermal processes are ongoing outside of the rift valley (e.g., Chambers et al., 2019). The rift axis of the MER is underlain by very slow seismic velocities, interpreted as accumulated melt residing for a long time in the crust (Chambers et al., 2019). The intensification of magmatic activity from 2Ma has produced a range of differing volcanic products in the MER, most notably large silicic calderas formed in the last 0.5Ma (Woldegabriel et al., 1990). These volcanoes are key geothermal resources (Barbier, 2002). Aluto volcano in Ethiopia and Olkaria in Kenya currently have operating geothermal power plants, and there are plans to develop other sites such as Corbetti, Tendaho, Abaye, and Tulu Moye (Kebede, 2012).

1.3 The African Large Low Shear Velocity Province

It is well-established that a mantle thermo-chemical anomaly (or anomalies) of some kind underlies the AARS. Often referred to as ‘plumes’ or ‘superplumes’, anomalies such as these are influential on magmatism and tectonic processes around the globe (e.g., Hill, 1993).

Global seismic tomography models have identified a broad region of low shear-wave velocity, known as the African Large Low Shear Velocity Province (LLSVP), or the African ‘superplume’ (Ritsema et al., 1999; Ritsema and Van Heijst, 2000; Garnero and McNamara, 2008; Koelemeijer et al., 2017). The LLSVP originates at the core-mantle boundary beneath southern Africa and rises to the base of the lithosphere somewhere in Ethiopia (Figure 1.6), and is likely to be a long-lived mantle feature (e.g., Torsvik et al., 2008).

There is evidence that the upper mantle beneath the AARS has been affected by this LLSVP. It is broadly characterised by anomalously slow seismic wave velocities; the absolute shear wave delay times are some of the slowest in the world, evidencing a significant hot anomaly in the upper mantle across the Arabian Shield, East African and Ethiopian plateaus (e.g., Nyblade et al., 2000; Benoit et al., 2006; Park et al., 2007; Bastow et al., 2008; Park et al., 2008; Hansen et al., 2012; Tang et al., 2019; Lim et al., 2020; Corbeau et al., 2014; Chambers et al., 2019; Civiero et al., 2022). Studies of mantle transition zone structure across Afar and the Main Ethiopian Rift have indicated that compositional and/or chemical heterogeneities exist at mantle transition zone depths, with chemically distinct upwellings connecting the upper mantle low velocities to the LLSVP in the lower mantle (Cornwell et al., 2011; Julià and Nyblade, 2013; Civiero et al., 2015; Thompson et al., 2015). Elevated mantle potential temperatures (Rooney et al., 2012) and isotopic anomalies (Manetti et al., 1991; Pik et al., 2006; Halldórsson et al., 2014) in the region are also consistent with the presence of material derived from deep in the mantle. Geodynamic models have used the global seismic tomography models to successfully predict the pattern and amplitude of plateau uplift in the AARS (e.g., Daradich et al., 2003; Moucha and Forte, 2011). The study of Chen et al. (2015) shows low effective elastic

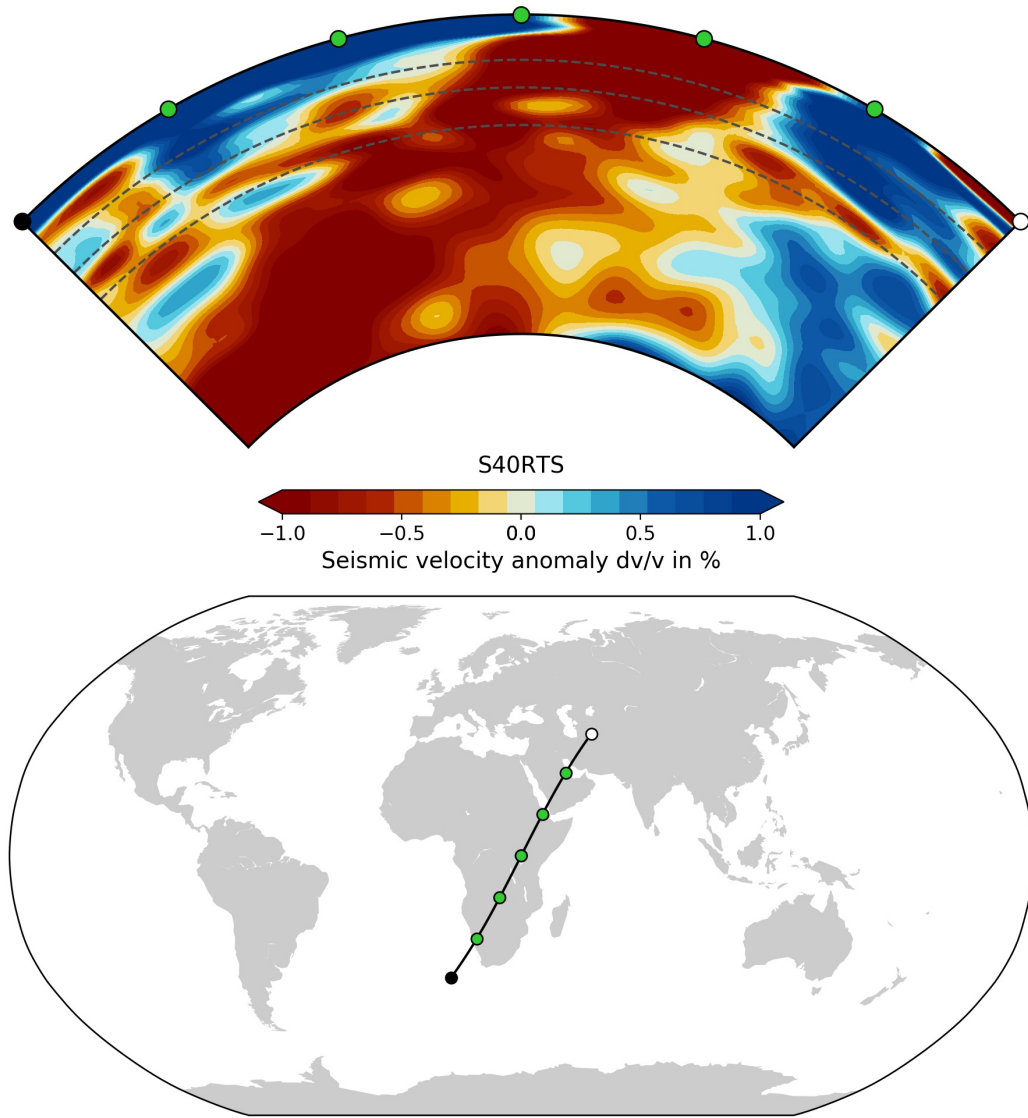


Figure 1.6: Whole mantle cross-section centred on the AARS, from the global S-wave tomographic model S40RTS (Ritsema et al., 2011). Cross-section is taken as a vertical 2D section along a great circle path at the Earth's surface and through its centre. The depths of the dashed lines are 410 km, 660 km and 1000 km from the surface. Figure produced using SubMachine (Hosseini et al., 2018).

thickness along both the Red Sea and Gulf of Aden, matching the low seismic velocity trends from seismic tomography studies. This implies that these regions are being dynamically supported by hot upwelling mantle.

Despite consensus over the presence of a broad low velocity anomaly existing in some form beneath the AARS, there is an active debate over the manner in which deep, hot mantle material upwells from the core-mantle boundary across the region. Three types of model have been proposed: a) a broad superplume; b) one or multiple distinct diapiric plumes; c) multiple diapiric plumes arising from the same common source.

The first model attributes the anomalous upper mantle velocity structure beneath the AARS to a continuation of the lower-mantle LLSVP anomaly into the upper mantle (e.g., Bastow et al., 2008; Park et al., 2007, 2008; Hansen et al., 2012). By contrast, the second model invokes plume head material ~ 100 -200 km thick, fed by a narrow (~ 100 -200 km diameter) plume tail. There is little consensus over how many plumes exist beneath the AARS. Ebinger and Sleep (1998) propose a single plume beneath southern Ethiopia, guided laterally by thin topography on the lithosphere-asthenosphere boundary (LAB), whilst others suggest two plumes located beneath Afar and East Africa (Montelli et al., 2006), three plumes beneath Ethiopia, Afar and central Kenya (Burke, 1996), two beneath Ethiopia and Kenya (George et al., 1998; Boyce et al., 2021), one beneath Arabia (Camp and Roobol, 1992). The difficulty of establishing exactly where plume heads arise from led to the third series of models, which argue for isolated mantle upwellings in the upper mantle, all fed from the common source of the African LLSVP (Furman et al., 2006b; Civiero et al., 2015, 2022).

The African LLSVP is central to debates over the driving forces behind the rifting in the AARS. Kendall and Lithgow-Bertelloni (2016) calculate stresses induced by mantle flow, crustal structure, and topography using two different models. In one model, mantle flow is exclusively driven by distant forces from ‘ridge-push’ or ‘slab pull’, while in the other it is derived from a shear wave tomographic model. They find that the latter model produces more accurate dynamic topography and

so conclude that mantle structure beneath Africa plays a key role in providing the radial and horizontal tractions necessary for rifting'. Even so, the combined stresses are still not enough to rupture thick, cold, dry continental lithosphere; and therefore Kendall and Lithgow-Bertelloni (2016) suggest the presence of melt as a mechanism for reducing the strength of the African plate. The abundant evidence for strain localisation and focused melt upwellings within the EARS and Main Ethiopian Rift (Section 1.2.5) are further evidence for magma-assisted models of rifting between Nubia and Somalia. Yet the mechanisms driving the opening of the Red Sea and the Gulf of Aden are less clear. Numerical modelling by Bellahsen et al. (2003) shows that intraplate stresses are related to the evolution of the northern convergent margin of the African plate, which changed from collision to active subduction in the Neogene. However, they also emphasise that weakening of the Arabian lithosphere from a hot mantle upwelling must have played a role in localising strain and ultimately leading to the formation of the Red sea Rift and the Gulf of Aden. A whole mantle convection cell is hypothesised by the geodynamic models of Faccenna et al. (2013) to explain the kinematics of the Arabia-Anatolia-Aegean system, with slab pull exerted in the Aegean combined with a mantle upwelling beneath Afar acting to drive Arabian plate motion.

1.4 Magmatism in the Afro-Arabian Rift System

Surface expressions of magmatism in the AARS occur throughout the Cenozoic (Figure 1.7a). The earliest magmatism predates the opening of the rifts. Basalts were first emplaced in southern Ethiopia and northern Kenya (Turkana) in the Eocene, 40–45 Ma (Figure 1.7b; Ebinger et al., 1993; George et al., 1998; Rooney, 2017). A second phase of magmatic activity occurred in the Oligocene, when ~ 2 km of flood basalts were erupted over a short time period, beginning ~ 30 Ma (Hofmann et al., 1997; Ukstins et al., 2002; Coulié et al., 2003). These basalts are often referred to as the Ethiopian (or Ethiopia-Arabian) Traps, extending predominantly over an area of at least $600,000\text{km}^2$ from southwestern Ethiopia into the Yemen highlands (Figure 1.7b; Mohr, 1983) and causing substantial uplift (Barberi et al.,

1975). The Oligocene basalts are also observed to the south in Turkana, to the west in Sudan, and to the east on the margin of the Somali plateau (Figure 1.7b; Rooney, 2017). Mantle potential temperature estimates from these basalts reveal significant elevation in temperature over ambient mantle values (Rooney et al., 2012), indicating that the flood basalt episode was coincident with a significant thermal anomaly within the mantle. This interpretation is also corroborated by the rapidity of emplacement of the basalts (Hofmann et al., 1997). Following the Oligocene flood basalts, a dramatic decrease in volcanism occurred from ~ 25 –20 Ma (Ukstins et al., 2002). There was a resurgence in basaltic volcanism in the early Miocene (~ 26 –22 Ma), manifesting largely on the Somali plateau, the northwestern Ethiopian Plateau and in Turkana (Figure 1.7b; Kieffer et al., 2004; Furman et al., 2006b; Brown and McDougall, 2011).

The origin of these pulses of magmatism are widely considered to be linked to the African LLSVP, although the exact form of the ascension of material from deep in the mantle is debated. As described in Section 1.3, explanations vary from a single plume (Ukstins et al., 2002; Rooney et al., 2012, e.g.), two or more plumes with different origins (e.g., George et al., 1998; Pik et al., 2006; Nelson et al., 2008), two plumes from a common underlying source (e.g., Furman et al., 2006a; Furman, 2007), or a broad region of mantle upwelling with no distinct plume heads (e.g., Kieffer et al., 2004).

Magmatism continued through the Miocene in the Main Ethiopian Rift, Kenya Rift and Turkana Depression, co-eval with continental breakup processes (Wolfenden et al., 2004). See Rooney (2020a) and Rooney (2020b) for a comprehensive synthesis of geophysical and geodynamic studies of these episodes of magmatism. Meanwhile, Yemen also experienced a mid-Miocene resurgence in magmatic activity, which is temporally consistent with activity occurring in the northern Main Ethiopian Rift and Afar (Capaldi et al., 1983, 1987; Gass and Mallick, 1968; Mallick et al., 1990). Isotopic signatures of the volcanic products in Yemen show a strong similarity to Ethiopian plateau basalts, the youngest Main Ethiopian Rift basalts and the Afar basalts (Manetti et al., 1991). These are cut by abundant dykes, regionally

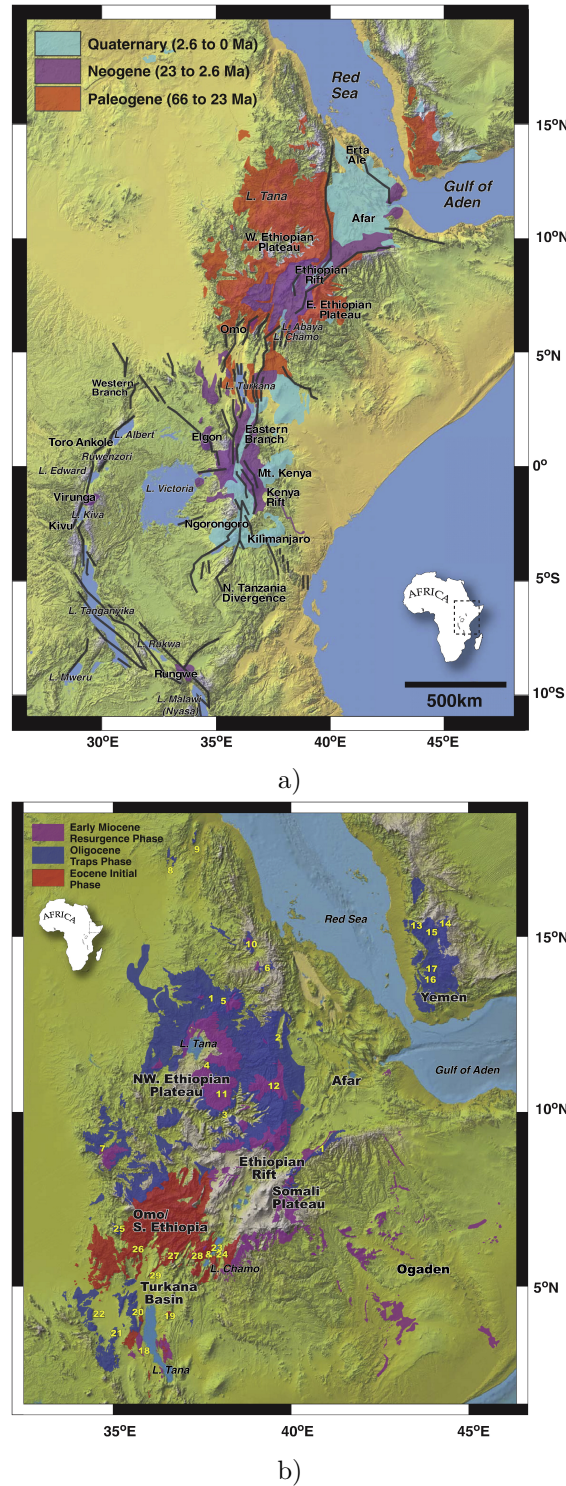


Figure 1.7: a) Cenozoic magmatic activity in East Africa, coloured by age: Paleogene (66 to 23 Ma); Neogene (23 to 2.6 Ma); Quaternary (2.6 Ma to today). Major volcanic provinces are labeled. Background image is a colored SRTM (NASA) digital elevation model (PIA04965). Major faults are labelled as grey lines. b) Magmatic activity in Africa and parts of the Arabian Peninsula in the Paleogene (45 to 22 Ma) and divided into three broad periods of magmatic activity: (A) Eocene Initial Phase (45–34 Ma), (B) Oligocene Traps phase (~33.9–27 Ma), (C) Early Miocene resurgence phase (26.9 Ma to 22 Ma). Background image is a colored SRTM (NASA) digital elevation model (PIA04965). Figures and captions from (Rooney, 2017), data from references therein.

striking NNW-SSE to NW-SE, associated with Red Sea opening (Mohr, 1991; Davison et al., 1994). A third episode of magmatic activity in Yemen developed during the Quaternary and has continued until recent times, with a significant shift towards more alkaline compositions (e.g., Camp et al., 1987; Mattash et al., 2014; Manetti et al., 1991).

The Oligocene marked the opening of the Gulf of Aden and Red Sea rifts—the initiation of Arabia separating from Africa, as discussed in Section 1.2.1. The Red Sea syn-rift strata are all associated with basaltic magmatism, mostly in the form of dykes striking parallel to the Red Sea margins (Baldrige et al., 1991; Coleman, 1993; Sebai et al., 1991). These dykes were emplaced from the Sinai Peninsula through Arabia down to Yemen. Magmatism was less voluminous further north in Arabia, as compared to Yemen (Baldrige et al., 1991). However, abundant intraplate alkali basalts were emplaced in Arabia from 30Ma, forming the harrats (Coleman et al., 1983; Sebai et al., 1991; Camp and Roobol, 1992). These cover a large spatial extent and have been active as recently as 2009 (Pallister et al., 2010). There is no counterpart to the harrats on the western margin of the Red Sea (Duncan and Al-Amri, 2013). The origin of the harrats is generally accepted to be a response to Red Sea rifting, but the exact mechanisms are debated: a plume upwelling directly beneath the Arabian shield (Moufti and Hashad, 2005); radial flow from a single plume beneath Afar (Camp and Roobol, 1992; Lim et al., 2020); melting of lithospheric mantle (Shaw et al., 2003). The continental margins of the Gulf of Aden are volcanic and characterised by seaward dipping reflector sequences dated from 40–16Ma and high heat flow at the ocean-continent transition (Lucazeau et al., 2009; Leroy et al., 2010). Major off-axis volcanic structures are also observed by Leroy et al. (2010), who interpret the source of melt as channelised flow from the Afar hotspot along the Aden ridge.

The evolution of magmatism in Afar from the onset of Arabia-Africa separation shows its progression from a rift margin to an active spreading centre. As the rift margin developed during the Miocene, extension along the western Afar margin dominated the tectonics (Wolfenden et al., 2005; Stab et al., 2016). Coincidentally,

silicic magmatic activity occurred, such as the dominantly rhyolitic Mabla formation, emplaced from 29 to 26 Ma (Varet, 2017; Wolfenden et al., 2005; Barberi et al., 1980). The next major fissural basaltic episode occurred between 8 and 4Ma, known as the Dalha Series (Barberi et al., 1975; Barberi and Varet, 1978). Following this, magmatic activity localised into rift-marginal basins (Chernet et al., 1998; Wolfenden et al., 2005). The most extensive magmatic event in Afar was the Afar Stratoid Series. Erupted between 3.9–1.6 Ma, the Stratoid Series comprised a fissural flow that erupted over an area of 55,000 km², reaching thicknesses of 1 km (CNR-CNRS, 1973; Barberi et al., 1975; Varet and Gasse, 1978). It has long been associated with extensional processes (Barberi et al., 1980). Overlapping with the Stratoid Series is the Gulf Series lavas, located around the Gulf of Tadjoura in Djibouti (Daoud et al., 2010; Rooney, 2020c). These are similarly associated with extensional processes, specifically the propagation of strain from the western Gulf of Aden (Tadjoura Rift), resulting in a more localised magmatic expression (Rooney, 2020c).

The Gulf Series marked the commencement of focused magmatic activity in Afar. Quaternary to modern volcanism is generally located in discrete magmatic segments, coincident with graben formation, known as ‘axial volcanic ranges’ (Tazieff et al., 1972; Mohr, 1978). These magmatic segments are made up of faulted volcanic ranges with aligned chains of basaltic cones and fissural flows (Hayward and Ebinger, 1996; Barberi and Varet, 1977). The localisation of strain into these focused zones of magmatism and deformation close to the triple junction has produced overlapping extensional rift sectors parallel to the Red Sea, Gulf of Aden, and Main Ethiopian Rift trends (Manighetti et al., 2001; Wolfenden et al., 2005; Rooney, 2020c).

The temporal evolution of magmatism in the AARS reflects the broader tectonic processes affecting the mantle and crust, and thus is an important focus of study to understand the progression of continental rifting.

1.4.1 Recent volcanic activity in Afar

One of the Quaternary focused magmatic segments in Afar, the ~60 km long Dabbahu segment, was part of the 2005–2011 Dabbahu-Manda-Harraro dyke

intrusions and rifting episode. This began in September 2005, when a $\sim 2.5 \text{ km}^3$ volume of magma was injected into the upper 10 km of crust beneath the entire length of the Dabbahu segment (Wright et al., 2006). The dykes were partially fed by the Dabbahu and Gab’ho volcanoes, but this only accounted for $\sim 20\%$ of the total dyke volume (Keir et al., 2009b). This volume imbalance is explained by an additional source located in the middle to lower crust or upper mantle, identified by Keir et al. (2009b) after subsequent dyke emplacement episodes in June and July 2006. The sequence of dyke intrusions continued until 2010 (Hamling et al., 2009; Wright et al., 2012).

Another segment is the Erta Ale Volcanic Segment, consisting of six volcanic centres. Recent volcanic activity includes a November 2010 overflow of the lava lake at Erta Ale (Field et al., 2012), a November 2008 fissure eruption on Alu Dalafilla, sourced from a shallow, 1 km deep, elongated magma chamber (Pagli et al., 2012) and a January 2017 eruption at Erta Ale, fed by a sill located ~ 1.4 km below the caldera (Xu et al., 2017). Geodetic observations such as the latter two studies provide evidence for the existence of shallow axial magma lenses along the length of this magmatic segment. Dallol, located slightly to the north of the Erta Ale volcanic centres, experienced a dyke intrusion in October 2004, fed by a 2.4 km deep active magma reservoir (Nobile et al., 2012). These examples illustrate the shallow depths of magma storage regions along these rift-related magmatic segments.

1.4.2 The Nabro Volcanic Range and Nabro volcano

The Nabro Volcanic Range (NVR) is a ~ 110 km lineament of volcanoes, trending SSW-NNE from Bara’ Ale volcano in Afar to the Kod Ali formation in the Red Sea (Wiar and Oppenheimer, 2005). The NVR bisects the Danakil microplate and aligns with a marked trend in seismic activity (Goitom, 2017). Unlike the other volcanic segments in Afar, the NVR lies oblique to the dominant rifting trends, and shows little surface evidence of faulting or extensional influence (Wiar and Oppenheimer, 2005). It is thus referred to as an “off-rift” volcanic lineament. Geochemically, the NVR also shows a distinct compositional difference in erupted products to the

on-rift volcanoes in Afar, with basalts of strongly alkaline composition containing peridotite inclusions that are not observed elsewhere in Afar (Barberi et al., 1974). A conclusive explanation for the NVR’s extensive off-rift magmatism remains elusive. Theories for off-rift volcanism (e.g., Maccaferri et al., 2014) often focus on isolated volcanic centres, rather than an aligned chain of volcanoes. Potential theories include the reactivation of an older, pre-rift structure (Tazieff et al., 1972), or localised diapiric upwellings from depth (Hammond et al., 2013).

Nabro is the largest volcano in the NVR, with a summit at 2248 m above sea level and a caldera of maximum diameter 8 km (Wiart and Oppenheimer, 2005). Together with the neighbouring caldera of Mallahle, it forms the Bidu Volcanic Complex. Historically, Nabro has not been well-documented. Prior to 2011, the only geological study of Nabro was that of Wiart and Oppenheimer (2005). Their geological mapping revealed the presence of large ignimbrite deposits, trachytic lava flows and rhyolite obsidian flows and domes, indicating a history of silicic volcanism at Nabro and Mallahle. Also present are lavas with a more basaltic composition, erupted from fissures between the calderas and aligned perpendicular to the NVR trend. These are interpreted as being sourced differently to the silicic volcanism (Wiart and Oppenheimer, 2005).

1.4.2.1 June 2011 eruption

On 12 June 2011, between 20:27 and 20:40 UTC, Nabro volcano underwent its first eruption on historical record. The lack of any local or regional geophysical surveillance networks and no known historical eruptive activity meant that the eruption began with very little warning, but a rapid response from emergency services enabled the evacuation of the settlements in and around the caldera (Goitom et al., 2015). In total, approximately 12,000 people were displaced and seven people lost their lives during the course of the eruption (Goitom et al., 2015).

Precursors to the eruption were 7 seismic events of local magnitudes 2.8 – 4.5 in the period 23 February to 11 June 2011, with the largest occurring within Nabro’s caldera on 31 March 2011 (Lapins, 2021a; Ogubazghi et al., 2014). This

event left visible tensional cracks trending NW-SE, parallel to the Red Sea rift trend (Ogubazghi et al., 2014). In the hours prior to the eruption, a seismic swarm and sustained tremor began, which were detectable regionally (Lapins, 2021a). Two ‘notable’ earthquakes occurred around the onset of eruptive activity, and infrasound signal was detected at distances as far as 1708 km away (Goitom et al., 2015; Fee et al., 2013).

The eruption was notable for the vast amount of SO_2 erupted. It released 4.5 Tg within the first 15 days, producing the largest stratospheric aerosol perturbation since the 1991 Pinatubo eruption and the single largest SO_2 column ever retrieved from space (Clarisse et al., 2012; Theys et al., 2013; Fromm et al., 2014). The increased fumarolic activity in the months following the eruption indicates that Nabro could be an important source of geothermal energy in the future (Yohannes, 2012).

The lava flows emplaced by the eruption are highly crystalline and have a composition of trachybasalt to trachybasaltic andesite (Goitom et al., 2015). The longest lava flow extended 17.5 km to the NW of Nabro (Goitom et al., 2015). Petrological analysis identifies two distinct batches of magma with zoning patterns that suggest a complicated storage history (Donovan et al., 2018). One batch is more primitive, while the other is higher in sulfur and water content. The latter batch likely underwent isobaric crystallisation in a storage region at $\sim 5\text{--}7$ km below sea level (b.s.l.), with the former batch rising rapidly to the shallow crust from depth (Donovan et al., 2018).

Observed volcanic tremor remained at a steady rate until 16 June 2011 (Goitom et al., 2015), when it fell to almost nothing. The next day, the largest event during the entire syn-eruptive period (5.9 local magnitude) occurred beneath Mallahle and tremor resumed (Goitom et al., 2015). Other event epicentres on 17 June follow a rough alignment with the NVR of SSW-NNE (Goitom et al., 2015); subsequent to this, the orientation of event locations rotates to a NW-SE trend. The seismicity likely relates to changes in the local stress regime due to post-eruptive subsidence or faulting (Goitom et al., 2015; Hamlyn et al., 2014, 2018).

In the months following the eruption, Nabro experienced subsidence at a slowly decaying rate. The source of this subsidence is modelled as a deflating Mogi source at a depth of 6.4 ± 0.3 km b.s.l. (Hamlyn et al., 2014, 2018). The Mogi source is thought to relate to outgassing or removal of magma from a reservoir, or the relaxation of a visco-elastic shell (Hamlyn et al., 2018).

A temporary seismic array was established around Nabro, operational from 31 August 2011 until October 2012. From this seismic data, Lapins et al. (2021) calculated hypocentre locations, local and moment magnitudes, path/site attenuation effects and b-values. This analysis revealed that the seismicity beneath Nabro delineates an aseismic zone, inferred to be a magma storage region due to its consistency with the modelled Mogi sources from previous studies (Goitom et al., 2015; Hamlyn et al., 2014, 2018) and the magma storage depth proposed by petrological analysis (Donovan et al., 2018). Patterns of seismicity indicate that fluids are continuing to migrate in the subsurface, potentially triggering movement on faults (Lapins, 2021a).

1.5 Seismic methods

Seismology provides us with the unique opportunity to probe the Earth’s interior on multiple scales. Earthquakes release energy in the form of seismic body waves and surface waves. Body waves are those which propagate directly through the Earth: compressional P-waves and transverse S-waves. As these seismic waves travel, they accumulate information about the different mediums they pass through—for example, a change in velocity or a conversion from one form of body wave to another can indicate a compositional difference between two layers. When the seismic signals are recorded on the surface, the information can be decoded through several different innovative techniques, revealing the internal structure and composition of the Earth.

Teleseismic earthquakes are those occurring at distances greater than 1,000 km from the measurement (seismograph) site. The sensitivity of modern seismographs means that energy from large magnitude ($M_b \geq 5$) earthquakes can be detected at sites all across the globe. This is useful, because it means that seismic events do

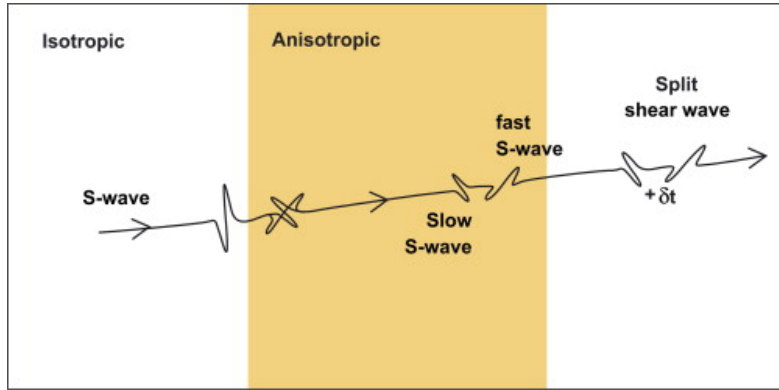


Figure 1.8: Shear-wave splitting occurs for a shear wave traveling through an anisotropic medium (yellow). When an incident shear wave arrives at an anisotropic medium, it splits into two shear waves of perpendicular polarisation, along seismic fast and slow directions, respectively. Traveling through the anisotropic medium, the two waves accumulate a delay time δt . Shear-wave splitting techniques invert for δt and the fast polarization direction ϕ . Figure and caption from (Wüstefeld et al., 2008).

not have necessarily have to occur locally to a region in order for seismic methods to be used to investigate its substructure.

1.5.1 Shear-wave splitting

Anisotropic media show a directional dependence of certain parameters. In the case of seismic anisotropy, this parameter is shear wave velocity. When a polarised seismic shear wave enters an anisotropic medium, it splits into two almost-orthogonal shear waves, which travel at different velocities (Figure 1.8). This process is known as shear-wave splitting and is quantified by two parameters: ϕ , the polarisation direction of the fast shear wave, and δt , the time lag between the fast and slow shear waves.

1.5.1.1 SKS phase and calculation of splitting parameters

The SKS phase propagates as an S-wave from the source to the core-mantle boundary, where it transforms into a P-wave to pass through the liquid outer core. Upon emerging from the core, it reverts back to an S-wave. The transformation into a compressional P-wave has the effect of removing anisotropy encountered by the wave between the source and the core. Therefore, the shear-wave splitting observed for the SKS phase solely describes the anisotropic medium (or media) encountered by the wave between the core and the receiver.

Seismometers measure motion in three orthogonal directions: vertical, radial and horizontal. The vertical component is directed vertically from the seismometer. The radial component is directed along the great circle path between the earthquake epicentre and the seismometer. The transverse component is the horizontal direction orthogonal to the radial component.

The transformation into a compressional P-wave has an effect on the polarisation of the final upgoing S-wave. The particle motion of this S-wave will be entirely within the vertical plane defined by the ray path, a polarisation denoted by “SV”. If the ray path between the core and the receiver is isotropic, the SKS phase will continue to exhibit linear particle motion perpendicular to propagation in the “SV” direction and consequently will only be visible on the radial and vertical components of a seismometer. However, passage through an anisotropic medium will split a shear wave into two orthogonal components travelling at different velocities and producing elliptical particle motion. In the case of the SKS phase, this will produce measurable energy on the transverse component of the seismometer; i.e., particle motion will no longer be confined to the vertical plane defined by the ray path.

To reverse the effect of anisotropy, the horizontal components must be rotated by $-\phi$ and their relative positions shifted by δt . This will linearise the particle motion and minimise the energy on the transverse component. The eigenvalue minimisation method, as formulated by Silver and Chan (1991) and automated by Teanby et al. (2004), searches for the combination of ϕ and δt which best linearise the SKS motion within a particular time window.

For certain measurements, no shear-wave splitting of the SKS phase is observed. These are known as null events, and they provide useful information: either that the structure beneath a station is close to isotropic or that the initial polarisation of the shear wave is aligned with the fast or slow axis of anisotropy.

1.5.1.2 Causes of seismic anisotropy

Interpreting shear-wave splitting observations can be challenging, because it is difficult to localise the depth of the anisotropic medium. The lower mantle is

considered isotropic, aside from the lowermost D'' layer (Kendall and Silver, 1996). The upper mantle, by contrast, contains large amounts of anisotropic olivine crystals, which adopt preferred orientations under conditions of strain (Nicolas and Christensen, 1987). Anisotropy can also arise from periodic thin layering of contrasting materials (e.g., Kendall, 2000). It is generally assumed that anisotropy is mainly accrued in the upper mantle, with additional contributions from the D'' layer and crustal anisotropy. However, there are often multiple mechanisms acting to produce this strain-induced orientation or layering of materials, and it is difficult to discriminate between lithospheric and/or asthenospheric sources in the upper mantle. Dynamic processes such as asthenospheric flow, mantle upwellings and aligned melt intrusion zones within the lithosphere can all produce anisotropic fabric. This fabric can also be frozen into the lithosphere from past deformation events (e.g., Al-Lazki et al., 2013). Thus, shear-wave splitting observations have to be considered alongside other geophysical observations to deduce the probable cause(s) of seismic anisotropy.

1.5.1.3 Previous measurements in the Afro-Arabian Rift System

Previous anisotropy measurements around the Afar Triple Junction and the Arabian plate reflect the tectonic complexity of the region. On the Western Ethiopian plateau, the anisotropy is consistent with pre-rift orogenic fabric, with two layers observed in regions with significant aligned melt (Hammond et al., 2014b; Gashawbeza et al., 2004). Large parts of the Afar Depression show only one layer of anisotropy, with fast directions oriented NE; two layers become ubiquitous moving south into the Main Ethiopian Rift. Like Afar, the lower layer in the Main Ethiopian Rift has fast directions oriented NE, which is linked to lattice preferred orientation of olivine crystals due to flow from the LLSVP beneath Africa, rather than a localised conduit-like plume (Bastow et al., 2010; Hammond et al., 2014b). Local deviations from this overall flow direction are observed close to locations of upper mantle diapiric upwellings. The upper layer of anisotropy in the Main Ethiopian Rift is associated with oriented melt pockets (Ayele et al., 2004; Kendall et al., 2005; Bastow et al., 2010; Hammond et al., 2014b). Overall, the anisotropy results can be modelled as

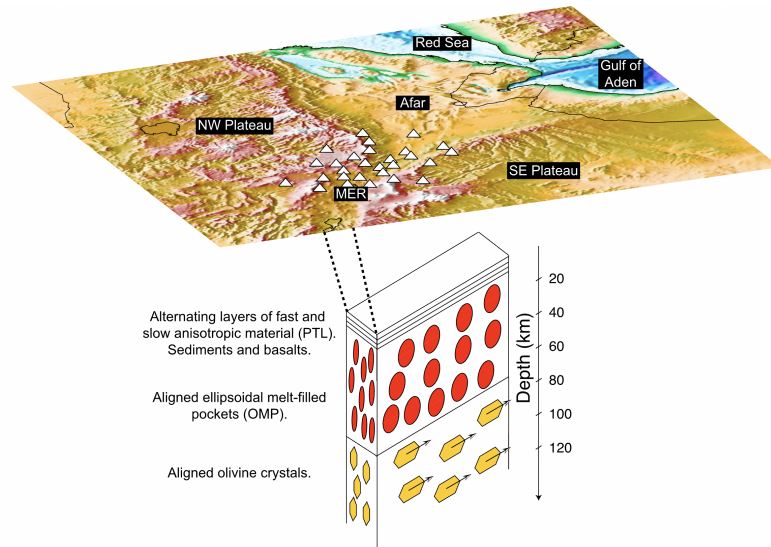


Figure 1.9: Suggested anisotropy mechanisms of the Ethiopian rift mantle: layering of seismically fast and slow layers in the top ~ 10 km and underlying oriented melt pockets due to melt intrusion into the Main Ethiopian Rift lithosphere. Lattice preferred orientation in the asthenosphere due to large-scale mantle flow associated with the superplume. Figure and caption from Bastow et al. (2010).

low aspect ratio melt inclusions (e.g., dykes and veins) intruding into an extending plate during late stage breakup (Figure 1.9; Bastow et al., 2010).

Studies using local earthquakes have also quantified anisotropy of the upper crust in Afar and Ethiopia. Keir et al. (2011a) map the strain field around the Afar Triple Junction using shear-wave splitting of local earthquakes. They find that deformation-related structures control anisotropy measurements across Afar. A dramatic increase in splitting magnitude and concomitant rotation of the fast direction is observed into the rift axis, which Keir et al. (2011b) propose is due to an increase in density of dyke-induced faulting combined with oriented melt pockets beneath active volcanic segments. Illsley-Kemp et al. (2017) extend this crustal anisotropy analysis further into northern Afar. Crustal anisotropy measurements within the rift are generally oriented perpendicular to the rift axis (NE-SW), changing to a W-E orientation at the rift margin onto the Western Ethiopian Plateau. However, Illsley-Kemp et al. (2017) observe regions north and south of the Danakil Depression where seismic anisotropy is rift-perpendicular. These regions are further from the rift axis and so the stress field is dominated by the gravitationally induced stress

field originating from variations in crustal thickness. Crust within the rift axis is characterised by rift-aligned structures and melt inclusions, implying that the rift axis is currently accommodating tectonic extension. Further south, Keir et al. (2005) investigate anisotropy of the upper crust beneath the northern Main Ethiopian Rift. Along the rift axis, the fast polarisation is oriented parallel to Quaternary-Recent faults, aligned volcanic cones and the direction of maximum horizontal stress. The magnitude of splitting is greatest in regions of shallow partial melt, and decays away from the rift axis, consistent with reduced melt on the rift margins.

On the eastern side of the Red Sea rift, one layer of anisotropy is observed across Arabia, with the fast direction of seismic anisotropy oriented in a N-S direction (Hansen et al., 2006). This is interpreted as the combination of two different asthenospheric flows. One is directed NE, driving the absolute plate motion of Arabia, and the other is oriented NW, associated with a channelised upwelling travelling along the Red Sea rift. South of the Gulf of Aden, Ali et al. (2023) identify two distinct layers of anisotropy: a crustal layer, associated with the opening of the rift in Oligocene times; and a deeper mantle layer, with anisotropy aligned parallel to the regional Nubia-Arabia absolute plate motion.

North of the Gulf of Aden, different origins of anisotropy are proposed for observations made in Yemen. Hammond et al. (2014b) observe two layers of anisotropy, with upper layer orientations that are attributed to fossilised fabric frozen into the pre-Cambrian lithosphere, whilst the lower layer indicates that mantle flow is being deflected along the Gulf of Aden trend. Gallacher (2016), however, explains the prevalence of null measurements in his study as a consequence of either strong anisotropic heterogeneity or a vertical axis of mantle flow, most likely in connection with a combination of small-scale convection and edge driven convection occurring between the mid-ocean ridge axis and the margins of the Gulf of Aden. Further east in Oman, Al-Lazki et al. (2013) interpret observations of distinct abrupt lateral anisotropy variation as representative of fossil lithospheric mantle anisotropy. Across the Arabian plate, Qaysi et al. (2018) observe a systematic decrease in the magnitude of anisotropy from beneath the Arabian Shield to the Persian Gulf.

1.5.2 Receiver functions

When a P-wave impinges on a sharp velocity contrast, part of its energy is converted into an S-wave. These are referred to as ‘Ps’ phases, with the lower-case ‘s’ indicating that the S-wave is travelling upwards after conversion. The Ps phases are observed following the arrival of a direct P-wave. Also observable are subsequent reverberations within a layer: the PpPs phase and the co-arriving PpSs + PsPs phases (Figure 1.10a).

Receiver functions are calculated from three-component seismograms to capture these Ps conversions and their reverberations (Figure 1.10; Langston, 1979). The vertical component of motion can be used as a good approximation of the event source (e.g., Jenkins et al., 2020) due to the incidence angle of teleseismic waves being close to vertical (27° or lower). To isolate the response from local crustal structure along this incoming, near-vertical raypath, the vertical component is deconvolved from the radial components. Deconvolution removes information common to both: the instrument response, path and source effects. The resulting time-series are known as receiver functions, containing the phase arrivals of the Ps conversions and reverberations (Figure 1.10b). The amplitudes of the arrivals depend on the P-wave incidence angle (ray parameter), the size of the velocity contrast, and whether the wave passes from a high velocity layer to lower velocity layer, or vice versa. The arrival times of the converted phase and reverberations depend on the depth of the velocity contrast, the P- and S-wave velocities between the contrast and the surface, and the ray parameter.

Receiver functions are most commonly used to investigate the velocity contrast of the Mohorovičić discontinuity (the Moho), which marks the boundary between the crust and the mantle (e.g., Zhu and Kanamori, 2000; Stuart et al., 2006; Thompson et al., 2010; Gallacher and Bastow, 2012; Ogden and Bastow, 2022). The H- κ technique, developed by Zhu and Kanamori (2000), uses the travel times of the Moho Ps conversion and its reverberations to determine the bulk crustal properties beneath a seismograph station. These bulk crustal properties are crustal thickness (H) and the velocity ratio of P- to S-waves (V_P/V_S ; κ). The stacking function

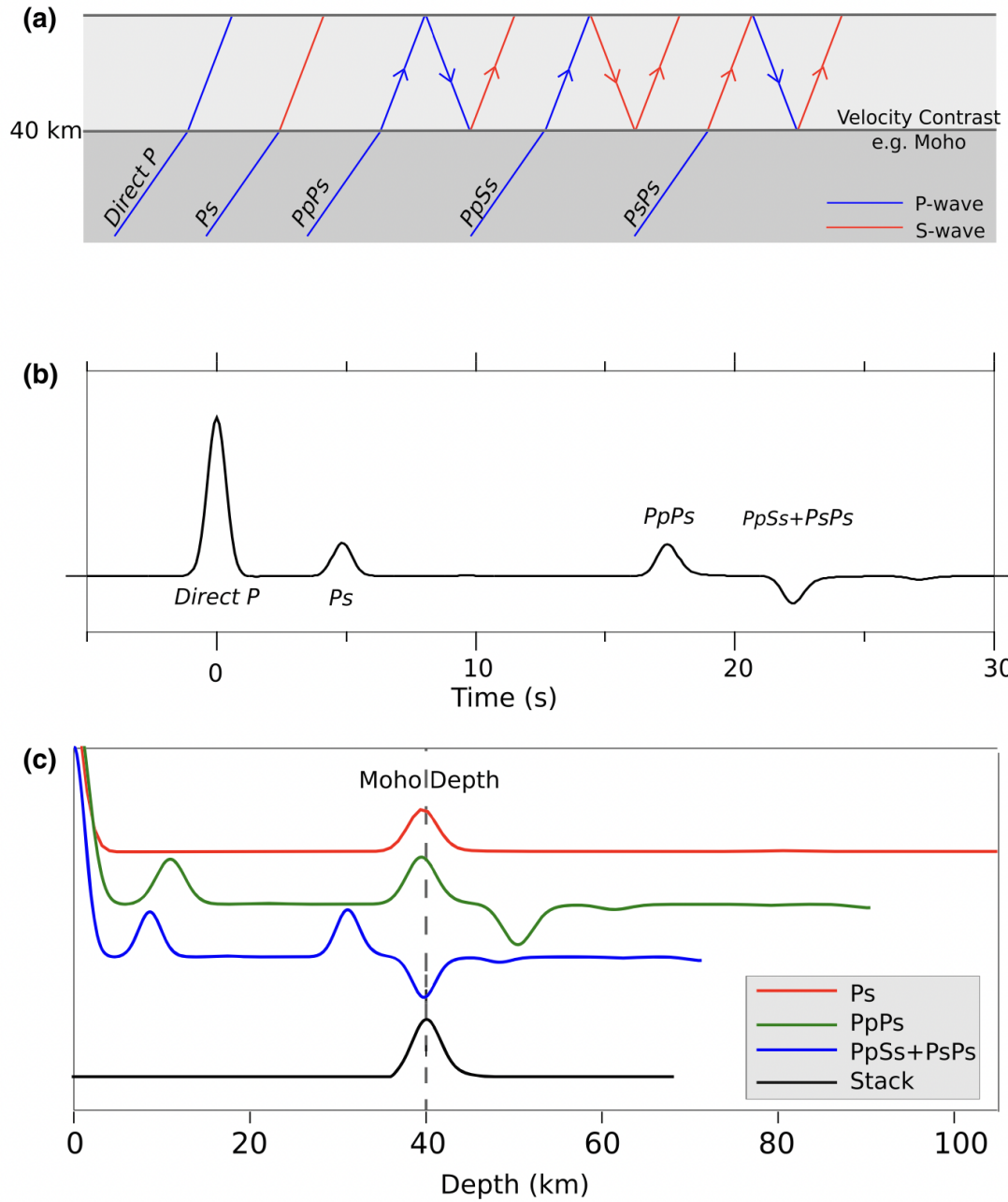


Figure 1.10: (a) P-to-S converted phases of interest in crustal imaging. (b) Example of a synthetic RF highlighting phases shown in (a). (c) Example of a multiphase time-depth conversion and multiphase stacking procedure. Figure and caption from (Jenkins et al., 2020).

$s(H, \kappa)$ is determined for plausible values of crustal thickness and V_P/V_S ratio using a grid search according to:

$$s(H, \kappa) = \sum_{j=1}^N w_1 r_j(t_1) + w_2 r_j(t_2) - w_3 r_j(t_3) \quad (1.1)$$

where $r_j(t_i)$ are the receiver function amplitudes at the predicted arrival-times of the Ps conversion (t_1) and subsequent reverberations (PpPs, t_2 ; PpSs + PsPs, t_3) for the j th receiver function. w_1, w_2, w_3 are stacking weights for phases Ps, PpPs and PpSs + PsPs, respectively, with $\sum_{i=1}^3 w_i = 1$. N is the total number of receiver functions used. The traveltimes are calculated as follows:

$$t_1 = H \left[\sqrt{\frac{1}{V_S^2} - p^2} - \sqrt{\frac{1}{V_P^2} - p^2} \right] \quad (1.2)$$

$$t_2 = H \left[\sqrt{\frac{1}{V_S^2} - p^2} + \sqrt{\frac{1}{V_P^2} - p^2} \right] \quad (1.3)$$

$$t_3 = 2H \sqrt{\frac{1}{V_S^2} - p^2} \quad (1.4)$$

where p is the ray parameter, or slowness (s km^{-1}). The stacking function reaches its maximum as the three phases stack coherently and thus represents the best estimate of H and κ at a certain station.

Ogden et al. (2019) showed that H - κ stacking can be unreliable in regions of complex tectonic structure where the Moho is gradational or intracrustal structure exists, such as near-surface sedimentary layers (Figure 1.11). A rigorous exploration of H - κ parameter space (chosen V_P , phase weights, choice of stacking procedure, receiver function frequency content) is therefore necessary to produce accurate results.

1.5.2.1 Previous P-to-S receiver function studies

P-to-S receiver function studies in Afar and Ethiopia have revealed considerable crustal heterogeneity across the region (Figure 1.12). Results show a variation in crustal thickness, from 40–45 km beneath the western Ethiopian plateau, decreasing to 20–26 km in central Afar and then thinning northward into the Danakil Depression to ~ 17 km (Berckhemer et al., 1975; Makris and Ginzburg, 1987; Mackenzie et al., 2005; Dugda et al., 2005; Maguire et al., 2006; Stuart et al., 2006; Cornwell et al.,

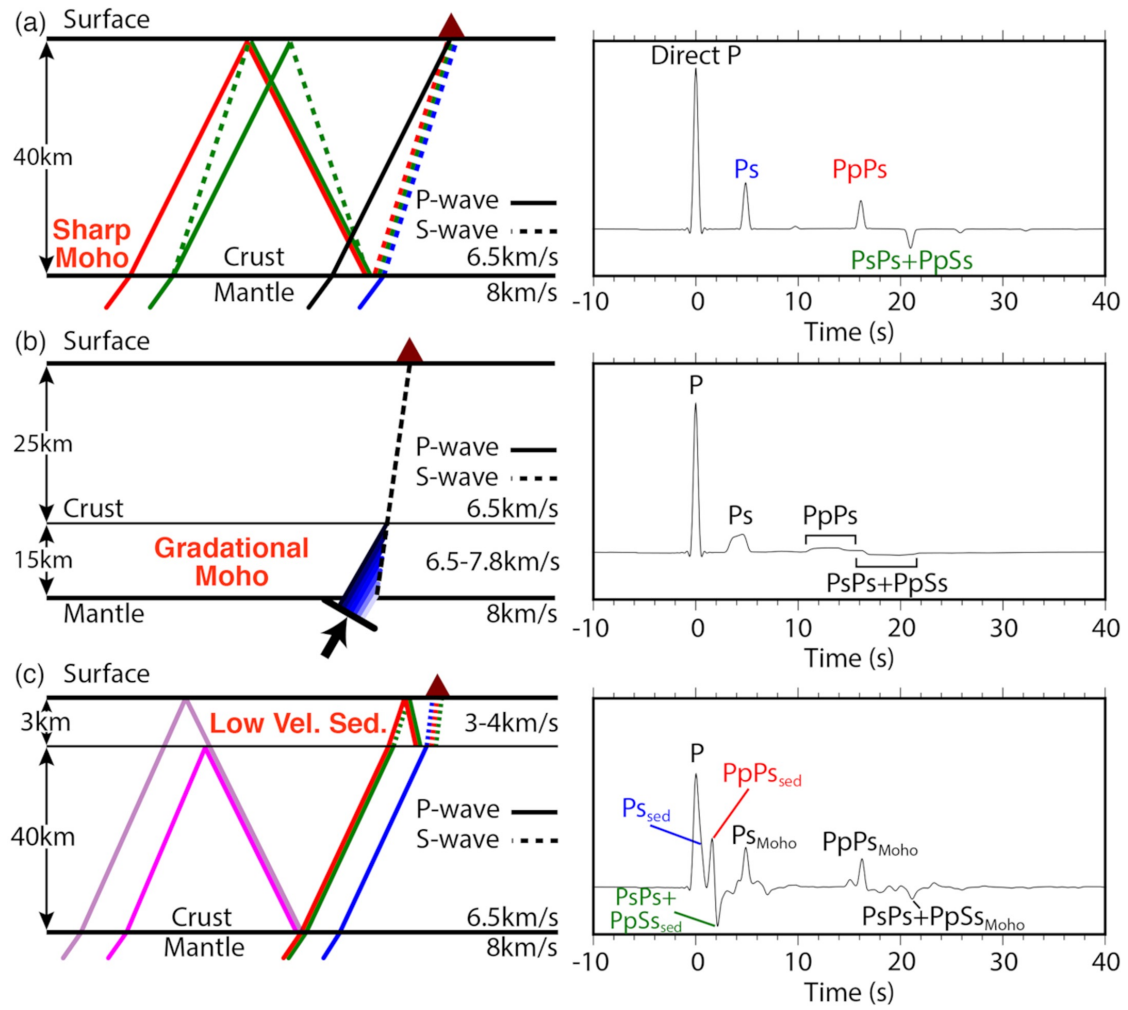


Figure 1.11: The impact of crustal structure on receiver functions. (a) Sharp Moho with high amplitude, impulsive P-to-S conversions. (b) A gradational Moho, for which P-to-S conversions occur over a large depth range. Receiver function signals weaker and more diffuse. (c) When near-surface layers exist, resulting P-to-S conversions can distort Moho signals. Figure and caption from (Ogden et al., 2019).

2010; Hammond et al., 2011c; Reed et al., 2014; Ahmed et al., 2022). High values of V_P/V_S ratio are observed across the region. Values of 1.7–1.9 beneath the plateau are higher than typical silicic continental crust V_P/V_S ratios, thus evidencing mafic alteration of the crust (Hammond et al., 2011c; Ahmed et al., 2022). Further elevated V_P/V_S ratios of > 2.0 within central Afar indicate widespread partial melting due to magmatic intrusions (Hammond et al., 2011c; Reed et al., 2014; Ahmed et al., 2022). The melt is likely stored in interconnected stacked sills in the lower crust (Hammond, 2014).

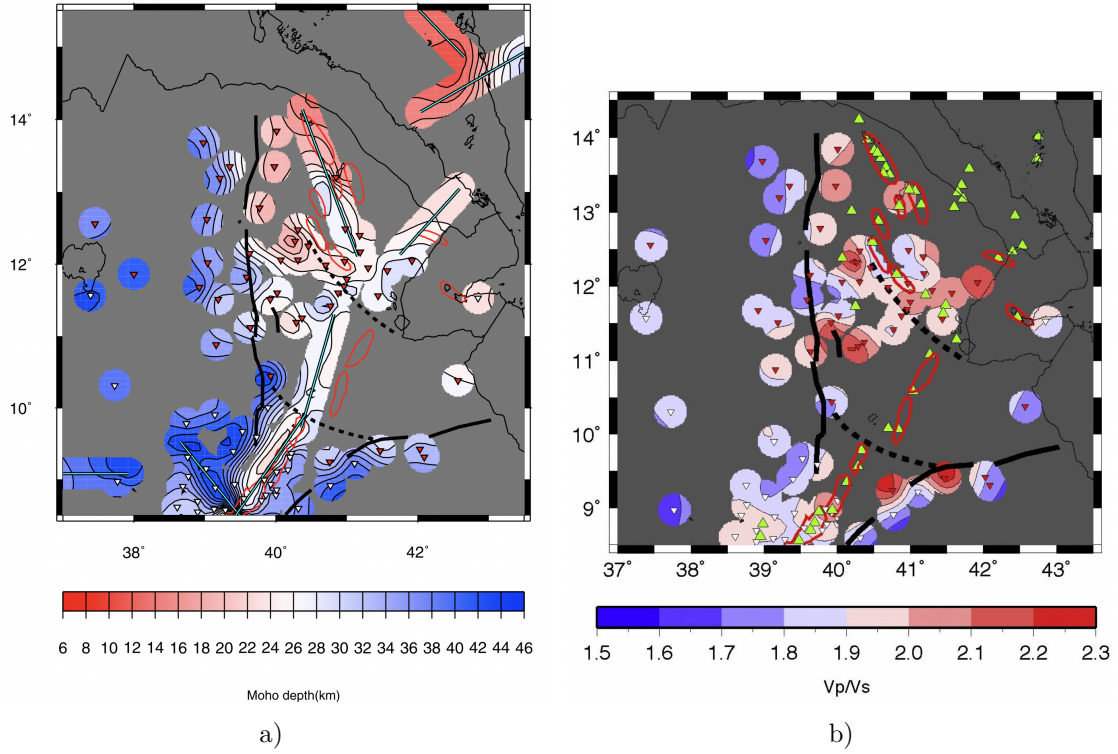


Figure 1.12: Moho depth and V_P/V_S ratios across Afar and Ethiopia, from the study of Hammond et al. (2011c). Black lines show the border faults separating Afar from the western and southeastern plateaus. The dashed line shows the Tendahao-Gob'a discontinuity (northern line) and arcuate accommodation zone (Tesfaye et al., 2003) (southern line). The red lines demarcate the Quaternary volcanic segments. Regions shaded grey mask regions with no data coverage. Green triangles show the location of active volcanoes. a) A map of Moho depth based on receiver function results from Hammond et al. (2011c) and previous receiver function studies (Dugda et al., 2005; Dugda and Nyblade, 2006; Stuart et al., 2006) and controlled source work (Berckhemer et al., 1975; Makris and Ginzburg, 1987; Egloff et al., 1991; Mackenzie et al., 2005; Maguire et al., 2006). Triangles indicate the location of receiver function Moho depth estimates (red: the study of Hammond et al. (2011c); white: previous studies), and the cyan lines show the location of controlled source Moho depth estimates. b) A map of V_P/V_S based on receiver function results from this and previous receiver function studies (Dugda et al., 2005; Dugda and Nyblade, 2006; Stuart et al., 2006). Inverted triangles indicate the location of receiver function V_P/V_S estimates (red: the study of Hammond et al. (2011c); white: previous studies).

1.5.3 Seismic body-wave velocity tomography

Seismic body-wave velocity tomography is a technique that inverts seismic data to constrain the 2D or 3D P- and S-wave velocity structures and the ratio between them (V_P/V_S ratio). The specific data used are arrival times of seismic body waves from earthquakes.

The seismic tomography method involves the calculation of the ‘forward problem’ and then solving the ‘inverse problem’. The forward problem is predicting how long it takes for a seismic wave to travel from source to receiver (traveltimes), given an initial velocity model. If the initial velocity model is correct, the arrival times observed in the data will be exactly the same as the predicted ones. However, if an anomalous velocity structure is present in the subsurface (such as a magma body, or a region of hydrothermal fluid), the observed data will differ. Seismic tomography uses these differences in traveltimes to reconstruct anomalies. An image is built up by numerous seismic raypaths traversing the same anomaly, each with their own difference to the predicted arrival time depending on the anomaly structure, as illustrated in Figure 1.13.

Thus, the inverse problem is that of finding a new velocity model such that the differences between the observed and predicted arrival times (known as traveltimes residuals) are minimised. It is an inherently non-linear problem, because the traveltimes are determined from earthquake locations, which themselves depend on knowledge of the seismic velocity structure. The raypaths themselves also rely on the model parameters that are being solved for. The relationship:

$$\mathbf{d} = \mathbf{g}(\mathbf{m}), \quad (1.5)$$

forms the basis of any tomographic method, where $\mathbf{d}$ is an array of data (e.g., traveltimes), $\mathbf{m}$ is an array of model parameters (e.g., velocity), and $\mathbf{g}$ is some functional relating the model to the data. The form of $\mathbf{g}$ depends on the chosen representation of Earth structure.

Typically, the inverse problem is linearised and then iterated several times to converge onto (hopefully) the correct non-linear solution. The FMTOMO code

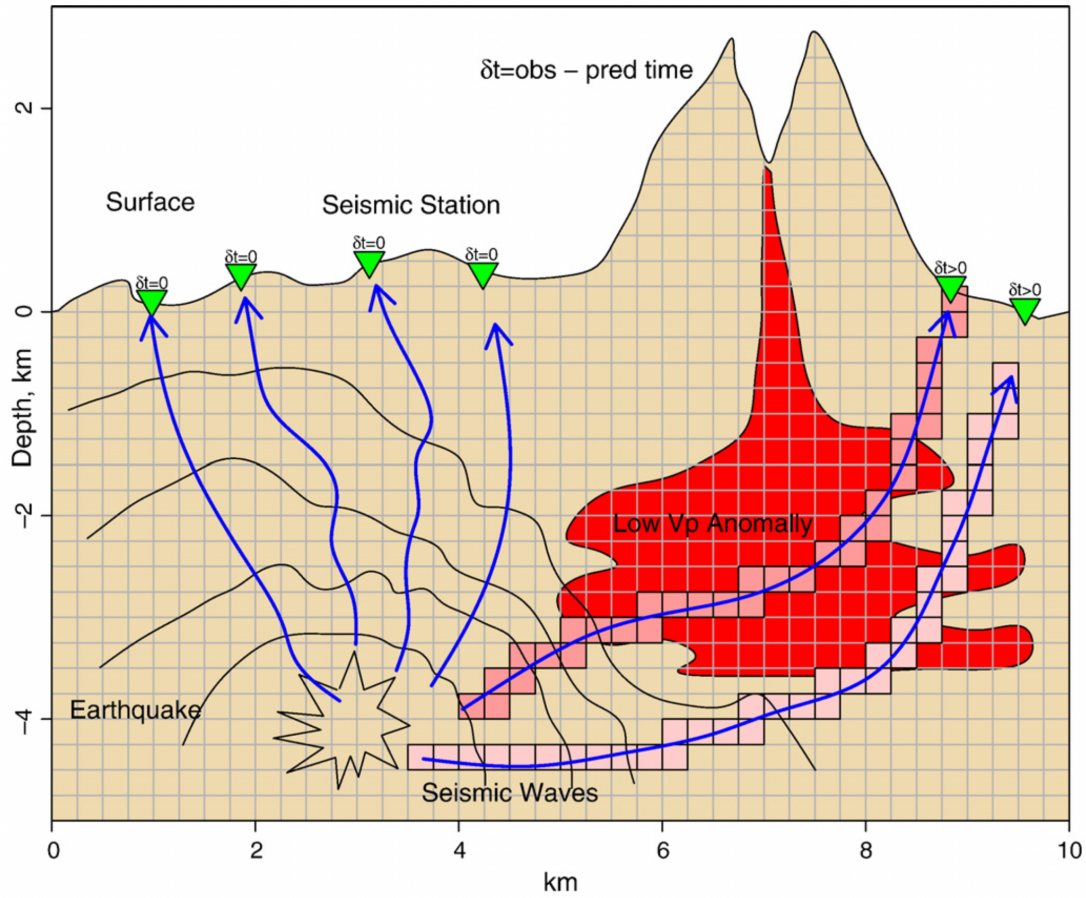


Figure 1.13: Sketch showing earthquake raypaths used for a hypothetical tomographic inversion. Raypaths that do not penetrate the anomalous body have a small residual and do not contribute significantly to changing the model. Those ray paths that pass through a low velocity magma region will have a positive anomaly, $\delta t > 0$. The travel-time difference can be projected back along the raypath by distributing the residual along the path. Colored blocks show two different rays intersecting the anomalous region, each with a different level of perturbation. Where paths intersect back-projections will constructively interfere and image reconstruction is attained. Figure and caption from (Lees, 2007).

developed by Rawlinson and Urvoy (2006) and modified by Pilia et al. (2013) is an example of a fast, robust software package that recovers 3D P-wave, S-wave and V_P/V_S velocity structure, as well as relocating earthquake hypocenters. It parameterises the Earth into a grid of nodes, each associated with a certain velocity, which control the values of a cubic B-spline continuum. To produce a 3D seismic tomography model, multiple iterations are performed, each of which involves:

- the solution of the forward problem through the current velocity model;

- an inversion for velocity parameters using a subspace inversion scheme (Kennett et al., 1988);
- hypocenter relocation based on both P- and S-wave arrival times, using a fully non-linear grid search in the presence of the updated velocity model.

The process is then repeated with the updated velocity model as the new input for the solution of the forward model. This continues until convergence is achieved: when the traveltime residuals change only a minimal amount between inversions.

The tomographic inversion problem is under-determined (not all model parameters are well-constrained by the data alone), and therefore solutions are highly non-unique. To reduce the non-uniqueness of the solution, regularisation terms are included. These act to find a solution model that is similarly smooth and as close to the initial model as possible, whilst still satisfying the data. FMTOMO introduces the smoothing and damping parameters, η and ϵ , which can be tuned to minimise the data fit, model variance and model roughness.

Despite regularisation constraints, many different models are still able to satisfy the data, meaning that it is imperative to assess the solution robustness before evaluation of the estimated model. This is commonly done using synthetic resolution tests. Parameterisations that involve continuous variations in seismic velocity typically do this using resolution tests that attempt to reconstruct a synthetic model using the same source-receiver configurations as the observed data. Examples of synthetic models are “checkerboard tests”, where the synthetic model consists of alternating regions of high and low velocity perturbations on different length scales; or “spike tests”, where an isolated spike is placed at a region where a velocity anomaly is observed in the solution model. The resolution tests are assessed on the basis of the similarity between the recovered model and the initial synthetic model. The smallest wavelength structure that is recovered from the synthetic model indicates the limit of resolution of anomalies in the solution model.

1.5.3.1 Regional seismic tomography studies in the AARS

Many regional studies throughout the AARS have used teleseismic body waves to conduct seismic tomography (e.g., Benoit et al., 2006; Park et al., 2007; Bastow et al., 2008; Park et al., 2008; Hansen et al., 2012; Tang et al., 2019; Lim et al., 2020; Corbeau et al., 2014; Civiero et al., 2022; Koulakov et al., 2016; Chang et al., 2011; Korostelev et al., 2014; Civiero et al., 2015, 2016; Hammond et al., 2013). The resolving power of these seismic tomography models varies, but typically resolution can be achieved at depths of at least ~ 600 km, with some models extending deeper into the mantle. The models are therefore capable of describing broad-scale mantle velocity variations and anomalies. Of particular interest in the AARS is the extent and influence of the African LLSVP and how this deep, hot mantle material upwells from the core-mantle boundary. Thus, seismic tomography studies have played a large role in the debate over the form and number of mantle plumes in the region (see Section 1.3).

Seismic tomography models have revealed broad regions of low P- and S-wave velocities in the mantle beneath the Main Ethiopian Rift and Afar (e.g., Benoit et al., 2006; Bastow et al., 2008; Hammond et al., 2013), the Red Sea and Arabian Shield (e.g., Koulakov et al., 2016; Park et al., 2007, 2008; Hansen et al., 2012; Tang et al., 2019; Lim et al., 2020), and the Gulf of Aden (e.g., Corbeau et al., 2014; Civiero et al., 2022). The Arabian Platform is associated with a high-velocity anomaly imaged down to 200–250 km depth (Koulakov et al., 2016). Smaller zones of low velocity extending through the upper mantle are observed throughout Afar and the Main Ethiopian Rift, interpreted as zones of localised asthenospheric upwelling feeding magmatism in the region (e.g., Hammond et al., 2013; Civiero et al., 2015, 2016).

1.5.3.2 Seismic velocity variations at volcanoes

Seismic tomography has frequently been applied on a local scale to deforming volcanoes, in order to better understand the subsurface active magmatic processes responsible for volcanic unrest (e.g., Patane et al., 2002; Chiarabba and Moretti, 2006; Korger and Schlindwein, 2014; Greenfield et al., 2016; Wilks et al., 2020; Koulakov

et al., 2021). Active volcanoes are often well-monitored and have a high rate of seismic activity, meaning there is plenty of local earthquake traveltime data available for seismic tomography inversions. See Lees (2007) for an overview of seismic tomography inversions for magmatic structure in a variety of tectonic settings.

Mapping seismic velocity variations gives information on a variety of physical parameters, including rock characteristics (composition, porosity, fractures, mineralogy), saturation conditions, presence of fluids (gases or liquids), temperature, and pressure. These physical properties often interplay or co-exist, making it a challenge to interpret observed seismic velocity anomalies. The V_P/V_S ratio is often used alongside P- and S-wave models, because it is sensitive to the type of fluid present and can distinguish between regions of partial melt or hydrothermal fluids, both of which are encountered beneath volcanoes. In saturated or partially-saturated rocks, the content and physical state of fluids has a greater effect on P-wave velocities than S-wave velocities (Vanorio et al., 2005). Fluid phase transitions induce changes in fluid compressibility and thus bulk modulus (Ito et al., 1979; Wang and Nur, 1986). Shear moduli are little affected by fluid phase transitions and hence S-wave velocities change insignificantly due to a density effect, meaning that low V_P/V_S ratios tend to characterise gas-bearing rocks (i.e., those with high fluid compressibility) whereas liquid-bearing rocks (with low fluid compressibility) are characterised by high V_P/V_S ratios (Vanorio et al., 2005).

The only tomography study done on a volcanic system in the AARS is that of Wilks et al. (2020) at Aluto volcano, located in the Main Ethiopian Rift. They identify a large low velocity zone with high V_P/V_S ratio at depths of 4–9 km below sea level (b.s.l.), interpreted as a melt-bearing ductile region. A shallower low seismic velocity, very low V_P/V_S zone is interpreted as a hydrothermal system, containing gases exsolved from the magma storage zone through a fault and fracture zone. Their findings demonstrate the coupled nature of magma and hydrothermal systems beneath this deforming volcano.

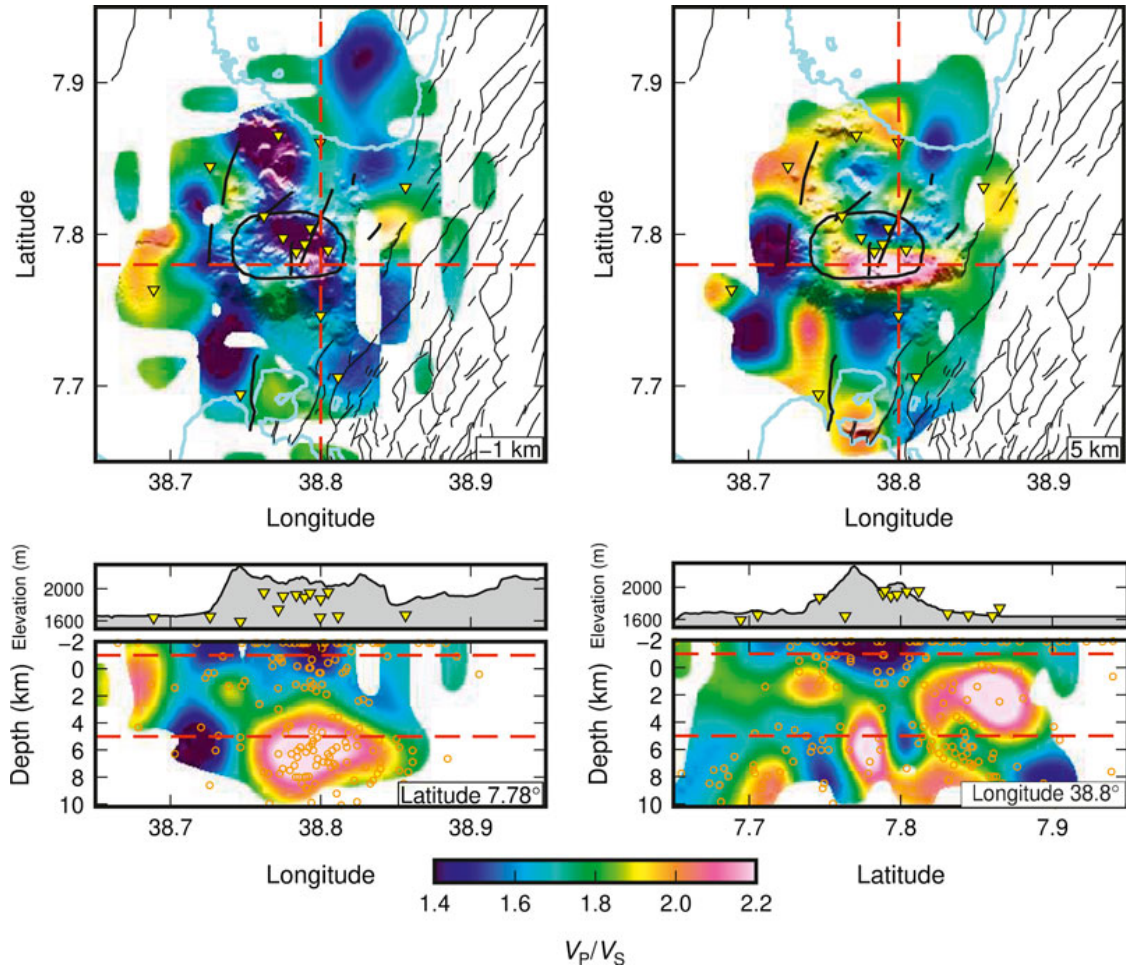


Figure 1.14: The Aluto seismic tomography model for V_P/V_S ratio. Depth slices are plotted at 1 km above sea level (top left) and 5 km below sea level (top right). East–west and north–south cross-sections are also shown (bottom left and right). Yellow triangles indicate station locations, black lines on the depth slices indicate mapped fault locations and small open circles on the cross-sections indicate event locations. Topography is shown above the cross-sections and indicated by hillside shading on the depth slices. Figure from Wilks et al. (2020).

1.6 Thesis overview

In this thesis, I examine the below-surface structure near the Afar Triple Junction at three different depths: the upper mantle (> 200 km), the crust (> 50 km) and the upper crust (> 15 km). I investigate deep mantle flow and crustal properties on a regional scale to place constraints on the driving forces of rifting in the mantle and understand how these tectonic processes manifest themselves within the crust. I also carry out a study of Nabro volcano to analyse magmatism on a local scale.

Chapter 2 uses shear-wave splitting of the SKS phase to investigate seismic anisotropy in northeastern Afar and southwestern Arabia. I aim to:

- elucidate mantle flow dynamics by constraining the orientation and magnitude of seismic anisotropy in these areas;
- quantify the influence of density-driven flow from the African LLSVP on the Afar Triple Junction;
- investigate the coherence of mantle flow along the three rift branches of the Afro-Arabian Rift System;
- understand the impact of asthenospheric flow on Arabian plate motion and continental breakup in the region.

In Chapter 3, I calculate receiver functions across northeastern and central Afar, using the H- κ stacking method. The aims of this chapter are to:

- constrain bulk crustal properties (crustal thickness and bulk V_P/V_S ratio);
- determine the crustal structure of the Danakil block;
- investigate the existence of partial melt across the region;
- quantify regional topographic dynamic support from the mantle;
- further our understanding of continental breakup.

Chapter 4 presents the results of a seismic tomography study examining the subsurface structure of Nabro volcano in Eritrea. I aim to:

- use a catalogue of earthquakes located by machine learning methods to produce 3D seismic models of P-wave, S-wave and V_P/V_S structure beneath Nabro volcano;
- identify different regions of the magmatic and hydrothermal plumbing systems by differentiating between fluids;

- understand the post-eruptive state of this off-rift volcano.

Chapter 5 summarises the main results of this thesis and discusses future research directions.

2

Channelized mantle flow through the Afar Triple Junction and around the Arabian plate

Contents

2.1	Introduction	47
2.2	Data and methodology	48
2.3	Results	50
2.4	Discussion	51
2.4.1	Asthenospheric flow in northeastern Afar	51
2.4.2	Bifurcation of asthenospheric flow around the Arabian Plate	51
2.5	Conclusion	55

Abstract

Mantle flow is shown to be coherent across the three branches of the Afro-Arabian Rift System (the Main Ethiopian, Red Sea and Gulf of Aden rifts). We produce the first estimates of seismic anisotropy that reveal there is a contiguous conveyor belt of density-driven flow extending from the Main Ethiopian Rift through to Afar and diverting around the Arabian Platform. The deep mantle flow arises from the African Large Low Shear Velocity Province. Our shear-wave splitting results of the

SKS phase show density-driven flow traveling through the upper mantle beneath the Afar Triple Junction and northeastern Afar from Ethiopia. Rotation of the fast direction in southwestern Yemen suggests that the flow then bifurcates around the thick, cold, dense lithosphere of the Arabian Platform and is channeled into two different thin lithospheric corridors: along western Arabia and the Red Sea Rift, contributing to the uplift and tilt of the Arabian Plate, and along the Gulf of Aden ridge. Our results directly support geodynamic models proposing that mantle upwelling pushes on the Arabian plate and drives its convergence with Eurasia. We posit that density-driven mantle flow is the principal driving force influencing plate kinematics and continental rifting in the region.

2.1 Introduction

The Afar Triple Junction is the meeting point of the East African, Red Sea and Gulf of Aden rifts (Figure 2.1). Known as the Afro-Arabian Rift System (AARS), this complex tectonic region encompasses all stages of continental rifting, from initial breakup to seafloor spreading. However, studies of the underlying mantle present a simpler picture. Global and regional seismic tomography models identify persistent low seismic velocities in the mantle beneath the AARS, rising from the core-mantle boundary beneath south Africa and directed NE-wards: the African Large Low Shear Velocity Province (LLSVP; e.g., Ritsema et al., 1999). Seismic anisotropy measurements reveal the dynamic pattern of mantle flow through the Main Ethiopian Rift (MER) with largely NE-SW fast directions indicating a uniform flow channeled along the rift (Ayele et al., 2004; Gashawbeza et al., 2004; Kendall et al., 2005; Gao et al., 2010; Hammond et al., 2014b), possibly impacting on broader plate kinematics (Faccenna et al., 2013; Civiero et al., 2023). Further north and east of the triple junction, fast directions are aligned more N-S and E-W, suggesting the large scale change in rift orientation from the MER to the Red Sea Rift (RSR) and Gulf of Aden (GOA) respectively may control the flow (Hansen et al., 2006; Qaysi et al., 2018). To date, limited data exists close to the

triple junction in and around the RSR and the GOA, meaning that it is unclear whether the same mantle flow is channeled along both rifts.

In this study, we examine shear-wave splitting in SKS phases from seismic data in Eritrea and Yemen. We show that shear-wave splitting systematically varies in this region, largely following gradients in lithospheric topography. In particular, we observe a distinct rotation in splitting direction at the southwestern tip of Yemen from E-W to N-S, indicating that density-driven flow from the mantle bifurcates around the Arabian plate. We suggest that the mantle upwelling from the African LLSVP and channeled along the MER is largely unaffected by the shallower crustal tectonics within the Afar Depression but is deflected by the thick lithosphere of the Arabian Plate, causing the flow to split and travel down the GOA and RSR. This provides an additional force in the same direction as the slab pull from subduction in the Aegean, acting to drive Arabian plate motion (Faccenna et al., 2013; Civiero et al., 2023).

2.2 Data and methodology

The seismic data were recorded on the Eritrea Seismic Project (ESP) network (red inverted triangles, Figure 2.1b), operational from June 2011 to October 2012, and the Young Conjugate Margins Lab in the Gulf of Aden (YOCMAL) network (orange inverted triangles, 2.1b), operational between March 2009 and April 2011.

We search the ISC earthquake catalogue for seismic events with magnitude $M_b \geq 5.8$, occurring at epicentral distances 85° – 130° , to capture the SKS phase distinctly. 101 teleseismic earthquakes produced usable splitting results or null results: 59 for the Eritrea Seismic Project stations and 42 for the YOCMAL stations network (red and orange dots respectively in map inset, Figure 2.1b).

Passage through an anisotropic medium will split a polarised shear wave into two orthogonal components travelling at different velocities, producing elliptical particle motion. Here we measure anisotropy of the SKS phase, which propagates as an S-wave through the mantle, transforms into a P-wave at the core-mantle boundary (CMB), and then reverts to an S-wave propagating towards the surface.

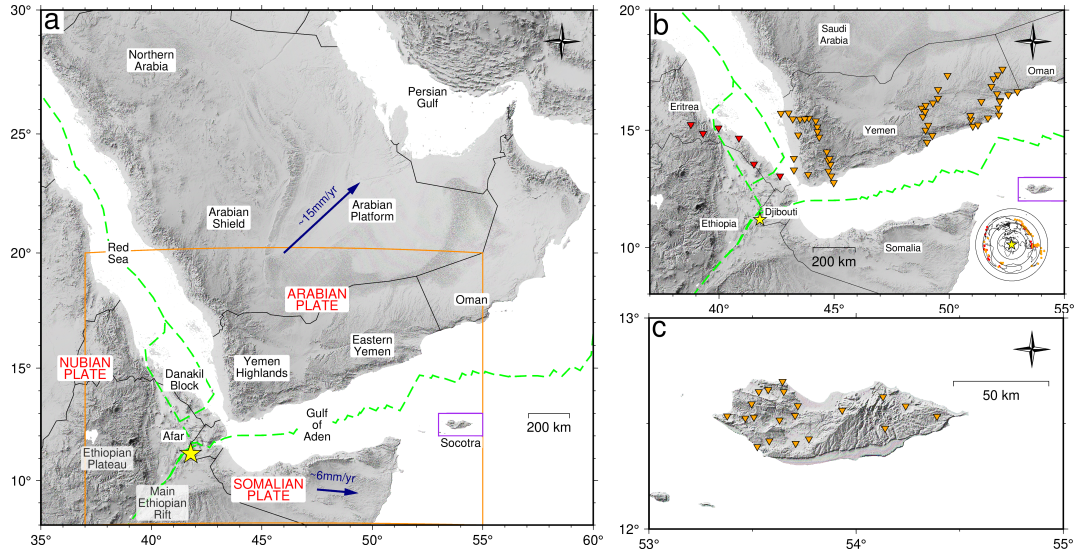


Figure 2.1: Topographic map and seismic networks. a) Afar Triple Junction and its surrounding plate boundaries (Red Sea rift, Gulf of Aden rift and Main Ethiopian Rift), outlined with dashed green lines Viltres et al. (2020); Bird (2003). The orange and purple boxes encloses the maps plotted in b) and c), respectively. The three tectonic plates are labelled with red text, and other geographic domains are labelled in black text. b) The seismic networks used in this study. Eritrea Seismic Project network stations are plotted as red inverted triangles, and YOCMAL network stations are plotted as orange inverted triangles. The bottom right inset shows the locations of teleseismic earthquakes that produced usable results or null results (red dots for the ESP network, orange dots for the YOCMAL network), with a yellow star marking the study area. c) Socotra Island stations.

Thus, any effect of anisotropy between the source and the CMB is removed when the shear wave becomes a compressional wave. The shear-wave splitting observed for the SKS phase solely describes the anisotropic medium encountered by the wave between the CMB and the receiver. The splitting is quantified by ϕ , the orientation of the fast component, and δt , the time delay between the fast and slow components, assuming a single dominant layer of anisotropy.

To reverse the effect of anisotropy, we use the eigenvalue minimization method to search for the combination of ϕ and δt which best linearize the SKS motion within a particular time window (see Silver and Chan, 1991). We deploy the automated technique of Teanby et al. (2004), as coded in the SHEBA software (Wuestefeld et al., 2010). The code enables the stacking of error surfaces associated with each measurement, resulting in an overall ‘stacked’ measurement at each station.

The procedure is illustrated in Supplementary Figures 2–4 for a good result used in the analysis, a poor result that was discarded, and a null result. Null results are measurements for which no shear-wave splitting of the SKS phase is observed. They provide useful information: either that the structure beneath a station is close to isotropic or that the initial polarisation of the shear wave is aligned with the fast or slow axis of anisotropy.

2.3 Results

We produce 162 usable SKS splitting observations across the ESP network (15–45 per station, Figure 2.2). We observe little backazimuthal variation in the shear-wave splitting parameters (Supplementary Figure 1), evidence that anisotropy beneath the stations is present in a uniform layer with a horizontal symmetry axis (e.g., Ayele et al., 2004). Measurement error surfaces are stacked, producing an overall estimate of ϕ and δt at each station (Supplementary Table 1). The orientation of the stacked fast directions shows a consistent NNE-SSW bearing, with an average $\delta t = 1.4$ s. We find 33 null results across the network. In each case, the backazimuthal direction of the null event aligns well with the stacked fast direction, so we explain these nulls as an alignment of event polarisation with a symmetry axis of anisotropy.

The YOCMAL network produces 153 usable SKS splitting results and 162 null results, with 15 stations producing solely null measurements. 4 stations produce no results or nulls. The backazimuthal coverage for the stations is not comprehensive enough to observe any variation in splitting parameters with backazimuth. We cannot definitively say that there is one uniform layer of anisotropy underlying these stations, but in the absence of further constraints we stack the error surfaces at the 29 stations which have ≥ 2 splitting measurements (Supplementary Table 1). The stations in the southwest Yemen highlands display bimodal variation in ϕ . Stations on the west coast have ϕ aligned NNW–NW, while stations in the south are aligned NE–NEE. Stations in Eastern Yemen show a more consistent eastward ϕ , which rotates slightly towards Oman. The average δt for Yemen stations is 1.1s. On Socotra Island, the average $\delta t = 0.85$ s.

The 15 stations with only null results tend to show consistent backazimuths. We conclude that these likely reflect the initial polarisation of the shear wave aligning with the fast or slow axis of anisotropy, but without better backazimuthal coverage, we cannot constrain this further. Some stations have null results for event backazimuths that are not well-aligned with the stacked result. We attribute this to the fact that there are less usable data for the YOCMAL network due to high amounts of noise, and therefore fewer splitting results per station, leading to greater uncertainty in the stacked results. Our stacked results are, however, consistent with previous estimates in our study region (Hammond et al., 2014b; Gallacher, 2016; Elsheikh, 2019; Kumar et al., 2021).

2.4 Discussion

2.4.1 Asthenospheric flow in northeastern Afar

The orientation of the fast direction across north-eastern Afar is coherent with previous anisotropy studies further south in Afar and the MER, where the NE-SW orientation is attributed to asthenospheric flow (Kendall et al., 2005; Hammond et al., 2014b). In concurrence with previous work, we see no evidence in our results for a conduit-like plume upwelling beneath the Afar Triple Junction (e.g., Gashawbeza et al., 2004). We observe no correlation between pre-rift shear zones and our results, indicating that fossilized lithospheric fabric has little effect on the anisotropy. Our results also show no deflection in flow caused by a thicker lithosphere associated with isolated continental material within the Afar region, such as the Danakil microplate (e.g., Mohr, 1970). We suggest that asthenospheric flow originating from the African LLSVP is the dominant mechanism for the anisotropy we observe in north-eastern Afar.

2.4.2 Bifurcation of asthenospheric flow around the Arabian Plate

The two groups of fast orientation directions in the Yemen highlands (Figure 2.2b) indicate that the asthenospheric flow coming from Afar and the MER bifurcates

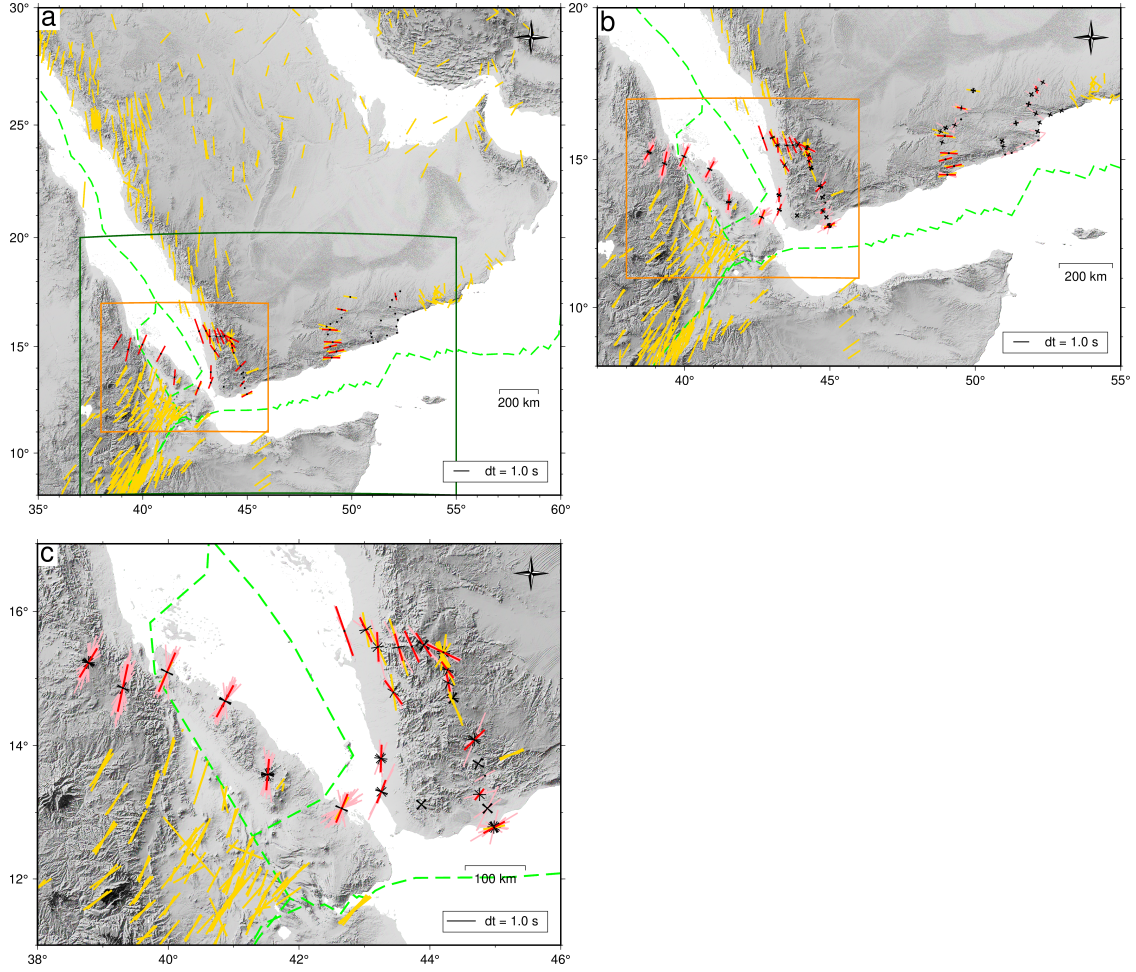


Figure 2.2: Map of the SKS splitting parameters calculated in this study alongside previous estimates. Plate boundaries are plotted as dashed green lines (Viltres et al., 2020; Bird, 2003). The individual results from this study are plotted as pink lines, with the stacked estimates plotted in red. Null results are plotted as black crosses. Previous SKS results are plotted as yellow lines (Ayele et al., 2004; Gashawbeza et al., 2004; Hansen et al., 2006; Al-Lazki et al., 2013; Hammond et al., 2014b; Gallacher, 2016; Elsheikh, 2019; Qaysi et al., 2018; Kumar et al., 2021; Kaviani et al., 2021; Ali et al., 2023). The orange, green and purple boxes outline the map sections plotted in b), c) and d) respectively.

at the southwestern tip of Arabia. Stations in the west, close to the RSR, show fast directions aligned with previous results from western and northern Arabia (Hansen et al., 2006; Qaysi et al., 2018). Stations in the south show fast directions aligned with the GOA trend.

We propose that density-driven flow from the LLSVP continues to flow through the upper mantle beyond the Afar Triple Junction. When it reaches the Arabian plate, it is diverted from its previous NE-SW trend due to the thick, cold lithosphere

seen in tomographic studies lying beneath the Arabian Platform (e.g., Koulakov et al., 2016) and corroborated by measurements of high effective elastic thickness (Chen et al., 2015). The Arabian Platform could therefore act as a backstop which stops the asthenospheric flow from continuing to travel northeastwards. The density-driven flow is then guided by topography on the lithosphere-asthenosphere boundary along channels of thin lithosphere beneath western Arabia and the GOA (Ebinger and Sleep, 1998; Hansen et al., 2007; Civiero et al., 2022).

Our proposal has been posited previously as an explanation for other observations in the region, but our seismic anisotropy observations now constrain this hypothesis. Asthenospheric flow from the African superplume has been invoked to explain: the high topography of the Arabian Shield (Daradich et al., 2003); low-velocity anomalies beneath western Arabia and the GOA (e.g., Hansen et al., 2012; Corbeau et al., 2014); deep mantle isotopic anomalies in Yemeni volcanics (e.g., Manetti et al., 1991) and off-axis volcanism in the GOA (Leroy et al., 2010). In particular, our results support geodynamic models that require a mantle upwelling to push the Arabian plate northwards (Faccenna et al., 2013; Civiero et al., 2023). Thus, the bifurcation of asthenospheric flow as it impinges on the cold, dense lithosphere of the Arabian platform is a driving force behind the absolute plate motion of Arabia, its convergence with Eurasia and the subsequent extrusion of the Anatolian plate. Flow from the African LLSVP has been influential on regional kinematics and played a direct role in the formation and migration of the Afar Triple Junction.

We plot our SKS results and those from previous studies over the tomography model of Celli et al. (2020b) in Figure 2.3. The low velocity anomalies match the bifurcated pattern of the seismic anisotropy very well, particularly at the southwestern corner of Arabia. The low velocity anomaly in the RSR branch is stronger and extends to greater depths than the GOA branch. This is reflected by the decreased δt and lack of coherent fast direction at Yemeni and Socotra Island stations east of 50°E, indicating that fossil fabric frozen into the lithosphere has started to dominate the anisotropy signal we observe, right as the low velocity anomaly from the density-driven flow through the GOA dies away. These results

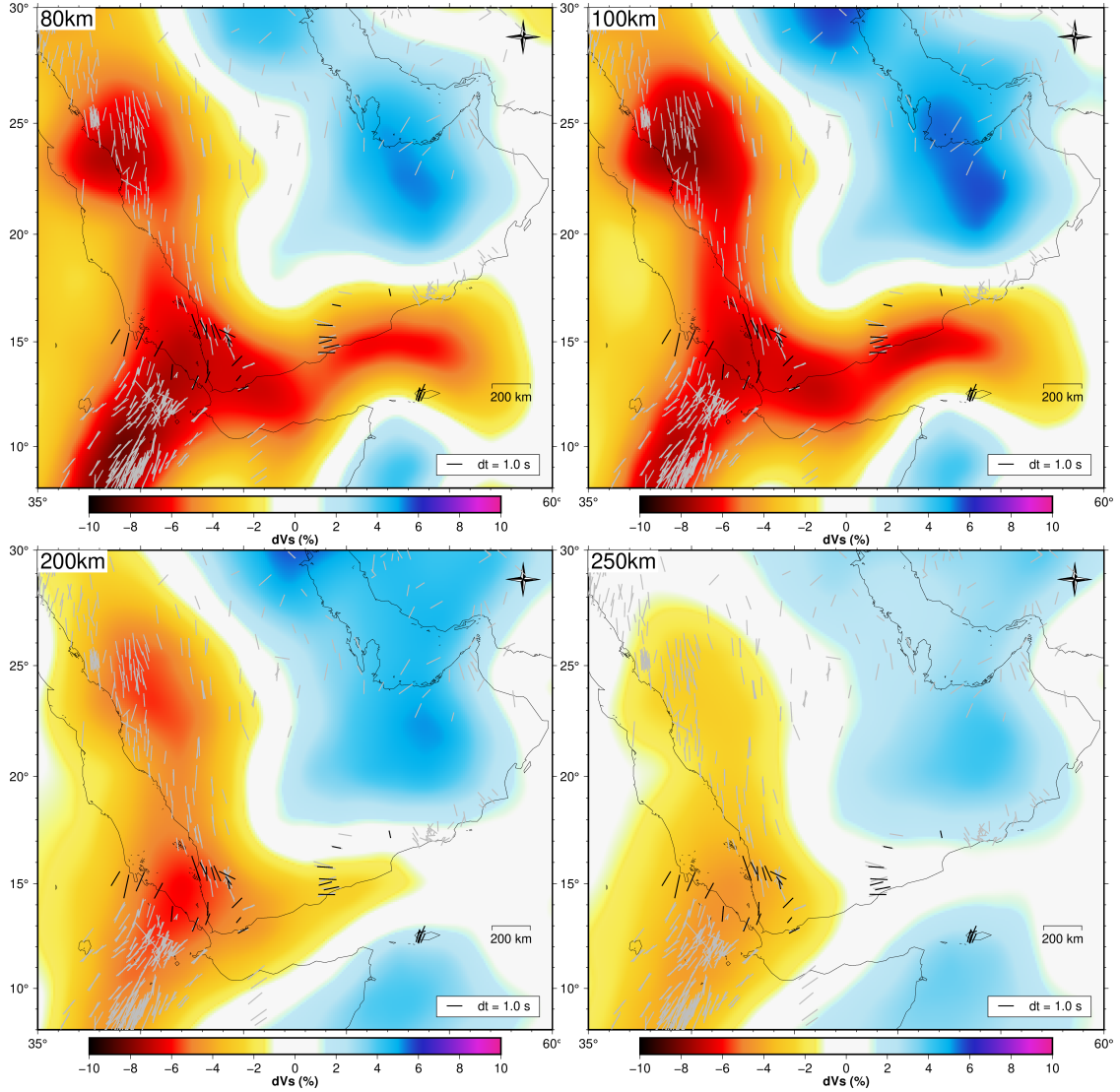


Figure 2.3: Map of the SKS splitting measurements from this study (black lines) and previous studies (grey lines) (Ayele et al., 2004; Gashawbeza et al., 2004; Hansen et al., 2006; Al-Lazki et al., 2013; Hammond et al., 2014b; Gallacher, 2016; Elsheikh, 2019; Qaysi et al., 2018; Kumar et al., 2021; Kaviani et al., 2021; Ali et al., 2023) plotted over four depth slices through the S-wave tomography model of Celli et al. (2020b).

are coherent with observations in Oman, the conjugate margin to Socotra Island on the other side of the GOA, where seismic anisotropy is due to fossilized lithospheric fabric and average $\delta t = 0.6s$ (Al-Lazki et al., 2013).

2.5 Conclusion

We have produced the first seismic anisotropy measurements that conclusively show that asthenospheric flow from the African LLSVP is ubiquitous beneath the three rift branches from the Afar Triple Junction. Flow is channeled through the Afar Triple Junction from the East African Rift System, then splits to flow along the Gulf of Aden and Red Sea rift trends. We observe a consistent fast shear-wave polarization oriented NE-NNE in north-eastern Afar, with a large average splitting magnitude $\delta t = 1.4\text{s}$. In southwestern Yemen, there is a distinct and dramatic rotation in fast direction over a short distance. Stations close to the RSR are aligned NNW–NW, while stations on the margin of the GOA have fast directions aligned NEE. Further to the east, stations show E–W trends that mirror the GOA trend, before losing coherence closer to Oman. The average $\delta t = 1.1\text{s}$ in Yemen. On the other conjugate margin of the GOA, Socotra Island stations show a smaller splitting magnitude ($\delta t = 0.85\text{s}$), reflecting fossilized lithospheric fabric.

We suggest that when the hot mantle flow meets the cold, thick, dense lithosphere of the Arabian Platform, it bifurcates and is directed along corridors of thin lithosphere down the GOA and north beneath western Arabia and the RSR. This supports geodynamic models which invoke ‘plume push’ driving Arabia’s current plate motion and its convergence with the Eurasian plate.

Our study provides the first direct data from this key region, confirming the fundamental connection between dynamic mantle flow, plate kinematics and continental breakup. We conclude that the density-driven mantle flow from the African LLSVP has been a driving force behind regional kinematics, as well as controlling and influencing rifting processes in the AARS.

Open Research

The seismic data used in this study are from the Eritrea Seismic Project and the Young Conjugate Margins Lab in the Gulf of Aden network, publicly available through IRIS Data Services (<http://service.iris.edu/fdsnws/dataselect/1/>). The

SHEBA software (Wuestefeld et al., 2010) is available online at <https://github.com/jwookey/sheba>. The full SKS results are archived on Zenodo at [doi/10.5281/zenodo.11058890](https://doi.org/10.5281/zenodo.11058890).

Acknowledgments

The Eritrean Seismic Project data were collected with funding from the Natural Environment Research Council (NERC) projects NE/J012297/1, NE/E007414/1, and NE/D008611/1. The UK seismic instruments and data management facilities were provided under loan number 976 by SEIS-UK (NERC, Agreement R8/H10/64) at the University of Leicester. Author MG was supported by a studentship from NERC [NE/S007474/1].

3

The dynamic crust of northern Afar and adjacent rift margins: evidence from receiver functions

Contents

3.1	Introduction	59
3.2	Data and Methodology	63
3.2.1	Seismic Network	63
3.2.2	Teleseismic Dataset	64
3.2.3	Methods	66
3.3	Results	69
3.4	Discussion	73
3.4.1	Stretching of the Danakil block	73
3.4.2	Crustal sills beneath Nabro volcano	74
3.4.3	Implications for rifting through the Gulf of Zula	75
3.4.4	Topographic elevation and crustal structure	76
3.5	Conclusion	83

Abstract

Afar is undergoing the final stages of continental rifting and hosts the triple junction between the Red Sea, Gulf of Aden and Main Ethiopian rifts. To better understand the nature of the crust and continental breakup in the region, we calculate teleseismic

receiver functions across northeastern Afar and the Danakil microplate, using new data from a regional deployment in Eritrea. We estimate the Moho depth and bulk crustal V_P/V_S ratio using the H- κ stacking method. The heterogeneity of our crustal thickness estimates ($\sim 19\text{--}35$ km) indicates that the Danakil microplate has undergone stretching and crustal thinning. By investigating the relationship between crustal thickness and topographic elevation, we estimate the regional crustal bulk density as $\rho_c \approx 2850 \pm 20 \text{ kg m}^{-3}$, which is higher than expected, given the crustal thickness of the region. We show that topography is 1.5 ± 0.4 km higher than would be expected due to crustal isostasy alone. We propose that this topography is supported by the same hot mantle upwelling suggested to be responsible for the onset of rifting in East Africa. Uplift is generated due to the presence of a hot thermal anomaly beneath the plate and by thinning of the lithospheric mantle. Our results are consistent with a number of independent constraints on the thermal structure of the asthenospheric and lithospheric mantle. Evidence of melt within the crust is provided by anomalously high V_P/V_S ratios of > 1.9 , demonstrating that magma-assisted extension continues to be important in the final stages of continental breakup.

Plain Language Summary

Afar is an area of northern Ethiopia that extends into Eritrea. It hosts three tectonic plate boundaries that are pulling apart from one another (rifting) as continental breakup is occurring. These rifting processes have led to a complicated tectonic history; isolating a small microcontinent (the Danakil) and giving rise to volcanism across the region. To better understand the nature of the crust, we study seismic data to estimate the crustal thickness and the ratio of seismic wave speeds. Our results indicate that the crust shows substantial variation in thickness, meaning that the Danakil microplate has undergone crustal thinning. We use our results to determine that dynamic mantle processes are responsible for supporting the elevation of the region. We also show that partially molten rock (magma) is likely

to be present in the crust beneath northeastern Afar and the Danakil microplate, which is evidence that magma assists with continental breakup.

Key Points

- Receiver functions produce the first estimates of Eritrean bulk crustal properties
- The crust of Afar and the Danakil microplate is denser than global average and highly heterogeneous
- Evidence found for melt within crust and for support from hotter mantle propping up topography beneath Afar and the Danakil microplate

3.1 Introduction

The Afro-Arabian Rift System is an ideal setting to investigate the temporal and spatial evolution of continental rifting. It consists of a network of rifts that developed asynchronously and are currently at different stages of the rifting process. Initial continental breakup is occurring in the south of the EARS in Mozambique, while further to the north, full seafloor spreading has been established in the Gulf of Aden and the Red Sea (e.g., Manighetti et al., 1998; Tesfaye et al., 2003). Elsewhere, the Main Ethiopian Rift displays the characteristics of a continental rift in its immediate stage of development, while Afar is undergoing the final stages of continental breakup as it develops into oceanic rifting in the transitional Red Sea zone (e.g., Makris and Ginzburg, 1987; Kogan et al., 2012). It is currently a matter of debate whether full seafloor spreading has begun on land in Afar. Magnetic anomalies have been identified in the region, analogous to those observed at oceanic spreading centers (e.g., Courtillot et al., 1980), which could indicate the onset of seafloor spreading, or could be explained as a result of magmatism during late stages of continental rifting (Bridges et al., 2012).

The Afar region is a diffuse extensional province which hosts a triple junction formed by the meeting of three diverging plate boundaries: the Main Ethiopian Rift,

the Gulf of Aden and the Red Sea, separating the Nubian, Somalian and Arabian tectonic plates respectively (Figure 3.1). The Gulf of Aden and the Red Sea do not directly connect in Afar, but are actively propagating through the region via the subaerial development of a series of disconnected, propagating rift segments (e.g., Manighetti et al., 1998, 2001). Instead of directly connecting to the Gulf of Aden through the Bab el-Mandeb Strait, the Red Sea bifurcates into two branches south of 16°N (Barberi and Varet, 1977; McClusky et al., 2010). This process has isolated the Danakil microplate (or Danakil block), a sliver of continental crust rotating independently of Nubia (Figure 1a; Sichler, 1980; Courtillot et al., 1984; Manighetti et al., 2001).

The coherent rotation of the Danakil block has been established by several geodetic models and is associated with the formation of the Danakil Depression, a region of very low topography in northern Afar (e.g., Eagles et al., 2002; McClusky et al., 2010; Viltres et al., 2020), but there are fewer measurements around the proposed borders of the microplate, meaning the exact boundaries are not well-constrained by the models. Continental deformation models have two end-members: the block model, where deformation is confined to major block-bounding faults, with little internal deformation, and the continuum model, where deformation is diffused and quasi-continuous across a wider region (Thatcher, 2007). The transition between block and continuum models is gradational and has many intermediate stages. Thus, despite the observed coherent rotation of the Danakil block, it may have undergone internal deformation and stretching. Kinematic models have produced different estimates for the extent of the Danakil block. Viltres et al. (2020) highlight the diffuse character of the boundary between the Danakil and Nubia in the northeast, where inter-rifting deformation extends over a wide deformation zone. Varet (2017) place the southern boundary of the Danakil microplate near the Nabro Volcanic Range. This boundary choice is supported by the distribution of regional seismicity along this trend (Goitom, 2017). As a result of geographic remoteness, very few studies have been undertaken in northeastern Afar and the Danakil block, meaning

there is a lack of information on the microplate and its role in partitioning strain between the two branches of the Red Sea Rift.

The nature of the crust beneath Afar has been widely debated. Some authors argue that new oceanic crust is being created in the region (e.g., Mohr, 1989), whilst others conclude that the area is entirely underlain by intruded, stretched continental crust (e.g., Makris and Ginzburg, 1987). An alternative proposal is that the overall crustal composition is transitional between continental and oceanic (e.g., Redfield et al., 2003; Hammond et al., 2011c). These studies suggest that the region west of the current rift axis is predominantly made up of mafic material with oceanic affinities, whereas continental material exists to the east, potentially connected to the Danakil block.

Continental breakup involves extensional faulting, ductile plate stretching and intrusive and extrusive magmatism (e.g., Barton and White, 1997; White and McKenzie, 1989; Bastow and Keir, 2011). It is crucial to elucidate the extent to which mechanical- or magma-assisted extension processes dominate at each stage of continental breakup as it transitions into seafloor spreading. Indeed, this transition process may be more complex than a smooth evolution from mechanical- to magma-dominated extension into oceanic spreading. Previous studies in the Main Ethiopian Rift have shown that upper-crustal extension throughout much of Ethiopia is currently accommodated primarily by dyke intrusion along narrow, magmatic segments; strain has migrated away from the border faults and localised to rift-aligned magmatic intrusions since around 20 Ma (e.g., Ebinger and Casey, 2001; Wolfenden et al., 2004). However, the Danakil Depression to the north, which has been affected by magmatism since rift onset, now hosts significantly thinned continental crust, coupled with marked subsidence and thick evaporite deposition (Hutchinson and Engels, 1972; Wolfenden et al., 2005; Keir et al., 2013). The seismic reflection study of Hurman et al. (2023) found that upper crustal axial faulting in a magma-poor area of the depression accommodates 30% of the total extension. Thus it has been proposed that the final stages of continental breakup

in northern Afar may be characterised by a return to ductile plate stretching and upper-crustal brittle faulting (Bastow et al., 2018).

Hammond et al. (2011c) comment that the lack of data from Eritrea limits overall regional interpretations of crustal structure. Eritrea hosts the Eritrean plateau (an extension of the western Ethiopian plateau), the Gulf of Zula, the Danakil Depression, the Danakil Alps and substantial off-rift volcanism in the Nabro Volcanic Range. The strike of the Nabro Volcanic Range is almost parallel to the trend of the Hanish-Zukur islands in the Red Sea and perpendicular to regional Red Sea rift trends. The presence of these volcanic centers across the Danakil block has led to proposals that they represent the surface expression of a ‘leaky’ transform fault accommodating strain transfer from the Red Sea into Afar (e.g., Barberi and Varet, 1977; Viltres et al., 2020).

Crustal thickness (i.e., depth to the base of the Mohorovičić discontinuity) and the ratio of P-wave velocity to S-wave velocity (V_P/V_S ratio) provide key constraints on crustal composition and the amount of partial melt hosted by the crust (e.g., Zandt and Ammon, 1995; Dugda et al., 2005; Stuart et al., 2006). Previous studies of crustal structure in the region have shown that the crustal thickness varies from 40–45 km beneath the western Ethiopian plateau, decreasing to 20–26 km in central Afar, and becoming as thin as 16 km in the Danakil Depression (Berckhemer et al., 1975; Makris and Ginzburg, 1987; Mackenzie et al., 2005; Dugda et al., 2005; Stuart et al., 2006; Maguire et al., 2006; Cornwell et al., 2010; Hammond et al., 2011c; Reed et al., 2014; Ahmed et al., 2022). The V_P/V_S ratio also shows spatial variation, from 1.7–1.9 beneath the plateau to > 2.0 near magmatic segments where the crustal thickness is less than 26 km; the high values of V_P/V_S are interpreted as the result of aligned melt stored in interconnected stacked sills in the lower crust (Hammond, 2014).

Finally, it is widely accepted that a large low velocity province exists within the mantle beneath the African plate (e.g., Ritsema et al., 1999; Koelemeijer et al., 2017). It has been long-argued that this long-lived thermo-chemical anomaly has provided the impetus necessary to incite rifting processes in the region, as well as inducing

mantle flow that dynamically supports the excess elevation of Africa (Lithgow-Bertelloni and Silver, 1998; Ebinger and Sleep, 1998; Kendall and Lithgow-Bertelloni, 2016). However, the exact form of mantle upwelling from the core-mantle boundary to the surface is still a matter of debate, particularly in northeast Africa (e.g., Civiero et al., 2015). Hence the regional effect of a sub-crustal thermal anomaly upon northeastern Afar and the Danakil block merits investigation. The primary controls on topographic elevation are the isostatic effects of variations in crustal thickness and density (e.g., Tiberi et al., 2005); however, the density structure of the mantle, particularly that of a hot thermal anomaly, can affect the pattern of topography observed at the surface (e.g., Bird, 1978; Hager and Richards, 1989; Stephenson et al., 2021; Bertelloni and Gurnis, 1997). By examining the relationship between topographic elevation and crustal structure in the study region and comparing it to the global relationship between these parameters, we can deduce information about mantle structure in northeastern Afar and the plateaus within the Danakil block.

In this paper, we exploit a new dataset from a temporary regional seismic deployment, the Eritrea Seismic Project, to calculate teleseismic receiver functions. Using the H- κ method developed by Zhu and Kanamori (2000) and modified by Ogden et al. (2019), we calculate the crustal thickness (H) and bulk V_P/V_S ratio (κ) at each station. We supplement our analysis with data from another temporary deployment in central Afar, AFAR0911. The use of the AFAR0911 deployment enables us to update previous work by Hammond et al. (2011c), Reed et al. (2014) and Ahmed et al. (2022), as well as validating the method of Ogden et al. (2019). Finally, we investigate the relationship between elevation and crustal thickness in order to understand the dominant control on topography within the region.

3.2 Data and Methodology

3.2.1 Seismic Network

To calculate receiver functions, we use seismic data from the Eritrea Seismic Project, a temporary regional seismic network deployed across the country over a period of 16 months from June 2011 until October 2012 (green inverted triangles

in Figure 3.1b (Hammond et al., 2011a)). Six broadband seismic systems were provided by SEIS-UK (five CMGESP and one CMG3T). Stations were placed on the Eritrean plateau (EITE, CAYE), near the Gulf of Zula (DOLE), at the foot of the Danakil Alps (TIOE), near the Nabro Volcanic Range (FAME) and close to the Bab el-Mandeb strait (ASSE).

We supplement seismic data from this regional network with data collected by a temporary local seismic network deployed in the aftermath of Nabro volcano’s 2011 eruption (plotted collectively as station NAB in Figure 3.1b (Hammond et al., 2011b)). Eight 3-component broadband 30 s Güralp seismometers (five CMG-6TD and three CMG-40TD) were provided by SEIS-UK to monitor Nabro’s post-eruptive state. The network was fully operational from 31 August 2011 until October 2012. The data were all initially recorded at 100 Hz sample frequency and then switched to 50 Hz sample frequency early in October 2011. We also calculate receiver functions for stations from the AFAR0911 deployment, which ran from 2009 until 2013 (yellow inverted triangles in Figure 3.1b (Keir and Hammond, 2009)). These stations are 3-component broadband 60 s CMG-ESP Güralp seismometers, recording at 50 Hz sample frequency. Due to the lack of cultural noise, all the data exhibit a generally high signal-to-noise ratio, with excellent data quality (Hammond et al., 2014a).

3.2.2 Teleseismic Dataset

We search the USGS earthquake catalogue for seismic events with body-wave magnitude $M_b \geq 5$, occurring at an epicentral distance greater than 30° . Events are then sorted into two epicentral distance ranges, $30^\circ - 90^\circ$ and $> 90^\circ$. The former range is standard for calculating P-to-S receiver functions, as it eliminates the effects of core phases and upper mantle triplications. To increase our azimuthal coverage, we also calculate PP-to-S receiver functions from events in the distance range $> 90^\circ$.

The data are filtered using a Butterworth bandpass filter with corner frequencies 0.04 and 3 Hz. Seismograms are inspected manually to remove poor quality traces, and then analysed based on the signal-to-noise ratio. Source-receiver pairs are kept for further analysis if the P-wave signal-to-noise ratio is > 3 . This process

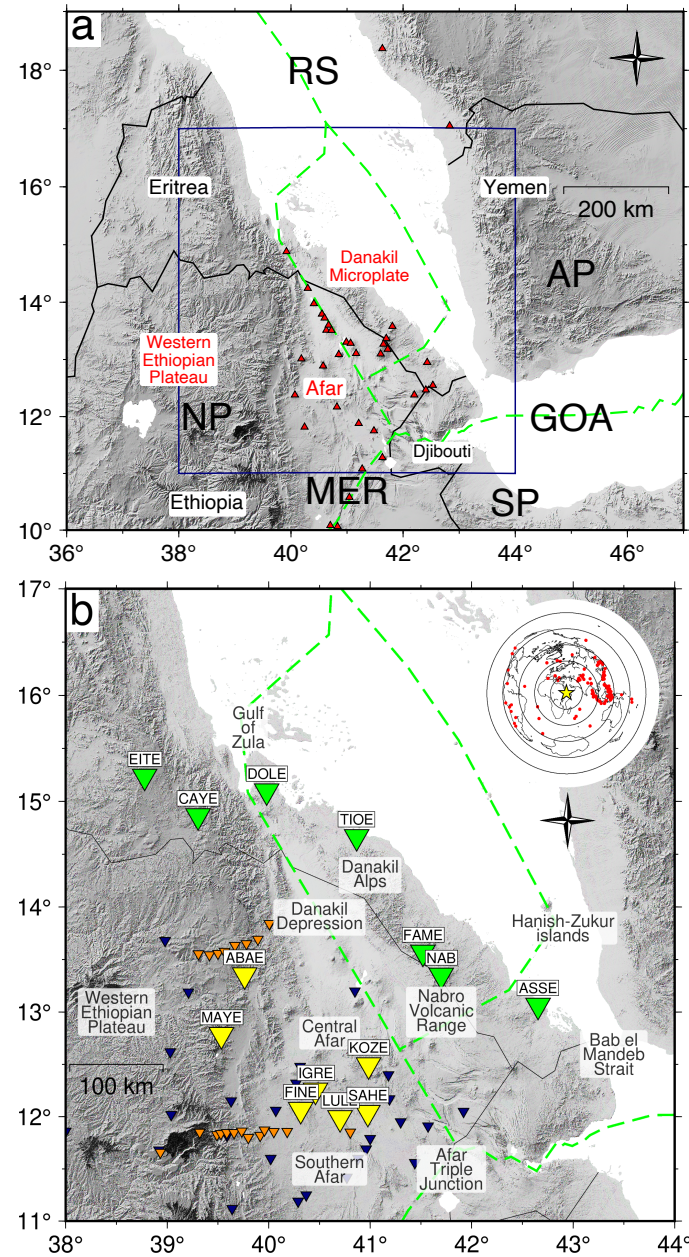


Figure 3.1: Topographic map. The boundaries of the Danakil microplate and other plate boundaries are outlined by a dashed green line (Viltres et al., 2020). a) Countries are labelled with black text and solid black lines demarcate national borders. Other geographic domains are labelled with red text. The navy box encloses the region plotted in b). RS: Red Sea, GOA: Gulf of Aden, MER: Main Ethiopian Rift, NP: Nubian Plate, AP: Arabian Plate, SP: Somalian Plate. b) Seismic stations used in this study are plotted as inverted green and yellow triangles and labelled. Seismic stations from previous studies are plotted as inverted blue triangles (Hammond et al., 2011c) and inverted orange triangles (Ahmed et al., 2022). Volcanoes from the Smithsonian Institution Global Volcanism Program are plotted as red triangles (Smithsonian, 2023). Inset shows the locations of the teleseismic earthquakes used in this study, with a yellow star marking the study area.

results in the selection of 102 earthquakes for receiver function construction and processing (inset in Figure 3.1b).

3.2.3 Methods

3.2.3.1 Receiver Function Calculation

Receiver function analysis is a well-established technique for investigating velocity contrasts in the crust and upper mantle beneath three-component seismic stations (Langston, 1979). When a P-wave impinges on a velocity contrast, part of the energy is converted into a S-wave. In the absence of anisotropy and dipping boundaries, these P-to-S conversions are preferentially recorded on the radial components of a seismogram, whilst direct P-wave arrivals are recorded on the vertical component. For teleseismic waves, the incidence angle of the incoming wave will be close to vertical (27° or lower), meaning that we can use the vertical component of motion as a good approximation of the event source. Deconvolution of the vertical component from the radial components removes information common to both (i.e., the instrument response, path and source effects). This isolates the response from the local crustal structure along the incoming, near-vertical raypath from a teleseismic earthquake. The resulting receiver functions contain phase arrivals that represent the P-to-S conversion and subsequent reverberations within a layer (the positive polarity PpPs phase, and the negative polarity co-arriving PpSs + PsPs phases, see Figure 3 in Jenkins et al. (2020)). The amplitudes of the arrivals depend on the P-wave incidence angle (i.e., the ray parameter), the size of the velocity contrast, and whether the wave passes from a high velocity layer to lower velocity layer, or vice versa. The arrival times of the converted phase and reverberations depend on the depth of the velocity contrast, the P-wave and S-wave velocity between the contrast and the surface, and the ray parameter.

We calculate receiver functions using the iterative time-domain deconvolution method of Ligorria and Ammon (1999). This technique iteratively builds receiver functions in the time domain from Gaussian pulses of a set width. For our analysis, we use a range of Gaussian width factors at 0.2 intervals from 0.8–4.0 to produce a

set of 17 individual receiver functions for each source-receiver pair. Since periods less than the Gaussian width are not resolvable, our calculated receiver functions have corresponding low-pass filters between 0.4–2.0 Hz. The radial receiver function is then convolved with the vertical component and cross-correlated with the original radial component to evaluate the iterative deconvolution variance (IDVAR). If this variance is $\leq 80\%$, the receiver function is rejected. We also undertake a final manual inspection of the receiver function traces, removing poor quality traces (e.g., those where the maximum peak arrival is not the direct P-wave arrival, those with anomalous long wavelength features, and those which are significantly different from others that sample the same region). This results in a range of 9 to 63 receiver functions per station (Table 3.1).

3.2.3.2 H- κ Stacking

We use the H- κ stacking procedure of Zhu and Kanamori (2000) to estimate bulk crustal properties beneath each seismic station (see Supplementary Information Text S1 for further details on the technique). This method has produced accurate estimates of these properties in regions where the crustal structure is simple (e.g., Thompson et al., 2010). However, our study region is tectonically complex region, and we therefore anticipate some potential difficulties in receiver function calculation and interpretation.

The presence of thick sedimentary layers in the study area has been well-established (see Rime et al. (2023) for a comprehensive geological map). The Danakil Depression hosts a sedimentary basin likely containing up to ~ 3 –5 km of Pliocene-Pleistocene-Recent evaporites (Hutchinson and Engels, 1972; Bastow and Keir, 2011; Bastow et al., 2018; Rime et al., 2023). Extensional basins in southern and central Afar are filled with Pliocene-Holocene lacustrine, alluvial and volcanoclastic sediments, exceeding 200 m in places (e.g., Renne et al., 1999). Although parts of southern and central Afar are covered by the Stratoid basalts, this cover is lacking to the north in the Danakil Depression and on the Danakil microplate, which are dominated by sedimentary rocks (Barberi et al., 1980; Varet,

2017; Tortelli et al., 2022; Rime et al., 2023). Shallow sedimentary layers produce significant P-to-S converted energy that can mask the signal from the Moho and its crustal reverberations, increase the time delay of the direct P-arrival, and enhance the ‘ringing’ nature of the receiver function signal (e.g., Zelt and Ellis, 1999; Reed et al., 2014; Ogden et al., 2019).

Due to the complicated tectonic history of Afar and the Danakil block, we expect significant local differences in the amount of extension and/or thickening due to underplating and intrusion. This can be seen in the variability in previous Moho depth estimates across Afar (e.g., Dugda et al., 2005; Hammond et al., 2011c; Reed et al., 2014; Lavayssière et al., 2018; Ahmed et al., 2022). It is distinctly observable at single stations with backazimuthal variation in Moho depth, due to the receiver functions from different directions sampling crust of different thickness on either side of the station (e.g., Dugda et al., 2005; Hammond et al., 2011c). This is particularly pronounced for stations in Afar that are close to the large border faults, such as ABAE (Hammond et al., 2011c). Regions of lower crustal intrusions show a gradational transition between the crust and mantle, as observed by Lavayssière et al. (2018) beneath the western Ethiopian plateau. Frequency content of receiver functions is also an important consideration for reliable detection of the Moho (e.g., Frassetto et al., 2011; Ogden et al., 2019). A gradational Moho—one with an increased width including a series of small steps in velocity gradient—may cause short period receiver functions to convert at different depths off the incremental layers, causing scattered Ps conversions with low amplitudes (Gallacher and Bastow, 2012). This scattering means that H- κ stacking becomes less reliable. To address this issue, Ogden et al. (2019) observe from synthetic data that a correct H- κ result can be returned if longer period receiver functions are used (Figure 5 in Ogden et al., 2019), which is then validated experimentally with a study of the Ethiopian plateau. The longer period receiver functions, with their increased wavelengths, show less scattered Moho Ps conversions with greater amplitudes. Using longer period data therefore allows the H- κ stacking process to resolve the gradational Moho discontinuity more reliably.

For H- κ stacking, V_P is held constant for the whole crust and has to be either known *a priori* or assumed. Previous receiver function studies in southern and central Afar use a wide range of values for average crustal P-wave velocity, from $4.65 - 6.5 \text{ kms}^{-1}$, based on criteria such as wide-angle seismic reflection studies or by examining the theoretical relationship between partial melt fraction and seismic velocities. Some authors allow for variability in V_P between stations (e.g., Reed et al., 2014), whilst others assume fixed values for all stations in their study (e.g., Dugda et al., 2005; Hammond et al., 2011c; Ahmed et al., 2022).

To produce robust estimates, we therefore employ the cluster analysis approach of Ogden et al. (2019). This methodology rigorously explores the H- κ parameter space, including frequency content of the receiver functions and assumed average crustal V_P , to account for sensitivity of the method to complex crustal structure. The quality control criteria applied to the receiver functions and the cluster analysis provides quantitative evidence for the reliability of H- κ stacking at a station. The number of quality control criteria failed by a station gives an indication of why H- κ stacking has produced an unsatisfactory result. In regions where the Moho is gradational, the ability to adjust the frequency content of the receiver functions means that H- κ stacking remains reliable. The final result will successfully identify the centre of the Moho depth range if the Moho layer thickness is in the range of 0 km to 13 km (Ogden et al., 2019). See the Supplementary Information (Text S2) for more detail on the procedure and the associated uncertainty estimates.

3.3 Results

Seven stations from the Eritrean Seismic Project, including the composite station NAB (made up of the six Nabro array stations), and seven stations from the AFAR0911 network produce reliable results using our H- κ stacking procedure (Table 3.1, Figure 3.2). Ten quality control criteria assess the quality of these results, and are described in the Supplementary Information (Text S3 for the Eritrean Seismic Project, Text S4 for AFAR0911). Diagnostic result figures for each station are plotted in Figures S1 – S19 in the Supplementary Information. One

of the stations on the Eritrean plateau (EITE) shows strong variation in the Ps arrival time with back-azimuth, a phenomenon previously observed in the region by Dugda et al. (2005) and Hammond et al. (2011c). Following these examples, we split the receiver functions at EITE into those arriving from the east and those arriving from the west, carrying out H- κ stacking on each set separately. These results are plotted as EITE_E and EITE_W, respectively, at the first Moho bounce point of the PpPs multiple for the dominant back azimuth (Figure 3.2). We follow a similar procedure for the central Afar stations ABAE and MAYE, as the receiver functions are almost exclusively from the east for these stations (Figures S11 and S18 in the Supplementary Information). In total, we calculate fifteen estimates of crustal thickness and V_P/V_S ratio .

Almost all of the receiver functions show an obvious Ps Moho arrival between 2.5 and 5 s, corresponding to a crustal thickness range of 18.9 – 34.8 km across all stations. This arrival is not always the first positive arrival, which is evidence of a sedimentary layer at the surface. In general, the receiver functions show additional P-to-S arrivals as well as the Moho crustal reverberations, indicating the presence of complex intracrustal structure leading to these subsidiary conversions. The PpPs Moho arrival is observable in most of the receiver functions between 8.5 and 15.5 s, whereas the final arrival of PpSs + PsPs is more ambiguous, occurring between 12 and 20 s. The stations show a range of V_P from 5.9 – 6.5 kms^{-1} , which we determine to be reasonable, given previous controlled source studies (Makris and Ginzburg, 1987; Mackenzie et al., 2005; Maguire et al., 2006).

Within error, the crustal thicknesses we observe are consistent with previous estimates by Hammond et al. (2011c), who carried out H- κ stacking at stations ABAE, FINE, IGRE, KOZE, LULE, MAYE and SAHE. Our V_P/V_S results are also generally consistent with the values found by Hammond et al. (2011c), although they tend to estimate a slightly higher V_P/V_S than our results. This consistency gives confidence to the H- κ stacking procedure we employ, as well as our selected values of V_P .

Table 3.1: H- κ and V_P Results

Station	Long. (°N)	Lat. (°E)	Elev. (km)	H (km)	Error (km)	κ	Error	V_P (kms ⁻¹)	RFs
<i>Eritrea Seismic Project stations</i>									
ASSE	13.06	42.65	0.019	24	1	1.83	0.03	5.9	14
CAYE	14.86	39.31	2.435	33	2	1.84	0.03	6.1	28
EITE (W)	15.24	38.28	2.171	35	2	1.80	0.03	6.2	9
EITE (E)	15.24	39.12	1.471	26	1	2.05	0.03	6.2	36
DOLE	15.10	39.98	0.088	27	2	1.70	0.03	6.5	9
FAME	13.57	41.52	0.622	20	2	1.84	0.08	6.4	26
NAB	13.35	41.7	1.272	25	1	1.77	0.03	5.9	63
TIOE	14.67	40.87	0.043	19	1	2.04	0.04	6.0	20
<i>AFAR0911 stations</i>									
ABAE	13.35	39.76	1.447	19	1	2.10	0.02	6.2	53
FINE	12.07	40.32	0.782	23	2	1.95	0.03	6.3	23
IGRE	12.25	40.46	0.675	25	2	1.75	0.05	5.9	42
KOZE	12.49	40.98	0.543	28	1	1.81	0.04	5.9	34
LULE	11.99	40.70	0.594	26	4	2.04	0.05	6.5	20
MAYE	12.78	39.53	2.440	21	2	1.91	0.023	6.5	26
SAHE	12.04	40.98	0.365	22	1	1.99	0.02	5.9	23

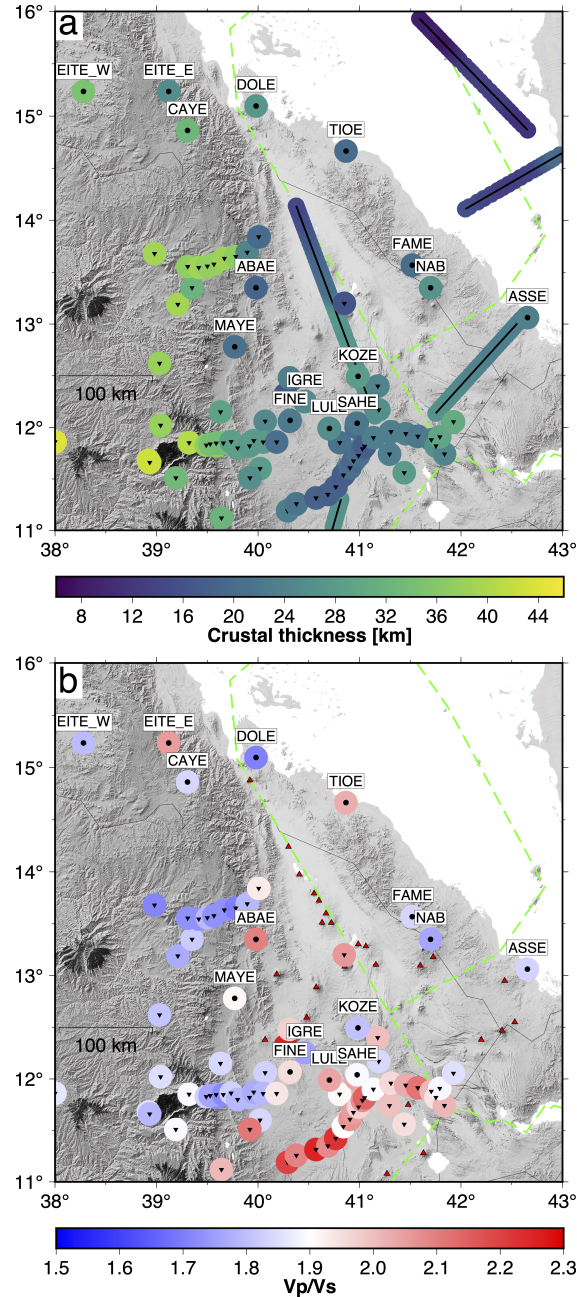


Figure 3.2: Regional map showing the results of this study (Table 3.1) alongside previous estimates of crustal thickness and V_P/V_S ratio (see Table S1 in the Supplementary Information for a complete list). Seismic stations from this study are labelled and plotted as black circles. Seismic stations from previous studies are plotted as black inverted triangles. The boundaries of the Danakil microplate and other plate boundaries are outlined by a dashed green line (Viltres et al., 2020). Dashed black lines demarcate national borders. Volcanoes from the Smithsonian Institution Global Volcanism Program are plotted as red triangles (Smithsonian, 2023). a) The crustal thickness estimates. Thick black lines mark the profiles for controlled source work by Berckhemer et al. (1975); Makris and Ginzburg (1987); Egloff et al. (1991); Mackenzie et al. (2005); Maguire et al. (2006). Other estimates of crustal thickness come from Hammond et al. (2011c), Reed et al. (2014) and Ahmed et al. (2022). b) V_P/V_S ratio estimates from this study and from Hammond et al. (2011c), Reed et al. (2014) and Ahmed et al. (2022).

3.4 Discussion

3.4.1 Stretching of the Danakil block

It is well-established that continental basement is exposed on the Danakil microplate, particularly within the Danakil Alps (e.g., Freund, 1970; Bosworth et al., 2005; Varet, 2017; Rime et al., 2023). Drury et al. (1994) propose that this crustal segment was shortened and thickened by transpression in late Neoproterozoic to Cambrian times. Our crustal thickness estimates for the stations located on the Danakil block are lower than the global average value for extended continental crust of 30.5 km (Christensen and Mooney, 1995). They are also 10–15 km thinner than estimates for stations found on the western Ethiopian plateau (Figure 3.2). Indeed, stations TIOE and FAME show thicknesses that are similar to those found in the rift valley of Afar. While we lack constraints from the interior of the Danakil Alps, we nevertheless propose that the Danakil is not ‘relatively unstretched’ as has been previously claimed (Wolfenden et al., 2005), but has undergone a significant amount of deformation and crustal thinning. Two-dimensional area balancing in a previous plate kinematic reconstruction study supports this conclusion, suggesting that the Danakil block has undergone a minimum of 200% stretch since the onset of rifting (Figure 3, Redfield et al., 2003). A geometric estimate by Morton et al. (1975) also gave a stretching factor (i.e., ratio of stretched width to original width) of up to 2.5 in the Danakil Alps.

Station EITE (E) has a V_P/V_S ratio of 2.054, which is significantly higher than the average value for continental crust of 1.768 (Christensen, 1996). V_P/V_S ratios above 1.9 cannot be accounted for by differences in rock composition (Thompson et al., 2010). Instead, these elevated values are typically explained by the presence of fluid (e.g., melt) in the crust, which decreases the S-wave velocity without significantly affecting the P-wave velocity (e.g., Watanabe, 1993). This effect is enhanced in the presence of aligned melt due to the influence of anisotropy (Hammond, 2014). EITE (E) is located on the edge of the western plateau. It shows strong similarity to stations MAYE and ABAE, which are positioned on

the rift shoulder in central Afar and also show high V_P/V_S ratios. We therefore propose that partial melt in the crust is present along the edge of the western plateau in central Afar and persists northward. Another very elevated V_P/V_S value of 2.042 is observed at station TIOE. In general, crustal thickness in the region is negatively correlated with the bulk V_P/V_S ratio (see Figure S20 in the Supplementary Information), implying that the thinnest crust is also the most intruded. This indicates that the stretched, thinned nature of the crust beneath the Danakil block has resulted in the presence of partial melt and focused magmatism, similar to what is observed by Dixon et al. (1989) in the Red Sea. It is possible that this melting and volcanism was what initially caused the onset of crustal extension, as postulated by Drury et al. (1994). Our observations suggest that it is likely that magmatic extension continues to play a dominant role in the final stages of continental breakup in Afar. Stations ASSE, CAYE and FAME have V_P/V_S values that are greater than 1.8, which are closer to the expected values for oceanic crust (Christensen, 1996). The most likely explanation for this is the presence of magmatic intrusions causing the continental crust to be more mafic in composition.

3.4.2 Crustal sills beneath Nabro volcano

Our V_P/V_S result at Nabro volcano (station NAB, Figure 3.2) is relatively low (1.773). Nabro erupted in 2011 and therefore the crust beneath the volcano might be expected to have an elevated V_P/V_S ratio, reflecting melt storage. However, it is important to consider the pore geometry of melt storage and its influence on seismic velocities (Takei, 2002). Hammond (2014) demonstrated that the relative orientation and alignment of melt has a significant effect on H - κ estimates. If melt is vertically aligned, the shear wave conversion will be split as it passes through, with each of the split waves producing different κ estimates due to their different velocities. When there is good back-azimuthal coverage at a station, the overall stacking procedure will be dominated by the slow shear wave velocity, and thus produces a higher V_P/V_S estimate. This effect becomes less pronounced as melt becomes more horizontally aligned. The petrological study of Donovan et al. (2018) concluded that Nabro hosts

distinct batches of magma in sills throughout the crust, which might go some way towards explaining the low bulk V_P/V_S ratio estimate produced by H- κ stacking.

Other geophysical studies have proposed crustal magma reservoirs taking the form of sub-horizontal lenses. Ground deformation observed at the Rhenish Massif in Central Europe has been explained as melt accumulation in sub-horizontal sill-like structures (Silverii et al., 2023). A seismic reflection study of the Juan de Fuca Ridge concludes that sub-horizontal stacked magma lenses are present in multiple locations beneath the ridge, providing sources for eruption events (Carbotte et al., 2021). The volcanic plumbing system around Mayotte island includes a subhorizontal magma reservoir lying between a new seamount and Mayotte, as suggested by the deformation study of Cesca et al. (2020). A local velocity model for Mayotte found an average V_P/V_S ratio of 1.6, an even lower value than our estimate for Nabro, which the authors attribute to the presence of low aspect-ratio cracks filled with gas-rich magma (Lavayssière et al., 2022).

However, the low V_P/V_S ratio cannot be solely explained by horizontally-aligned melt. Alternatively, the overall crust at Nabro could be melt-depleted, with melt travelling directly up from the mantle to shallow storage regions just prior to eruption. Nabro’s high crustal thickness could be due to these magmatic intrusions and additions occurring through time. There is petrological evidence for a significant amount of old eruptive products in the crust beneath Nabro, due to the presence of xenocrystic material in the erupted magmas Donovan et al. (2018). Unfortunately, the back-azimuthal coverage is not comprehensive enough to carry out a crustal anisotropy study at Nabro to investigate these proposals further (e.g., Liu and Niu, 2012; Hammond, 2014).

3.4.3 Implications for rifting through the Gulf of Zula

Afar is a diffuse extensional province; Courtillot et al. (1980) note that it is difficult to identify plate boundaries in Afar due to the homogeneous stretching of the region. Manighetti et al. (2001) find that strain within Afar is localized in disconnected rift segments, currently opening and propagating through numerous active faults,

open fissures and volcanism. The active rift axis passes from the Red Sea through the Gulf of Zula to the west of the Danakil block and becomes more distributed further south in Afar, eventually spanning distances of ~ 175 km (Figure 26 in Manighetti et al., 2001; Kogan et al., 2012). The rotation of the Danakil block towards Arabia favours this propagation of rifting on land in Afar rather than a southward propagation of the Red Sea Rift beyond 16°N (Sani et al., 2017).

We note that station DOLE, despite its proximity to the rift axis through the Gulf of Zula, has a relatively thick crust of 27.2 km. However, it is important to notice that the receiver functions mainly sample the crust to the north-east of the station (and thus further from the rift), due to the back-azimuthal coverage. Therefore, the crustal thickness recorded at this station likely reflects a portion of unstretched continental crust on the Danakil microplate.

Viltres et al. (2020) conclude from their GPS study that inter-rifting deformation is not localised directly in the Gulf of Zula. Instead it is accommodated by volcanic vents and distributed faults 20 km west of the Gulf of Zula. Station EITE (E) lies to the west of this area of active deformation, with a crustal thickness of 25.5 km and a relatively high V_P/V_S ratio of 2.05. We postulate that extension may continue to shift westwards if the crust is rheologically weak and magma-rich at the edge of the plateau.

3.4.4 Topographic elevation and crustal structure

Topographic elevation can primarily be attributed to the isostatic effects of crustal thickness and density variations (e.g., Airy, 1855; Pratt, 1855). However, a secondary control on continental topography is the density structure of the mantle (e.g., Hager et al., 1985; Lithgow-Bertelloni and Silver, 1998; Hoggard et al., 2021; Sembroni et al., 2016; Stephenson et al., 2021). Our results can therefore place indirect constraints on crustal and mantle density structure. We investigate the relationship between topographic elevation (ϵ) and crustal thickness for central and northeastern Afar and the Danakil block (Figure 3.3a).

There is a demonstrable linear relationship between H and ϵ , suggesting that 3 km of variation in the topographic elevation in the study area can be ascribed to variation in crustal thickness (Figure 3.3a). A similar relationship between topographic elevation and crustal thickness is widely accepted from global observations (see Figure 5 in Stephenson et al., 2024; Lamb et al., 2020) and plotted in Figure 3.3a for comparison). However, two characteristics of the distribution of H and ϵ in our regional study provide a rare opportunity to quantify the regional properties of the crust and uppermost mantle. First, the slope of the relationship between H and ϵ is on the lower end of globally averaged estimates (e.g., Lamb et al., 2020). This difference is not unexpected given the rift setting of the region since the value of the slope is largely controlled by the density of the crust, although it has seldom been quantified (e.g., Ebinger et al., 1989a; Tiberi et al., 2005). Furthermore, the intercept crustal thickness value is shifted upwards, which implies significant levels of sub-plate topographic support which must arise from processes operating in the mantle beneath the region. In other words, topographic elevation is generally higher than would be expected for a given value of crustal thickness if crustal isostasy were the only control on topography (Airy, 1855; Pratt, 1855). Here, we use these observations to provide new constraints on lithospheric and sub-plate density structure.

3.4.4.1 Bulk crustal density

We can quantify the difference in average crustal density between our study region and the global average by using the gradient of the plot in Figure 3.3a to estimate crustal density beneath central and northeastern Afar and the Danakil block (e.g., Lamb et al., 2020; Stephenson et al., 2024). Assuming Airy isostatic equilibrium, the slope of the relationship between crustal thickness, H and topographic elevation, ϵ , is given by on the ratio $(\rho_a - \rho_c)/\rho_a$, where ρ_c is crustal density and ρ_a is asthenospheric mantle density. This relationship further assumes that topography is compensated in the asthenospheric mantle and that lithospheric mantle structure and crustal density are both consistent across the region of interest (c.f. Lamb et al.,

2020). Hence if we can estimate the local value of ρ_a , it is possible to place a bound on the average density of the crust in our study region.

To estimate the value of ρ_a , we must calculate the effects of temperature and pressure upon mantle rocks given local constraints on asthenospheric temperature structure. Mantle density as a function of pressure, P , and temperature, T , is given by

$$\rho(P, T) \approx \rho_{m_0} \exp\left(\frac{P}{K} - \alpha T\right), \quad (3.1)$$

where $\alpha = (3.28 \pm 0.25) \times 10^{-5} \text{ K}^{-1}$ is the thermal expansivity, $K = 140 \pm 20 \text{ GPa}$ is the bulk modulus of the mantle and $\rho_{m_0} = 3330 \text{ kg m}^{-3}$ is mantle density at surface pressure and temperature. In this way, we place constraints on T and P in the asthenosphere beneath our study region.

Ball et al. (2021) use inverse modeling of rare earth element concentrations in mafic volcanic rocks to estimate that the sub-plate potential temperature beneath north-eastern Ethiopia is $\sim 1442 \pm 50 \text{ }^\circ\text{C}$ and lithospheric thickness $a_1 = 30 \pm 20 \text{ km}$. Shear wave velocity can also be converted into temperature using the experiment-based parameterisation of Yamauchi and Takei (2016), as carried out by Priestley and McKenzie (2013), Richards et al. (2020) and Hoggard et al. (2020). This approach minimises the weighted misfit between: predicted and observed values of seismic velocities as a function of oceanic plate age and depths $\leq 112.5 \text{ km}$; the predicted shear wave velocity as a function of the mantle isentrope between 225 and 400 km depth; the globally averaged shear-wave attenuation as a function of depth; and finally published and calculated values of mantle viscosity (Richards et al., 2020; Hoggard et al., 2020). When applied to the recent tomographic model of Celli et al. (2020a), this approach yields an asthenospheric potential temperature of $T_P \approx 1440 \pm 30 \text{ }^\circ\text{C}$. To extract a lithospheric thickness, this temperature model is contoured at the 1175°C isotherm, which approximates the temperature at which convection begins. We find that the lithosphere beneath Afar and the Danakil microplate is between ~ 40 and 50 km thick. To ensure that we place the depth of isostatic compensation within the asthenospheric mantle, we choose

a depth to the base of the lithosphere of 50 ± 20 km. Assuming that the mantle is isentropic, we assume a mantle potential temperature beneath the study region of $T_P = 1440 \pm 50$ °C. Hence, using

$$T_a = (T_P + 273) \exp\left(\frac{g\alpha z}{C_p}\right) - 273, \quad (3.2)$$

we obtain asthenospheric temperature at this depth of $T_a = 1465 \pm 52$ °C, given acceleration due to gravity, $g = 9.81\text{ms}^{-2}$ and heat capacity, $C_p = 1187 \pm 50\text{J}^\circ\text{C}^{-1}$ (Mckenzie and Bickle, 1988).

These values for asthenospheric temperature and lithospheric thickness are corroborated by several other studies. Petrogenic modelling by Ferguson et al. (2013) finds a mantle potential temperature of 1450°C , corroborated by the geodynamic models of Armitage et al. (2015). Stephenson et al. (2023) use major element thermobarometry of mafic magmas to estimate a lithospheric thickness of 40-60 km beneath northern Afar (see methodology of McNab and Ball, 2023; Plank and Forsyth, 2016). Using a similar approach as we have followed above, Fishwick (2010) also obtains a lithospheric thickness of about 50 km. Finally, Ebinger and Hayward (1996) find that the elastic thickness of the plate in the Danakil Depression is between 3 and 8 km. Although they obtain a higher value on the Ethiopian plateau of 56 km, this value is still lower than the stable cratonic elastic thicknesses of 64–79 km observed elsewhere on the African plate by Ebinger et al. (1989a). These results suggest that the lithosphere is anomalously weak beneath our study region and therefore consistent with thin lithosphere and elevated sub-plate potential temperatures.

We use the relationship $P_1 = \bar{\rho}ga_1$ to find the pressure at the base of the modern-day lithosphere, where $\bar{\rho} \approx 3000 \text{ kg m}^{-3}$ is overburden density, and a_1 is depth to the base of the lithosphere. We find that $P_1 \approx 1.3 \pm 0.5$ GPa. Therefore $\rho_a \approx 3210 \text{ kg m}^{-3}$. In Figure 3.3a we apply a non-biased linear regression (i.e., a total least squares regression; see Supplementary Information Text S4) that assumes uncertainty in both topography and crustal thickness (± 30 m and ± 3 km, respectively) and find that $(\rho_a - \rho_c)/\rho_a = 0.11$, which is on the lower end of the

global average value of around 0.11–0.13 (Lamb et al., 2020; Stephenson et al., 2024). Using our estimated value of ρ_a and by propagating uncertainties in T_a and P_1 , we find that in our study region, $\rho_c \approx 2850 \pm 20 \text{ kg m}^{-3}$. In our study region, crustal thickness is $26 \pm 10 \text{ km}$ (i.e., median and interquartile range). Stephenson et al. (2024) provide a function to predict average bulk crustal density as a function of crustal thickness, which for this range yields a predicted mean crustal density of $2800 \pm 10 \text{ kg m}^{-3}$, which indicates that the crust in our study region is somewhat denser than the global average. We compare the result to the results of Tiberi et al. (2005), who carry out inverse modeling of gravity anomalies in Afar. In the upper 10–20km beneath northern Afar, they find a large positive lateral density contrast compared to other areas of their study region. This large change in density is interpreted to be a result of a combination of crustal thinning and the addition to the crust of dense mafic material due to magmatic intrusions.

In summary, this somewhat elevated crustal density likely reflects the substantial magmatic intrusions and additions to the crust by mafic volcanic rocks that have taken place in this region. It furthermore provides a new and independent constraint that adds weight to our earlier conclusion that the high V_P/V_S ratios we observe are likely to be due to magmatic activity. Indeed, it is probable that these density estimates represent a lower bound for the thinnest crystalline crust in Afar and the Danakil microplate, much of which is overlain by thick sedimentary layers. In the Danakil Depression, the layers of sediment reach 5 km in extent (Hutchinson and Engels, 1972; Bastow and Keir, 2011; Bastow et al., 2018; Rime et al., 2023). If crystalline crust of thickness 15 km is overlain by 5 km of sediment of density 2700 kg m^{-3} , and if the bulk density of both layers combined is 2850 kg m^{-3} , it follows that the density of the crystalline crust is $\sim 2900 \text{ kg m}^{-3}$. This value is consistent with significant mafic intrusion.

3.4.4.2 Sub-plate topographic support

Next, we turn our attention to the topographic effects of elevated mantle potential temperatures beneath Afar and the Danakil block. It is well-established that mantle

upwelling from a large low velocity province has had a significant effect on the tectonics and topography of Africa (e.g., Ebinger and Sleep, 1998; Ritsema et al., 1999; Kendall and Lithgow-Bertelloni, 2016; Nyblade and Robinson, 1994). While the magnitude of sub-crustal support from the mantle has been estimated in the Main Ethiopian Rift, a paucity of data means that the extent of the hot mantle swell northwards towards Afar and the Danakil block has not been well constrained (Stephenson et al., 2024; Holdt et al., 2022; Sembroni et al., 2016). From the relationship between crustal thickness and elevation, it is clear that topography is anomalously high given the crustal thickness in our study region. It is notable that, for $H \approx 10\text{--}20$ km, the topography is approximately at sea level. However, globally, the average crustal thickness is about 32.1 km for regions at sea level (Lamb et al., 2020; Stephenson et al., 2024). To illustrate this point, take station CAYE, which has crustal thickness of $H = 32.7$ km, which is very close to the thickness of crust that is globally expected to reside at sea level. Instead, CAYE is located at an elevation of about $\epsilon = 2.1$ km. This marked offset implies a significant level of regional sub-crustal support that sustains the topography in Afar and the Danakil microplate.

Using our results, we can for the first time quantify this dynamic support regionally. Some proportion of this locally anomalous topography will be supported by the flexural rigidity of the lithosphere. However, the size of our study region exceeds the expected elastic half-wavelength given an elastic thickness of 5–49 km that would result in regional elastic support (Ebinger and Hayward, 1996). It is therefore possible to assess the extent of regional excess topography by using the regression of H against ϵ . The difference between the mean intercept values of the local and global regressions of H against ϵ indicates that the study region is on average 1.5 ± 0.4 km (i.e., 2 standard errors on the mean) higher than would be expected as a result of crustal thickness variations alone (Figure 3.3a).

A hot mantle upwelling can generate uplift in three interconnected ways (Hager et al., 1985; Hager and Richards, 1989; Molnar et al., 2015; Hoggard et al., 2021). First, the upward flow of hot mantle produces viscous forces that cause vertical

tractions on the base of the plate (e.g., Forte et al., 1993; Molnar et al., 2015). Secondly, the buoyancy of the hot mantle material generates isostatic uplift at the surface (e.g., Rudge et al., 2008; Richards et al., 2020). Finally, erosion of the base of the lithosphere by heat and/or melt and fluid infiltration can thin the lithospheric mantle (e.g., Bird, 1979). Here, we explore the effects of the second and third processes in our study region by estimating the uplift that is generated by the combined effects of a hot thermal anomaly beneath the plate and by thinning of the lithospheric mantle.

Uplift above a hot asthenospheric anomaly of thickness H and excess temperature ΔT is given by

$$U = \frac{\alpha T_1}{1 - \alpha T_1} \left(\frac{a_1^2}{2a_0} + \frac{a_0}{2} - a_1 + \frac{\Delta T}{T_1} h \right), \quad (3.3)$$

where a_0 and a_1 are initial and final lithospheric thicknesses and T_1 is ambient mantle temperature. We assume an initial lithospheric thickness of $a_0 = 120$ km, and take an asthenospheric channel to be $h = 100$ km thick (McKenzie, 1978; Rudge et al., 2008). The effects of using different values of a_0 and H are reported in the Supplementary Information in Text S5 and Figure S21.

Next, we return to the independent estimates of mantle temperature and lithospheric thickness determined from tomographic and geochemical constraints. We carry out a grid search by sweeping through the range of reported values for a_1 and ΔT using Equation (3.3) (see, e.g., Hoggard et al., 2020; Ball et al., 2021; Stephenson et al., 2024; Ebinger and Hayward, 1996)). The residual topographic estimates calculated from our simple regression between H and ϵ in Figure 3.3a are consistent with the magnitude of uplift generated above a temperature anomaly of around 110°C residing in an asthenospheric channel of 50–150 km thickness beneath lithosphere that has been thinned by around 80 km (i.e., from 120 to 40 km; compare hatched area with circles and star on Figure 3.3b). These values are corroborated by the aforementioned independent estimates for sub-plate temperature and lithospheric thickness. It is important to emphasise that this signal of sub-plate topographic support is regional, extending beneath regions of both thick and

thin crust, necessitating the emplacement of a hot sub-plate thermal anomaly and removal of the lithospheric mantle beneath both rift and plateau.

These results are consistent with the presence of a major mantle upwelling beneath Afar and the Danakil microplate. We note that our results do not attempt to estimate the mantle flow component of sub-crustally supported topography, nor do they attempt to quantify the component arising from below the uppermost 200 km of the mantle. Furthermore, a potentially significant component of topographic support from the lithospheric mantle arises from lithospheric extension. However, the mantle swell is clearly a widespread feature that extends from the Main Ethiopian Rift into northern Afar and the Danakil microplate and sub-plate temperature anomalies appear to be required to explain the height of topography in the region.

3.5 Conclusion

We have produced the first estimates of crustal thickness and bulk crustal V_P/V_S ratio in Eritrea, across an area encompassing northeastern Afar and the Danakil microplate. We calculate receiver functions from a temporary regional seismic array and apply the modified H- κ stacking technique of Ogden et al. (2019) to constrain the bulk crustal properties at eight locations in Eritrea. To validate this procedure, we also carry out H- κ stacking at stations in central Afar, producing estimates that are consistent with previous studies in the region.

This study has revealed the heterogeneity of crust in Afar and the Danakil microplate. We calculate a variability in crustal thickness of ~ 16 km throughout the study area, ranging from ~ 19 – 35 km. This reveals that the Danakil microplate is not an undeformed, unified block, but rather has experienced extension and crustal thinning resulting in crust as thin as 18.9 km. We also see variation in V_P/V_S ratio across our study region. Some stations have values typical of continental crust (~ 1.77). Others show higher values of 1.8 – 1.86 , likely a consequence of alteration by magmatic intrusion or addition to the crust. Two stations have V_P/V_S greater than 2.0 , which we associate with the presence of partial melt, suggesting that magmatic extension continues to be important in the final stages of continental rifting.

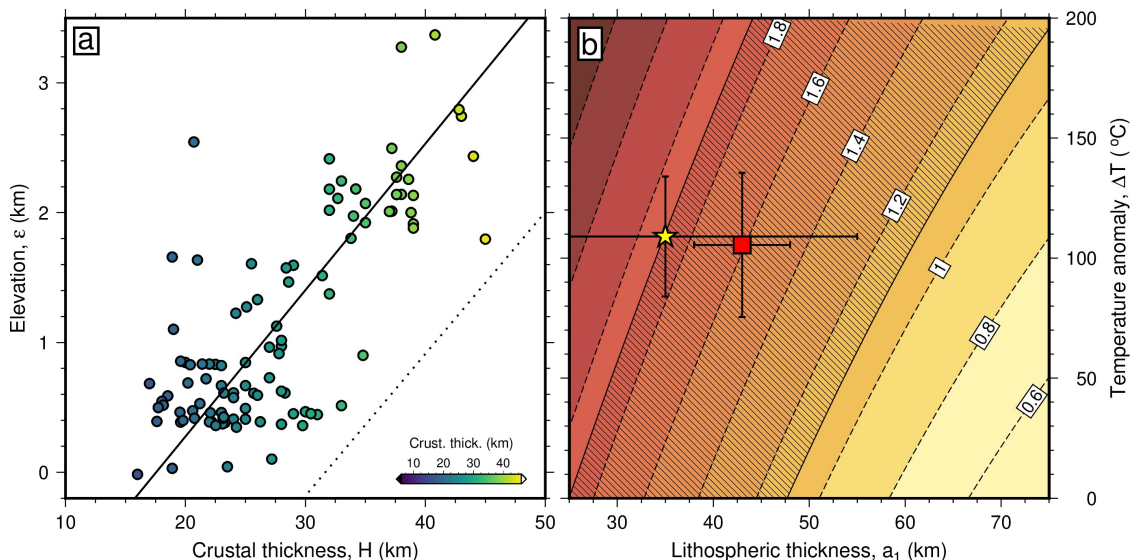


Figure 3.3: Crustal and mantle structure. (a) Topographic elevation plotted against crustal thickness estimates (coloured circles) for the study region, from this study and the previous receiver function studies of Hammond et al. (2011c); Reed et al. (2014); Ahmed et al. (2022). Elevation estimates are from the GMT Global Earth Relief grid, with the topography low-pass filtered for wavelengths $> 30\text{km}$ (Wessel et al., 2019). The black line is the best-fitting regional isostatic relationship where $(\rho_a - \rho_c)/\rho_a = 0.11$; the dotted black line is the average global isostatic relationship (Stephenson et al., 2024). (b) Sub-crustally supported topography as a function of asthenospheric mantle temperature and lithospheric thickness, assuming a temperature anomaly is constrained to a 100 km thick channel and equilibrium lithospheric thickness is 120 km. Contours indicate regional elevation anomaly, ϵ , generated by lithospheric thickness and sub-plate temperature. The shaded area represents the range of regional sub-crustal support ($\Delta\epsilon = 1.5 \pm 0.4$ km) estimated from the offset in intercept values between the regional crustal thickness-elevation relationship and the global one. The red square is plotted at the sub-plate temperature and lithospheric thickness given by the tomography model of Celli et al. (2020a) after conversion into temperature following the methodology of Hoggard et al. (2020) and Richards et al. (2020). The yellow star is plotted at the sub-plate temperature and lithospheric thickness given by inverse modelling of rare earth element compositions (Ball et al., 2021).

By examining the relationship between topographic elevation and crustal thickness, we observe a deviation from the global relationship between these parameters. We estimate a denser bulk crust (i.e., including both crystalline basement and sediment) across northern and central Afar and the Danakil microplate ($\rho_c \approx 2850 \pm 20 \text{ kg m}^{-3}$), which is evidence for intrusive magmatism altering the composition of the crust to be more mafic. Our results reveal that the topography is anomalously high for the observed crustal thicknesses in Afar and the Danakil microplate, implying a level of sub-crustal support from a hot thermal

anomaly beneath the crust.

Our observations of crustal properties address the previously existing data gap in northeastern Afar and the Danakil microplate, provide a useful base for further investigations of the crust and mantle within the Danakil microplate and at its boundaries, and give insights into the processes responsible for topographic elevation in the region.

Open Research

The seismic data used in this study are from the Eritrea Seismic Project (Hammond et al., 2011a), the Nabro Urgency Array (Hammond et al., 2011b), and the AFAR0911 network (Keir and Hammond, 2009), all publicly available through IRIS Data Services (<http://service.iris.edu/fdsnws/dataselect/1/>). When analysing the seismograms, we made use of the Seismic Analysis Code software (Helffrich et al., 2013) and the iterative deconvolution code “iterdecon”, available online at <http://eqseis.geosc.psu.edu/cammon/HTML/RftnDocs/thecodes01.html>. Figures and maps were plotted using Generic Mapping Tools (GMT) version 6 (Wessel et al., 2019) licensed under LGPL version 3 or later, available at <https://www.genericmapping-tools.org>.

Acknowledgements

The authors thank Anne Paul for her helpful editorial comments. We are also grateful for the comments and advice from two anonymous reviewers, which greatly improved this paper. The seismic data were collected with funding from the Natural Environment Research Council (NERC) projects NE/J012297/1, NE/E007414/1, and NE/D008611/1 and NSF grant EAR-0635789. The UK seismic instruments and data management facilities were provided under loan number 976 by SEIS-UK at the University of Leicester. The facilities of SEIS-UK are supported by NERC under Agreement R8/H10/64. Author MG was supported by a Doctoral Training Partnership studentship from NERC [NE/S007474/1]. We gratefully acknowledge

the cooperation we received from the Eritrea Institute of Technology, Eritrean government, Southern and Northern Red Sea Administrations, local sub-zones and village administrations. We thank the Department of Mines, Ministry of Energy and Mines for their continued support throughout the Eritrean project. We also thank Addis Ababa University, the Ethiopian Federal Government, Afar National Regional State Government and Ethioder tour and travel for vital help and support during the Afar deployment. Special thanks go to Zeraï Berhe, Mebrahtu Fisseha, Michael Eyob, Ahmed Mohammed, Kibrom Nerayo, Asresehey Ogbatsien, Andemichael Solomon and Isaac Tuum. We thank Alem Kibreab for vital help in facilitating the fieldwork. IRIS Data Services are funded through the Seismological Facilities for the Advancement of Geoscience (SAGE) Award of the National Science Foundation under Cooperative Support Agreement EAR-1851048.

4

Seismic tomography of Nabro caldera

Contents

4.1	Introduction	89
4.2	Data	92
4.2.1	Network and Data Collection	92
4.2.2	Data Selection	94
4.3	Methodology	96
4.3.1	Tomographic Method	96
4.3.2	1D Model Selection	100
4.3.3	Defining a Grid for 3D Inversions	102
4.3.4	Optimising Data-Model Fit	102
4.3.5	Earthquake Relocations	103
4.4	Results	104
4.4.1	Checkerboard Resolution Tests	104
4.4.2	Inverting for 3D Velocity Structure	106
4.5	Discussion	110
4.5.1	Defining a Local 1D Velocity Model at Nabro	110
4.5.2	Earthquake Detection Using U-GPD	113
4.5.3	Interpretation of V_P/V_S Variations	113
4.5.4	The Crustal Substructure Beneath Nabro Caldera	119
4.5.5	Comparisons with Other Volcanoes	120
4.5.6	Limitations and Future Work	121
4.6	Conclusion	122

Abstract

Understanding the crustal structure and the storage and movement of fluids beneath a volcano is necessary for characterising volcanic hazard, geothermal prospects and potential mineral resources. This study uses local earthquake traveltime tomography to image the seismic velocity structure beneath Nabro, an off-rift volcano located within the central part of the Danakil microplate near the Ethiopia-Eritrea border. Nabro underwent its first historically-documented eruption in June 2011, thereby providing an opportunity to analyse its post-eruptive state by mapping subsurface fluid distributions. We use a catalog of earthquakes detected on a temporary seismic array using machine learning methods to simultaneously relocate the seismicity and invert for the three-dimensional P- and S-wave velocity structures (V_P , V_S) and the ratio between them (V_P/V_S). Overall, our model shows higher than average P- and S-wave velocities, suggesting the presence of high-strength, solidified intrusive magmatic rocks in the crust. We identify an aseismic region of low V_P , low V_S and high V_P/V_S ratio at depths of 6–10 km b.s.l., interpreted as the primary melt storage region that fed the 2011 eruption. Above this is a zone of high V_S , low V_P and low V_P/V_S ratio, representing an intrusive complex of fractured rocks partially-saturated with over-pressurized gases. Our observations identify the persistence of magma in the subsurface following the eruption, and track the degassing of this melt through the crust to the surface. The presence of volatiles and high temperatures within the shallow crust indicate that Nabro is a viable candidate for geothermal exploration.

Plain Language Summary

Understanding the structure of the crust and the distribution and movement of fluids beneath a volcano allows for the assessment of volcanic hazard, geothermal potential and possible mineral extraction. To identify different regions of the crust and differentiate between fluids, we use the fact that the speed of seismic waves depends on the material they are travelling through. For example, seismic waves will travel through magma (molten, or liquid, rock) at lower speeds than in the

surrounding rock. The focus of this study is Nabro volcano in Eritrea, which erupted in 2011. We use earthquakes that have been automatically detected following the eruption to image the structure of the crust in the form of 3D variations in seismic wave speeds. This identifies a volume of magma stored at depths of 6—10 km below sea level, which fed the eruption. Above this, we observe a region of rocks that are likely remnants of earlier eruptions at Nabro, with fractures containing gases at high pressure. The source of this high pressure is the release of gas from the magma storage zone. The presence of hot fluids means Nabro could be used as a source of geothermal power in the future.

Key Points

- 3D seismic modelling reveals the structure of the magmatic and hydrothermal systems beneath Nabro volcano in Eritrea.
- The primary melt storage region feeding the 2011 eruption is located at depths of 6—10 km below sea level.
- Degassing from the magma storage zone causes overpressure in partially-saturated, fractured intrusive complex above.

4.1 Introduction

Modelling the crustal architecture of a volcano is of fundamental importance; interactions between magmatic and hydrothermal systems play a central role in volcanic unrest and eruption (e.g. Chouet and Matoza, 2013; Pritchard et al., 2019; Wilks et al., 2020). However, these systems are often poorly characterized, both due to their complexity and to the difficulty of probing the crust in sufficient detail. Further motivation for investigation is that volcanic-hosted hydrothermal systems may be a source of geothermal energy production (Reinsch et al., 2017) or metal-rich brines (Blundy et al., 2021).

Seismic body wave tomography is a powerful geophysical tool used to image the Earth’s interior on various scales. It has been applied locally to deforming volcanoes

in order to understand their subsurface active magmatic and hydrothermal processes (e.g. Patane et al., 2002; Chiarabba and Moretti, 2006; Korger and Schlindwein, 2014; Greenfield et al., 2016; Wilks et al., 2020; Koulakov et al., 2021). Seismic velocities of rocks are influenced by a multitude of factors, including lithology, fractures, temperature, the presence of fluids and gases, fluid saturation and porosity. Knowledge of the seismic velocity structure can therefore help to identify melt-bearing regions, hydrothermal fluids and over-pressurized gases beneath a volcano (e.g. Londoño and Sudo, 2003; Vanorio et al., 2005; Lin, 2013; Wilks et al., 2020). These observations are crucial to understanding volcanic unrest, characterising future volcanic hazard and assessing geothermal potential.

Using seismic data collected following the 2011 eruption of Nabro, we present high-resolution tomographic images of the crust with the aim of improving understanding of the post-eruptive state and dynamics at depth beneath the volcano. Nabro volcano is located on the central part of the Danakil microplate near the Ethiopia-Eritrea border in the Afar region. The Afar depression is part of the East African Rift System, an active continental rift system, and contains the triple junction between Arabian, Nubian and Somalian plates (Hammond et al., 2011c). Most of the active volcanism associated with continental rifting is found along the central rift axis. However, Nabro is offset from the axis of spreading and is thus known as an “off-rift” or “off-axis” volcano (Barberi et al., 1974; Wiart and Oppenheimer, 2005). Together with the neighbouring caldera of Mallahle, Nabro makes up the Bidu Volcanic Complex. Nabro and Mallahle are characterized by large calderas, thought to have been formed circa 130 and 295 ka, respectively (Oppenheimer et al., 2019). The alignment of the volcanic centers bears NE-SW, striking obliquely to the NW-SE trend of the Red Sea (Wiart and Oppenheimer, 2005; Goitom et al., 2015). Nabro is the largest volcano in the Nabro Volcanic Range (NVR), which runs in a NNE-SSW direction from Bara’Ale volcano in Ethiopia to the Kod Ali formation in the Red Sea (Wiart and Oppenheimer, 2005). Nabro’s summit is 2248 m above sea level, and its caldera reaches a diameter of 8 km (Wiart and Oppenheimer, 2005). It remains unclear how magma is supplied to off-rift volcanoes

such as Nabro; further, their role in accommodating extension is not well understood (Maccaferri et al., 2014). Indeed, the propagation of strain transfer from the Aden and Red Sea plate boundaries into the Afar region south of Nabro is complex and transient, with active faults distributed over hundreds of thousands of square kilometres (Manighetti et al., 2001). Possible explanations for the NVR’s extensive off-rift magmatism include reactivation of an older, pre-rift structure (Barberi et al., 1974) or localized diapiric upwellings from depth (Hammond et al., 2013).

On 12 June 2011, Nabro volcano underwent its first eruption on historical record—the last dated activity occurred within the caldera circa 23 ka (Oppenheimer et al., 2019). The volcano was unmonitored at the time of the eruption, with no geophysical surveillance networks operating in Eritrea. The eruption resulted in seven fatalities and displaced some 12,000 people (Goitom et al., 2015). The explosive activity generated significant tephra clouds and released 4.5 Tg of SO₂ into the atmosphere within the first 15 days, producing the largest stratospheric aerosol perturbation since the 1991 Pinatubo eruption (Theys et al., 2013; Fromm et al., 2014). Since the eruption, Nabro has been identified as one of the main geothermal prospects in Eritrea due to increased fumarolic activity at the surface (Yohannes, 2012).

Geodetic modelling suggests that a shallow, NW–SE-trending dyke fed the eruption, which triggered slip on parallel normal faults, consistent with the orientation of vents within the crater (Goitom et al., 2015). Petrological analysis by Donovan et al. (2018) identifies two distinct batches of magma, one more primitive and the other high in sulfur and water content. The authors propose that the latter batch underwent isobaric crystallisation in a storage region at $\sim 5\text{--}7$ km depth below sea level (b.s.l.), while the more primitive batch rose rapidly to the shallow crust from depth. In the months following the eruption, Nabro experienced subsidence at a slowly decaying rate (Hamlyn et al., 2014, 2018). By inverting the deformation field, Hamlyn et al. (2018) propose a best-fitting deflating Mogi source at 6.4 ± 0.3 km depth b.s.l..

A temporary seismic network was established around Nabro in the aftermath of the eruption, operational from 31 August 2011 until October 2012. Lapins et al.

(2021) trained and validated a novel deep learning model on this data in order to automatically detect phase arrivals. When deployed, the deep learning model significantly augmented the seismic catalog analysed in previous studies. From this catalog, Lapins (2021a) calculates hypocenter locations, local and moment magnitudes, path/site attenuation effects and b-values. A key result is that seismicity beneath Nabro lies above and below an inferred, aseismic magma storage zone at depths of 6–9 km b.s.l., consistent with the modelled Mogi source from previous studies (Hamlyn et al., 2014; Goitom et al., 2015) and petrological inferences about magma storage depths (Donovan et al., 2018). Events below the reservoir are thought to result from small pulses of magma or volatile migration, while events above the aseismic zone could reflect outgassing processes, migrations of fluid or melt into the reservoir or intense fracturing as a result of the observed subsidence (Lapins, 2021a). The patterns of seismicity to the northeast of Nabro indicate that deeper fluid or magmatic processes have triggered movement on a shallower fault, suggesting that fluids may play an important role in regional extensional processes (Lapins, 2021a).

Lapins (2021a) notes that one of the major limitations of their study is the lack of a well-constrained velocity model. Here, we apply seismic tomography methods to the dataset from Lapins et al. (2021) to derive a more accurate velocity model. We then use this new model to jointly carry out 3D P-wave (V_P), S-wave (V_S) and V_P/V_S tomography and earthquake hypocenter relocation, in order to yield further insight into the subsurface processes responsible for the seismicity and surface deformation at Nabro. By interpreting these tomographic images, we aim to characterize the migration and distribution of volcanic fluids in Nabro’s active magmatic system. These results have particular relevance for the assessment of Nabro’s geothermal energy potential, as well as its future seismic and volcanic hazard.

4.2 Data

4.2.1 Network and Data Collection

The seismic data used in the tomographic inversion were collected by a temporary local seismic network deployed in the aftermath of Nabro’s 2011 eruption. Eight

3-component broadband 30 s Gralp seismometers (five CMG-6TD and three CMG-40TD) were provided by SEIS-UK to monitor Nabro’s post-eruptive state (Hamlyn et al., 2014). The network was fully operational from 31 August 2011 until October 2012. However, one of the 40TD stations, NAB6, was damaged due to flooding and thus was inoperable, producing no useable data (Lapins, 2021a). NAB7 also had frequent data gaps but was still used for phase arrival picking and event location, and therefore we also use it in our tomographic inversions, along with the other six stations. The data were all initially recorded at 100 Hz sample frequency and then switched to 50 Hz sample frequency early in October 2011 (Lapins, 2021a).

Manually picking seismic phase arrivals is time-consuming, and can be especially difficult in volcanic settings due to the fact that volcano-tectonic earthquakes tend to be fairly low magnitude (< 4) events. A previous manual analysis of the Nabro seismic data only covered the time period 31 August – 31 December 2011 (Goitom, 2017), which left eight months of data unpicked. Therefore, a new deep learning model for automated phase arrival detection based on a convolutional neural network, known as U-GPD, was applied to the seismic data from the temporary network around Nabro (Lapins et al., 2021). The U-GPD model is trained and validated using 35 days of manually picked data from Goitom (2017). To overcome issues surrounding the use of a small training set, they use transfer learning on an existing deep learning model for phase arrival detection (Ross et al., 2018) trained using millions of phase arrivals from earthquakes in Southern California. The resulting U-GPD transfer learning model was shown to outperform two existing, comprehensively trained models, PhaseNet (Zhu and Beroza, 2019) and GPD (Ross et al., 2018), and the existing manual catalog in terms of pick error and number of phase arrivals detected (Lapins et al., 2021). When the automated phase arrival picks and original manual pick times are compared, the root mean square deviation for P-wave picks is 0.038 s and for S-wave picks it is 0.053 s (Figure 6 in Lapins (2021a)). The model is far more efficient than manual phase picking, processing the 14 months of seismic data in less than 4 hours. The three output channels of the U-GPD model give the probability of a P-wave arrival, S-wave arrival or neither

(noise), respectively. A P or S prediction probability must exceed a threshold value of 0.4 to be identified as a true phase arrival detection (Lapins et al., 2021).

Events are then located using NonLinLoc, a probabilistic, nonlinear hypocenter location package (Lomax et al., 2000). The 1D starting velocity model used in the NonLinLoc inversion is based on the crustal structure of the Afar region deduced from wide-angle controlled-source seismology and assuming a V_P/V_S ratio of 1.76, the approximate average for continental crust (Ginzburg et al., 1981; Goitom, 2017). This produces an initial catalog of 31,387 events with at least four P-wave arrivals and one S-wave arrival that are available for subsequent use in seismic tomography inversions (Lapins et al., 2021).

4.2.2 Data Selection

Seismic tomography is highly reliant on accurate traveltime picks, and therefore we restrict the catalog produced by U-GPD and located in NonLinLoc based on the inversion statistics and event properties. We select earthquakes with at least four P and four S phases, azimuthal gaps that are less than 180° and location errors of less than 2 km, reducing the catalog to 11,319 earthquakes (Figure 4.1).

The U-GPD deep learning model does not include explicit pick uncertainties, but following Lapins (2021a), we associate errors to the picks based on the probability of being a true arrival. If this probability exceeds 0.85, a pick error of 0.05 s is assigned to it. Pick arrivals with probabilities $0.7 - 0.85$, $0.55 - 0.7$, and $0.4 - 0.55$ are assigned pick errors of 0.1 s, 0.2 s, and 0.3 s, respectively. The tomography results are independent of the absolute values of these errors due to the application of regularization (see Section 4.3.4); rather, it is the relative difference in the pick errors that matters, giving less weight to the picks we are less confident in during the tomographic inversion.

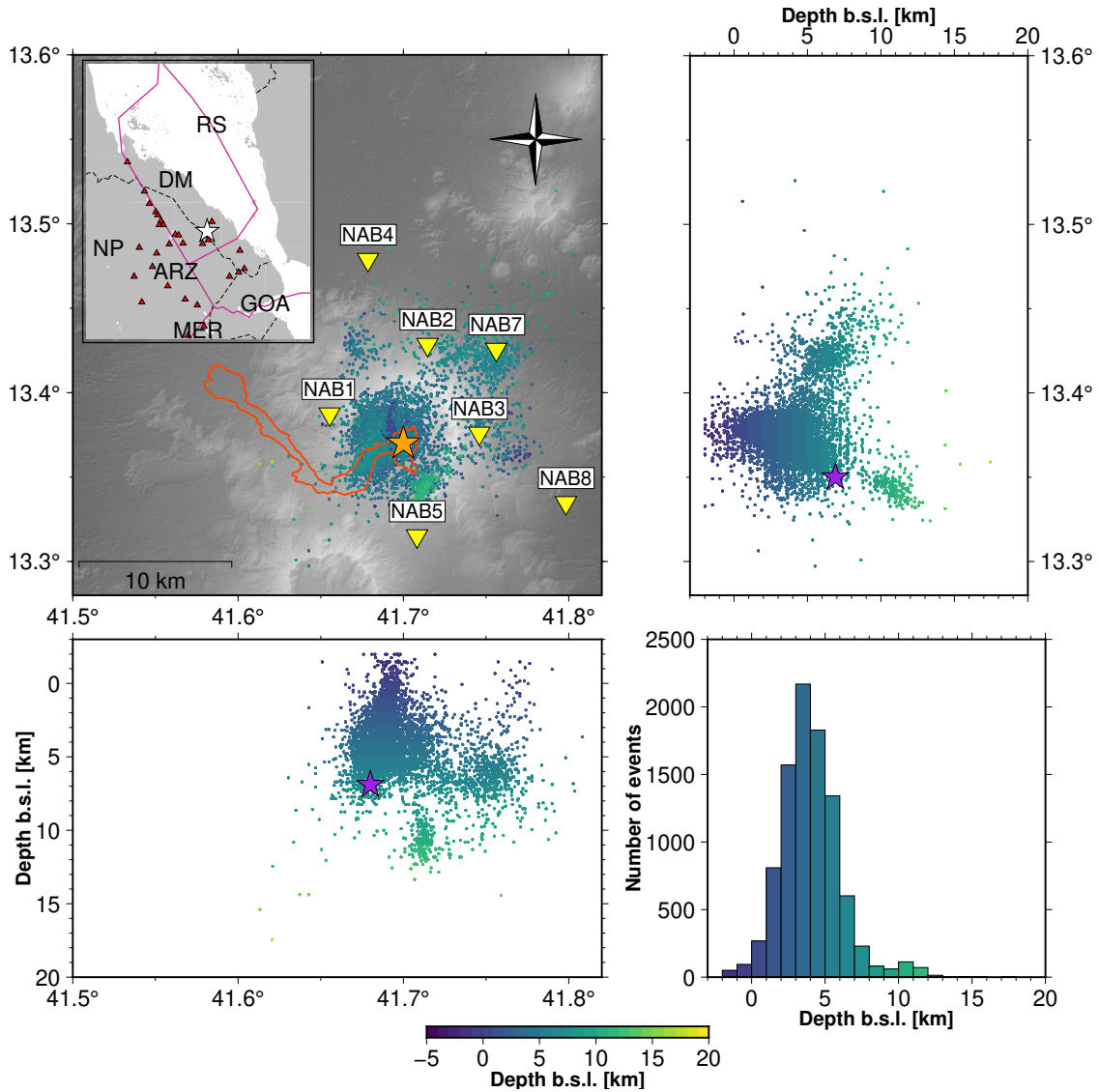


Figure 4.1: The final event catalog after relocation in a joint inversion for velocity structure and earthquake location. Thin black lines represent the caldera rims of Nabro and Mallahle calderas. Seismic stations from the temporary seismic network are plotted as inverted yellow triangles, excluding the inoperational station (NAB6). The dark orange line shows the extent of the 2011 eruption lava flow, and the orange star shows the location of the vent region (Hamlyn et al., 2014). The cross-sections show the catalog projected into the longitude-depth and latitude-depth planes, with the Mogi source from Hamlyn et al. (2014) represented by a purple star. The seismicity is coloured by depth below sea level (b.s.l.) and the histogram uses bins of 1 km depth b.s.l.. The inset shows a regional map of the Afar Triple Junction. Red triangles represent Holocene volcanoes recorded in the Smithsonian catalog 'Volcanoes of the World' database. The white star indicates the location of Nabro. Dashed lines are political borders. DM: Danakil Microplate, RS: Red Sea Rift, GOA: Gulf of Aden Rift, MER: Main Ethiopian Rift, ARZ: Afar Rift Zone, NP: Nubian Plate.

4.3 Methodology

4.3.1 Tomographic Method

To investigate the subsurface velocity structure at Nabro, we make use of an iterative nonlinear tomographic inversion package, FMTOMO (Fast-Marching TOMOgraphy) (Rawlinson and Sambridge, 2004a,b; de Kool et al., 2006). FMTOMO inverts seismic traveltime data to constrain 3D V_P and V_S structure. The package has been adapted by Pilia et al. (2013) to permit the fully nonlinear relocation of hypocenters and to solve directly for V_P/V_S structure. It has been applied in a variety of tectonic settings, using either passive or active source datasets, or a combination of the two (e.g. Rawlinson et al., 2006; Rawlinson and Kennett, 2008; Brikke, 2010; Pilia et al., 2013; Korger and Schlindwein, 2014; Zenonos et al., 2019; Wilks et al., 2020).

The key innovation of FMTOMO is the use of an efficient, consistent and robust grid-based eikonal solver known as the fast marching method (FMM) (Sethian, 1996; Sethian and Popovici, 1999) to solve the forward problem of predicting traveltimes in a 2D or 3D heterogeneous, layered medium. Since the subsurface structure beneath volcanoes is often highly heterogeneous, this makes FMTOMO an appropriate choice for our study of Nabro. Furthermore, FMTOMO can invoke traveltime reciprocity when solving the forward problem. The FMM source points are interchanged with the receivers, and the eikonal solver computes traveltimes from each receiver location to all the other grid points, so that the complete traveltime field for each receiver is available rather than for each source. Typically, most of the computing time of FMTOMO is dedicated to calculating these traveltime fields, and a large ratio between the number of sources and receivers means that invoking the reciprocity principle can lead to a significant increase in efficiency. We refer the reader to de Kool et al. (2006) for a more detailed overview of FMM.

FMTOMO defines the seismic velocity field with a regular 3D grid of nodes, which are used as the control vertices of a mosaic of cubic B-spline volume elements. Cubic B-spline functions preserve continuity of the second derivative whilst also being defined in terms of local basis functions, meaning that changing the velocity

value of one node will only affect the velocities at nodes in the immediate vicinity. This creates a smoothly varying, locally-controlled velocity continuum. Cubic B-spline functions can also be rapidly evaluated, which is useful, since the multi-stage FMM requires several evaluations of the spline function.

The next step of the algorithm solves the linearized problem of matching observed and predicted traveltimes, i.e., finding model parameters that best satisfy the data. In this case, the data are the arrival time residuals, and the unknowns are the grid of vertices which control the pattern of the cubic B-spline velocity field. FMTOMO implements the gradient-based subspace inversion scheme of Kennett et al. (1988), which minimizes the objective function:

$$S(\mathbf{m}) = \frac{1}{2} \left[(\mathbf{g}(\mathbf{m}) - \mathbf{d}_{obs})^T \mathbf{C}_d^{-1} (\mathbf{g}(\mathbf{m}) - \mathbf{d}_{obs}) + \epsilon (\mathbf{m} - \mathbf{m}_0)^T \mathbf{C}_m^{-1} (\mathbf{m} - \mathbf{m}_0) + \eta \mathbf{m}^T \mathbf{D}^T \mathbf{m} \right], \quad (4.1)$$

where the vector $\mathbf{m}$ represents the model vector of unknown velocity parameters that are adjusted during the inversion process, $\mathbf{g}(\mathbf{m})$ are the predicted traveltime residuals associated with the model defined by $\mathbf{m}$, $\mathbf{d}_{obs}$ are the observed residuals, $\mathbf{m}_0$ is the reference model, $\mathbf{C}_d$ is the data covariance matrix and $\mathbf{C}_m$ is the *a priori* model covariance matrix. $S(\mathbf{m})$ is minimized when the model traveltimes most closely resemble the observed traveltimes. The subspace method locally minimizes the objective function by projecting the quadratic approximation of $S(\mathbf{m})$ onto an n -dimensional subspace of the full model space. In this case we choose a maximum of $n = 20$ orthogonal search directions, with singular value decomposition used to reduce the size of the subspace based on the magnitude of the singular values. Regularization constraints are applied to address solution non-uniqueness: the damping term encourages the search for models to remain within the vicinity of the reference model, whilst the smoothing term minimizes the amount of structural variation required to satisfy the observational constraint. Further information on the inversion scheme and how it is implemented can be found in Rawlinson et al. (2006).

Many local earthquake tomography algorithms rely on a linearized approach to the tomographic inversion (e.g. Evans et al., 1994). However, the hypocenter location problem is more strongly nonlinear than the velocity recovery problem,

meaning that the linearized approximation leads to poor results in regions where there is significant velocity heterogeneity and/or where source locations are not well constrained (Pilia et al., 2013). Since the computational cost of having a fully nonlinear inversion scheme for both velocity structure and source location would be huge, we opt for a compromise approach using a fully nonlinear source relocation algorithm, which exploits the grid-based nature of FMM. The availability of the complete traveltimes field for each receiver means that a fully nonlinear grid search for the best source location can be done efficiently, regardless of how complex the velocity model is (Pilia et al., 2013). The objective function minimized in the grid search is given in the Supplementary Information (Text S1).

Although the source relocation algorithm is fully nonlinear, the use of a linearised velocity inversion scheme means that an iterative approach is needed to account for the trade-off between velocity variations and hypocenter locations. The source and velocity inversions are done sequentially. Sources are first relocated using P- and S-arrival times via the nonlinear grid search method. Next, V_P and V_S are updated using the new locations, which involves two steps: 1) the solution of the forward problem using FMM; 2) an inversion for velocity parameters using the subspace inversion scheme. This is undertaken separately for P-wave and S-wave velocity structure, but both V_P and V_S models must be updated between each relocation as both P- and S-arrival times are used to constrain hypocenter location. We then repeat the entire process of source relocation and velocity inversion. In this case, an acceptable level of convergence is attained after six iterations.

Following this, we use the final hypocenter locations (as determined by the "joint" inversion for V_P , V_S and earthquake location) in the modified FMTOMO algorithm developed by Pilia et al. (2013) in order to calculate V_P/V_S . This procedure inverts S-P differential traveltimes for V_P/V_S structure along the ray paths from the S-wave model. We assume that 1) each S-wave path between two points has a corresponding P-wave path; 2) the P- and S-wave paths taken between two different points are identical and have similar Fresnel zones. Under these assumptions, the inverse problem is linear, as any lateral heterogeneity will cause a

divergence of the P- and S-ray paths (Walck, 1988; Thurber, 1993; Eberhart-Phillips and Reyners, 2012). The method requires common P- and S-arrival times, so rays with only S-phases or P-phases will be removed. In our case, this does not result in significant data loss between the calculation of the V_P and V_S models as compared to the V_P/V_S model, as the U-GPD model phase association method has already discarded rays which only have S-arrivals (Lapins et al., 2021). See the Supplementary Information (Text S2) for more detail on how the problem is formulated within a linear framework, and Pilia et al. (2013) for a full description of the direct inversion of S-P differential traveltimes.

We choose to directly invert S-P differential traveltimes rather than dividing the P-wave model by the S-wave model to obtain V_P/V_S . S-wave data coverage tends to be poorer than P-wave data coverage, and is usually noisier, due to S-wave arrivals being more difficult to pick. The imposition of relatively arbitrary regularization constraints on the amplitude of anomalies means that the resulting S-wave solution models are comparatively smoother than P-wave models. For interpretation of individual P- and S-wave models, the absolute amplitude being correct is less relevant than the overall pattern of anomalies. However, when dividing the models to obtain V_P/V_S , the amplitude of P- and S-wave velocity anomalies directly influences the V_P/V_S model. If the S-wave model is smoother, the final V_P/V_S model obtained from direct division can inherit smaller wavelength features from the P-wave model, as shown by Pilia et al. (2013).

One potential drawback of our approach is that the V_P/V_S model cannot be derived explicitly from the V_P and V_S models but due to solution non-uniqueness, we argue that our inversion produces the optimum model of each type. Any inconsistencies must be viewed in the context of model uncertainty, which is unavoidable when undertaking an inversion with noisy data that is unevenly distributed. Indeed, synthetic tests show that the assumptions inherent to this technique have less effect on the results than ad hoc regularization choices (Pilia et al., 2013).

A flow chart detailing the full tomographic workflow can be found in the Supplementary Information (Supplementary Figure S10).

4.3.2 1D Model Selection

Under the assumption of weak nonlinearity, velocity perturbations cannot move too far from the unperturbed model. Hence, the starting reference model should ideally be as close to the true solution as possible to avoid the inversion becoming trapped in a local minimum that is far from the correct model. Initial test 3D inversions using a basic three-layer velocity model from Ginzburg et al. (1981) as a starting reference model do not resolve much detail, further demonstrating the need for a reference model that more closely reflects the local velocity structure at Nabro. Therefore we use FMTOMO in a ‘quasi-1D’ inversion to develop and refine a suitable 1D velocity model to use as a reference model for subsequent 3D inversions, following the method of Wilks (2016).

Because FMTOMO requires there to be at least two velocity grid nodes in any particular direction when the grid is defined, we cannot explicitly invert for 1D velocity structure. Instead, we perform ‘quasi-1D’ inversions using a simplified velocity grid, with two nodes defined in both latitude and longitude, spanning $13.15 - 13.55^\circ\text{N}$, $41.5 - 41.9^\circ\text{E}$. 23 nodes are defined in the depth direction, and the grid spans $-3 \text{ km} - 20 \text{ km b.s.l.}$, resulting in velocities defined at $\sim 1 \text{ km}$ depth increments. We then invert for V_P and V_S and relocate hypocenters, and calculate the average velocity at each nodal depth across the four nodes. FMTOMO automatically generates a boundary layer of two additional nodes at the grid limits, so it is important to exclude the nodes that make up this padding when averaging over the nodes.

When a velocity model with a number of discrete, homogeneous layers is used to locate earthquakes, they tend to cluster at the velocity discontinuities (Hamlyn et al., 2014). Thus we smooth out the sharp discontinuities in the three-layer regional model with a Gaussian filter and use the result as a starting model. The resulting P- and S-wave ‘quasi-1D’ velocity models are plotted in Figure 4.2.

For the P-wave velocity inversion, the data variance is reduced from 0.0997 s^2 to 0.0273 s^2 . For the S-wave velocity inversion, the variance is reduced from 0.264 s^2 to 0.0469 s^2 . This suggests that this new 1D model is ‘closer’ to the solution

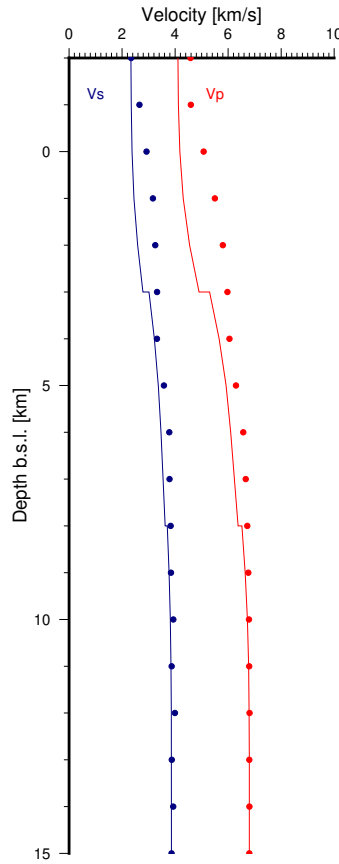


Figure 4.2: The V_P and V_S velocity model output from a quasi-1D inversion (red circles and navy circles respectively), compared with the smoothed 3-layer starting models (red and navy lines respectively).

model and represents the seismic structure at Nabro more accurately than the three-layer regional model from Ginzburg et al. (1981).

To investigate how sensitive the 3D solution model is to the initial model, we perturb each value of the new 1D model randomly by up to 10% prior to inversion and carry out 3D tomographic inversions. We find that the solution models show broadly similar structure, with differences only occurring outside of the data resolution limits determined by synthetic resolution tests described in Section 4.4 (i.e., the models differ in small-scale structure or in regions of poor data coverage). Results of these perturbation tests are plotted in Supplementary Figure S1 and described in the Supplementary Information (Text S3). Analysis of the inversion statistics shows the same or higher traveltime residuals and variances for the solution models using the perturbed 1D model compared to their starting model, as expected, with

percentage deviations that are less than 16% as shown in Supplementary Table S1. Therefore we conclude that our choice of 1D starting models is robust, and use these models as the starting model in all of the subsequent V_P and V_S 3D inversions.

4.3.3 Defining a Grid for 3D Inversions

Before undertaking 3D inversions, we first define an inversion grid, which describes the velocity model in terms of cubic B-spline functions. We also define a propagation grid, which represents a discrete sampling of the velocity field for use in the grid-based eikonal solver employed during the forward step and during the nonlinear relocation of events. Both grids are comprised of a 3D set of nodes that span $13.15^\circ\text{N} - 13.55^\circ\text{N}$, $41.5^\circ\text{E} - 41.9^\circ\text{E}$, and $-3.0\text{ km} - 20.0\text{ km}$ in depth below sea level. The node spacing for the propagation grid is chosen to be $\sim 0.5\text{ km}$ and for the inversion grid it is $\sim 1\text{ km}$. The relocation code also carries out a sub-cell search, dicing the initial cells by a factor of 10, and therefore 50 m is the smallest separation distance between the relocated earthquakes. The inversion grid spacing is sufficiently small to capture features that are constrained by the data, noting that smoothing is applied to control the wavelength of recovered features. At half the spacing of the inversion grid, the propagation grid is sufficiently fine to render errors in the forward prediction of traveltimes sufficiently small that they will not influence the inversion results. The 1D model determined in Section 4.3.2 is used as a starting model.

4.3.4 Optimising Data-Model Fit

The aim of the tomographic inversion procedure is to find a model that is similarly smooth and as close to the initial model as possible (to satisfy local linearity), whilst still satisfying the data. Therefore, three parameters need to be minimized: data fit, model variance and model roughness. The solution model variance is a measure of the difference between the starting model and the final model, and model roughness is a measure of how much complexity exists in the final model itself (based on the second spatial derivative). We optimise these using the smoothing and damping parameters, η and ϵ . These regularization parameters also affect the data fit, i.e.,

the difference between the observed data and the final solution model predictions, quantified by the variance of the traveltime residuals.

Through numerous inversions for 3D velocity structure using different values of the regularization parameters, we plot trade-off curves to find the damping and smoothing parameters that give the best compromise between data fit, model variance and model roughness. This process is done separately for the V_P , V_S and V_P/V_S models; see the Supplementary Information (Text S4) for further detail. For V_P the parameters are $\epsilon = 3$ and $\eta = 10$, for V_S they are $\epsilon = 10$ and $\eta = 50$ and for V_P/V_S , $\epsilon = 100$ and $\eta = 20$ (see Supplementary Figure S2).

4.3.5 Earthquake Relocations

After an initial inversion using the optimal damping and smoothing parameters, we examine the output relocated seismicity, plotted in Supplementary Figure S3. We find that many events have been relocated substantial distances. A considerable number of earthquakes migrate to the inversion grid boundaries and above the topography line, which indicates that the locations of these earthquakes were poorly constrained in the initial catalog produced by NonLinLoc, perhaps due to picking uncertainties from the U-GPD model. The mean relocation offset is 1.61 km. Therefore, we identify events that are relocated by distances greater than 3 km, and remove them from the catalog. We then repeat the inversion procedure with this reduced catalog of 8,893 events in order to improve the inversion stability.

The inversion results using the new subset of events show improved data fits over the full catalog. For V_P the data variance of the final solution model is reduced from 0.0218 s^2 using the full catalog to 0.0101 s^2 using the subset of events. The data variance of the final V_S solution model decreases from 0.0298 s^2 to 0.0170 s^2 . V_P/V_S also experiences a reduction in the variance from 0.0386 s^2 to 0.0283 s^2 . Furthermore, events are now relocated by reduced offsets: the mean relocation offset for the reduced catalog is 0.91 km. We note that some earthquakes remain located above the topography line after this process. We attribute this to shallow events being more poorly constrained and thus erroneously relocated in the air.

The average depth uncertainty across all events located above sea level is ± 4.99 km, calculated following Wilks et al. (2020) and described in the Supplementary Information (Text S1). The shallowest depth an earthquake is relocated to is 2.49 km above sea level. Therefore, within error, the location of these earthquakes is consistent with their true locations being in the very shallow subsurface.

We therefore use this reduced catalog in the following section to carry out resolution tests and produce the final solution model.

4.4 Results

4.4.1 Checkerboard Resolution Tests

The tomographic inversion problem is non-unique, with many different models able to satisfy the data. Different factors that help constrain the solution of this inverse problem include the path coverage of the data, data noise and the choice of implicit and explicit regularization. Thus, assessing the solution robustness is challenging, yet crucial for comprehensive evaluation of the spatial resolution of the estimated model.

Synthetic reconstruction tests are typically used to investigate the robustness of tomographic models. During these tests, a heterogeneous synthetic model is formulated and the forward problem is solved using this model, with the identical source-receiver configurations to the observational dataset. This produces a synthetic traveltime dataset. The same inversion method used for the experimental data is applied to the synthetic dataset in order to reconstruct the synthetic model. The most common input structure is a ‘checkerboard’ structure overlain on the starting model, with alternating positive and negative velocity anomalies making up the ‘checkers’ (e.g. Hearn and Clayton, 1986; Glahn et al., 1993; Rawlinson et al., 2003). Regions where the checkerboard pattern is recovered are considered to be well resolved.

The parameters we choose for our checkerboard tests are: V_P perturbations set to ± 0.5 km/s from the initial model ($\pm 7.35 - 11.7\%$) and V_S perturbations set to ± 0.2 km/s from the initial model ($\pm 5.18 - 8.69\%$). It is instructive to generate checkerboards with different scale lengths of perturbation—this amounts to altering

the number of grid nodes, N , that are perturbed simultaneously. For example, a checkerboard of size $N = 2$ will have nodes perturbed in pairs. Increasing the value of N will increase the size of each checkerboard element. In Figure 4.3, we present checkerboard tests with size $N = 2, 4, 8$, corresponding to scale lengths of ~ 2 km, ~ 4 km and ~ 8 km respectively, to assess the model resolution at different scales.

If no noise is added to the synthetic dataset, the result will give an indication of the optimal spatial resolution. However, since data noise is present in all seismic datasets, noise with a Gaussian distribution is often added, with a standard deviation equal to that of noise estimates obtained from the data. However, it is important to note that estimating data uncertainty is often subjective, and it is not clear that the actual noise distribution takes a Gaussian form (Rawlinson and Spakman, 2016). Bearing this in mind, we add Gaussian noise with a standard deviation of 0.05s, representing half a cycle of the dominant frequency of the microseisms (~ 10 Hz).

For our synthetic checkerboard test, we carry out six iterations for V_P/V_S and source location, with results plotted in Figure 4.4. The output checkerboards of scale length ~ 4 km and ~ 8 km show that V_P/V_S anomalies on these scales are well-resolved within the seismic network in map view (Figure 4.4a and b). Both checkerboards are resolvable down to 10 km depth b.s.l.. The finest scale checkerboard (~ 2 km, Figure 4.4c) is much less well-defined.

These tests indicate the limits of resolution of our dataset—they demonstrate that V_P/V_S anomalies can be robustly detected above depths of 10 km within the seismic array and on scale lengths greater than ~ 2 km. Similar resolution is observed in the results of synthetic inversions for V_P and V_S structure (see Supplementary Figures S5 and S7 and Text S5).

The lack of recovery of anomalies at depth can be explained by 99% of the seismic events in our dataset occurring between the surface and ~ 10 km depth b.s.l.; thus, there are very few paths available to resolve structure below this depth. The amplitude recovery of the input perturbations varies—in places the amplitude of anomalies is underestimated, whereas it is overestimated in certain regions (e.g.,

the negative anomaly in the longitude cross-section of Figure 4.4a). Outside the seismic array, the amplitude recovery is particularly weak, as expected.

4.4.2 Inverting for 3D Velocity Structure

4.4.2.1 Inversion Statistics

Table 4.1: Inversion Statistics for the Final V_P , V_S and V_P/V_S Inversions

RMS residuals			
Parameter	Starting model	Solution model	Reduction (%)
V_P	0.166 s	0.104 s	37.3
V_S	0.182 s	0.130 s	28.6
V_P/V_S	0.192	0.168	12.5
Variance			
Parameter	Starting model	Solution model	Reduction (%)
V_P	0.0276 s ²	0.0110 s ²	60.1
V_S	0.0332 s ²	0.0170 s ²	48.8
V_P/V_S	0.0364	0.0283	22.3
χ^2			
Parameter	Starting model	Solution model	Reduction (%)
V_P	4.59	1.39	69.7
V_S	9.60	4.66	51.5

The inversion statistics in Table 4.1 show that the RMS arrival time residuals and the data variance are reduced for the final P-wave, S-wave and V_P/V_S solution models as compared to the starting models. The normalized χ^2 value is the result of a statistical test for how well a model compares to actual observed data. In theory, it should be equal to 1 if all the data are satisfied to the level of the noise. For the V_P solution model, it is reduced to $\chi^2 \approx 1$, but for V_S it is only reduced to $\chi^2 = 4.66$. However, as detailed by Rawlinson et al. (2010), tomographic inversions usually do not have $\chi^2 = 1$ due to a) estimation of data uncertainties being difficult to quantify; b) the use of a regular and smooth model parameterisation; c) application of *ad hoc* regularization to stabilize the inversion, which suppresses some structures that are needed to satisfy the data; d) the assumptions and approximations made when solving the forward problem. Thus the range of models that can be retrieved is

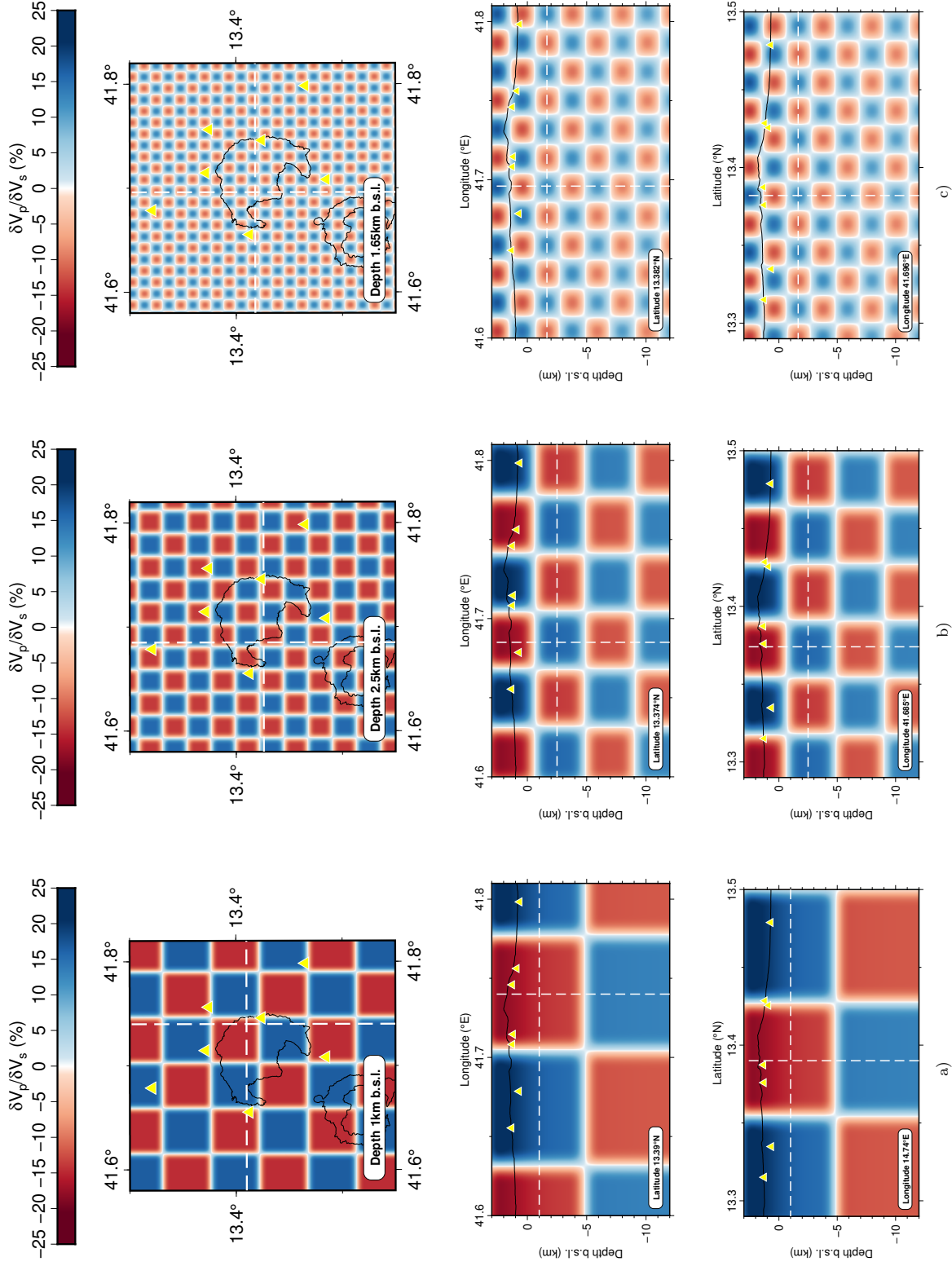


Figure 4.3: Input V_p/V_s checkerboard models of differing scale lengths: a) ~ 8 km; b) ~ 2 km. Slices are taken through the maximum perturbations of each checkerboard in latitude, longitude and depth, as indicated by the white dashed lines. Seismic stations are plotted as yellow triangles.

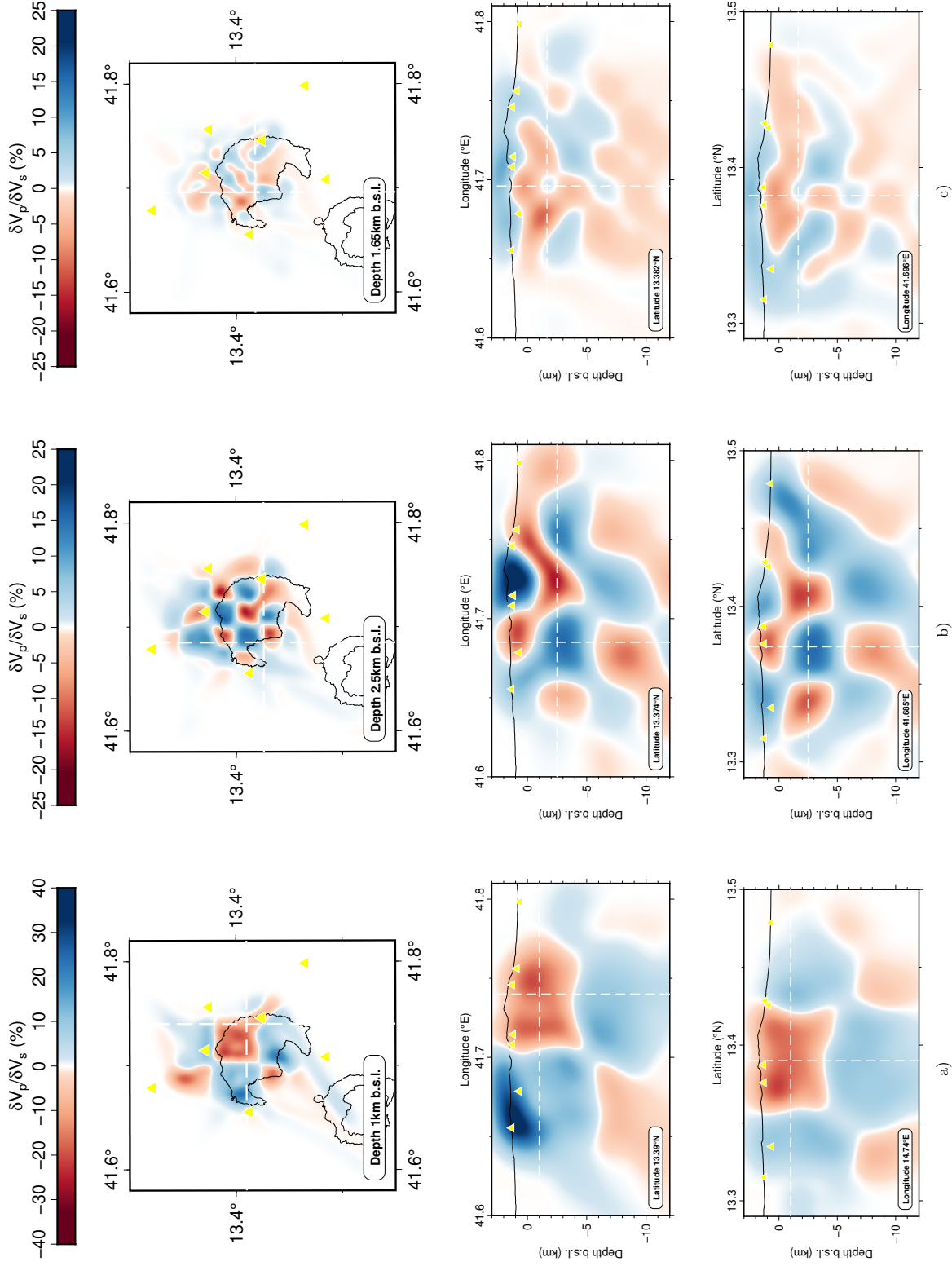


Figure 4.4: Output V_p/V_s checkerboard models of differing scale lengths: a) ~ 8 km; b) ~ 4 km; c) ~ 2 km. Slices are taken through the maximum perturbations of each checkerboard in latitude, longitude and depth, as indicated by the white dashed lines. Seismic stations are plotted as yellow triangles.

limited. Despite this, the final data fit is a significant improvement on the starting 1D models for all the solution models, indicating that recovered lateral heterogeneities are generally required by the data and hence are physically meaningful within the limits of data resolution, as estimated from the synthetic tests.

4.4.2.2 Velocity Structure

Figure 4.5 shows east-west and north-south cross-sections through the final V_P , V_S and V_P/V_S solution models. The cross-sections are taken through the center of the caldera, passing directly through the vent location of the 2011 eruption (Hamlyn et al., 2014). For clarity, we only plot the anomalies and seismicity below the topography line. The original figures can be seen in the Supplementary Information (Supplementary Figure S8). Depth slices are plotted in Figure 4.6.

The dominant feature in the V_S solution model (Figure 4.5c–d, Figure 4.6e–h) is a region of high V_S and high levels of seismicity within the caldera outline, extending downwards from the surface to 6 km b.s.l.. At 6 km b.s.l., an aseismic region of low V_S is observed, extending downwards to ~ 10 km b.s.l.. Close to the surface, the high V_S region is surrounded to the north and south by very low V_S areas, and to the east and west by less pronounced low V_S areas, all of which are aseismic.

The V_P model shows more heterogeneity compared to the V_S model. The difference is particularly noticeable in the perpendicular east-west and north-south cross-sections taken through the center of the caldera (Figure 4.5a–b). A low V_P structure extending from $41.7 - 41.8^\circ\text{E}$ dips from east to west from a depth of 4 km b.s.l. to 10 km b.s.l.. Above this, there is a region of high V_P . This region contains two low V_P anomalies extending $\sim 2-3$ km in depth and $\sim 0.2^\circ$ in longitude, which is around the limit of data resolution. In the depth slice at 1 km b.s.l., a low V_P region extends from $13.44 - 13.48^\circ\text{N}$, aligned N-S (Figure 4.6a). This region extends in depth down to 10 km b.s.l., as observed in the longitude cross-section (Figure 4.5a), and is seismically active.

Following the joint inversion for V_P and V_S structure, we calculate the average ratio of the V_P and V_S models across all velocity grid nodes, which is 1.77. We

use this reference value to adjust our colour scale in the V_P/V_S plots in Figures 4.5 and 4.6; red colours correspond to ratios higher than the reference and blue colours to ratios lower than the reference.

At depths of 6–10 km b.s.l., a region of high V_P/V_S ratio (as high as 1.9) is observed in the longitude and latitude cross-sections (Figure 4.5e–f). This region correlates with low V_S and low V_P anomalies, and is aseismic. Above this high V_P/V_S region, there is an area of high seismicity and very low V_P/V_S ratio (as low as 1.5), extending from depth 0 km b.s.l. down to 6 km b.s.l. and lying within the caldera outline (Figure 4.6i–l). The strongest low V_P/V_S ratios are seen between 0–2 km b.s.l. and correspond to high V_S values. Close to the surface, high V_P/V_S ratios are observed again, with the ratio reaching 2.0 in places. These high V_P/V_S regions all exhibit low levels of seismicity.

As described in Section 4.3, we obtain the V_P/V_S solution model by directly inverting S-P differential traveltimes. We also plot the solution model obtained from simply dividing the V_P solution model by the V_S solution model (Supplementary Figure S9). The direct division model shows largely the same pattern of velocity anomalies within the data resolution limits determined by our synthetic tests (Section 4.4). Differences in amplitude are seen, but this is expected due to solution non-uniqueness and the imposition of regularization constraints. Thus, we are confident that our method produces a model that is consistent with direct division, and that our subsequent interpretations of these V_P/V_S ratio anomalies would remain the same if we had chosen to directly divide the V_P solution model by the V_S solution model instead.

4.5 Discussion

4.5.1 Defining a Local 1D Velocity Model at Nabro

Compared to the three-layer velocity model developed for the Afar region in previous studies, our refined local 1D velocity model has faster P- and S-wave velocities in the uppermost 10 km. High crustal velocities in a volcanic setting are typically attributed to the presence of solidified, high-strength intrusive magmatic rocks,

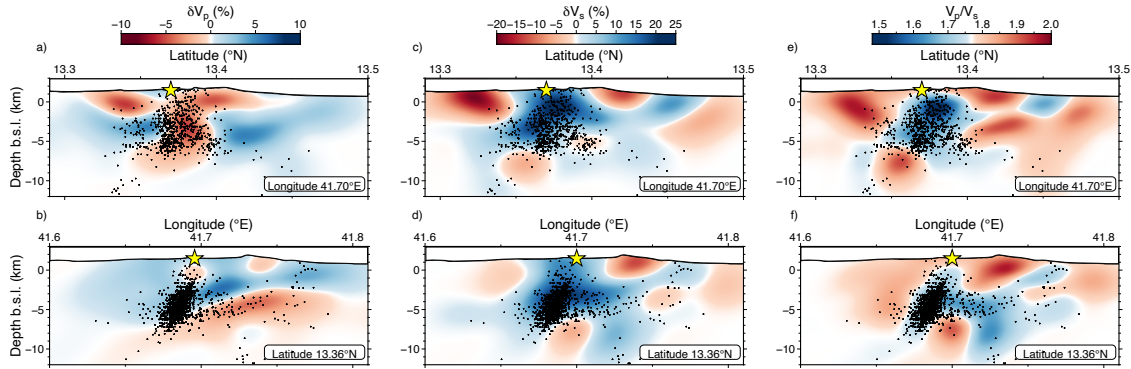


Figure 4.5: Cross-sections through the final V_P (a, d), V_S (b, e) and V_P/V_S (c, f) solution models at a longitude of 41.7°E and latitude of 13.36°N (the center of the caldera). The V_P and V_S models are plotted as percentage deviations from the initial 1D model. The V_P/V_S ratio model is plotted as absolute values, with the center of the colour bar corresponding to the reference V_P/V_S value—the average ratio of the V_P and V_S models across all velocity grid nodes. Earthquakes within ± 1 km of the displayed section are indicated by black dots. The yellow stars mark the vent location of the 2011 eruption. As discussed in the text, we only show the earthquakes and anomalies below the topography line for the purposes of clarity, with the full solution plotted in Supplementary Figure S8.

such as the cumulates and dykes at Mount Etna (Aloisi et al., 2002), a plutonic body at Mount St. Helens (Lees, 1992), an old lateral dyke system at Tungurahua volcano (Molina et al., 2005) and the solid andesitic cores of the volcanic complexes of Soufrière and Centre Hills, Montserrat (Paulatto et al., 2010; Shalev et al., 2010). At Nabro, analysis of inclusions in erupted products from 2011 suggests that these are derived from older and more primitive basalt (Donovan et al., 2018). The presence of xenocryst material in the erupted magmas leads Donovan et al. (2018) to conclude that the subsurface crustal structure beneath the caldera is composed of a series of sills and older eruptive products. This provides supporting evidence that the elevated crustal velocities directly beneath Nabro reflect intrusions, potentially remnants of earlier episodes of magmatism.

Below 10 km, the refined model shows negligible variation from the regional model, as expected due to the fact that the vast majority of seismic events originate above 10 km b.s.l.. Using this refined local 1D model as the starting model for tomographic inversions results in solution models that better fit the traveltime data.

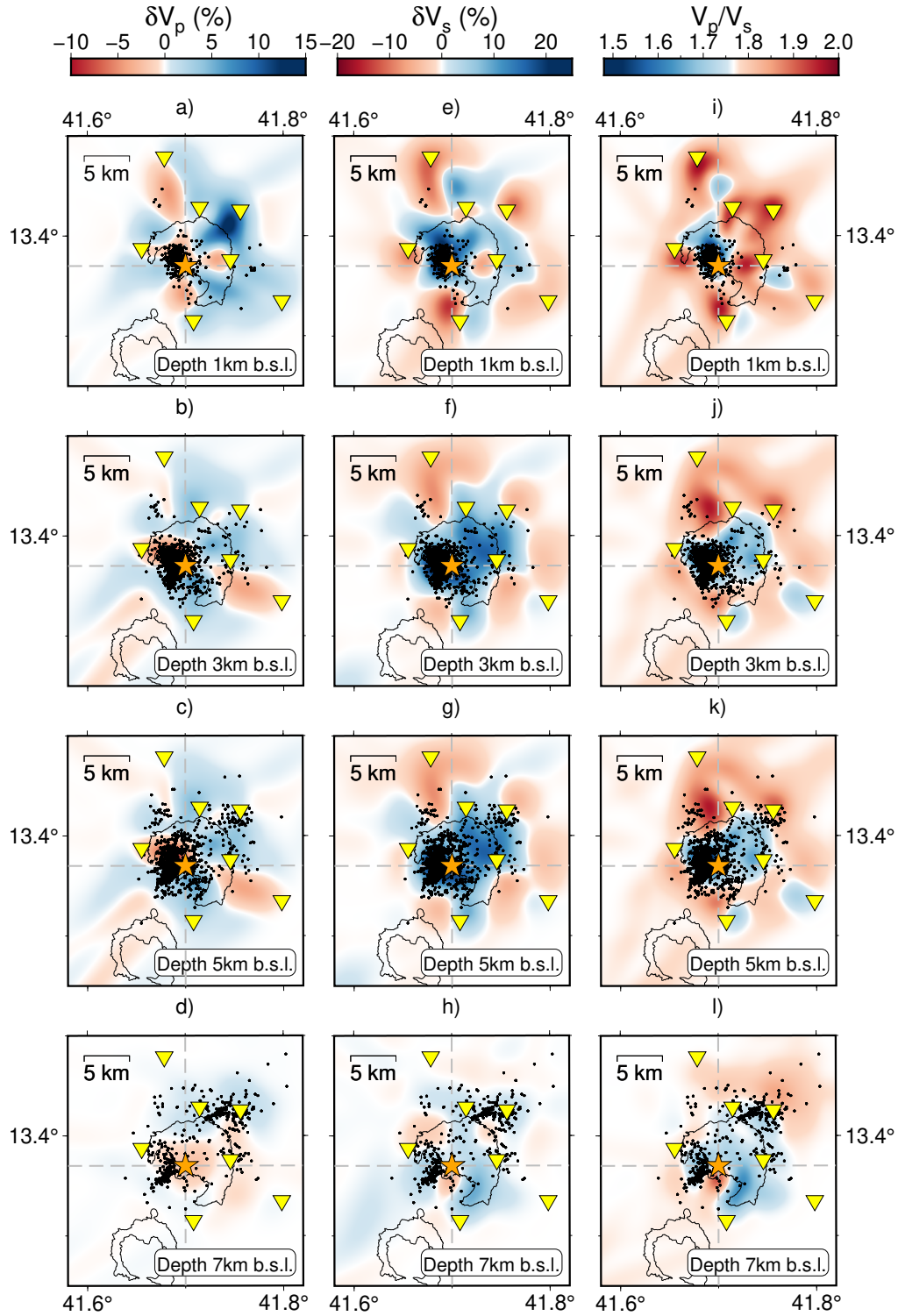


Figure 4.6: Cross sections at depths of 1 km, 3 km, 5 km and 7 km below sea level (b.s.l.) through the final (a–d) V_P , (e–h) V_S , and (i–l) V_P/V_S ratio velocity models. The V_P and V_S models are plotted as percentage deviations from the initial 1D model. The V_P/V_S ratio model is plotted as absolute values, with the center of the colour bar corresponding to the reference V_P/V_S value—the average ratio of the V_P and V_S models across all velocity grid nodes. Earthquakes within ± 0.5 km of the displayed section are plotted as black dots and seismic stations are plotted as yellow inverted triangles. The orange stars mark the vent location of the 2011 eruption. The grey dashed lines represent the latitude and longitude cross-sections depicted in Figure 4.5.

4.5.2 Earthquake Detection Using U-GPD

This study presents the first seismic tomography results from an earthquake catalog detected using machine learning methods. The inversion statistics from this study (RMS residuals, variance and χ^2) are of the same order of magnitude as those from previous FMTOMO studies that relied on manually picked catalogs of earthquakes (e.g. Wilks et al., 2020; Pilia et al., 2013), demonstrating that our use of the deep learning model U-GPD to pick the seismic arrivals has not adversely affected the stability and robustness of the inversion. The method therefore has the ability to detect seismic events with sufficient accuracy to be used successfully in an FMTOMO tomographic inversion. This has implications for future tomographic studies at volcanoes: the efficiency of U-GPD’s phase arrival picking method means that far more events can be detected than previously possible, enabling more timely exploitation of such data for the purposes of seismic tomography.

4.5.3 Interpretation of V_P/V_S Variations

Seismic velocity variations reflect a variety of physical parameters, including rock characteristics (composition, porosity, fractures, mineralogy), saturation conditions, presence of fluids (gases or liquids), temperature, and pressure. The interplay of these diverse influences makes it difficult to interpret observed seismic anomalies. Considering the ratio of compressional velocity to shear velocity, V_P/V_S , enables greater constraint to be placed on the cause of seismic velocity variations. This is particularly helpful when attempting to constrain the location of fluids in the subsurface of a volcano, because the V_P/V_S ratio is sensitive to the type of fluid present and can distinguish between regions of partial melt or hydrothermal fluids, both of which are encountered beneath volcanoes. In saturated or partially-saturated rocks, the content and physical state of fluids has a greater effect on P-wave velocities than S-wave velocities (Vanorio et al., 2005). Fluid phase transitions induce changes in fluid compressibility and thus bulk modulus (Ito et al., 1979; Wang and Nur, 1986). Shear moduli are little affected by fluid phase transitions and hence S-wave velocities change insignificantly due to a density effect, meaning that low

V_P/V_S ratios tend to characterize gas-bearing rocks (i.e., those with high fluid compressibility) whereas liquid-bearing rocks (with low fluid compressibility) are characterized by high V_P/V_S ratios (Vanorio et al., 2005).

4.5.3.1 High V_P/V_S Anomaly at 6–10 km Depth

In Figure 4.5, the high V_P/V_S (> 1.9) region at 6–10 km b.s.l. coincides with pronounced low V_S ($< 3.8 \text{ km s}^{-1}$) and low V_P ($\sim 6.6 \text{ km s}^{-1}$) anomalies as compared to the starting model. This correlation between the P- and S-wave velocity models in a region of high V_P/V_S ratio suggests a region of elevated temperature (Sanders et al., 1995). Furthermore, the anomalous region is approximately aseismic. In an examination of a deep cluster of seismicity at Nabro, Lapins (2021a) finds that the highest attenuation is observed at station NAB1 where the raypaths travel directly through this aseismic, high V_P/V_S anomaly, suggesting that the region attenuates S-waves strongly. Anomalous low Q_S at depth is usually attributed to the presence of partial melt (Sanders et al., 1995). The region also coincides with the location of a Mogi source inferred by Hamlyn et al. (2014) to explain the observed post-eruptive surface subsidence at Nabro. Petrological analysis by Donovan et al. (2018) finds that most melt inclusions in erupted products from the 2011 eruption were entrapped at 5–10 km depth b.s.l.. This represents the storage location of an older body of melt which was remobilized and erupted when an intrusion of fresh melt rose through the crust and mingled with the older melt (Donovan et al., 2018). The estimated depth of this melt body is consistent with the depth of our observed high V_P/V_S anomaly.

Previous tomographic studies at volcanoes have interpreted similar regions with high V_P/V_S ratios as delineating magmatic storage zones. For example, an anomalous body with low P-wave velocity, low S-wave velocity and V_P/V_S ratio > 1.84 is observed by Lin et al. (2014) at 8–11 km depth beneath the Kīlauea volcano in Hawaii, and interpreted as a crustal magma reservoir beneath the volcanic pile. At Nevado del Ruiz volcano in Colombia, a region of high V_P/V_S ratios (> 1.80) at 2–10 km is inferred to be an intrusive body of magmatic origin that included

partial melt zones associated with low S-wave velocity anomalies (Londoño and Sudo, 2003). Greenfield et al. (2016) observe a pair of prominent anomalies with low P- and S-wave velocities and V_P/V_S ratios > 1.82 at depths of 5 km and 9 km b.s.l. beneath Askja volcano, Iceland, which are interpreted as the primary magma storage regions in the upper crust.

Thus, we interpret the high V_P/V_S anomaly as the storage location of melt that fed the 2011 eruption. Only a small fraction (5–20%) of the stored melt is typically erupted at the surface (Greenfield and White, 2015; White et al., 2019). This is seen in the results of a post-eruption seismic velocity study at Mount St. Helens, which finds that there is a persistent high V_P/V_S region at 4–13 km b.s.l., interpreted as the primary upper-middle crustal magma reservoir and indicating that a significant amount of melt remains in the crust (Kiser et al., 2016). Similarly, our observation of a high V_P/V_S region at depths 6–10 km b.s.l. suggests that melt is still stored at these depths, which could feed future eruptions.

4.5.3.2 Low V_P/V_S Region

Above the inferred magma storage region, our results show a region of low V_P/V_S ratio (1.5–1.7), colocated with high S-wave velocities ($> 3.6 \text{ km s}^{-1}$) and low P-wave velocities compared to the starting model (region B in Figure 4.7). This anomaly extends from depths of ~ 5 km b.s.l. to the surface. The high V_S values can be explained by lithology—high-strength, solidified intrusive magmatic rocks are expected to show high seismic velocities (e.g. Lees, 1992; Aloisi et al., 2002; Molina et al., 2005; Lees, 2007). However, the V_P/V_S ratio in intrusive igneous rocks is expected to be higher than typical continental crust (Christensen, 1996), whereas the V_P/V_S ratio we measure in this region is low—down to 1.5 in places—and so there must be another factor acting to reduce the V_P/V_S ratio. It has been shown that the velocities of P- and S-waves in rocks are strongly affected by the saturation conditions of the rock, particularly whether the rock is saturated with gas, liquid or a mixture thereof (Toksöz et al., 1976; Ito et al., 1979). A geothermal regime near the water-steam transition has low P-wave velocities but

normal S-wave velocities: the presence of gas reduces the bulk modulus and causes a decrease in P-wave velocity, without significantly altering the propagation of shear waves (Walck, 1988). The addition of a small amount of gas in a water-brine mixture can lower the velocity of P-waves significantly (Toksöz et al., 1976). We observe low P-wave velocities coincident with the lowest V_P/V_S ratio in our model, providing evidence for the presence of gas in the rocks in this region. Furthermore, calculations of seismic attenuation in P- and S-waves at Nabro show that P-wave attenuation is significantly higher than S-wave attenuation across all seismic stations (Lapins, 2021a), which is generally attributed to partial saturation of a compressible fluid in cracks, fractures or pores (Winkler and Nur, 1979; Hauksson and Shearer, 2006; Amalokwu et al., 2014). Further evidence in support of the existence of gases in the upper subsurface is the significant post-eruption fumarolic activity observed at Nabro (Yohannes, 2012). Petrological analysis of melt inclusions erupted in 2011 indicates that melt-fluid separation occurred at depths of up to 18 km b.s.l., generating CO_2 rich fluids (Donovan et al., 2018). The magma storage zone described in Section 4.5.3.1 coincides with a Mogi source; the deformation model invokes deflation, which is explained by the outgassing of magma at depth (Hamlyn et al., 2018). The low V_P/V_S region could reflect the degassing pathways between the magma and the surface fumaroles.

A velocity reversal is seen in the 1D V_S model between 3–4 km b.s.l., where velocity values decrease with depth (Figure 4.2). The trend in V_P does not reverse here, but the rate of change of velocity with depth decreases. The location of these reversals is coincident with the lowest V_P/V_S ratio seen in the model. Density, resistivity and sonic velocity logs that go through velocity reversals are generally interpreted as departures of the effective stress from normal compaction trends (Hottmann and Johnson, 1965; Pikington, 1988; Bowers, 2002). Overpressure is one explanation for this. If the fluid pressure is higher than the normal hydrostatic fluid gradient for a given depth, it prevents the effective stress from increasing with depth as it usually would (Vanorio et al., 2005). To be over-pressurized, the gas present in the low V_P/V_S region would have to be trapped and experiencing expansion, uplift,

compaction, temperature increase or a combination of these factors; all of which are possible in the context of a recently active volcano. Indeed, the overpressure could be driven by the magma storage zone directly below the low V_P/V_S region that extends from depths of ~ 5 km b.s.l. to the surface (Hamlyn et al., 2018).

Nabro’s caldera outline matches the region of high V_S and low V_P/V_S well (Figure 4.6), while the low V_P region is slightly smaller. Therefore, we propose that the crustal structure within the caldera is formed of intrusive rock, potentially with layers of stacked sills from previous eruptions. This would elevate the S-wave velocity.

During the June 2011 eruption of Nabro, magma ascended to the surface via a NW-SE-oriented dyke (Goitom et al., 2015). Volcanic conduits have associated damage zones, common to all shallow magmatic systems beneath volcanoes (Afanasyev et al., 2018), and contain fragmental infills related to prior intrusions, eruptions and steam explosions (Blundy et al., 2021). These conduits therefore have high porosities, within which fluids can be stored. Dyke emplacement also causes fracturing and creates permeable pathways for fluid transport and accumulation (e.g. Bakker et al., 2016; Brown et al., 2007). Our results show that this region is highly seismogenic, pointing to the presence of fractures and cracks enabling fluid migration that drives pore pressure increases and leads to abundant seismicity. Thus, the most likely route from the degassing magma storage region to the surface is along the conduit that fed the 2011 eruption, as it will be formed from highly permeable and damaged rock. This explains why the low V_P region (indicating the presence of gas) is less horizontally extensive than the high V_S region.

Similar conclusions are reached at Aluto volcano, where a region of low V_P/V_S ratio is interpreted as the signature of an over-pressurized gas volume within a hydrothermal system (Wilks et al., 2017, 2020). A seismogenic zone of low V_P/V_S ratios coincident with low P-wave velocities is also observed at Campi Flegrei, and explained as over-pressurized gases accumulating at the top of dyke intrusions (Vanorio et al., 2005; Chiarabba and Moretti, 2006). At Nevado del Ruiz volcano, the upper part of a high P- and S-wave velocity anomaly (0–2 km depth) is characterized by low V_P/V_S ratios (< 1.68) and described as a steam-dominated

geothermal system by Londoño and Sudo (2003). Multiple studies of Mammoth Mountain, California, have identified a low V_P/V_S region from depths of -3–2 km, attributed to the presence of CO_2 distributed in oblate spheroid pores, which supplies gas-rich thermal springs at the surface (Julian et al., 1998; Foulger et al., 2003; Lin, 2013; Dawson et al., 2016).

Our observations have shown that the degassing of partial melt influences the upper crustal substructure beneath Nabro. The coupling of the shallow heat source to volatile transport above, as well as the presence of fumaroles at the surface, suggests the possibility of a high-temperature geothermal system similar to those hosted by volcanic s in the Main Ethiopian Rift (Pürschel et al., 2013).

4.5.3.3 Shallow, High V_P/V_S Anomalies

In volcanic settings, high V_P/V_S values at depth are typically associated with the presence of melt. However, the high temperatures and pressures necessary for sustaining partial melt post-eruption are unlikely to prevail very close to the surface (depths 0–5 km below surface), and so other explanations have been proposed: for example, steam condensates that manifest at shallow depths off the main volcanic edifice, where temperatures are reduced (Aster and Meyer, 1988; Vanorio et al., 2005; Chiarabba and Moretti, 2006; Wilks et al., 2020). These condensates may form brines that migrate towards the surface along fracture networks, explaining extensive fumarolic activity at the surface (Hudson et al., 2022; MacQueen et al., 2021). Alternatively, high V_P/V_S anomalies have been attributed to the penetration of meteoric water into the volcanic cone through fractures (Bushenkova et al., 2019; Koulakov et al., 2021).

The shallowest part of our model shows high V_P/V_S anomalies (> 1.9) to the north and south of the caldera that coincide with low V_P and low V_S anomalies. These anomalies are in the very shallow subsurface, meaning that they are constrained by only one seismic station. Therefore, in the absence of further geophysical constraints, such as magneto-telluric surveys, fluid sampling and analysis or well-log data, it is possible that these anomalies are artefacts.

4.5.4 The Crustal Substructure Beneath Nabro Caldera

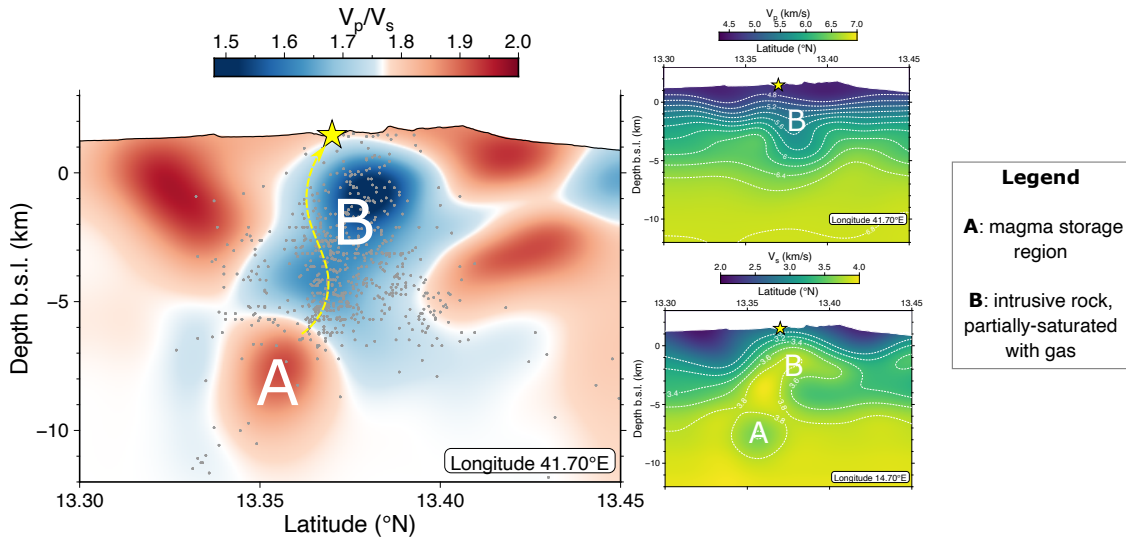


Figure 4.7: The V_P/V_S (left image), V_P (top right image) and V_S (bottom right image) solution models plotted along the 41.7°E cross-section beneath Nabro. To aid visualisation, contours are added to the V_P and V_S models for velocities greater than 4.8 km/s and 3.0 km/s respectively. A) The aseismic high V_P/V_S , low V_P and low V_S magma storage region is described in Section 4.5.3.1. B) The very low V_P/V_S and high V_S region is described in Section 4.5.3.2; we interpret this as a zone of intrusive rock, potentially in a stacked sill structure. The region hosts fractured, partially-saturated rocks with low V_P values. The yellow star marks the vent location of the 2011 eruption. Earthquakes located within ± 1 km of the cross-section are plotted as grey circles. The yellow dashed arrow indicates a potential degassing pathway from the region of partial melt to the surface, following the densest clusters of seismicity.

Nabro is an active volcano that experienced subsidence and seismicity following its 2011 eruption. Figure 4.7 provides an overview of the results and interpretations of the seismic anomalies identified in this study, as summarized below.

An aseismic region of low V_P , low V_S and high V_P/V_S ratio at depths of 6–10 km (region A in Figure 7) likely represents a storage region of partially molten material, and is consistent with the results of previous geodetic, petrological and seismic studies. It is likely that partial melt has remained stored here post-eruption, as only a small fraction of the total volume of melt stored in a reservoir is generally erupted at the surface (Greenfield and White, 2015; White et al., 2019). The cause of surface subsidence is likely degassing of volatiles from this magma storage region (Hamlyn et al., 2014; Donovan et al., 2018).

Above the zone of partial melt, we observe a region of abundant seismicity and high V_S (region B in Figure 7), which we interpret as a zone of intrusive rocks from previous eruptions. These could exist in a stacked sill structure, the fine details of which we are unable to resolve with our tomographic model. This region contains the conduit that fed the 2011 eruption, which is formed of fractured, cracked rocks partially-saturated with over-pressurized gases (a mixture of CO_2 and H_2O), leading to low V_P and very low V_P/V_S ratio. The magma storage region below is a likely cause of the overpressure in these gases. The influence of this shallow heat source on the outflow of gases through the subsurface means that Nabro is likely to have geothermal potential that may be exploited as an energy resource.

Previous studies are in agreement with our interpretations of the velocity structure at Nabro. By inverting the deformation field from satellite InSAR images, Hamlyn et al. (2018) explain the subsidence as the deflation of a Mogi source located at 6.4 ± 0.3 km depth b.s.l.. This coincides with Donovan et al. (2018)’s findings that most melt inclusions in erupted lava were trapped at 5–10 km depth b.s.l.. Their petrological study of erupted products from the 2011 eruption concludes that distinct batches of magma were stored in sills and mixed together prior to eruption (Donovan et al., 2018). A study of the post-eruption seismicity also identifies an aseismic magma storage zone at depths of 6–9 km b.s.l. (Lapins, 2021a). Numerous fumaroles have been observed at Nabro after the eruption, which has led to its identification as a region of geothermal interest (Yohannes, 2012).

4.5.5 Comparisons with Other Volcanoes

As an off-rift caldera in the under-studied East African Rift System that erupted months prior to the seismic deployment, there are no previous seismic tomography studies that allow for direct comparison to Nabro. However, it is still instructive to examine a few examples of volcanoes that share certain similar features with it.

Koryaksky volcano in Kamchatka erupted months before the seismic events used in the tomographic study of Bushenkova et al. (2019) were recorded. Despite the different tectonic setting, the tomographic images show a similar structure to Nabro.

At depth, a high V_P/V_S anomaly represents a magma storage region. Above this, there is a low V_P/V_S anomaly associated with a vertical seismicity cluster, marking the pathway of fluid ascent. Another actively erupting volcano, Kīlauea in Hawaii, shows elevated V_P and V_S at depth, interpreted as representing the high-velocity cumulates of the volcanic core (Lin et al., 2014). An anomalous body of low V_P , low V_S and high V_P/V_S at 8–11 km depth is explained as a crustal magma reservoir. Both of these features are also observed at Nabro. The Kīlauea images also show a region of low V_P/V_S above the magma reservoir, but this is not interpreted.

Aluto volcano is also located in the East African Rift System, situated in the Main Ethiopian Rift. The seismic tomography study of Wilks et al. (2020) finds a large low velocity, high V_P/V_S zone at depths of 4–9 km, interpreted as a more ductile and melt-bearing region. Away from the volcano, there are shallow, localized high V_P/V_S regions, representing steam condensates which may form brines that migrate to the surface. A hydrothermal system with very low V_P/V_S is observed at shallow depths, hosting gases exsolved from the deeper melt body. These features are broadly similar to what is seen beneath Nabro. The main difference is that Nabro’s low V_P/V_S region extends to greater depths. Despite a recent increase in surface deformation, Aluto has been quiescent for thousands of years. Therefore, the crustal substructure of Aluto could represent a ‘steady-state’ situation for volcanoes in the region, from which Nabro has been disturbed due to its recent eruption and the ascent of magmatic fluid from depth.

4.5.6 Limitations and Future Work

A fundamental limitation on our results is the resolution of the tomographic inversion, likely caused by the small size of the seismic network deployed at Nabro. Jointly inverting for velocity structure and event relocation with such a small seismic array is challenging. This is reflected in the large relocation offsets observed in particular for events outside of the aperture of the seismic network, which subsequently are removed from the catalog. Indeed, checkerboard sensitivity tests demonstrate that outside of the seismic array and deeper than ~ 10 km b.s.l., we cannot recover

synthetic velocity perturbations. Recovery of velocity anomalies on scales of < 2 km is also limited. Therefore, we restrict our interpretations of the tomographic images to velocity anomalies occurring on scales > 4 km, within the seismic network and in the uppermost 10 km of the crust.

We are also limited in our interpretations by the lack of other observations at Nabro. Studies such as magneto-telluric surveys, geochemical analyses of volcanic fluids and well-log data would all help to provide further constraints on the interpretations presented here. The resolution of seismic velocity structure is poor at depths greater than 10 km b.s.l., due to the distribution of seismicity being mostly located at shallower depths. Thus it is difficult to form broader conclusions about magmatic processes in the mid-lower crust.

Future work could involve attempts to probe the lower crust at Nabro, e.g., through receiver function analysis (e.g., Janiszewski et al., 2020; Hammond, 2014) or an investigation of seismic anisotropy using shear-wave splitting (e.g., Nowacki et al., 2018), in order to understand how off-rift magmatism at Nabro is sustained and supplied. The application of U-GPD to seismic datasets from other volcanoes, particularly those in a post-eruptive state, would also provide useful points of comparison to this study.

4.6 Conclusion

We use a seismic catalog created by a deep learning model for automating phase arrival detection to invert for the earthquake locations and the 3D velocity structure beneath Nabro caldera, an off-rift volcano in the Afar region. This has produced the first tomographic images of the volcano, which was unmonitored before its explosive eruption in June 2011.

The main findings of the tomographic study are: 1) an aseismic region of low V_P , low V_S and high V_P/V_S at depths of 6–10 km b.s.l., interpreted as the primary melt storage region that fed the 2011 eruption; 2) a region of high seismicity, very low V_P/V_S ratio and low V_P , representing a zone of partially-saturated rocks containing gases that are over-pressurized due to degassing from the magma storage zone

directly below; 3) general high V_P and V_S beneath the volcanic edifice, pointing to the existence of high-strength, solidified intrusive magmatic rocks.

Our results have demonstrated that deep learning models are an efficient way to obtain earthquake catalogs for the purposes of seismic tomography at volcanoes. Although our model cannot elucidate the origins of magma supply to Nabro at depths exceeding 10 km b.s.l., it does illustrate that this off-rift volcano has a similar shallow magmatic plumbing system to other hydrothermally active, restless volcanoes. The observations are consistent with the existence of a melt storage region at 6–10 km b.s.l. beneath Nabro. We have also uncovered a region that is high in volatile content, coupled to the degassing magmatic system, indicating that Nabro should be considered a region of geothermal interest. Our study highlights the need for further geophysical studies at Nabro.

Open Research Section

The raw seismic data used in this study are from the Nabro Urgency Array (Hammond et al., 2011b), publicly available through IRIS Data Services (<http://service.iris.edu/fdsnws/dataselect/1/>). Full code to reproduce the U-GPD transfer learning model, perform model training, run the U-GPD model over continuous sections of data and use model picks to locate events in NonLinLoc (Lomax et al., 2000) are available at <https://github.com/sachalapins/U-GPD>, with the release (v1.0.0) associated with this study archived and available through Zenodo (Lapins, 2021b). The arrival time picks for the initial event catalog produced by the U-GPD model, as well as the station metadata, are also archived in a Zenodo repository (Lapins, 2022). The FMTOMO package is freely available to download at <http://rses.anu.edu.au/~nick/fmtomo.html> (Rawlinson and Sambridge, 2004a). Files containing the final V_P , V_S and V_P/V_S models, and the relocated event catalog are available through Zenodo (Gauntlett et al., 2023).

Figures and maps were plotted using Generic Mapping Tools (GMT) version 6 (Wessel et al., 2019) licensed under LGPL version 3 or later, available at <https://www.genericmapping-tools.org>.

Acknowledgements

The authors thank Michael Bostock and Greg Waite for their helpful editorial comments. We are also grateful for the comments from Ivan Koulakov and an anonymous reviewer, which helped improve this paper. The seismic data were collected with funding from the Natural Environment Research Council (NERC) project NE/J012297/1 (“Mechanisms and implications of the 2011 eruption of Nabro volcano, Eritrea”). The UK seismic instruments and data management facilities were provided under loan number 976 by SEIS-UK at the University of Leicester. The facilities of SEIS-UK are supported by NERC under Agreement R8/H10/64. Author MG was supported by a Doctoral Training Partnership studentship from NERC [NE/S007474/1]. Author SL was supported by a GW4+ Doctoral Training Partnership studentship from the Natural Environment Research Council (NERC) [NE/L002434/1]. Author BG was funded by the Engineering and Physical Sciences Research Council (EPSRC) and the School of Earth Sciences at the University of Bristol. We gratefully acknowledge the cooperation we received from the Eritrea Institute of Technology, Eritrean government, Southern and Northern Red Sea Administrations, local sub-zones and village administrations. We thank the Department of Mines, Ministry of Energy and Mines for their continued support throughout the project. Special thanks go to Zerai Berhe, Mebrahtu Fisseha, Michael Eyob, Ahmed Mohammed, Kibrom Nerayo, Asresehey Ogbatsien, Andemichael Solomon and Isaac Tuum. We thank Alem Kibreab for vital help in facilitating the fieldwork. We would also like to acknowledge helpful comments from Tarje Nissen-Meyer and Lara Wagner on the initial results of this study. The tomographic inversions were carried out using the University of Oxford Advanced Research Computing service; we thank Andrew Walker for his assistance with this. IRIS Data Services are funded through the Seismological Facilities for the Advancement of Geoscience (SAGE) Award of the National Science Foundation under Cooperative Support Agreement EAR-1851048.

5

Conclusion

Contents

5.1	Summary	126
5.1.1	Dynamic mantle flow constrained by seismic anisotropy	126
5.1.2	Heterogeneous crustal structure in northern Afar and dynamic support from the mantle	127
5.1.3	Magma storage and fluid transport through an intrusive complex beneath Nabro volcano	128
5.2	Future work	129
5.2.1	Further investigation of Nabro volcano and the Nabro Volcanic Range	129
5.2.2	Other seismic tomography studies in the AARS	131
5.2.3	Improving understanding of the crustal structure beneath northern Afar	132
5.2.4	Regional mantle dynamics	133
5.3	Final remarks	134

In this thesis, I have used different seismological observations to study the crust and upper mantle beneath the Afro-Arabian Rift System (AARS). I find that the crustal structure throughout the region is highly complex and heterogeneous. Despite this, large scale mantle dynamics underlying the AARS are remarkably simple. Extensional processes in the region have facilitated magmatism throughout the crust and upper mantle, producing volcanic episodes such as Nabro volcano's June 2011 eruption.

5.1 Summary

5.1.1 Dynamic mantle flow constrained by seismic anisotropy

In Chapter 2, I present the first evidence from seismic anisotropy that there is a contiguous conveyor belt of density-driven mantle flow extending from the Main Ethiopian Rift through to Afar and departing around the Arabian Platform. The fast polarisation direction (ϕ) of SKS shear-wave splitting measurements is evidence of the alignment of anisotropic minerals in the mantle and thus shows the directionality of dynamic mantle processes. I show that ϕ is aligned NE-SW in northeastern Afar, coherent with observations further south in the Main Ethiopian Rift.

My measurements in southwestern Yemen show a clear bimodality in fast shear-wave polarisation direction. Stations close to the Gulf of Aden have ϕ aligned with the rift trend, which rotates through southwestern Yemen to align with the Red Sea rift trend for stations on the Red Sea margin. Further east, stations on Socotra Island and close to Oman show less coherence in ϕ and weaker splitting magnitudes.

I propose that density-driven flow in the mantle passes through Afar to meet the thick, cold lithosphere of the Arabian Platform, where it bifurcates and flows along the thin lithospheric channels of the Red Sea and Gulf of Aden Rifts. This bifurcation exerts a force on the Arabian plate, driving its motion and convergence with Eurasia. Although the influence of mantle upwelling on the regional kinematics of Arabia has been suggested previously by geodynamic models and integrated geophysical observations, until now it has not been demonstrated with seismic anisotropy measurements.

Comparison with recent regional velocity tomography models reifies these suggestions. The Arabian Platform is underlain by thick, seismically fast lithosphere with a high elastic thickness, and thus could act as a backstop to the density-driven flow. By contrast, low shear velocities beneath Western Arabia and the Red Sea and the Gulf of Aden align well with the fast direction of my measurements, implying that hot mantle is flowing beneath these regions. The tomography model also shows that the Gulf of Aden low shear velocity anomaly is shallower than the Red

Sea velocity anomaly, and has a smaller lateral extent. This supports my results close to Oman and Socotra, where fossilised fabric in the lithosphere dominates the anisotropic signal. It is potentially an indication that the Gulf of Aden rift branch may fail in favour of the Red Sea and Main Ethiopian Rift branches.

5.1.2 Heterogeneous crustal structure in northern Afar and dynamic support from the mantle

Chapter 3 presents the results of my teleseismic receiver function analysis across northern Afar and the Danakil microplate. I use data from a temporary seismic network deployed from June 2011 until October 2012, the Eritrea Seismic Project, to produce estimates of bulk crustal properties for the first time in northern Afar and the Danakil microplate. Stations further south in Afar from the AFAR0911 network are also used, in order to validate the method by comparing my results to previous measurements at these stations. I calculate teleseismic receiver functions and estimate the Moho depth (H) and bulk crustal V_P/V_S ratio (κ) for each station, applying an innovative method that rigorously explores H - κ parameter space to give robust and reliable results.

My results constrain the crustal structure and composition of northern Afar and the Danakil microplate, providing information on the progression of continental rifting in this diffuse extensional domain. I find that thin crust underlies parts of the Danakil microplate, proving that it has undergone significant stretching and crustal thinning, rather than being an ‘undeformed block’. My results show varied V_P/V_S ratios, some greater than 1.9. These elevated values can only be explained by the presence of partial melt, thus demonstrating that magma-assisted extension continues to be important in the final stages of continental breakup.

By investigating the relationship between crustal thickness and topographic elevation across the region of study, I estimate the regional crustal bulk density as $\rho_c \approx 2850 \pm 20 \text{ kg m}^{-3}$, which is higher than expected, given the crustal thickness of the region. I show that topography is $1.5 \pm 0.4 \text{ km}$ higher than would be expected due to crustal isostasy alone. I propose that this topography is supported by the

same hot mantle upwelling suggested to be responsible for the onset of rifting in Africa. Uplift is generated due to the presence of this hot thermal anomaly beneath the plate and by thinning of the lithospheric mantle.

5.1.3 Magma storage and fluid transport through an intrusive complex beneath Nabro volcano

In Chapter 4, I explore the crustal subsurface beneath Nabro caldera, an off-rift volcano located on the Danakil microplate. Nabro underwent its first eruption on historical record in June 2011; following this, a temporary network of seismometers was deployed there from August 2011–October 2012. To characterise the movement and storage of fluids in the subsurface following the eruption, I use local earthquake traveltimes tomography to image the seismic velocity structure beneath Nabro.

I jointly relocate the seismicity and invert for the three-dimensional P- and S-wave velocity structures (V_P , V_S) and the ratio between them (V_P/V_S). My observations identify the continued presence of magma in the subsurface following the eruption, stored within an aseismic region of low V_P , low V_S and high V_P/V_S ratio at depths of 6–10 km b.s.l.. I track the degassing of this melt through an intrusive complex in the crust, defined by high seismicity, very low V_P/V_S ratio and low V_P . This intrusive complex contains overpressurised gases, which are released at the surface. The presence of volatiles and high temperatures within the shallow crust provide evidence that Nabro is a viable candidate for geothermal exploration.

The novelty of these results lies firstly in the use of a seismic catalogue produced by a deep learning model, which has never been done before in a seismic tomography study. I prove that deep learning models are an effective method to obtain earthquake catalogues for the purposes of seismic tomography at volcanoes. Furthermore, it is rare to observe the post-eruptive state of a volcano, particularly one in the understudied Afro-Arabian Rift System. Nabro was almost completely unmonitored prior to its eruption. Other than this work, no seismic tomography studies exist for any volcano in the ~110 km long Nabro Volcanic Range. I thus provide

crucial insights into the subsurface structure of a recently erupted volcano in an understudied tectonic setting.

Nabro is located in the Nabro Volcanic Range, an off-rift volcanic lineament trending obliquely to the dominant rifting trends in Afar. My receiver function results in Chapter 3 do not show marked crustal thinning around Nabro, nor do the measurements indicate a high degree of partial melt. I postulate that the Nabro Volcanic Range may be a pre-existing weakness in the lithosphere that has been exploited as rifting processes developed in the wider Afar region. The broad, coherent mantle flow across the region, as identified in Chapter 2 may lead to a patchwork of terranes as the three rift branches compete for dominance. Ultimately, the region could evolve to look something like the modern-day Seychelles.

5.2 Future work

5.2.1 Further investigation of Nabro volcano and the Nabro Volcanic Range

My seismic tomography study of Nabro volcano has revealed its post-eruptive subsurface structure. However, my tomography images are limited by their depth resolution, due to the vast majority of seismic events occurring above 10 km b.s.l.. Nabro's deeper crustal structure remains elusive. Particularly puzzling is why I observe such a low bulk V_P/V_S ratio beneath Nabro in the H- κ results from Chapter 3. This result implies that either partial melt is stored in horizontal stacked sills that reduce the observed V_P/V_S ratio through the H- κ stacking method, or that the crust lacks partial melt overall. The latter could have interesting implications for the formation of the Nabro Volcanic Range (NVR); as an oblique off-rift volcanic lineament, its origin remains mysterious. It is not well-understood how the NVR accommodates extensional processes, nor how magma is supplied to it (Hamlyn et al., 2014). Understanding the low bulk V_P/V_S ratio observed at Nabro would naturally lead to a more comprehensive study of the subsurface structure and tectonic processes that formed the NVR.

While investigating receiver functions at the Nabro stations, I attempted to estimate crustal anisotropy beneath the volcano, following the method of Liu and Niu (2012). Crustal anisotropy can constrain melt geometries in the subsurface (e.g., Hammond, 2014). Unfortunately, the backazimuthal coverage of the receiver functions was not sufficient to enable calculation of anisotropy in the crust. This could be addressed by studying shear-wave splitting in local earthquakes to quantify anisotropy in the upper crust (e.g., Keir et al., 2005, 2011a; Illsley-Kemp et al., 2017). Crustal anisotropy is crucial to understanding the pattern of stress and strain in the crust, which is of particular interest at Nabro due to its unusual post-eruptive rotation of the regional maximum compressive stress axis (Hamlyn et al., 2014). Mapping crustal stress field orientations at volcanoes could be a potential eruption forecasting technique (e.g., Roman and Cashman, 2006). Investigating the volcanic stress field response at Nabro using crustal anisotropy could thus provide insights into the origins of the NVR and the magma supply to this off-rift volcanic lineament.

I also studied the Ps arrivals of the receiver functions at stations around Nabro, hoping to observe a systematic variation in arrival time, as seen by Janiszewski et al. (2020) at Cleveland Volcano. Using this variation in arrival time, Janiszewski et al. (2020) image a mid-crustal magma reservoir beneath the volcano. However, I was again limited by the backazimuthal coverage of events at the Nabro stations (see Appendix D). I therefore suggest that future work should also focus on attempting to image the transcrustal magmatic architecture of Nabro. A longer seismic deployment could provide the backazimuthal coverage necessary for studies like those of Janiszewski et al. (2020) or Liu and Niu (2012). Electrical conductivity models, produced by magnetotelluric images, could also be used to characterise deep magmatic transport and storage beneath the volcano (e.g., Samrock et al., 2018). Alternatively, Journeau et al. (2022) use sources of seismovolcanic tremor to delineate the deep magma plumbing system beneath the Klyuchevskoy Volcanic Group in Kamchatka. A reanalysis of the seismic data collected following the eruption, with a focus on characterising long-period (LP) events and tremor, could provide new insights into the deeper magmatic architecture beneath Nabro.

New detection methods for identifying tremor and LP events are currently being developed and deployed at other volcanoes (e.g., Permana et al., 2022; Fee et al., 2023; Steinmann et al., 2024).

To fully understand the hydrothermal system at Nabro, both field and remote studies are required. A magnetotelluric study could characterise geothermal potential at Nabro by mapping volatile transport and delineating hydrothermal systems in the crust (e.g., Siniscalchi et al., 2010; Samrock et al., 2018). Other suggestions for investigating the geothermal system at Nabro are: mapping volcanic vent distributions (e.g., Paulsen and Wilson, 2010), soil-gas CO₂ degassing surveys (e.g., Chiodini et al., 1998), LIDAR (Light Detection And Ranging) observations of digital elevation to map fault systems and observe hydrothermally altered rocks (e.g., Crowley and Zimbelman, 1997; Pyle and Elliott, 2006), and seismic attenuation tomography (e.g., Hudson et al., 2022). Comprehensive analyses such as these would both allow an assessment of Nabro’s potential as a geothermal resource, and indicate how it might best be exploited.

5.2.2 Other seismic tomography studies in the AARS

More than 12 volcanoes in the East African Rift have experienced deformation over the past three decades, with “significant deformation” occurring beneath four volcanic edifices in the Main Ethiopian Rift (Biggs et al., 2011). Of these four volcanoes, only Aluto has been investigated using seismic tomography (Wilks et al., 2020). Earthquake catalogues exist for two deforming volcanoes: Bora-Tulu Moya and Corbetti (Greenfield et al., 2019; Lavayssière et al., 2019). I suggest that carrying out seismic tomography at either of these volcanoes would be useful and instructive, both in understanding magma storage and transport beneath the edifices, and also as a comparison to the studies at Aluto and Nabro. I have begun a seismic tomography study at Bora-Tulu Moya and am in the process of producing initial results. I also propose seismic deployments should be carried out at other volcanoes in Afar, such as the Erta’Ale range or Alid, to further characterise and compare these extensional segments to the Nabro Volcanic Range.

The more models we have of rift and off-rift volcanoes in the AARS, the more we will begin to understand the origin and development of these volcanic systems, and their role in continental breakup processes.

5.2.3 Improving understanding of the crustal structure beneath northern Afar

This thesis has identified avenues for future investigations into the crustal structure of northern Afar and the Danakil microplate. Questions remain about the subsurface structure beneath the Danakil Alps; whether the diffuse extension in the Gulf of Zula will continue to migrate westwards; and what direction rifting is propagating in (north-to-south or south-to-north).

My receiver function analysis could be extended to produce more detailed and continuous images of the crustal structure. If a denser seismic network were deployed in northern Afar and across the Danakil microplate, the individual Moho measurements could be interpolated to produce a single continuous surface (e.g., Miller and Moresi, 2018). Common conversion point (CCP) stacking is another method that can provide enhanced images of continuous structures if the network is dense enough (e.g., Dueker and Sheehan, 1998; Angus et al., 2009; Bianchi et al., 2010; Hammond et al., 2011c; Miller et al., 2018). CCP stacking migrates each receiver function to depth along its incoming raypath within a 3D grid. Receiver function amplitude is added to all grid points within the ray Fresnel zone, which increases in width as depth increases, due to increased velocity and thus wavelength at depth (Jenkins et al., 2020). The procedure therefore constructs a broad-scale image of crustal structure.

A denser network could generate a local earthquake catalogue, enabling a shear-wave splitting study of local earthquakes to investigate crustal anisotropy (e.g., Keir et al., 2005, 2011a; Illsley-Kemp et al., 2017). If a local earthquake catalogue cannot be produced, a network with a longer runtime would produce a higher number of teleseismic observations, and crustal anisotropy could be investigated using receiver

functions (e.g., Liu and Niu, 2012). Knowledge of crustal anisotropy is essential for investigating the state of stress in the crust.

Ambient seismic noise (the background vibrations of the Earth's surface) can also be exploited as a seismic signal to image the subsurface (e.g., Shapiro et al., 2019). Cross-correlation of ambient seismic noise (or ambient noise interferometry) reconstructs signals of fundamental mode surface waves. This method can be used as a high-resolution tomographic imaging technique to model the crust and upper layers of the Earth (e.g., Shapiro et al., 2005; Yang et al., 2007; Bensen et al., 2007, 2008; Bao et al., 2015). Ambient noise tomography has been used to investigate the crust and uppermost mantle of the northern Main Ethiopian Rift and Afar (Chambers et al., 2019), but without a comprehensive analysis of azimuthal and radial anisotropy. New methods have emerged using ambient noise interferometry to invert for radial and azimuthal anisotropy of shear wave velocity structures, which constrains deformation patterns in the crust and upper mantle (e.g., Huang et al., 2010; Liu et al., 2019). If applied to Afar, ambient noise surface wave tomography could reveal the dynamic processes frozen into the fabric of the crust and uppermost mantle, supplementing the ambient noise tomography work of Chambers et al. (2019). Surface wave dispersion data can also be jointly inverted alongside receiver functions to provide an estimate of absolute seismic velocities alongside the velocity contrasts observed in receiver function analysis (e.g., Horspool et al., 2006; Obrebski et al., 2015). This method has proved successful in imaging the the deep crustal magmatic architecture of the Altiplano-Puna Volcanic Complex (Ward et al., 2014).

All of these suggestions would produce a deeper understanding of the crustal structure in Afar, in turn answering key questions on the progression and evolution of rifting in the region (e.g., Rime et al., 2023).

5.2.4 Regional mantle dynamics

Future regional deployments should address the data gap in Egypt, Sudan and Somaliland, where very few shear-wave splitting measurements have been reported. Somaliland is of particular interest due to its location on the southern margin of

the Gulf of Aden. An analysis of seismic anisotropy beneath Somaliland would provide crucial insights into the mantle flow dynamics beneath this continental margin. Similarly, the data gap in Egypt and Sudan is striking, with only four seismic stations available on IRIS in Egypt, and none in Sudan. This means that the seismic structure of the western margin of the Red Sea is not well-constrained. The western margin has no counterpart to the volcanic harrat fields in Arabia, but this is not well-understood from a seismic perspective. To fully understand the initial rifting processes that led to the separation of Arabia from Africa, it is crucial to address the absence of measurements in these areas.

I also suggest that a network deployed for a longer time period in Yemen would allow for an analysis of multilayer anisotropy (e.g., Hammond et al., 2014b). A large amount of data across a range of backazimuths is required for the shear-wave splitting inversion procedure of Wookey (2012). I believe the results of this would be instructive in distinguishing complex anisotropic signatures across the lithosphere from the simple asthenospheric anisotropy that I observe in Chapter 2, and provide further constraints on the extent of the density-driven mantle flow along the margin of the Gulf of Aden.

Surface wave azimuthal anisotropy tomography models have good depth resolution but poor lateral resolution, a direct complement to the poor depth resolution and relatively good lateral resolution of SKS studies (e.g., Becker et al., 2012; Faccenna et al., 2014). My SKS measurements across southwestern Arabia indicate a need for a regional tomography study that specifically incorporates anisotropy, either using surface waves (e.g., Becker et al., 2014) or body waves (e.g., VanderBeek et al., 2023). Both models have proved successful in modelling heterogeneity in the mantle and revealing large-scale patterns in mantle flow beneath tectonically complex regions (e.g., Long and Becker, 2010; Faccenda and VanderBeek, 2023; Rappisi et al., 2022).

5.3 Final remarks

This thesis has examined the dynamic crust and mantle structure beneath the Afar Triple Junction using three different seismological techniques on three different scales.

I reveal the complexity of the crust, underlain by a simple mantle flow pattern. This work forms part of the decades-long project by seismologists and other researchers to investigate the Afro-Arabian Rift System, beginning in the 1970s. Over the last twenty years, a considerable amount of work has gone into acquiring the datasets analysed in this thesis by deploying temporary seismic networks across key areas of the AARS, particularly Eritrea. My new observations provide critical insight across all scales into the driving forces and manifestations of continental breakup in the region, an essential foundation for future explorations of these complex processes.

Appendices



Supplementary information for “Channelized mantle flow through the Afar Triple Junction and around the Arabian plate”

Contents

A.1 Introduction	137
-----------------------------------	------------

A.1 Introduction

Figure A.1 plots the backazimuthal variation in shear-wave splitting parameters (ϕ and δt) for the Eritrea Seismic Project results. The SKS linearization procedure is illustrated in Figure A.2 for a good result recorded at station CAYE. Figure A.3 shows a poor result recorded at station TIOE that was discarded; the particle motion was not linearized effectively by the method. Figure A.4 shows an example of a null event recorded at station ASSE, where no shear-wave splitting of the SKS phase is observed. Table A.1 reports all the stacked shear-wave splitting estimates for the stations, plotted as red bars in Figure 2 in the main text.

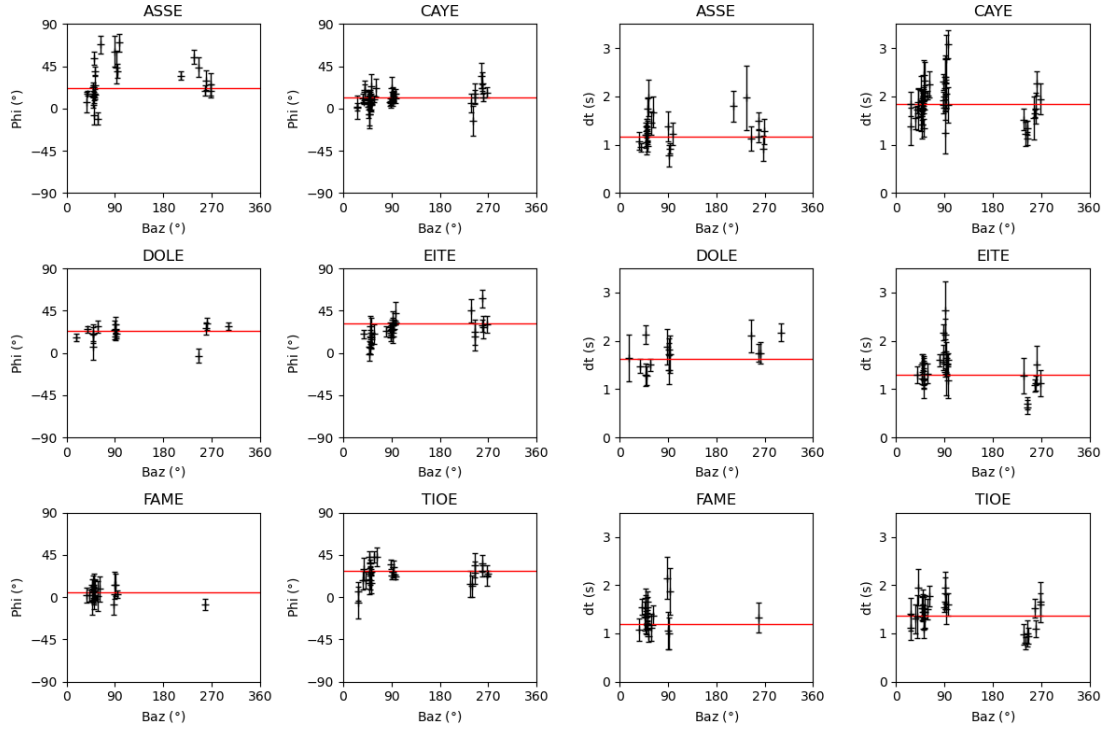


Figure A.1: Shear-wave splitting parameters (ϕ and δt) plotted as a function of source backazimuths for the Eritrea Seismic Project stations. Error bars are 1σ uncertainties.

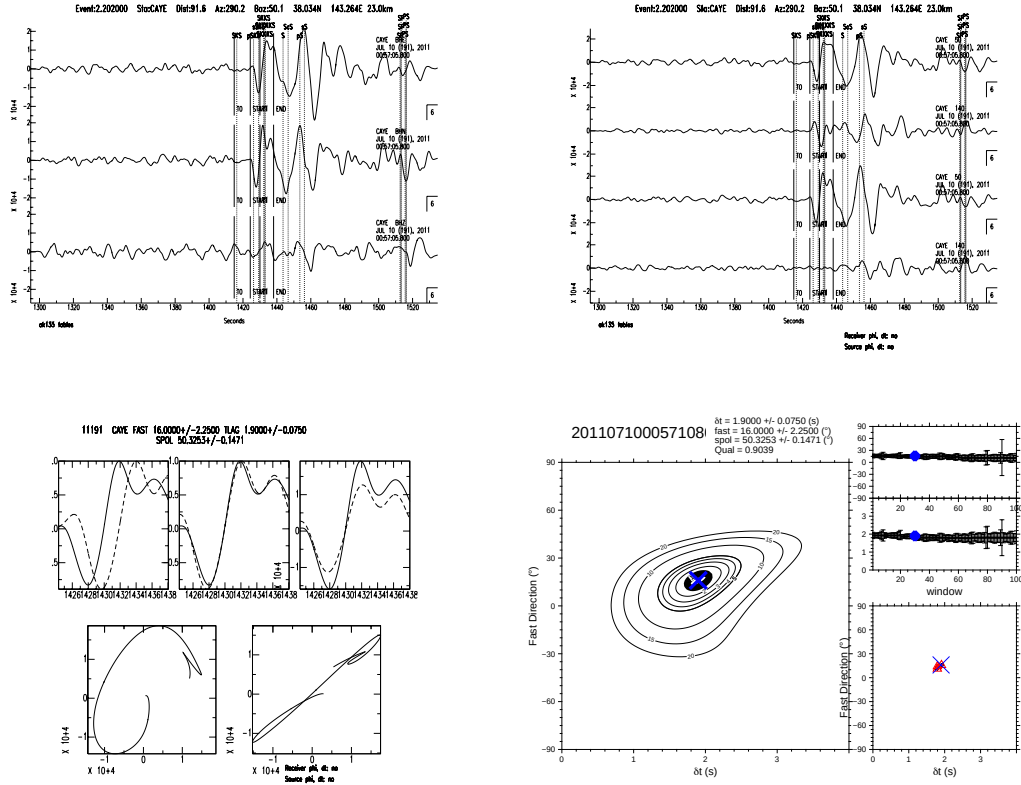


Figure A.2: SKS splitting analysis for an event recorded at station CAYE. Predicted arrival times for the shear-wave phases are calculated using the IASPEI-91 travel-time tables. Top left panel: original seismogram N, E and Z components. Start and end of the final choice of window are marked. Top right panel: (top two traces) original transverse and radial components; (bottom two traces) transverse and radial components after anisotropy correction. SKS energy has been minimized in the transverse component. Bottom left panel: (top) superposition of fast and slow shear waves, uncorrected (left, right) and corrected (center); (bottom) SKS-particle motion, uncorrected (left) and corrected (right). Corrections are made using the estimated values of $(\phi$ and δt). Bottom right panel: (main) contour plot of the error surface of $(\phi, \delta t)$, with the best splitting parameters marked by a blue cross; (upper inset) the splitting parameters for each window and associated error, diamonds mark the final choice of window; (lower inset) clustering analysis results, with the best splitting parameters marked by a blue cross.

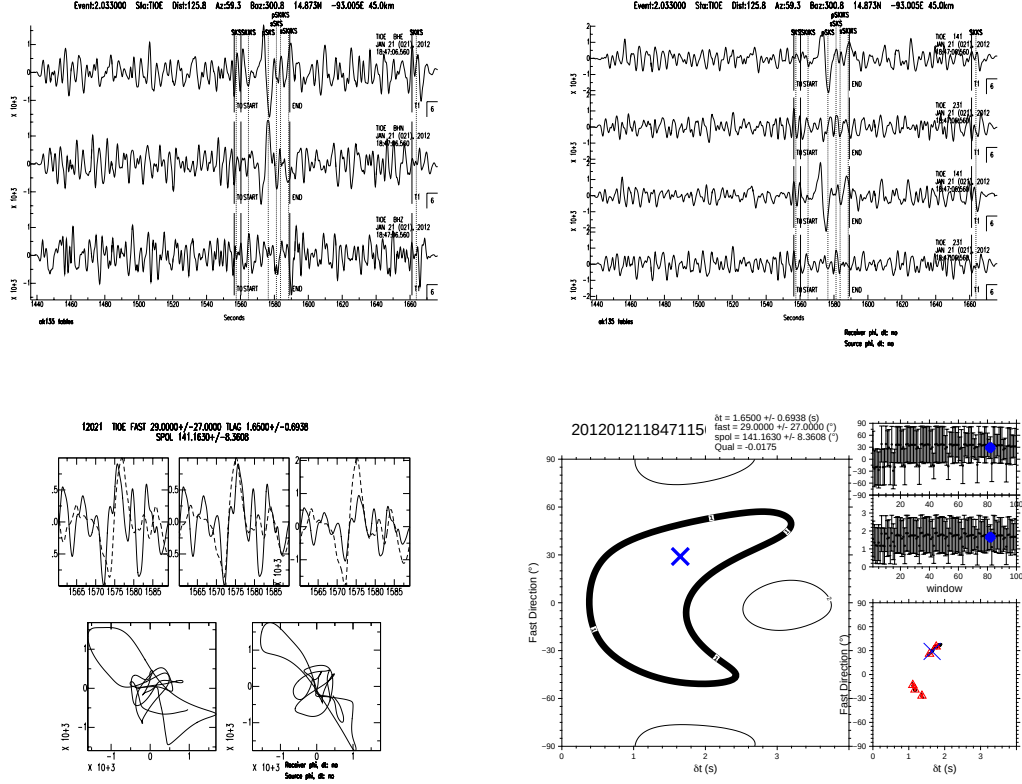


Figure A.3: SKS splitting analysis for a poor result recorded at station TIOE that was discarded from the results. The inversion process has not adequately linearized the particle motion or corrected the anisotropy. Predicted arrival times for the shear-wave phases are calculated using the IASPEI-91 travel-time tables. Top left panel: original seismogram N, E and Z components. Start and end of the final choice of window are marked. Top right panel: (top two traces) original transverse and radial components; (bottom two traces) transverse and radial components after anisotropy correction. SKS energy has been minimized in the transverse component. Bottom left panel: (top) superposition of fast and slow shear waves, uncorrected (left, right) and corrected (center); (bottom) SKS-particle motion, uncorrected (left) and corrected (right). Corrections are made using the estimated values of (ϕ and δt). Bottom right panel: (main) contour plot of the error surface of (ϕ , δt), with the best splitting parameters marked by a blue cross; (upper inset) the splitting parameters for each window and associated error, diamonds mark the final choice of window; (lower inset) clustering analysis results, with the best splitting parameters marked by a blue cross.

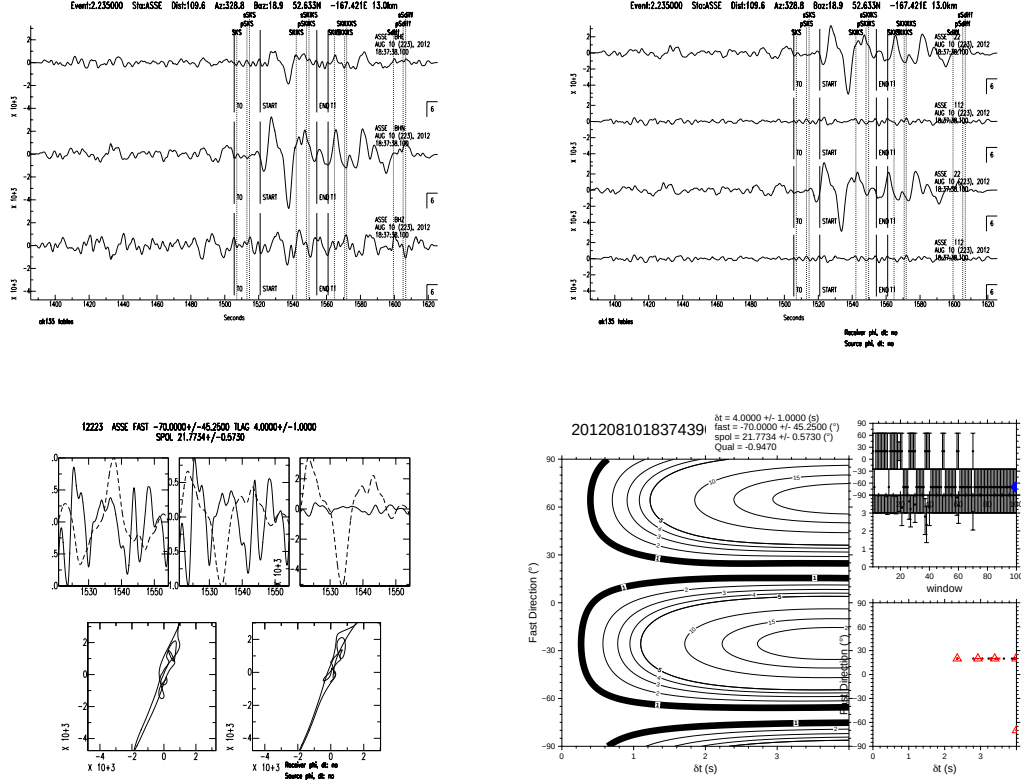


Figure A.4: SKS splitting analysis for a null event recorded at station ASSE; the backazimuth of the source is 18.9° , which is very close to the stacked ϕ result at the station, 22° . Predicted arrival times for the shear-wave phases are calculated using the IASPEI-91 travel-time tables. Top left panel: original seismogram N, E and Z components. Start and end of the final choice of window are marked. Top right panel: (top two traces) original transverse and radial components; (bottom two traces) transverse and radial components after anisotropy correction. SKS energy has been minimised in the transverse component. Bottom left panel: (top) superposition of fast and slow shear waves, uncorrected (left, right) and corrected (center); (bottom) SKS-particle motion, uncorrected (left) and corrected (right). Corrections are made using the estimated values of $(\phi$ and $\delta t)$. Bottom right panel: (main) contour plot of the error surface of $(\phi, \delta t)$, with the best splitting parameters marked by a blue cross; (upper inset) the splitting parameters for each window and associated error, diamonds mark the final choice of window; (lower inset) clustering analysis results, with the best splitting parameters marked by a blue cross.

Table A.1: Shear-wave Splitting Stacked Results.

Station	Long. (°N)	Lat. (°E)	ϕ (°)	Error (°)	δt (s)	Error (s)	Events
Eritrea Seismic Project stations							
ASSE	13.06	42.65	22	1.0	1.18	0.0250	25
CAYE	14.86	39.31	12	0.50	1.85	0.0188	45
DOLE	15.10	39.98	23	1.3	1.63	0.0438	15
EITE	15.24	38.78	31	1.0	1.30	0.0250	29
FAME	13.57	41.52	5	1.3	1.20	0.0125	20
TIOE	14.67	40.87	28	1.0	1.38	0.0188	28
YOCMAL stations							
<i>Yemen highlands</i>							
ADEN	12.78	44.98	61.0	3.0	0.925	0.0688	10
DAMT	14.09	44.68	45	15	1.10	0.356	2
DKUM	13.27	44.76	39	8.8	0.550	0.175	3
KHAW	13.80	43.25	2	5.5	1.05	0.0813	4
MAWI	15.47	43.52	-16.0	1.0	1.08	0.0375	9
MOKA	13.31	43.26	21	3.8	0.950	0.0500	5
SANA	15.39	44.21	-64	3.3	1.40	1.06	4
SHIB	15.50	43.91	-33	2.0	1.08	0.0688	4
SUGH	14.780	43.44	-36	1.5	1.13	0.025	3
TAWI	15.48	43.72	-24	3.0	1.33	0.106	2
UAYA	15.71	42.69	-19	2.3	2.00	0.110	2
YSLE	14.94	44.28	-11	4.0	0.675	0.0875	4
ZUWA	15.73	43.02	-26	1.5	1.05	0.0438	6
<i>Eastern Yemen</i>							
ADAH	12.65	53.65	24	5.5	1.23	0.263	3
EYUN	14.78	49.27	78	4.5	1.20	0.0563	3
GEAN	16.71	49.52	-79	3.0	0.675	0.0625	6
HAHY	15.21	49.09	-87	3.5	1.35	0.0688	2
HATT	17.32	52.11	-12	5.3	0.575	0.0500	4
MUKA	14.49	49.04	-89	3.3	1.30	0.0563	6
TAMR	15.78	48.95	-87	8.3	1.20	0.213	2
<i>Socotra island</i>							
ADUA	15.00	48.97	75	4.3	1.05	0.075	2
ANID	15.47	43.20	-3	2.3	1.13	0.0625	5
DABI	15.13	44.27	-36	4.3	0.775	0.131	2
DANA	12.39	53.52	18	1.8	0.675	0.0500	5
KABH	12.41	53.71	25	2.5	0.775	0.100	4
KADA	12.70	53.64	76	8.3	0.625	0.131	4
SAFN	12.53	53.46	16	10.3	1.03	0.2938	3
SHUA	12.59	53.50	24	10.3	0.675	0.163	3
WURE	12.42	53.58	18	1.8	0.900	0.0563	8

B

Supplementary information for “The dynamic crust of northern Afar and adjacent rift margins: evidence from receiver functions”

Contents

B.1	Introduction	143
B.2	Text S1: H-κ Stacking Method	144
B.3	Text S2: H-κ Stacking Result Quality Control and Uncertainty Estimates	145
B.4	Text S3: Result Quality Control–Eritrea Seismic Project Stations	147
B.5	Text S4: Result Quality Control–AFAR0911 Stations	148
B.6	Text S5: Isostatic calculations	149

B.1 Introduction

This supplementary information file gives further detail on the H- κ stacking method (Text S1) and the quality control criteria developed by (Ogden et al., 2019) (Text S2). Text S3 and S4 describe the results of these quality control tests on the Eritrea stations and Afar stations, respectively. Text S5 describes the calculation of residual topography using different values for asthenospheric channel thickness and initial

lithospheric mantle thickness (Equation 3.3 in main text). Table B.1 contains all H- κ estimates from this study and the studies of (Hammond et al., 2011c; Reed et al., 2014; Ahmed et al., 2022). Figures B.1 – B.10 are the diagnostic result figures for Eritrea, and Figures B.11 – B.19 are those for Afar. Figure B.20 plots the estimates of V_P/V_S ratio against crustal thickness from this study and the studies of (Hammond et al., 2011c; Reed et al., 2014; Ahmed et al., 2022). Figure B.21 plots parameter sweeps for different asthenospheric channel and initial lithospheric thickness values.

B.2 Text S1: H- κ Stacking Method

The H- κ technique, developed by (Zhu and Kanamori, 2000), uses the travel times of the Moho P-to-S conversion (Ps) and its subsequent crustal reverberations (PpPs, PpSs + PsPs) to determine the crustal thickness (H) and the velocity ratio of P-to S-waves (V_P/V_S , κ). The stacking function $s(H, \kappa)$ is determined for plausible values of crustal thickness and V_P/V_S ratio using a grid search according to:

$$s(H, \kappa) = \sum_{j=1}^N w_1 r_j(t_1) + w_2 r_j(t_2) - w_3 r_j(t_3) \quad (\text{B.1})$$

where $r_j(t_i)$ are the receiver function amplitudes at the predicted arrival-times of the Ps conversion (t_1) and subsequent reverberations (PpPs, t_2 ; PpSs + PsPs, t_3) for the j th receiver function. w_1, w_2, w_3 are stacking weights for phases Ps, PpPs and PpSs + PsPs, respectively, with $\sum_{i=1}^3 w_i = 1$. N is the total number of receiver functions used. The traveltimes are calculated as follows:

$$t_1 = H \left[\sqrt{\frac{1}{V_S^2} - p^2} - \sqrt{\frac{1}{V_P^2} - p^2} \right] \quad (\text{B.2})$$

$$t_2 = H \left[\sqrt{\frac{1}{V_S^2} - p^2} + \sqrt{\frac{1}{V_P^2} - p^2} \right] \quad (\text{B.3})$$

$$t_3 = 2H \sqrt{\frac{1}{V_S^2} - p^2} \quad (\text{B.4})$$

where p is the ray parameter, or slowness ($s\ km^{-1}$). The stacking function reaches its maximum as the three phases stack coherently and thus represents the best estimate of H and κ at a certain station.

B.3 Text S2: H- κ Stacking Result Quality Control and Uncertainty Estimates

The results of H- κ stacking are strongly sensitive to the choice of input parameters (the chosen V_P , values assigned to the phase weights and whether the stacking is linear or phase-weighted). Data parameters such as receiver function frequency content and the subset of receiver functions to be stacked also exert control on the final result. To mitigate this issue, we follow the procedure of (Ogden et al., 2019). This adapted H- κ stacking procedure repeats the process 1000 times, with H- κ input parameters and data parameters sampled randomly between geologically plausible limits. We allow H , κ and V_P to vary within ranges of 15 – 55 km, 5.9 – 6.6 kms^{-1} and 1.65 – 2.25 respectively. Cluster analysis selects a final solution from the individual estimates. Ten quality control criteria variously assess the final numerical result, the receiver function data set and the extent of clustering of results, as follows:

1. Cluster analysis selected H or κ lies within the boundaries of the H- κ grid space.
2. Errors for cluster analysis selected result are: $< \pm 2.5$ km in H and $< \pm 0.05$ in κ .
3. Standard deviation of H is $< \pm 2.5$ km.
4. Standard deviation of κ is $< \pm 0.05$.
5. Mean average amplitude comparison estimate (ACE) of the repeats with frequency less than the maximum frequency content of the receiver function results is > 3 . ACE is the ratio of the amplitude at the t_1 arrival time for each receiver function predicted by the H- κ solution with the root-mean-square amplitude of the receiver function between $t_1 + 2s$ and $t_2 - 2s$.

6. Mode H and κ (the values occurring most frequently) lie within the same cluster as the mean H and κ (the average of all values).
7. Summed amplitudes of Ps, PpPs and PsPs + PpSs are positive, positive, and negative, respectively.
8. Mean average cross-correlation of all possible pairs of same frequency receiver functions is > 3 .
9. Mean average iterative deconvolution variance (IDVAR) of the repeats with frequency less than the maximum frequency content of the receiver function results is $> 90\%$. A high IDVAR indicates that the deconvolution consistently produces accurate receiver functions.
10. Overall mean H and κ for the linearly stacked repetitions lie within one standard deviation of H and κ for phase-weighted stacked repetitions, and vice versa.

These quality indicators are used to assess the reliability of H- κ stacking at a station. Criteria 1 and 2 characterise the numerical quality of the final solution, with the error limits in criteria 2 determined by the resolvable Moho thickness and synthetic tests for a sharp Moho (Ogden et al., 2019). Criteria 3, 4 and 6 assess the extent to which the 1000 repetitions produce well-clustered H- κ results. Criteria 5, 7 and 9 analyse the receiver functions and determine whether the Ps arrival and its crustal reverberations are impulsive. In particular, criteria 5 assesses whether the station overlies a gradational Moho or complex intracrustal structure, both of which may produce additional Ps conversions between the Moho Ps and PpPs arrivals. Criteria 8 tests the effect of noise and variations in structure with backazimuth, and is an indication of the level of homogeneity of crustal structure beneath a seismic station. Criteria 10 compares linear and phase-weighted stacking H- κ strategies.

In general, stations that pass ≥ 9 of these criteria can be considered to be reliable. If < 6 criteria are passed, the result is likely to be unreliable and H- κ is not an appropriate way of analysing bulk crustal structure at that station. A

study involving 582 broadband seismic stations found that stations that pass ≥ 8 of the quality control indicators are almost always reliable (Ogden and Bastow, 2022). Stations with 6–8 criteria passed must be analysed more carefully to ascertain reliability—in particular, whether criteria 1 and 6 are passed is a crucial indication of the stability of a result.

The final result automatically selected by cluster analysis is useful in determining the quality of the receiver functions at a certain station (see quality control criteria 1, 2, 5 and 7). However, the numerical error determined from the $s(H, \kappa)$ surface for an individual H- κ measurement is often very small and therefore implausible (Ogden et al., 2019). Thus, we follow (Ogden and Bastow, 2022) and calculate the mean H and κ from the 1000 H- κ estimates using a bootstrapping algorithm, reporting the standard deviation of the ensemble as the associated error. This approach produces more realistic error estimates, particularly given the variations in H and κ that arise from changes in the H- κ input parameters, and also incorporates uncertainties in V_P . We therefore use these mean results and associated errors in our subsequent analysis and discussion.

B.4 Text S3: Result Quality Control—Eritrea Seismic Project Stations

As explained in Text S1, the quality control criteria allow the overall quality of a station to be characterised and visualised via diagnostic result figures (plotted in Figures B.1 – B.10). In total there are nine results, including one from the composite station made up of the Nabro Urgency Array (NAB), and two produced by station EITE after it is split by back-azimuth (EITE (W) and EITE (E)). We also present the result figure for station EITE prior to splitting by back-azimuth (Figure B.4) to indicate that the clustering of H- κ results produces a bimodal distribution, providing support for our decision to split the station into western and eastern back-azimuths. Station LIDE did not have enough reliable receiver functions associated with it to produce a reliable result (Figure B.9).

Two stations pass all ten criteria (ASSE and EITE (W)) and the results are considered to be stable and reliable. Three stations pass nine criteria. Stations CAYE and TIOE both fail criteria 5, which tests the ACE value, and we attribute this to intracrustal structure producing additional arrivals. Station EITE (E) fails criteria 2, which assesses the errors of the cluster-selected H - κ values, but only exceeds the κ error limit by 0.017 and is within the bounds of the H error limit. NAB also fails criteria 5 (an indication of intracrustal structure, particularly due to the volcanic setting of these stations) and criteria 2, again by a small margin in the κ error (0.07) only. We therefore decide that the results produced at these stations are reliable.

Station DOLE fails 3 criteria—5, 8 and 9. Failing criteria 8 indicates a less homogeneous crustal structure beneath the station, whilst failing criteria 9 implies that the receiver functions are of a lower quality. DOLE is close to the Gulf of Zula, where the Red Sea Rift passes into Afar. The area is dominated by broad inter-rifting deformation and mechanical extension (e.g., Viltres et al., 2020), which explains why criteria 5 and 8 indicate that there is complex, inhomogeneous structure beneath DOLE. Furthermore, the receiver functions are likely to be of lower quality at this station, and indeed fewer pass the earlier quality control and filtering—only 9 compared to 28 at CAYE. Since DOLE does produce a stable result as criteria 1 and 6 are passed (Ogden and Bastow, 2022), we accept this result.

Station FAME also fails 3 criteria—2, 4 and 6. From the result figure (Figure B.7), we observe that there is a wide distribution of κ values produced by the cluster analysis. Since criteria 2 is only failed by a slim margin in κ (0.018), we report the result from FAME, but caution that the κ value is more unreliable than other stations.

B.5 Text S4: Result Quality Control—AFAR0911 Stations

We carry out H - κ stacking on the ten AFAR0911 stations which have ≥ 6 associated receiver functions (Figures B.11 – B.19).

Station SAHE passes all ten criteria and its result is considered to be stable and reliable. Three stations (ABAE, FINE and KOZE) pass nine criteria, failing criteria 5. Criteria 5 assesses the ACE value, so we attribute this failure to the presence of complex intracrustal structure producing additional P-to-S conversions, and determine the results at these stations to be reliable. Two stations (IGRE and MAYE) pass eight criteria. MAYE fails criteria 9, indicating that the receiver functions are of lower quality, likely due to the complex intracrustal structure, which causes MAYE to also fail criteria 5. Since MAYE passes criteria 1 and 6, we accept this result, following (Ogden and Bastow, 2022). IGRE fails criteria 4 and 6, and its result figure (Figure B.15) shows that there is a wide spread of plausible κ values. Similarly, station LULE fails criteria 3, 4 and 5 and the result figure (Figure B.17) shows that the cluster analysis has a wide spread. However, since the final chosen results for IGRE and LULE are consistent with the results independently determined by (Hammond et al., 2011c), we accept them as reliable.

Stations HITE and BTIE fail 4 criteria. BTIE produces a result that is inconsistent with (Hammond et al., 2011c), and we therefore reject this result. HITE is also rejected due to the κ error on the mean result being > 0.1 , meaning it is difficult to interpret the reported value as the spread is so large. Station DALE fails 5 criteria and is also rejected as being unreliable.

B.6 Text S5: Isostatic calculations

In estimating crustal density we propagate uncertainties by sampling distributions of parameter values stated in the main text. We draw 10^5 samples from truncated normal distributions where uncertainties represent one standard deviation. Parameter values are prohibited from falling below zero. In the case of lithospheric thickness, a_1 , we impose a threshold of 15 km to prevent lithospheric thicknesses that are not consistent with our crustal thickness observations. Uncertainties in linear regressions are calculated using a total least squares regression calculated using principal component analysis. Before carrying out principal component analysis, data are centred on the mean and scaled by their standard deviation in both H and ϵ .

We carry out uplift calculations using different values for asthenospheric channel thickness, h , and initial lithospheric mantle thickness, a_0 (Figure S21). Results indicate that residual topography is consistent with independent constraints on sub-plate mantle temperature and lithospheric thickness if the lithosphere was ~ 100 - 140 km thick before being thinned to ~ 30 - 50 km and a temperature anomaly of $+110^\circ$ C exists in a sub-plate lithospheric channel that is ~ 50 - 150 km thick.

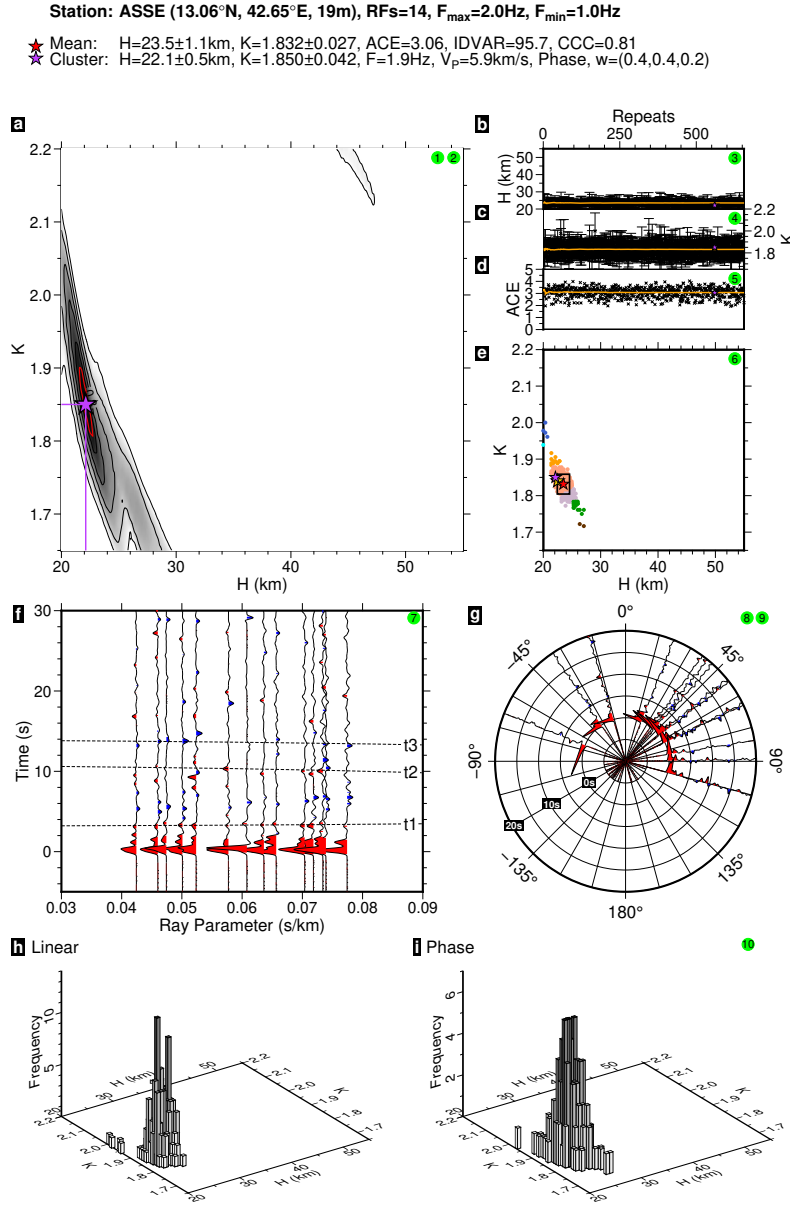


Figure B.1: H- κ result for station ASSE from the Eritrea Seismic Project deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H- κ solutions. (e) All 1000 H- κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

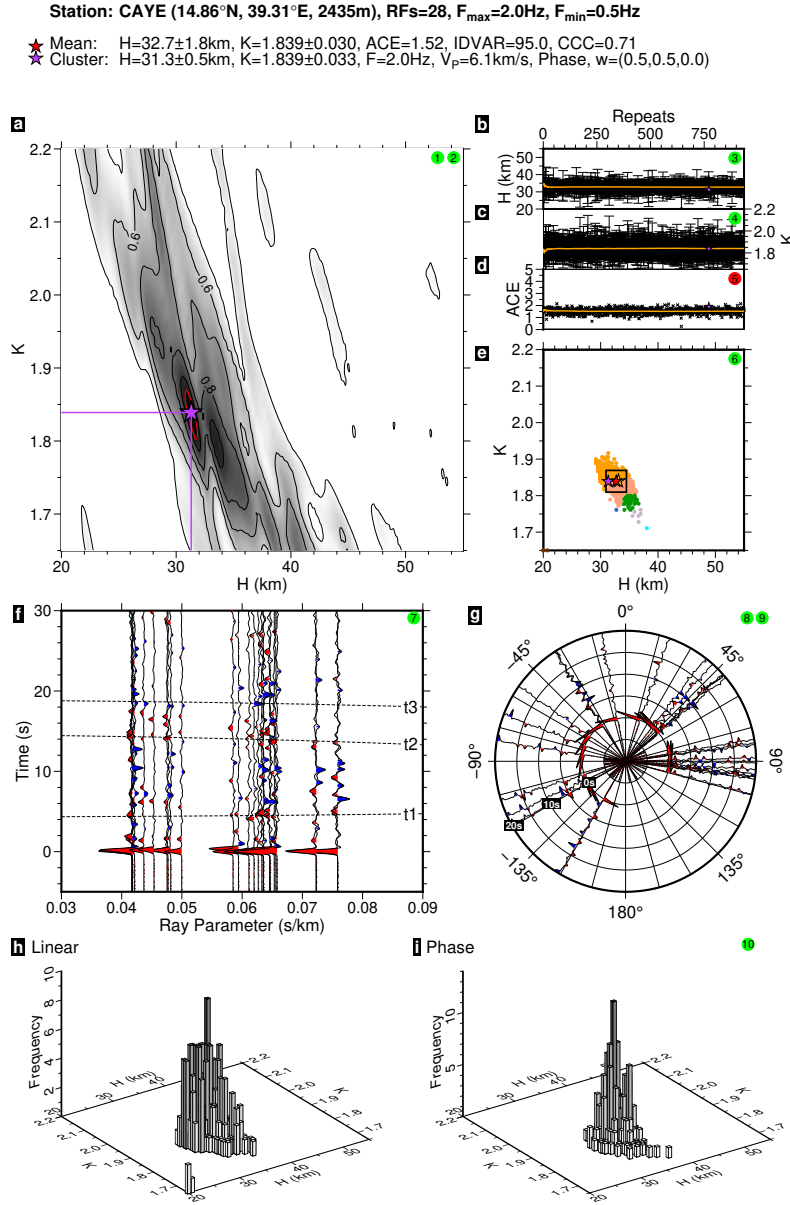


Figure B.2: H- κ result for station CAYE from the Eritrea Seismic Project deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H- κ solutions. (e) All 1000 H- κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

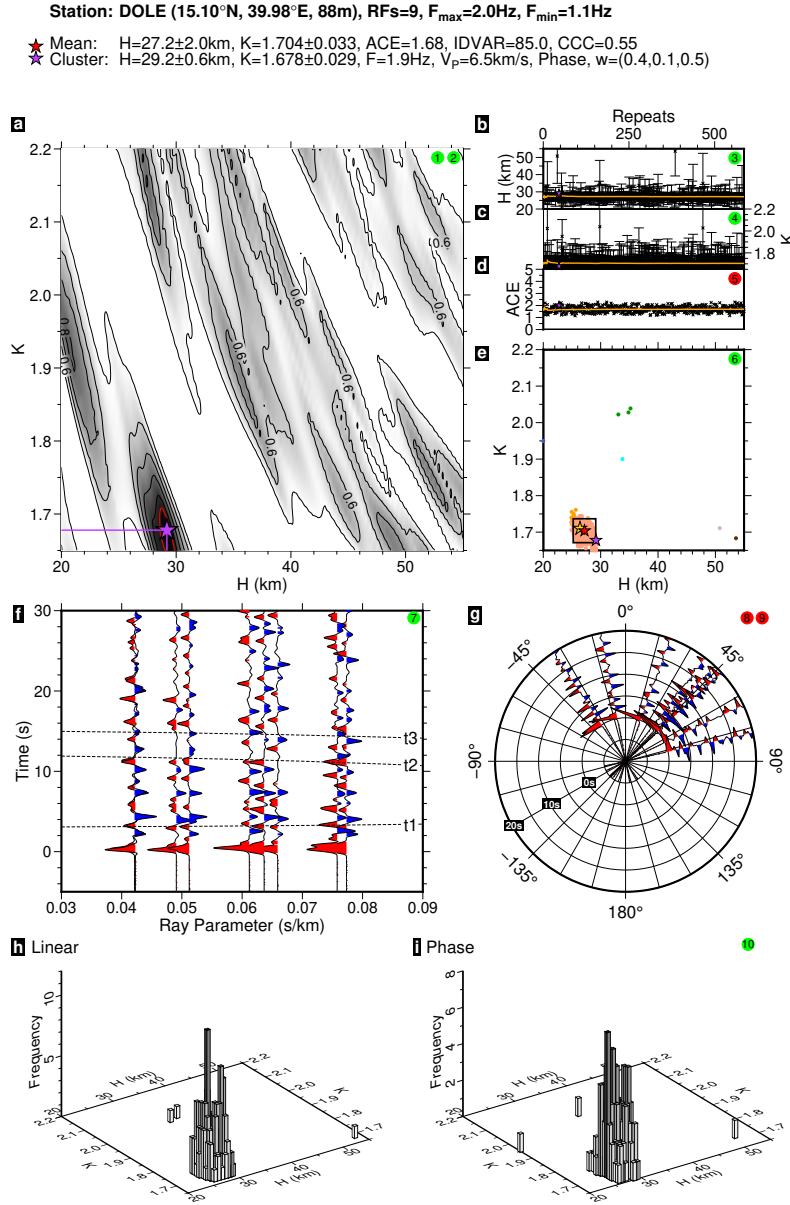


Figure B.3: H- κ result for station DOLE from the Eritrea Seismic Project deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H- κ solutions. (e) All 1000 H- κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

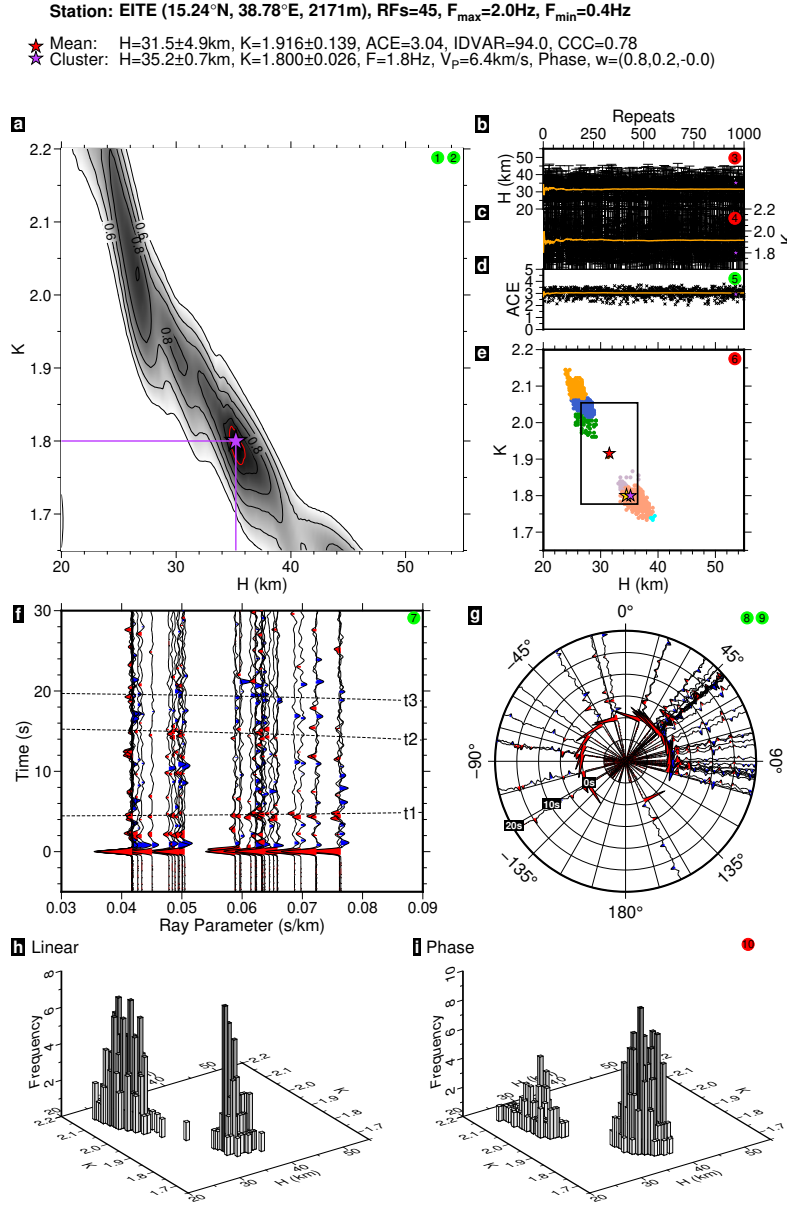


Figure B.4: H- κ result for station EITE from the Eritrea Seismic Project deployment (using all back azimuths). Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H- κ solutions. (e) All 1000 H- κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

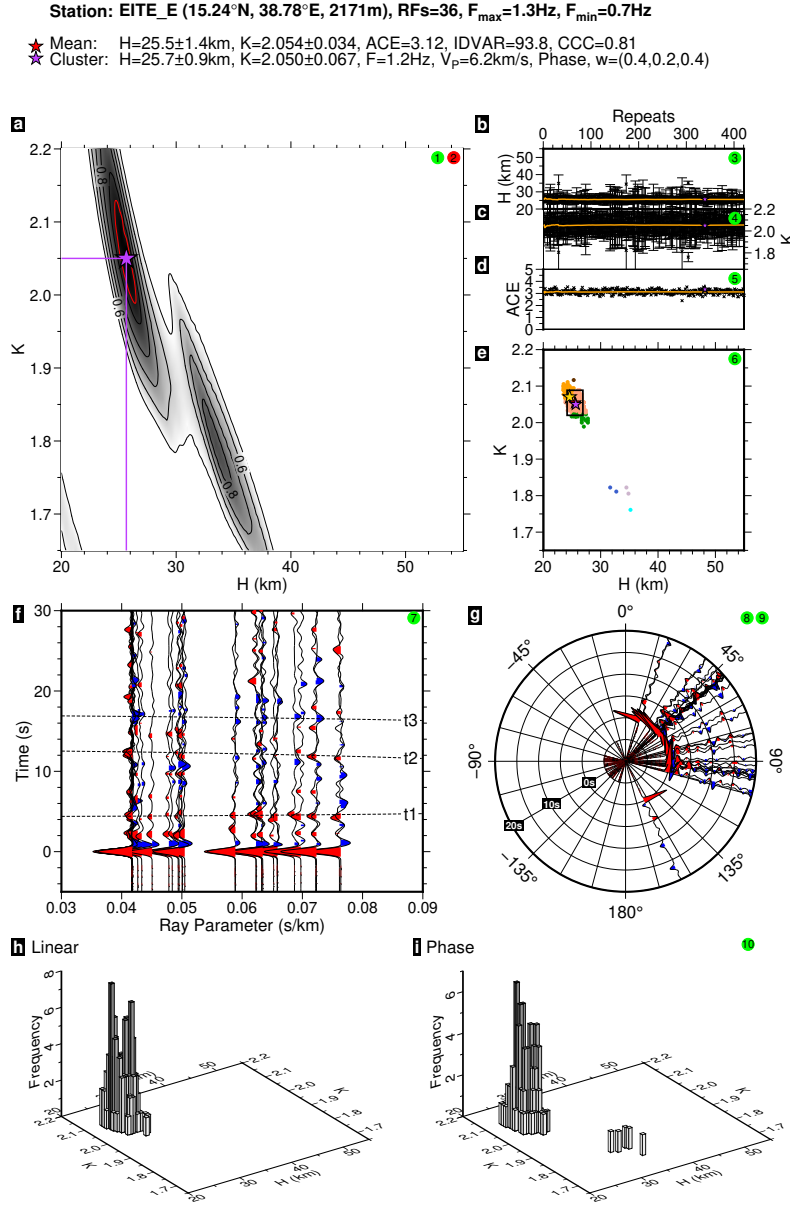


Figure B.5: H - κ result for station EITE from the Eritrea Seismic Project deployment (using only eastern back azimuths). Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H - κ solutions. (e) All 1000 H - κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

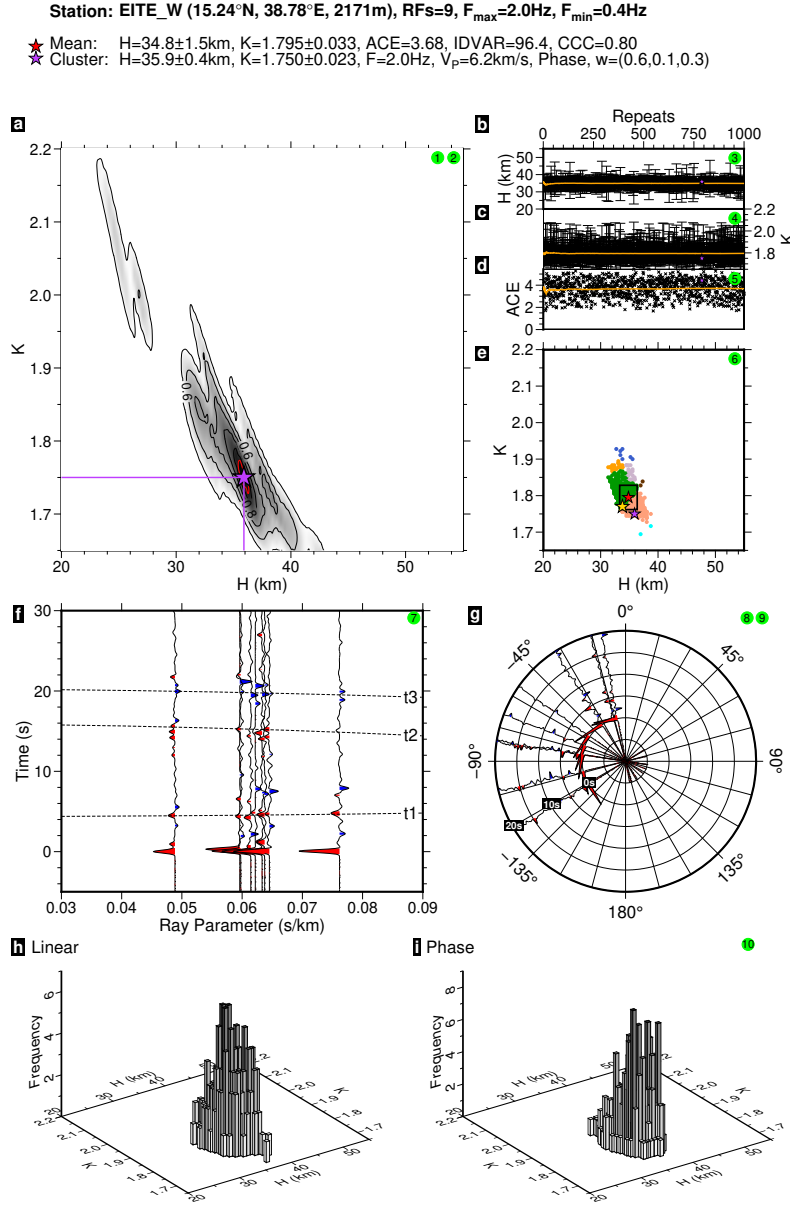


Figure B.6: H - κ result for station EITE from the Eritrea Seismic Project deployment (using only western back azimuths). Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H - κ solutions. (e) All 1000 H - κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

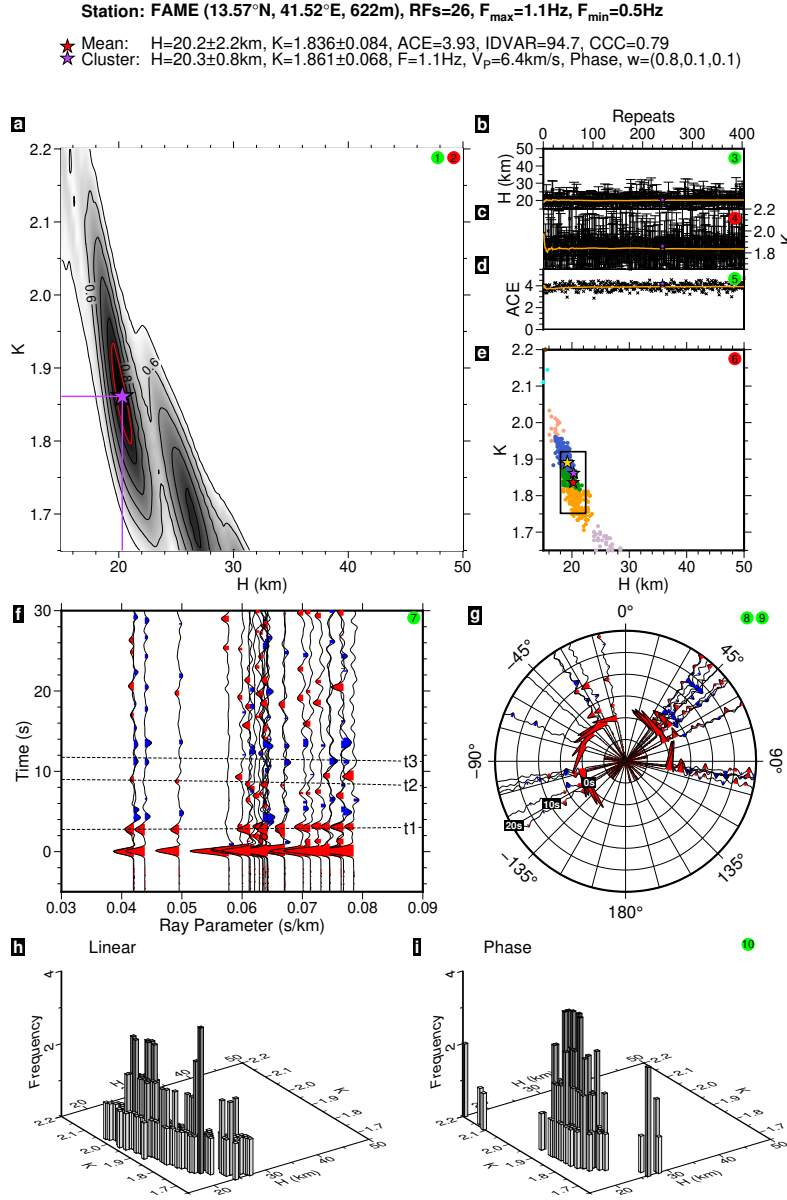


Figure B.7: H- κ result for station FAME from the Eritrea Seismic Project deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H- κ solutions. (e) All 1000 H- κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

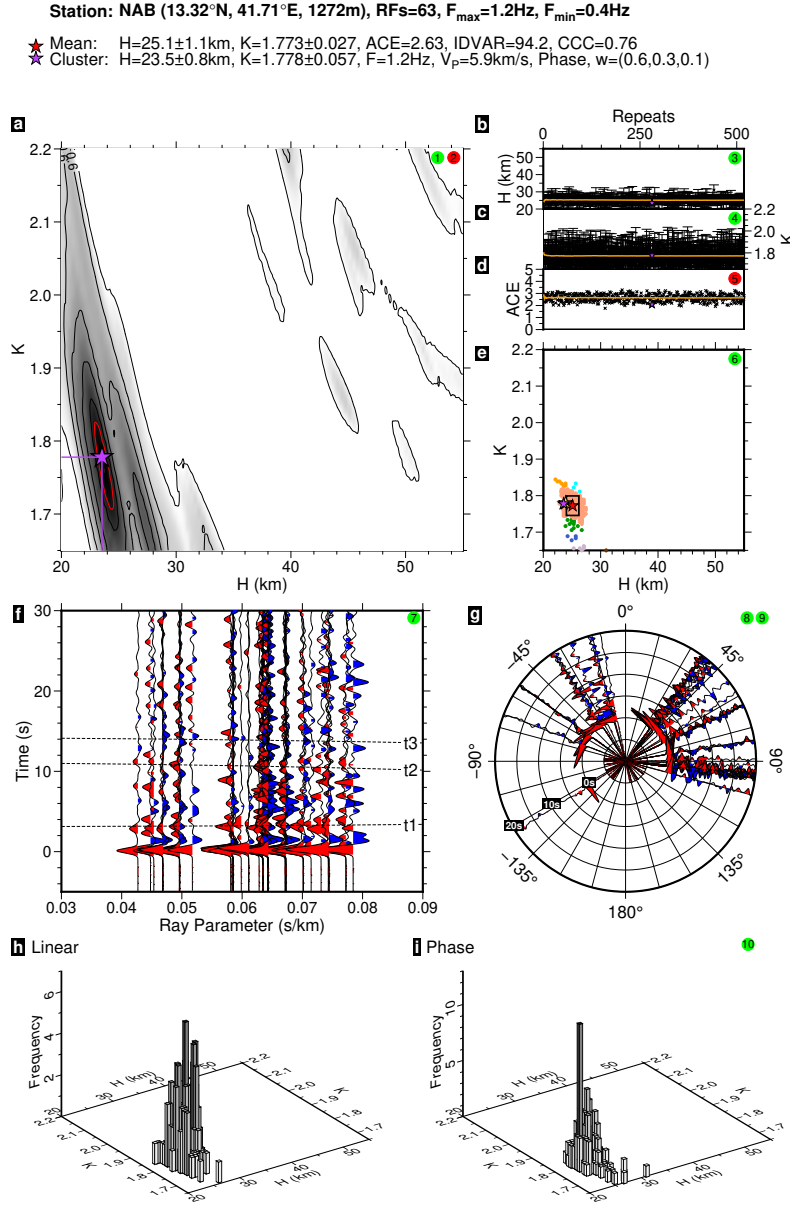


Figure B.8: H- κ result for composite station NAB. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H- κ solutions. (e) All 1000 H- κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

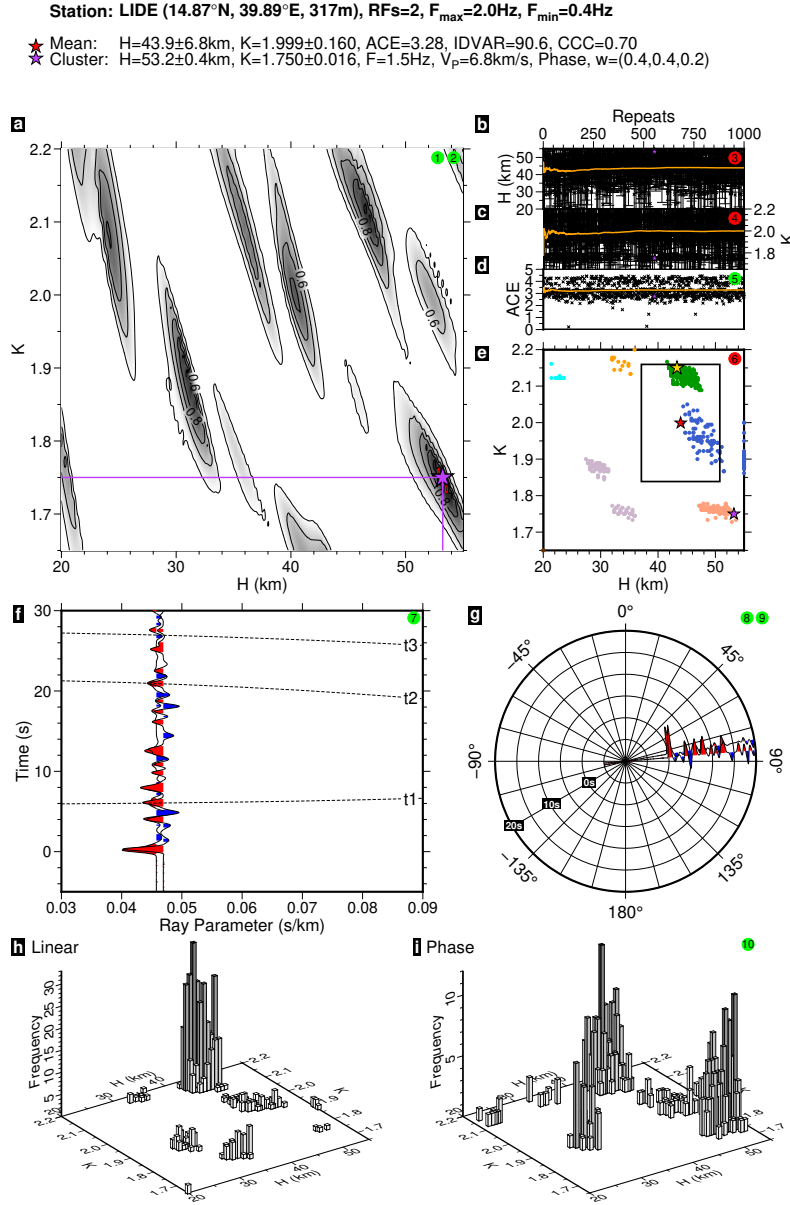


Figure B.9: H- κ result for station LIDE from the Eritrea Seismic Project deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H- κ solutions. (e) All 1000 H- κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

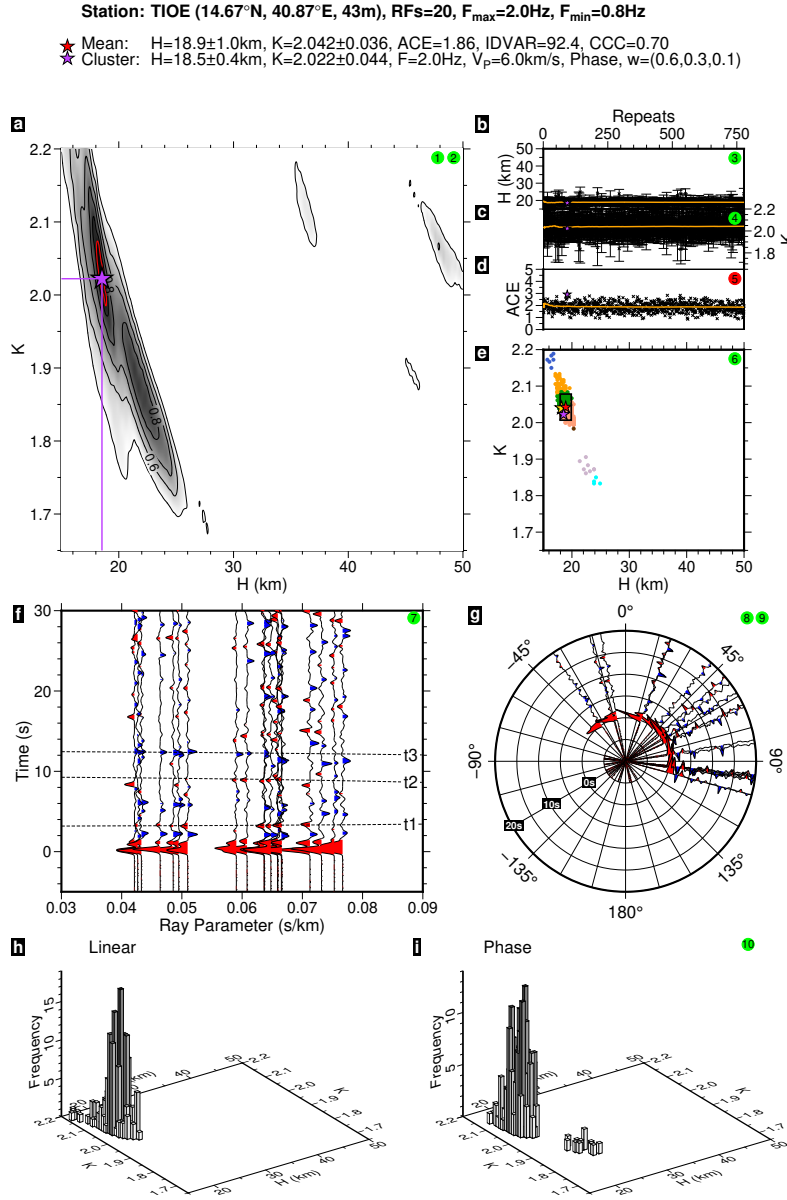


Figure B.10: H- κ result for station TIOE from the Eritrea Seismic Project deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H- κ solutions. (e) All 1000 H- κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

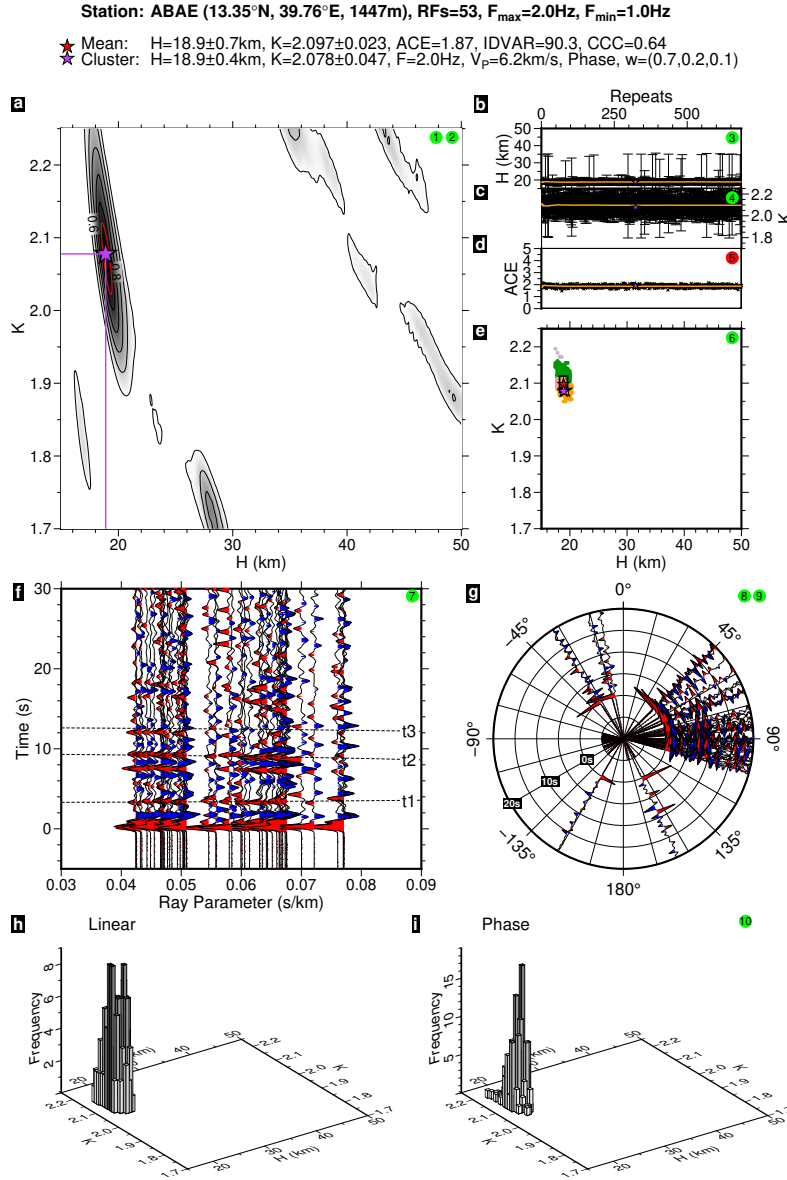


Figure B.11: H - κ result for station ABAE from the AFAR0911 deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H - κ solutions. (e) All 1000 H - κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

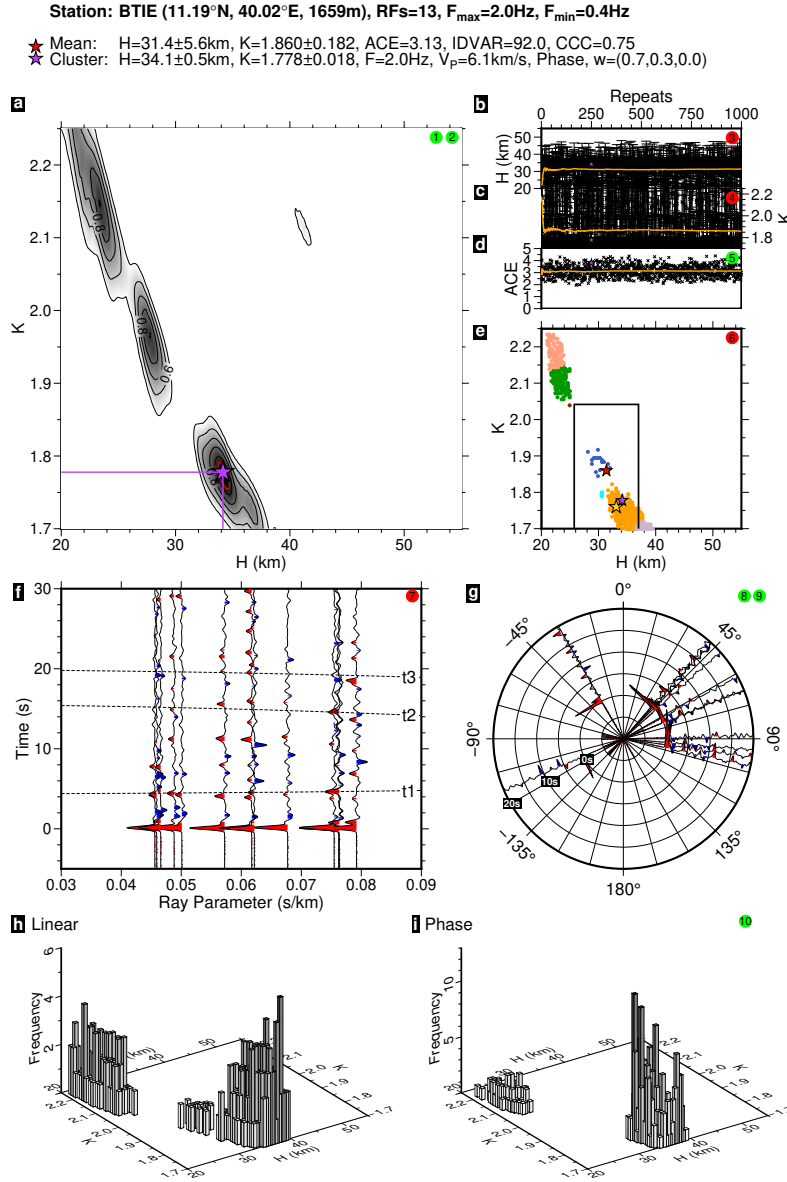


Figure B.12: H - κ result for station BTIE from the AFAR0911 deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H - κ solutions. (e) All 1000 H - κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

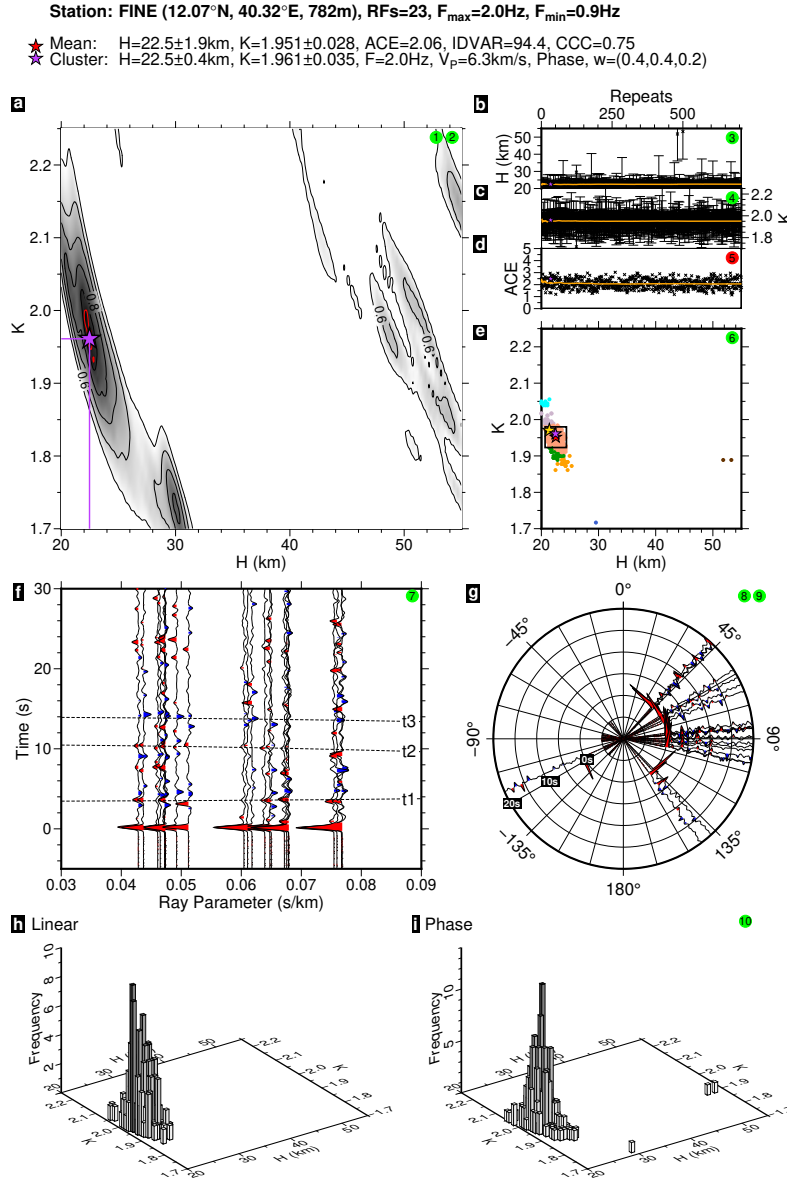


Figure B.13: H- κ result for station FINE from the AFAR0911 deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H- κ solutions. (e) All 1000 H- κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

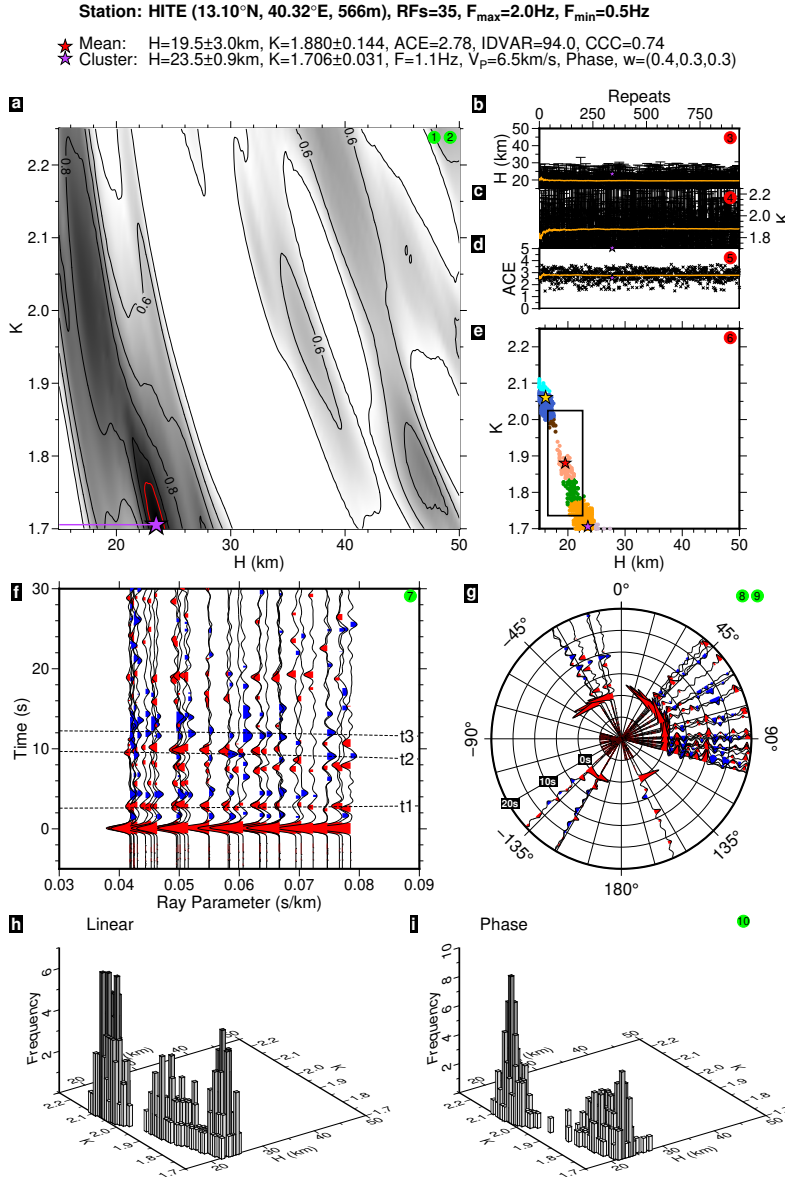


Figure B.14: H- κ result for station HITE from the AFAR0911 deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H- κ solutions. (e) All 1000 H- κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

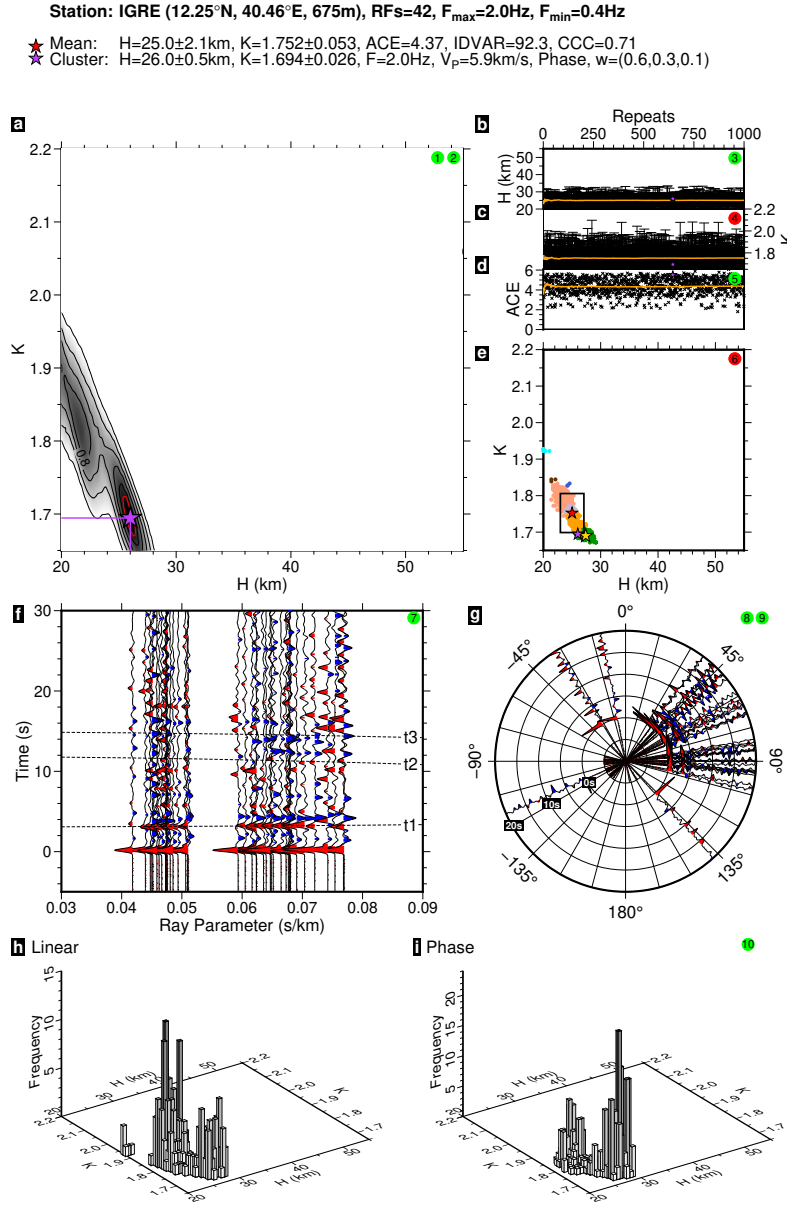


Figure B.15: H - κ result for station IGRE from the AFAR0911 deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H - κ solutions. (e) All 1000 H - κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

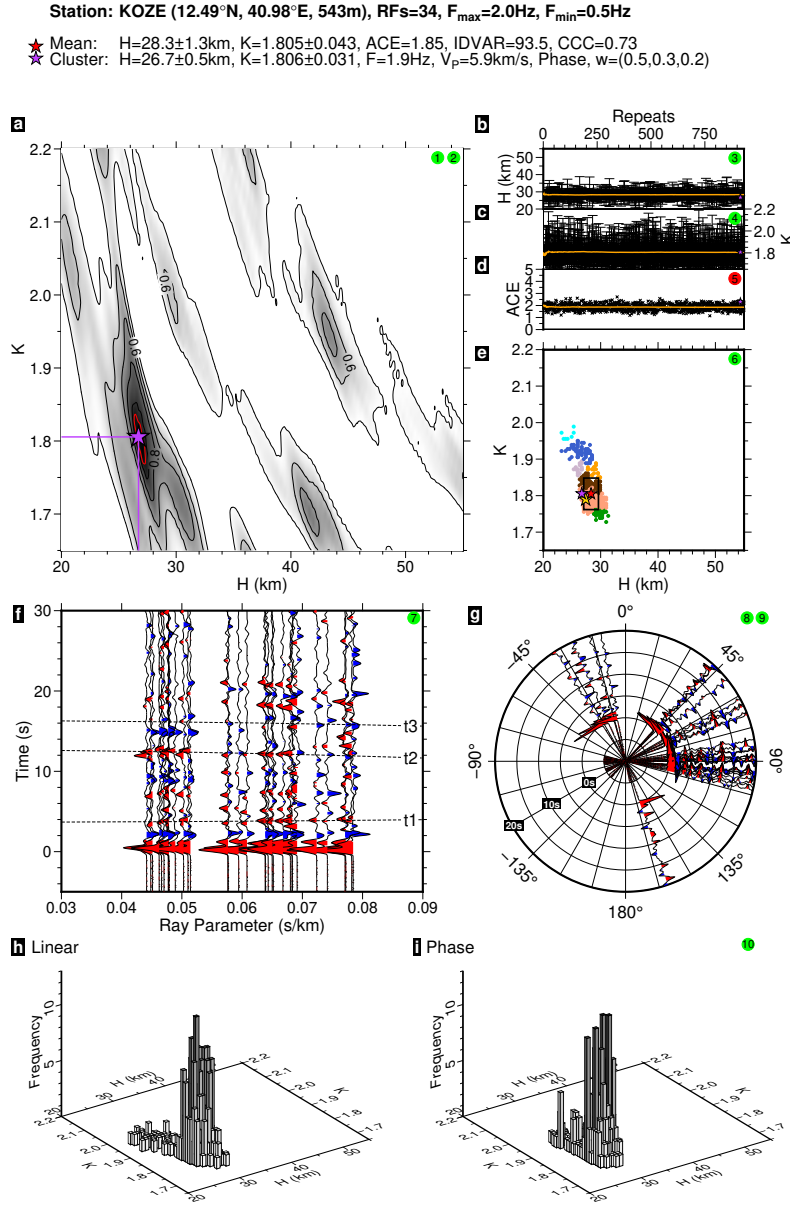


Figure B.16: H - κ result for station KOZE from the AFAR0911 deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H - κ solutions. (e) All 1000 H - κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

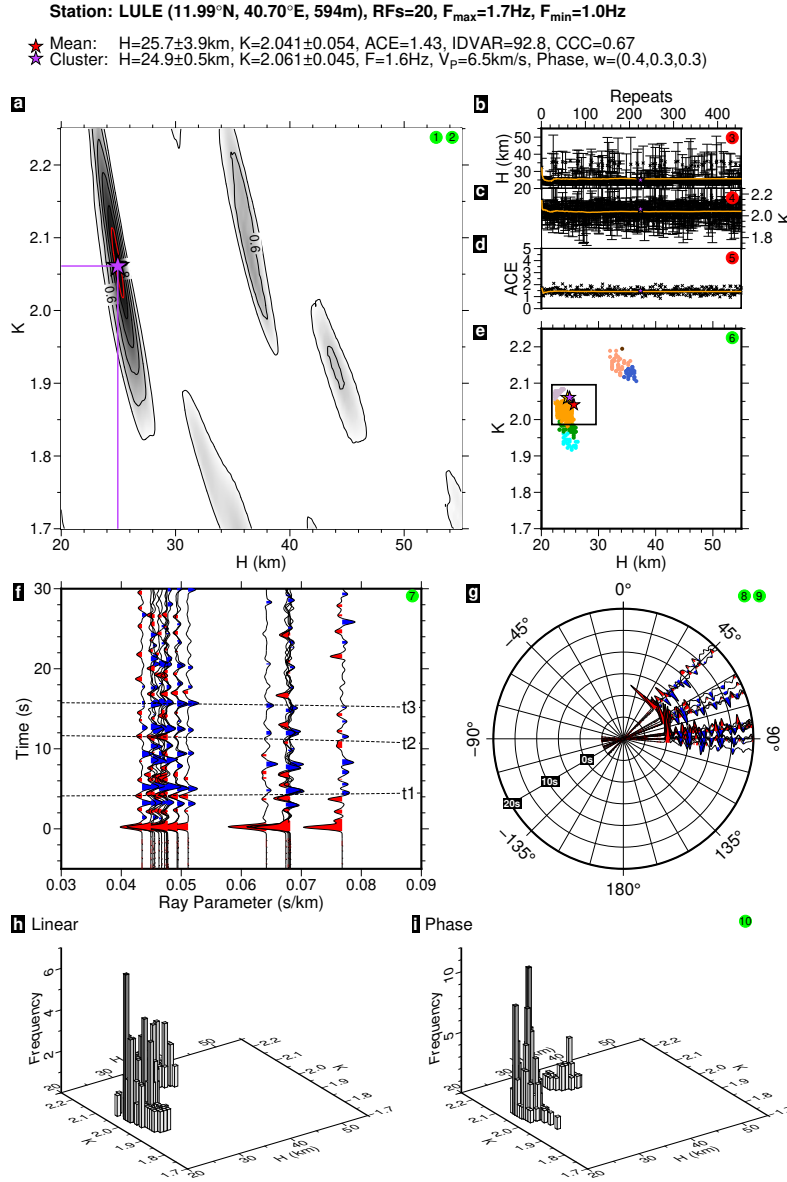


Figure B.17: H - κ result for station LULE from the AFAR0911 deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H - κ solutions. (e) All 1000 H - κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

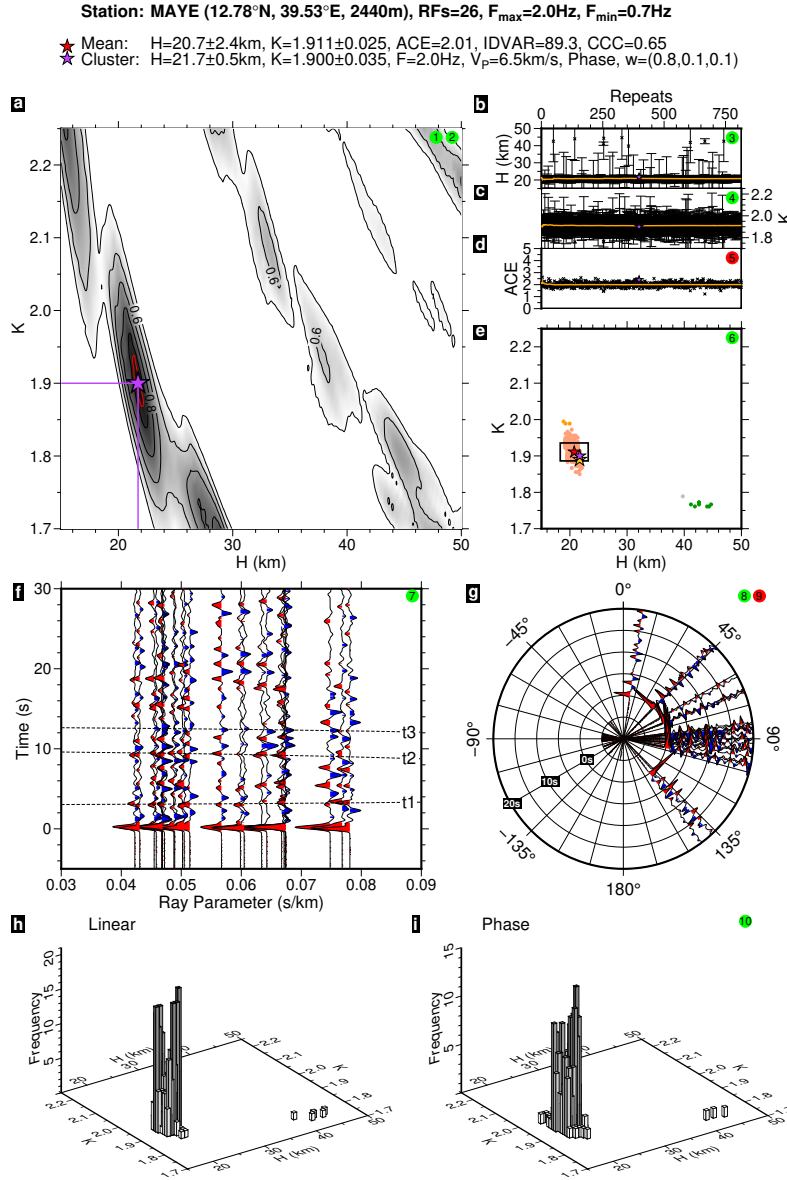


Figure B.18: H - κ result for station MAYE from the AFAR0911 deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H - κ solutions. (e) All 1000 H - κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

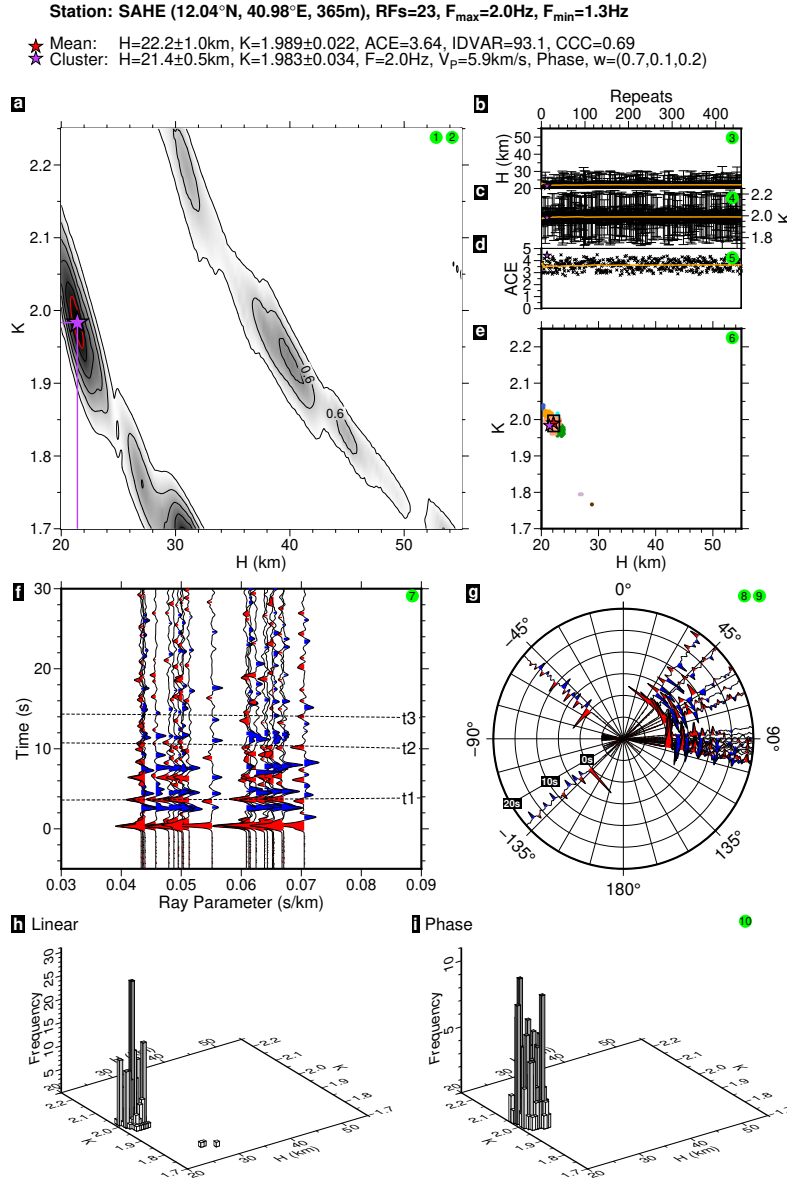


Figure B.19: H - κ result for station SAHE from the AFAR0911 deployment. Numbers mark which quality control criteria each sub-figure represents (discussed in Text S1), with red indicating the test has been failed, amber an inconclusive test result, and green indicating the test has been passed. (a) The $s(H, \kappa)$ stacking surface associated with the result selected using hierarchical cluster analysis. The (b) crustal thickness, (c) V_P/V_S ratio, and (d) ACE for each of the 1000 H - κ solutions. (e) All 1000 H - κ solutions, colored by the cluster that a result is assigned to. The red star represents the mean H and κ values from the 1000 repeats, the yellow star is the mode combination of H and κ , the purple star is the combination of H and κ selected by the cluster analysis (also plotted in (a)). The black box marks one standard deviation in H and κ . (f) Accepted receiver functions plotted by horizontal slowness, with t_1 , t_2 and t_3 denoting the predicted arrival times from the selected H and κ solution in (a). Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (g) Accepted receiver functions plotted by back-azimuth. Peaks and troughs are colored when their amplitude is $> 20\%$ of the P-arrival amplitude. (h) and (i) Distribution of linear and phase-weighted stacking results.

Table B.1: H- κ Results

Station	Long. (°N)	Lat. (°E)	Elev. (km)	H (km)	Error (km)	κ	Error	V_P (km s ⁻¹)	RFs
<i>This study</i>									
ASSE	13.06	42.65	0.019	24	1	1.83	0.03	5.9	14
CAYE	14.86	39.31	2.435	33	2	1.84	0.03	6.1	28
EITE (W)	15.24	38.28	2.171	35	2	1.80	0.03	6.2	9
EITE (E)	15.24	39.12	1.471	26	1	2.05	0.03	6.2	36
DOMLE	15.10	39.98	0.088	27	2	1.70	0.03	6.5	9
FAME	13.57	41.52	0.622	20	2	1.84	0.08	6.4	26
NAB	13.35	41.7	1.272	25	1	1.77	0.03	5.9	63
TIOE	14.67	40.87	0.043	19	1	2.04	0.04	6.0	20
ABAE	13.35	39.76	1.447	19	1	2.10	0.02	6.2	53
FINE	12.07	40.32	0.782	23	2	1.95	0.03	6.3	23
IGRE	12.25	40.46	0.675	25	2	1.75	0.05	5.9	42
KOZE	12.49	40.98	0.543	28	1	1.81	0.04	5.9	34
LULE	11.99	40.70	0.594	26	4	2.04	0.05	6.5	20
MAYE	12.78	39.53	2.440	21	2	1.91	0.023	6.5	26
SAHE	12.04	40.98	0.365	22	1	1.99	0.02	5.9	23
<i>Study of (Ahmed et al., 2022)</i>									
N001	13.5545	39.3077	2.148	37.2	2.0	1.73	0.04	–	45
N002	13.5453	39.4162	2.000	38.8	2.1	1.73	0.01	–	24
N003	13.5580	39.5007	2.080	37.6	0.5	1.79	0.02	–	47
N004	13.5775	39.5653	2.265	37.6	0.3	1.73	0.01	–	60
N005	13.6334	39.6677	2.270	38.6	1.6	1.73	0.04	–	36
N006	13.6509	39.7805	2.383	37.0	7.0	1.70	0.10	–	50
N007	13.6922	39.8938	0.970	24.2	1.4	1.79	0.02	–	29
N008	13.8406	40.0066	0.692	19.6	1.0	1.93	0.04	–	35
S001	11.6591	38.9328	2.904	42.8	1.7	1.79	0.01	–	23
S003	11.8511	39.3211	3.461	40.8	0.8	1.89	0.01	–	57
S004	11.8274	39.4903	2.375	37.2	4.0	1.83	0.06	–	15
S005	11.8387	39.5291	1.931	34.2	1.9	1.76	0.03	–	55
S006	11.8472	39.5874	2.057	34.0	0.4	1.76	0.02	–	58
S007	11.8478	39.6583	1.601	33.8	1.4	1.72	0.01	–	56
S008	11.8621	39.7348	1.491	28.4	0.3	1.83	0.01	–	47
S009	11.8001	39.7995	1.468	28.6	0.5	1.79	0.01	–	51
S011	11.8203	39.9164	1.075	27.6	1.4	1.79	0.01	–	58
S012	11.8717	39.9581	1.072	28.0	1.1	1.77	0.01	–	17
S013	11.8574	40.0506	0.981	27.8	1.4	1.79	0.01	–	26
S014	11.8584	40.1831	0.938	20.4	1.3	1.94	0.08	–	26
FTNE	12.0600	40.3100	0.782	21.4	0.4	1.99	0.02	–	72
S017	11.9856	40.7044	0.631	23.2	1.3	1.95	0.01	–	59
S018	11.8521	40.8044	0.441	21.2	1.0	1.91	0.01	–	15
S020	12.0019	40.9667	0.403	19.6	0.8	2.04	0.02	–	50
<i>Study of (Reed et al., 2014)</i>									
ELLE	11.258	40.378	0.672	21.720	0.110	2.092	0.008	5.10	107
AD01	11.315	40.571	0.580	18.530	0.048	2.254	0.010	4.80	64
AD02	11.349	40.688	0.523	18.040	0.097	2.082	0.010	5.10	119
AD03	11.422	40.766	0.488	18.150	0.053	2.230	0.011	4.85	190
AD04	11.537	40.841	0.483	17.733	0.082	2.400	0.016	4.65	38
AD05	11.608	40.906	0.445	20.600	0.000	2.066	0.000	5.15	287
AD06	11.675	40.940	0.460	19.544	0.194	2.050	0.033	5.20	187
AD07	11.727	40.990	0.395	20.750	0.173	1.967	0.013	5.40	42
AD08	11.792	41.028	0.391	19.800	0.221	2.076	0.016	5.10	297
AD09	11.817	41.050	0.385	17.620	0.228	2.145	0.037	5.00	100
AD10	11.900	41.137	0.370	23.290	0.120	1.891	0.010	5.65	73
AD11	11.741	41.296	0.370	23.100	0.131	1.997	0.011	5.30	42
AD13	11.958	41.310	0.352	23.140	0.361	1.958	0.016	5.45	142
AD14	11.939	41.455	0.422	23.190	0.238	2.030	0.016	5.20	47
DICE	11.914	41.574	0.464	22.070	0.392	2.116	0.023	5.05	125
AD15	11.880	41.707	0.088	22.520	0.319	2.010	0.024	5.25	20
AD16	11.820	41.753	0.118	26.243	0.181	1.944	0.011	5.45	56
AD17	11.742	41.837	0.150	24.250	0.387	2.012	0.026	5.25	106
AD18	11.908	41.791	0.452	29.770	0.189	1.887	0.019	5.70	114
AD19	12.054	41.926	0.435	30.456	0.532	1.853	0.018	5.90	121
<i>Study of (Hammond et al., 2011c)</i>									
ADYE	13.68	38.98	1.860	39	1	1.71	0.02	6.25	85
AKEE	10.88	39.16	3.235	38	4	1.91	0.06	6.25	50
DERE	11.12	39.64	2.545	32	4	2.01	0.12	6.25	90
GASE	11.68	38.92	2.972	43	6	1.85	0.11	6.25	18
GDR	12.56	37.45	2.103	39	5	1.83	0.15	6.25	7
KARE	10.44	39.93	1.774	45	2	1.78	0.05	6.25	49
LALE	12.02	39.04	2.422	38	–	1.85	–	6.25	30
SEKE	12.62	39.03	2.264	38	2	1.81	0.07	6.25	44
SMRE	13.19	39.21	1.977	39	2	1.77	0.07	6.25	72
YAYE	11.86	38.00	2.627	44	–	1.86	–	6.25	6
ALE	09.42	42.03	2.026	35	1	1.79	0.07	6.25	28
HYNE	09.31	42.09	1.982	35	1	1.77	0.01	6.25	81
BOBE	10.38	42.57	0.941	27	3	1.77	0.16	6.25	37
MIEE	09.24	40.76	1.349	32	5	2.22	0.28	6.25	41
QATE	09.40	41.47	2.148	32	6	2.27	0.21	6.25	67
ASYE	11.56	41.44	0.370	28	5	1.95	0.14	6.25	107
AWEE	12.06	40.07	0.836	25	1	1.83	0.04	6.15	78
BREE	12.17	41.19	0.576	26	1	1.84	0.02	6.15	95
CHIE	11.60	40.02	0.938	28	3	1.84	0.11	6.25	105
DAME	11.69	40.96	0.424	31	–	1.97	–	6.15	23
DICE	11.91	41.57	0.464	30	5	2.02	0.20	6.15	47
DIGE	12.32	40.27	0.675	17	1	2.26	0.06	6.15	66
ELLE	11.25	40.37	0.672	27	7	2.07	0.22	6.15	45
FINE	12.06	40.31	0.782	22	1	1.99	0.03	6.15	154
HARE	11.60	40.88	0.447	25	1	1.99	0.04	6.15	45
IGRE	12.25	40.46	0.675	23	5	1.85	0.20	6.25	45
KOBE	12.15	39.63	1.508	29	1	1.84	0.02	6.25	52
KOZE	12.49	40.98	0.543	28	2	1.84	0.07	6.15	46
LULE	11.99	40.70	0.594	24	2	1.98	0.07	6.15	39
LYDE	12.05	41.92	0.435	29	2	2.19	0.11	6.15	30
MILE	11.42	40.77	0.489	33	5	1.88	0.11	6.15	42
SEHE (SAHE)	12.04	40.97	0.358	22	1	1.98	0.02	6.15	146
TEND	11.79	41.00	0.419	25	5	2.02	0.15	6.15	114
SILE	12.40	41.18	0.472	24	1	2.00	0.03	6.15	66
SRDE	11.95	41.30	0.392	24	3	2.04	0.12	6.15	32
TRUE	12.48	40.31	0.381	23	1	1.97	0.03	6.15	115
AFME	13.20	40.85	0.046	16	1	2.06	0.02	6.25	161
HALE	13.84	40.01	0.797	20	2	1.98	0.02	6.15	17
ABAE (W)	13.35	39.35	1.447	32	6	1.82	0.20	6.25	14
WUCE (W)	11.51	39.19	1.912	33	7	1.89	0.17	6.25	9
BTIE (E)	11.19	40.29	1.659	23	3	2.2	0.14	6.25	126
MAYE (E)	12.78	39.77	2.440	21	1	1.93	0.03	6.25	80
WUCE (E)	11.51	39.92	1.912	26	8	2.11	0.25	6.25	93
ABAE (E)	13.35	39.98	1.447	19	1	2.09	0.05	6.25	93

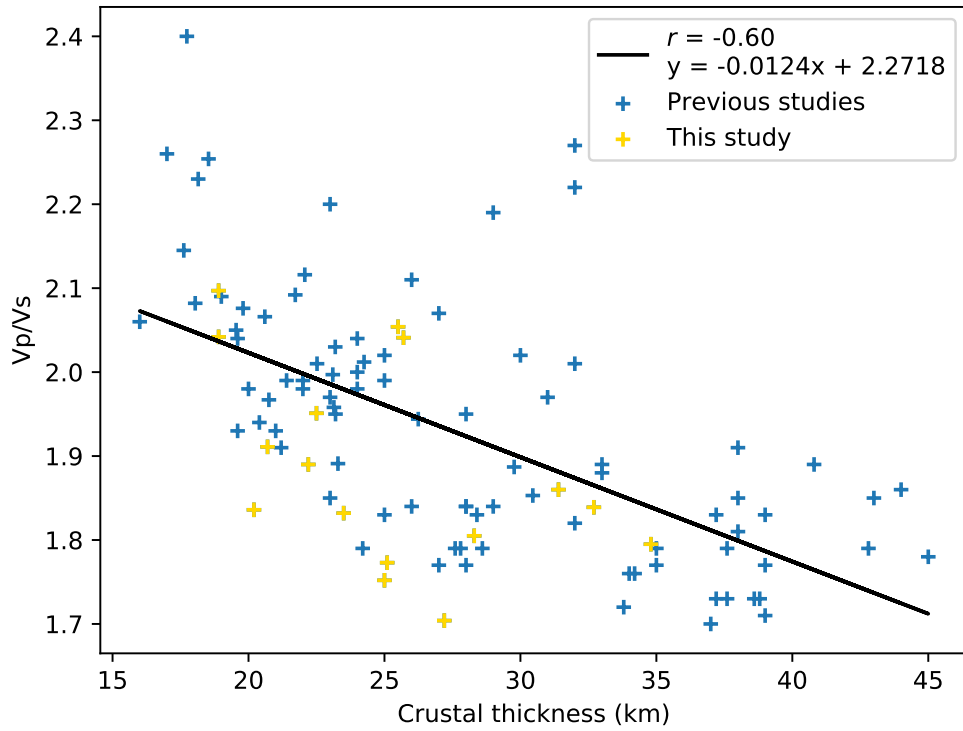


Figure B.20: Estimates of V_P/V_S ratio (κ) plotted against crustal thickness (H). Estimates calculated in this study are plotted as gold crosses; estimates from the studies of (Hammond et al., 2011c; Reed et al., 2014; Ahmed et al., 2022) are plotted as blue crosses. The black line is the optimal fit of unbiased linear regression with Pearson’s correlation coefficient $r = -0.60$.

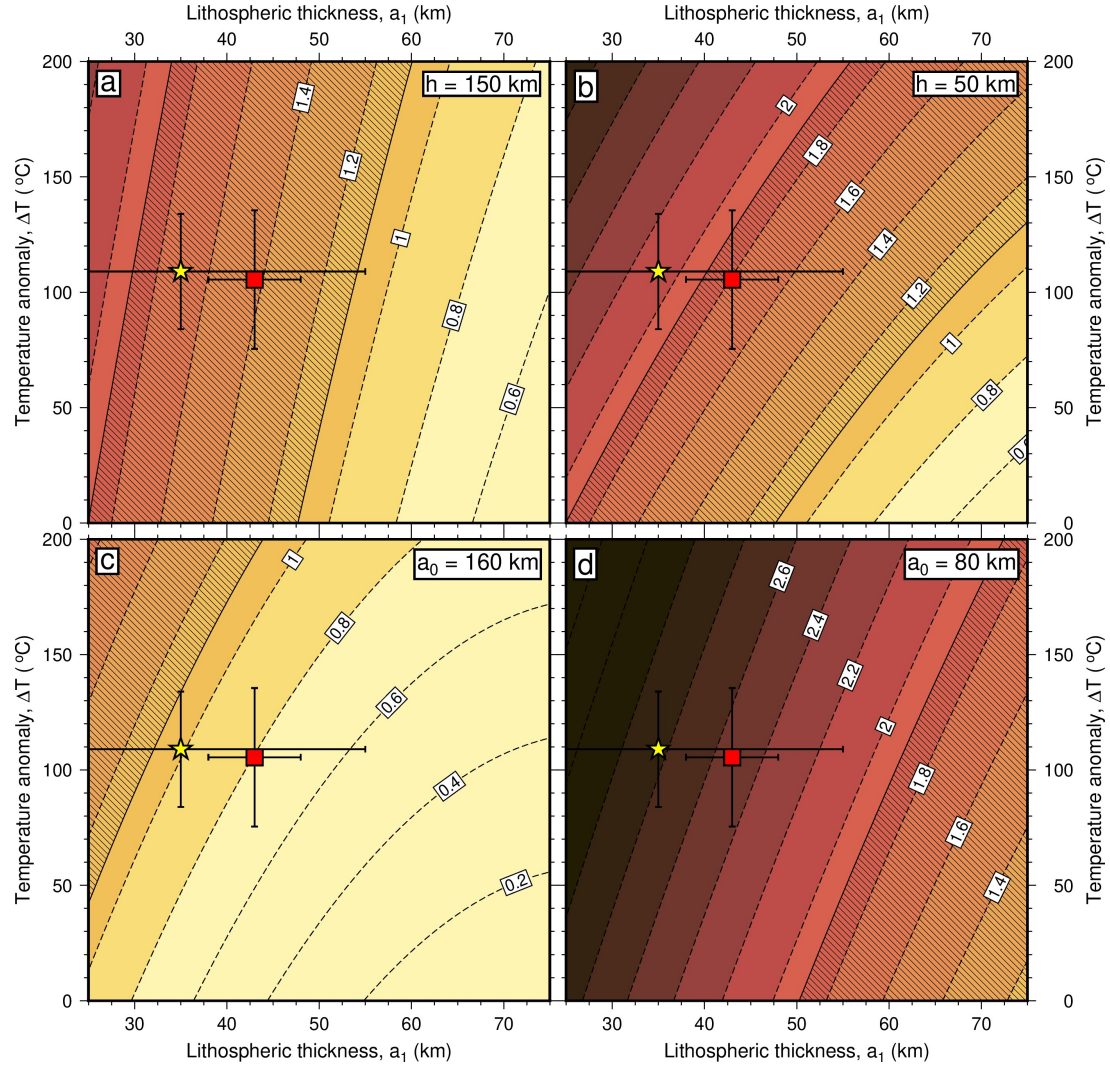


Figure B.21: Parameter sweeps for different asthenospheric channel and initial lithospheric thickness values. (a) Sub-crustally supported topography as a function of asthenospheric mantle temperature and lithospheric thickness, assuming a temperature anomaly is constrained to a 150 km thick channel and equilibrium lithospheric thickness is 120 km. Contours indicate elevation, ϵ . The shaded area represents the range of regional sub-crustal support $\Delta\epsilon = 1.5 \pm 0.4$ km estimated from the offset in intercept values between the regional crustal thickness-elevation relationship and the global one. The red square is plotted at the sub-plate temperature and lithospheric thickness given by the tomography model of Celli et al. (2020a) after conversion into temperature following the methodology of Hoggard et al. (2020) and Richards et al. (2020). The yellow star is plotted at the sub-plate temperature and lithospheric thickness given by inverse modelling of rare earth element compositions (Ball et al., 2021). (b) $h = 50$ km, $a_0 = 120$ km. (c) $h = 100$ km, $a_0 = 160$ km. (d) $h = 100$ km, $a_0 = 800$ km.

C

Supplementary information for “Seismic tomography of Nabro caldera”

Contents

C.1	Introduction	173
C.2	Text S1: Nonlinear Earthquake Relocation	174
C.3	Text S2: Formulation of the V_P/V_S Inversion	175
C.4	Text S3: Perturbation of Derived 1D Starting Model .	176
C.5	Text S4: Finding Regularisation Parameters	176
C.6	Text S5: Synthetic Resolution Tests	178

C.1 Introduction

This appendix gives further detail on the nonlinear earthquake relocation procedure and the tomographic inversion for V_P/V_S structure. It shows the results of tests perturbing the starting 1D velocity model, and explains the process of finding regularisation parameters for the tomographic inversions, including plots of the trade-off curves (Figure C.2). Figure C.3 shows the location of seismicity after inversion and earthquake relocation, demonstrating the migration of earthquakes to the inversion grid boundaries. The input and output checkerboards used to assess the resolution of the V_P and V_S solution models are plotted in Figures C.4,

C.5, C.6 and C.7. Inversion statistics for the synthetic models are given in Table C.2. Figure C.8 shows original longitude and latitude cross-sections through the final tomography model, with anomalies and earthquakes above the topography displayed. Figure C.9 plots the V_P/V_S model obtained by direct division of the final V_P model by the final V_S model, for comparison with our result from the inversion of differential S-P traveltimes. Figure C.10 describes the overall workflow. Figure C.11 presents the evolution of the mislocations of sources in the checkerboard tests.

C.2 Text S1: Nonlinear Earthquake Relocation

The objective function minimised during the nonlinear source relocation grid search is:

$$S(\mathbf{m}) = (\mathbf{d}_{obs}^r - \mathbf{d}_m^r)^T \mathbf{C}_d^{-1} (\mathbf{d}_{obs}^r - \mathbf{d}_m^r) \quad (\text{C.1})$$

where $\mathbf{C}_d^{-1}$ is the data covariance matrix, $\mathbf{d}_{obs}^r$ is a set of observed arrival times (from a given hypocenter) at a set of N stations and $\mathbf{d}_m^r$ is a set of predicted arrival times. Both the observed and predicted traveltimes have had the mean subtracted in order to eliminate the origin time as an unknown in the inversion. The origin time can be found after minimisation of $S(\mathbf{m})$ by using $\mathbf{d}_0 = \mathbf{d}_{obs} - \mathbf{d}^t$ where $\mathbf{d}^t$ are the observed traveltimes. The minimum of the objective function can be found by a simple grid search comparison of the difference between each observation and each grid point in the traveltime field for all receivers. Sub-cell discretisation based on interpolation of the traveltimes is implemented for greater accuracy (Pilia et al., 2013).

To obtain hypocenter uncertainty information, we follow Wilks et al. (2020), adopting the approach described by Sambridge and Kennett (1986). We define the 95% confidence region around the solution obtained from the grid search ($\hat{m}$) and use this boundary as a proxy for location uncertainty, assuming that data errors are Gaussian. To find this boundary, we identify the region of parameter space which satisfies:

$$T(m, \hat{m}) = \frac{S(m) - S(\hat{m})}{S(\hat{m})} \leq \frac{\chi_3^2(0.95)}{(N - 3)}, \quad (\text{C.2})$$

where N is the number of data picks, $\chi^2_3(0.95) = 7.815$ is the chi-squared statistic for a p-value of 0.05 and 3 degrees of freedom. We can formulate the problem like this since the objective function, with $[C_d^{-1}]_{ij} = \delta_{ij}\sigma_{ij}$, can be written as

$$S(m) = \sum_{i=1}^N \frac{[d_{obs}^i - d_m^i(m)]^2}{\sigma_i^2} \quad (C.3)$$

which has a chi-squared distribution with $N - M$ degrees of freedom, where $M = 3$ is the number of model parameters. This equation accounts for the effects of errors in the velocity model by rescaling S so that $S(\hat{m}) = N - 3$ (the expectation value of a chi-squared distribution with $N - 3$ degrees of freedom). This is achieved by multiplication by the factor $(N - 3)/S(\hat{m})$.

C.3 Text S2: Formulation of the V_P/V_S Inversion

The inversion for V_P/V_S structure is a linear inversion performed along the S-wave ray paths. It is assumed that 1) each S-wave path between two points has a corresponding P-wave path; 2) the P- and S-wave paths taken between two different points are identical and have similar Fresnel zones, meaning the reference V_P/V_S model is constant (Walck, 1988; Thurber, 1993; Eberhart-Phillips and Reyners, 2012).

Pilia et al. (2013) formulate the problem as follows:

$$\delta \left(t_s - \left(\frac{V_P}{V_S} \right)^0 t_p \right) = \int_{L_p} \frac{1}{V_P^0} \delta \left(\frac{V_P}{V_S} \right) dl \quad (C.4)$$

where t is the traveltimes for P- or S-waves along the S-wave path L_s , v is the velocity, l is the path length and $(V_P/V_S)^0$ is a reference model. This allows for V_P/V_S to be calculated without prior knowledge of V_P or V_S .

In this study, we use the final hypocenter locations determined by the “joint” inversion for V_P , V_S and earthquake location. We calculate the average ratio of the V_P and V_S models across all velocity grid nodes, which is 1.77, and use this as the reference V_P/V_S model, assumed constant throughout the model.

C.4 Text S3: Perturbation of Derived 1D Starting Model

To test the dependence of the final model on the 1D starting model, we perturb the velocity of our 1D model at each depth randomly by $\pm 10\%$ and repeat this 10 times. The V_P/V_S model produced by three different examples of this test are plotted in Figure C.1. The recovered pattern of anomalies are similar to those produced by the original model within the data resolution limits.

The initial and traveltime residuals for each of the perturbed models as compared to the original model are given in Table C.1.

C.5 Text S4: Finding Regularisation Parameters

The data fit of a solution model is defined as the variance of the traveltime residuals. The model variance σ^2 is a measure of how close the solution model ($\mathbf{m}$) is to the starting model ($\mathbf{m}_0$):

$$\sigma^2 = \sum_M^{i=1} \frac{(m^i - m_0^i)^2}{M}, \quad (\text{C.5})$$

where i is each velocity node of the grid and M is the number of parameters in the model. Model roughness is a finite difference measure of the second derivative, averaged across the entire model in 3-D. It is calculated for depth as follows:

$$P(m) = \frac{(m_{i+1,j,k} - 2m_{i,j,k} + m_{i-1,j,k})}{h^2}, \quad (\text{C.6})$$

where P refers to variations in depth and h is the separation between velocity nodes. The same second derivative calculation is carried out in depth, latitude and longitude and averaged to give a final estimate for each velocity node. An average is then taken across the final model to give the overall model roughness.

There exists a trade-off between data fit, model variance and model roughness—for example, a low value of model roughness is desirable, but this comes at the expense of the model fitting the data less well. Therefore, trade-off curves are used to find the optimum values of the damping and smoothing parameters. Fixing the

damping and varying the smoothing parameter leads to a trade-off between the data fit and the solution model roughness, whilst fixing the smoothing parameter and varying the damping means a trade-off between the data fit and the solution model variance.

Figure C.2 shows the results of numerous inversions for 3D velocity structure using different values of the damping and smoothing parameters. For the V_P inversions, we first keep the smoothing parameter, η , constant at 10, and vary the damping parameter, ϵ between 0.5 and 500, as shown in Figure C.2a. In order to minimise the variance of the traveltime residuals whilst also keeping the model variance as low as possible, we select $\epsilon = 3$ as the best compromise between data fit and model variance. Next ϵ is held fixed at 3 and η is varied between 2 and 500, as plotted in Figure C.2b. We conclude that $\eta = 10$ is still an acceptable compromise between optimising the data fit and reducing the solution model roughness. The regularisation parameters for the V_P inversions are therefore $\epsilon = 3$ and $\eta = 10$.

For the V_S inversions, we again begin with $\eta = 10$ and vary ϵ between 1 and 500 (Figure C.2c). We choose $\epsilon = 10$ as the value of the damping parameter that gives the best compromise between data fit and model variance. Then ϵ is held fixed at 50 and η is varied between 5 and 500, as shown in Figure C.2b. We conclude that $\eta = 50$ gives the best compromise between minimising the variance of the traveltime residuals and reducing the solution model roughness. The damping and smoothing parameters affect both model variance and roughness, so another trade-off curve is plotted with $\eta = 50$ to make sure that $\epsilon = 10$ produces a good compromise between data fit and model variance with this value of smoothing. Figure C.2e shows that $\epsilon = 10$ is still a good choice of smoothing parameter, and the process does not need to go through subsequent iterations. The regularisation parameters for the V_S inversions are $\epsilon = 10$ and $\eta = 50$.

These regularisation parameters are used in the individual P- and S-wave velocity inversions before we vary the regularisation parameters for the direct V_P/V_S inversion, as shown in Figures C.2f–h.

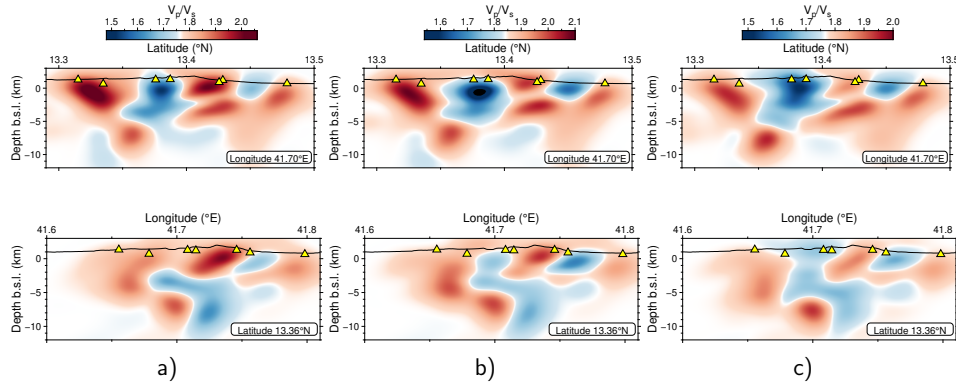


Figure C.1: Cross-sections along a longitude of 41.7°E and latitude of 13.36°N through the V_P/V_S ratio velocity models, each using a starting model that has been randomly perturbed by $\pm 10\%$ from the original 1D starting model. The V_P/V_S ratio model is plotted as absolute values, with the centre of the colour bar corresponding to the reference V_P/V_S value—the average ratio of the corresponding V_P and V_S models across all velocity grid nodes. Inversion statistics are tabulated in Table A.1.

C.6 Text S5: Synthetic Resolution Tests

Figures C.4, C.5, C.6 and C.7 show input and output checkerboards used to assess the resolution of the V_P and V_S solution models respectively. The formulation of these synthetic tests is discussed in the main text in Section 4.1. The anomalies for the 8 km and 4 km checkerboards are generally well-reconstructed within the bounds of the seismic array and to depths of 10 km b.s.l.. Recovery tends to be poor outside of this region. The recovered resolution is similar for the V_P and V_S models.

Table A.2 gives the inversion statistics for the synthetic models.

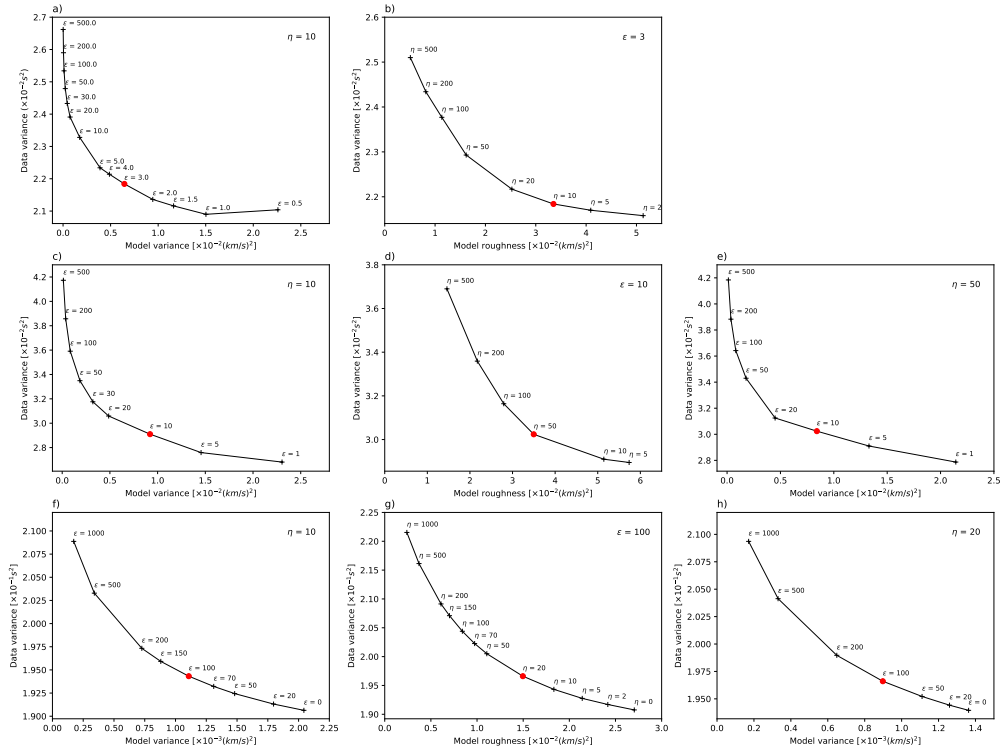


Figure C.2: Trade-off curves used to find the optimum values of the regularisation parameters, damping (ϵ) and smoothing (η), for V_P , V_S and V_P/V_S inversions. Red circles indicate the ‘best’ value of ϵ or η selected to be held fixed during the next set of inversions. (a) V_P inversion: varying the damping parameter while keeping the smoothing parameter fixed at $\eta = 10$. (b) V_P inversion: varying the smoothing parameter with the damping parameter fixed at $\epsilon = 3$. (c) V_S inversion: varying the damping parameter with the smoothing parameter held fixed at $\eta = 10$. (d) V_S inversion: varying the smoothing parameter with the damping parameter fixed at $\epsilon = 50$. (e) V_S inversion: varying the damping parameter with the smoothing parameter fixed at the best compromise value of $\eta = 50$. (f) V_P/V_S inversion: varying the smoothing parameter with $\eta = 10$. (g) V_P/V_S inversion: varying the damping parameter with $\epsilon = 100$. (h) V_P/V_S inversion: varying the smoothing parameter with $\eta = 20$.

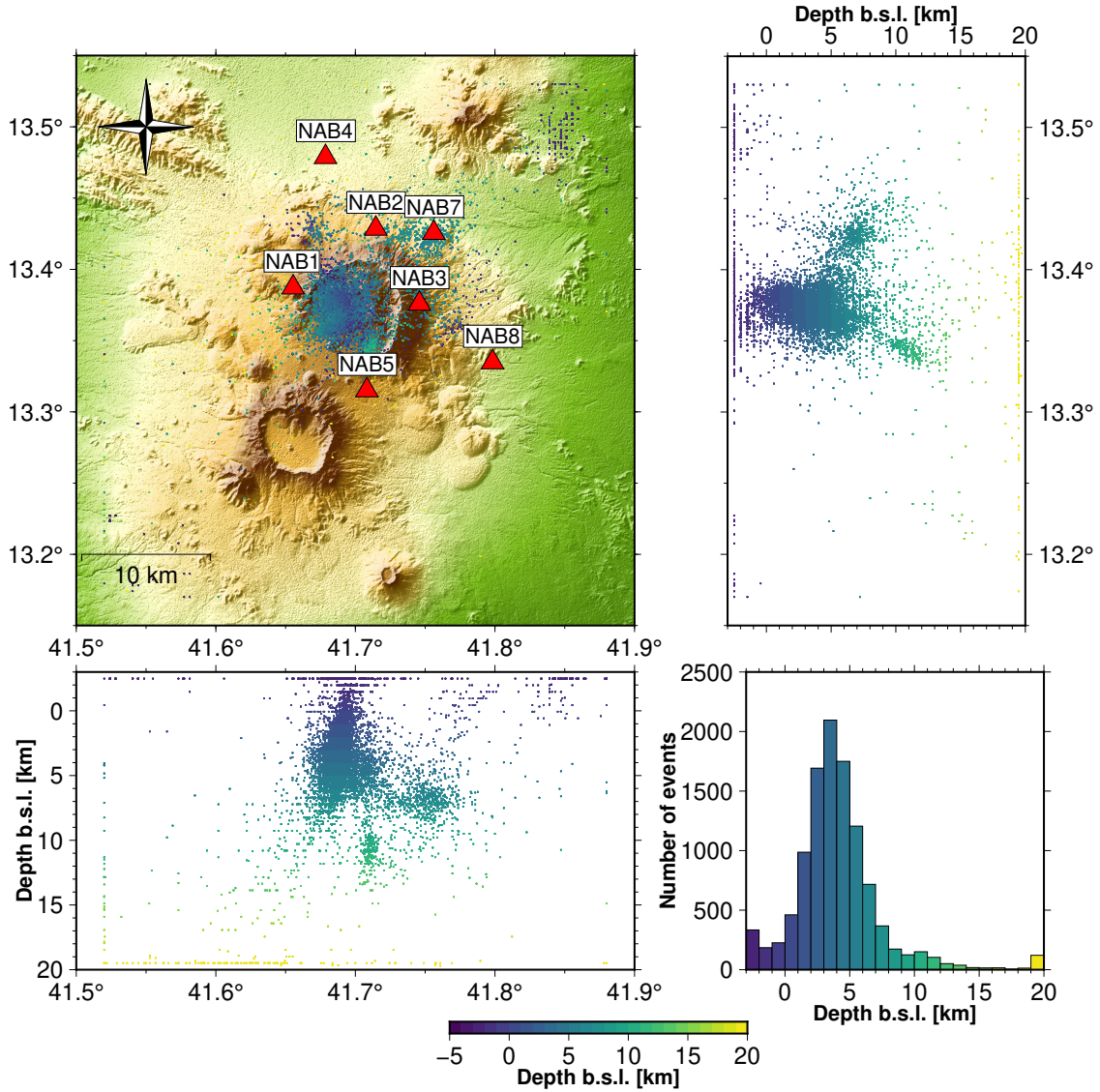


Figure C.3: The event relocations following six iterations of the FMTOMO inversion procedure plotted over a topographic map of Nabro volcano. The locations of the earthquakes often move considerable distances, and a considerable number of earthquakes migrate to the inversion grid boundaries, indicating that the initial locations of these earthquakes are poorly constrained and will be removed from the initial catalogue. Seismic stations from the temporary seismic network are plotted as red triangles, excluding the inoperational station (NAB6). The seismicity is coloured by depth below sea level (b.s.l.) and the histogram uses bins of 1 km depth b.s.l..

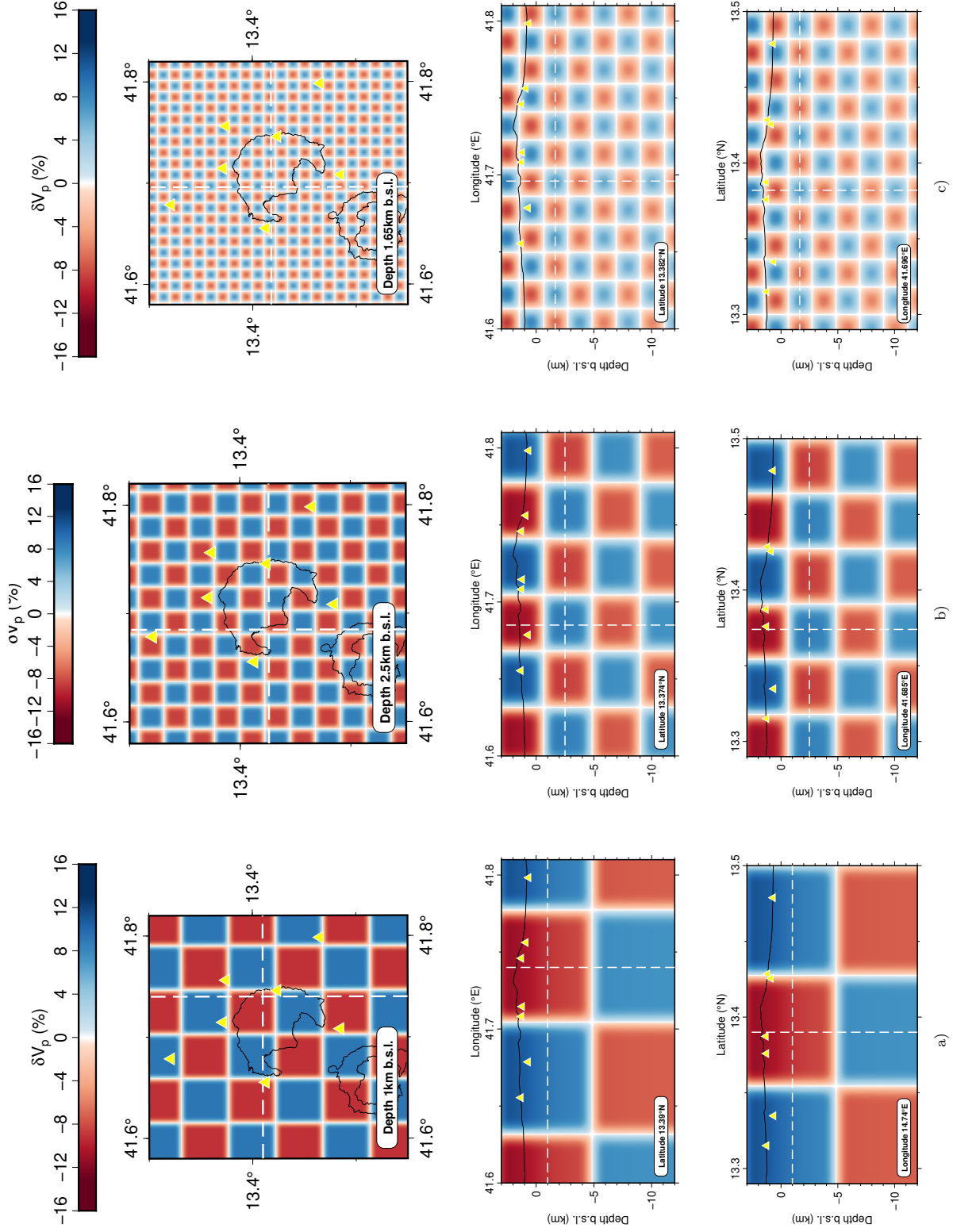


Figure C.4: Input V_p checkerboard models of differing scale lengths: a) ~ 8 km; b) ~ 4 km; c) ~ 2 km. Slices are taken through the maximum perturbations of each checkerboard in latitude, longitude and depth, as indicated by the white dashed lines.

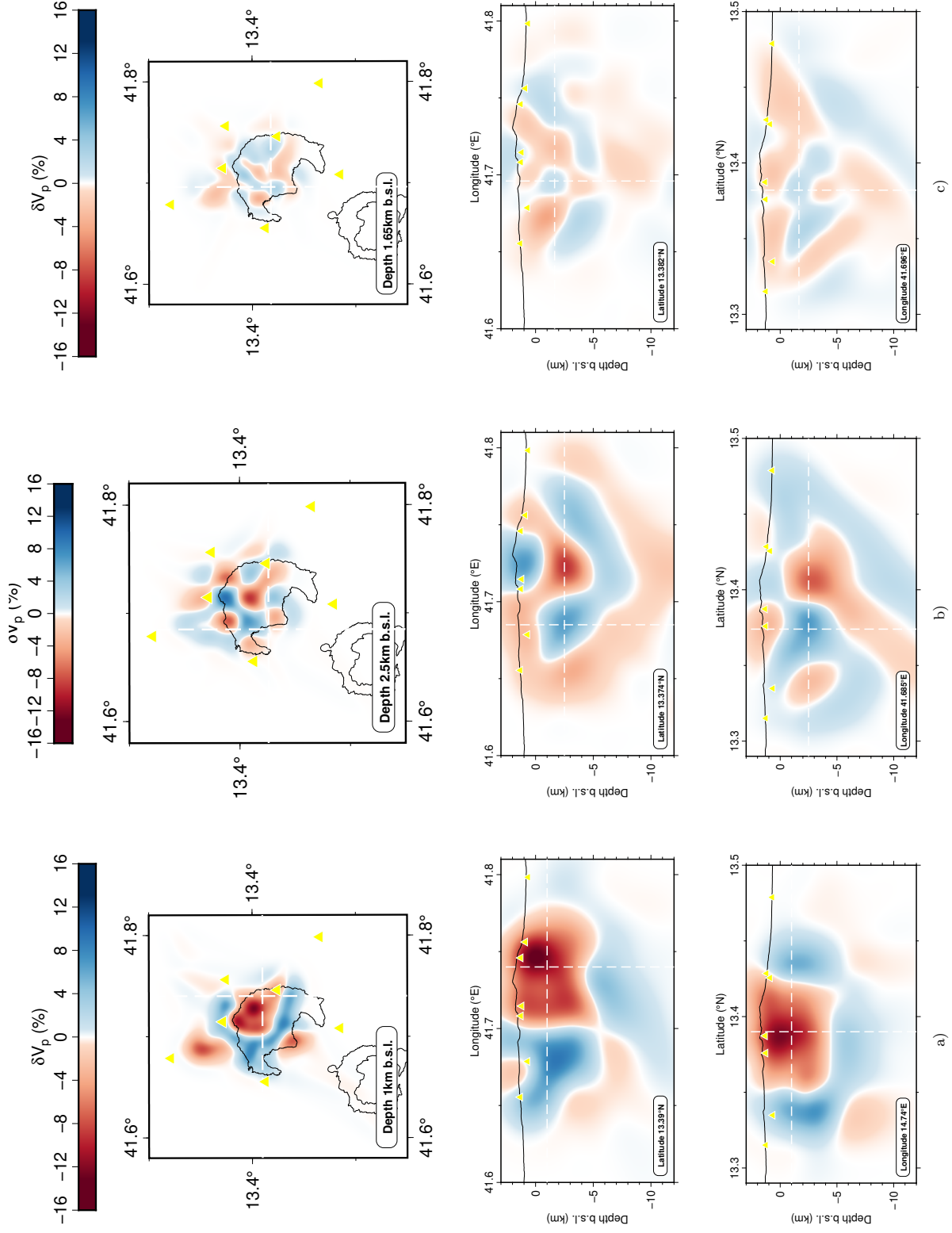


Figure C.5: Output V_p checkerboard models of differing scale lengths: a) ~ 8 km; b) ~ 4 km; c) ~ 2 km. Slices are taken through the maximum perturbations of each checkerboard in latitude, longitude and depth, as indicated by the white dashed lines.

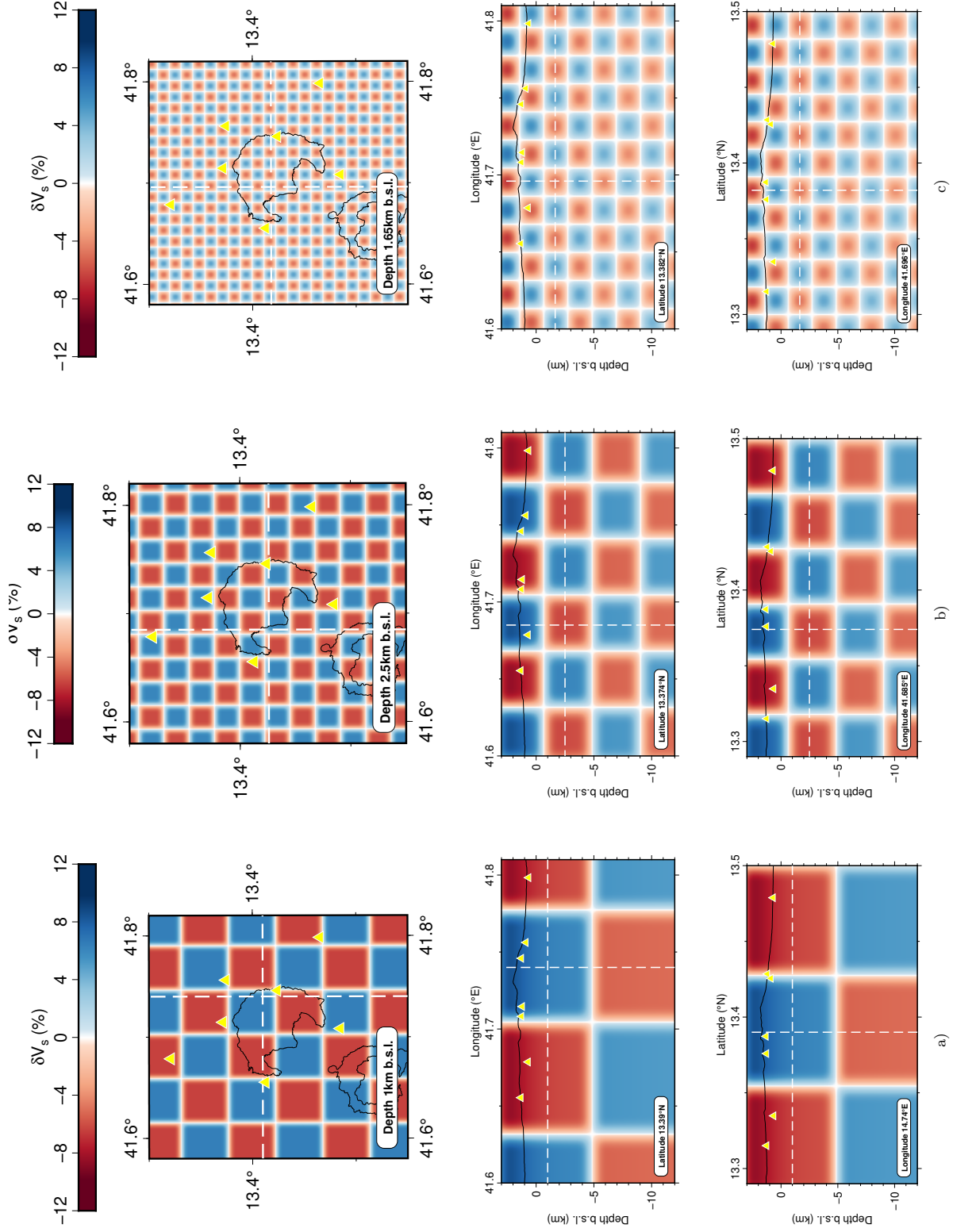


Figure C.6: Input V_s checkerboard models of differing scale lengths: a) ~ 8 km; b) ~ 4 km; c) ~ 2 km. Slices are taken through the maximum perturbations of each checkerboard in latitude, longitude and depth, as indicated by the white dashed lines.

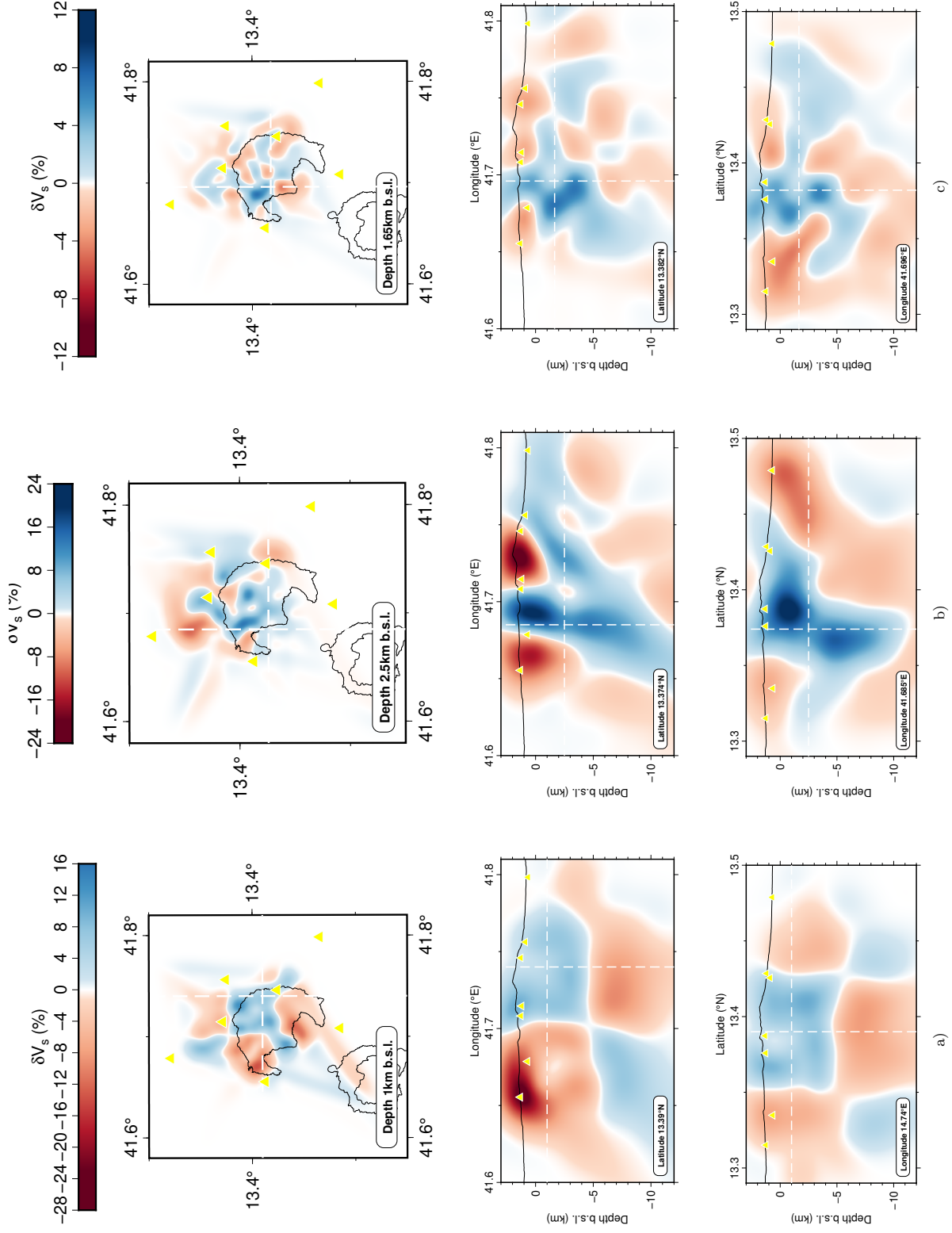


Figure C.7: Output V_s checkerboard models of differing scale lengths: a) ~ 8 km; b) ~ 4 km; c) ~ 2 km. Slices are taken through the maximum perturbations of each checkerboard in latitude, longitude and depth, as indicated by the white dashed lines.

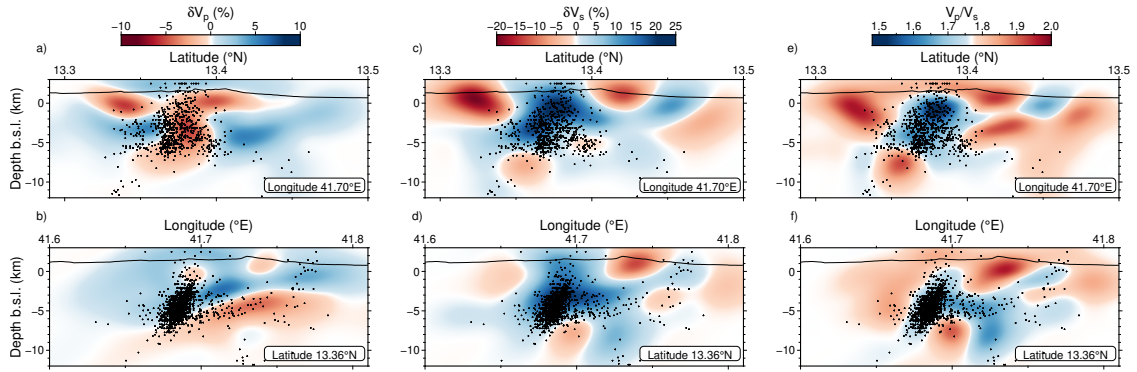


Figure C.8: Cross-sections through the final V_P (a, d), V_S (b, e) and V_P/V_S (c, f) solution models at a longitude of 41.7°E and latitude of 13.36°N (the center of the caldera). The V_P and V_S models are plotted as percentage deviations from the initial 1D model. The V_P/V_S ratio model is plotted as absolute values, with the center of the colour bar corresponding to the reference V_P/V_S value—the average ratio of the V_P and V_S models across all velocity grid nodes. Earthquakes within ± 1 km of the displayed section are indicated by black dots.

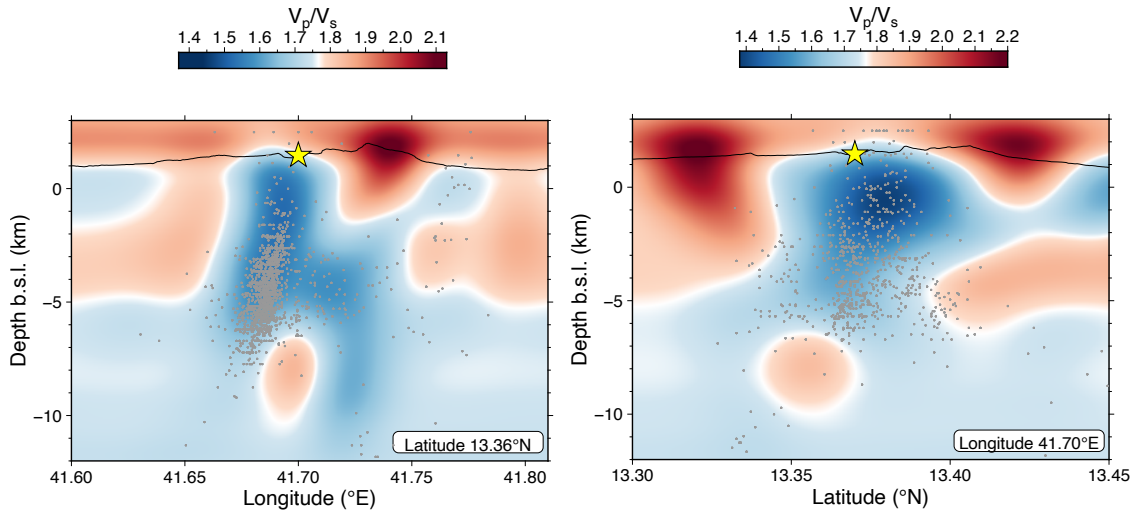


Figure C.9: Cross-sections through the V_P/V_S model obtained through direct division of the final V_P model by the final V_S model at a latitude of 13.36°N (left) and longitude of 41.7°E (right). The V_P/V_S ratio model is plotted as absolute values, with the center of the colour bar corresponding to the reference V_P/V_S value—the average ratio of the V_P and V_S models across all velocity grid nodes. Note the slightly different colour scale to the models plotted in Figure 5 and Figure 6, to account for the larger amplitude range. Earthquakes within ± 1 km of the displayed section are indicated by black dots and the yellow star marks the location of the vent from the 2011 eruption.

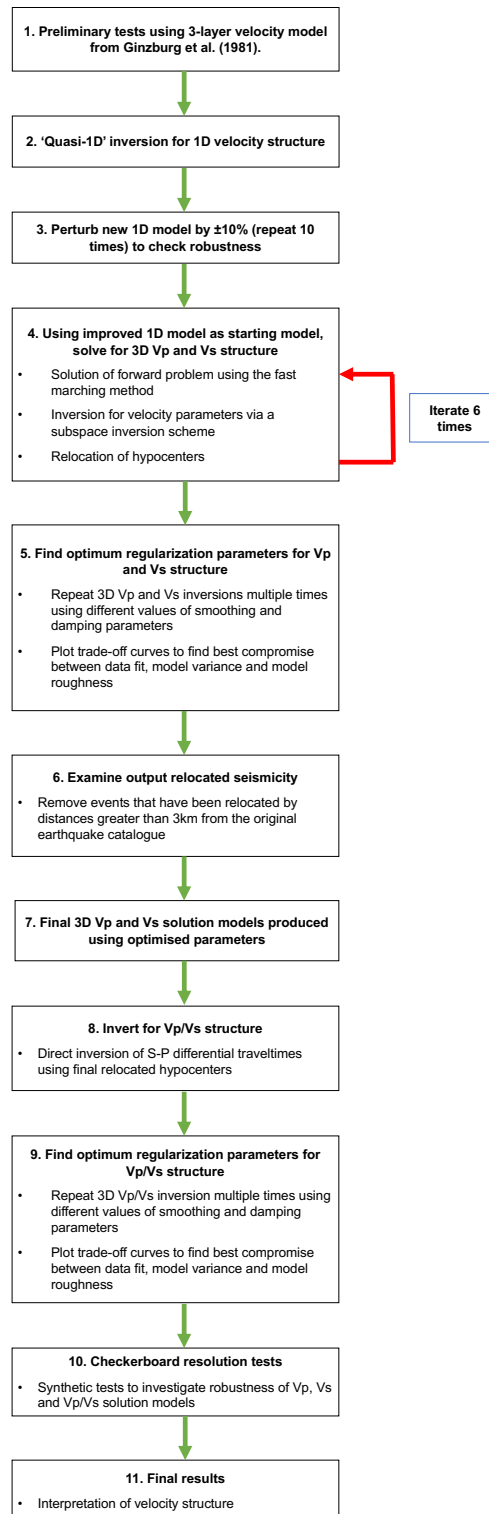


Figure C.10: The workflow followed in this paper to jointly invert for V_P , V_S and V_P/V_S structure and earthquake locations beneath Nabro volcano.

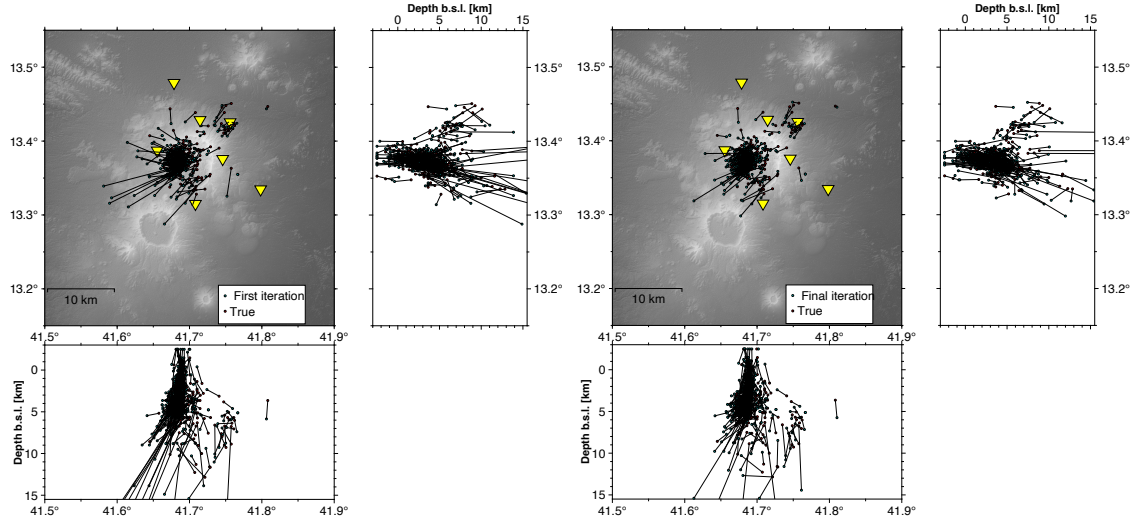


Figure C.11: Source mislocations during the synthetic checkerboard tests. A selection of 392 earthquakes from the overall catalogue is presented for ease of viewing. The red dots indicate the true locations determined in the data inversion, joined by black lines to the current locations (cyan dots). On the left are the location results after one relocation using the starting 1D velocity model, and on the right are the final locations after six iterations.

Table C.1: Inversion statistics for the V_P/V_S inversions using the perturbed 1D models as starting models, as compared to the original model.

RMS residuals			
1D Model	Starting model	Solution model	% difference from original model
Original	0.192	0.168	0
Perturbed (a)	0.223	0.173	2.98
Perturbed (b)	0.216	0.181	7.73
Perturbed (c)	0.193	0.168	0
Variance			
1D model	Starting model	Solution model	% difference from original model
Original	3.64E-2	2.83E-2	0
Perturbed (a)	3.73E-2	3.00E-2	6.00
Perturbed (b)	4.62E-2	3.27E-2	15.5
Perturbed (c)	3.71E-2	2.83E-2	0

Table C.2: Inversion Statistics for the Synthetic V_P , V_S and V_P/V_S Checkerboard Tests

RMS residuals				
Scale length	Parameter	Starting model	Solution model	Reduction (%)
2 km	V_P	0.292 s	0.0129 s	95.6
	V_S	0.497 s	0.029 s	94.2
	V_P/V_S	0.0625	0.0359	42.6
4 km	V_P	0.294 s	0.0140 s	95.2
	V_S	0.490 s	0.0319 s	93.5
	V_P/V_S	0.134	0.0403	69.9
8 km	V_P	0.307 s	0.0150 s	95.1
	V_S	0.491 s	0.0365 s	92.6
	V_P/V_S	0.225	0.042	81.3
Variance				
Scale length	Parameter	Starting model	Solution model	Reduction (%)
2 km	V_P	0.0854 s ²	0.00017 s ²	99.8
	V_S	0.248 s ²	0.00084 s ²	99.7
	V_P/V_S	0.00389	0.00129	66.8
4 km	V_P	0.0879 s ²	0.0002 s ²	99.8
	V_S	0.241 s ²	0.00102 s ²	99.6
	V_P/V_S	0.00180	0.000163	90.9
8 km	V_P	0.0943 s ²	0.00022 s ²	99.8
	V_S	0.242 s ²	0.00133 s ²	99.5
	V_P/V_S	0.0502	0.00176	96.5
χ^2				
Scale length	Parameter	Starting model	Solution model	Reduction (%)
2 km	V_P	213	0.417	99.8
	V_S	619	2.11	99.7
4 km	V_P	217	0.502	99.8
	V_S	601	2.54	99.6
8 km	V_P	236	0.559	99.8
	V_S	604	3.33	99.4

D

Differential Ps times from receiver functions

Contents

D.1 Introduction	189
D.2 Plotting differential Ps times at Nabro stations	190

D.1 Introduction

The differential time between direct P-waves and Ps converted waves can be used to examine crustal velocity structure (Janiszewski et al., 2020). Janiszewski et al. (2020) observe a systematic variation in differential time at stations around Cleveland volcano, with early Ps arrivals seen for backazimuths directed away from the edifice, and late arrivals directed towards the edifice. This variability implies either a sharp change in Moho depth over a short distance, or a different shear wave velocity directly beneath the edifice. The authors then use a forward modelling approach to delineate a columnar low velocity zone beneath the volcano, extending to mid-crustal depths (~ 10 – 20 km depth). Their technique is independent of deep crustal seismic activity and provides a novel approach to imaging crustal magma storage regions.

D.2 Plotting differential Ps times at Nabro stations

I attempt to carry out the same method at Nabro volcano. I calculate teleseismic receiver functions at each station in the Nabro Urgency Array network. Descriptions of the receiver function method and of the seismic network can be found in Chapter 3 and Chapter 4 respectively. For each station, I plot the receiver functions as a function of their backazimuth (Figure D.1). I do not observe an obvious systematic variation in Ps arrivals for any station, but nevertheless I proceed to calculate the Ps time residuals. I use the predicted Ps arrival time from the stacked receiver functions used to calculate H- κ at the composite NAB station (including all the Nabro Urgency Array stations). See Chapter 3 and Appendix B for further detail. Ps arrival time varies with slowness (the reciprocal of seismic velocity), which in turn varies from station to station, and so I extract the predicted Ps arrival times for a variety of slowness values. For every station, I subtract the Ps arrival time observed in the receiver function from the predicted Ps time at the correct slowness value. The results are plotted in Figure D.2. Few ray paths traverse the volcanic edifice, and no systematic variation in Ps differential time is observed.

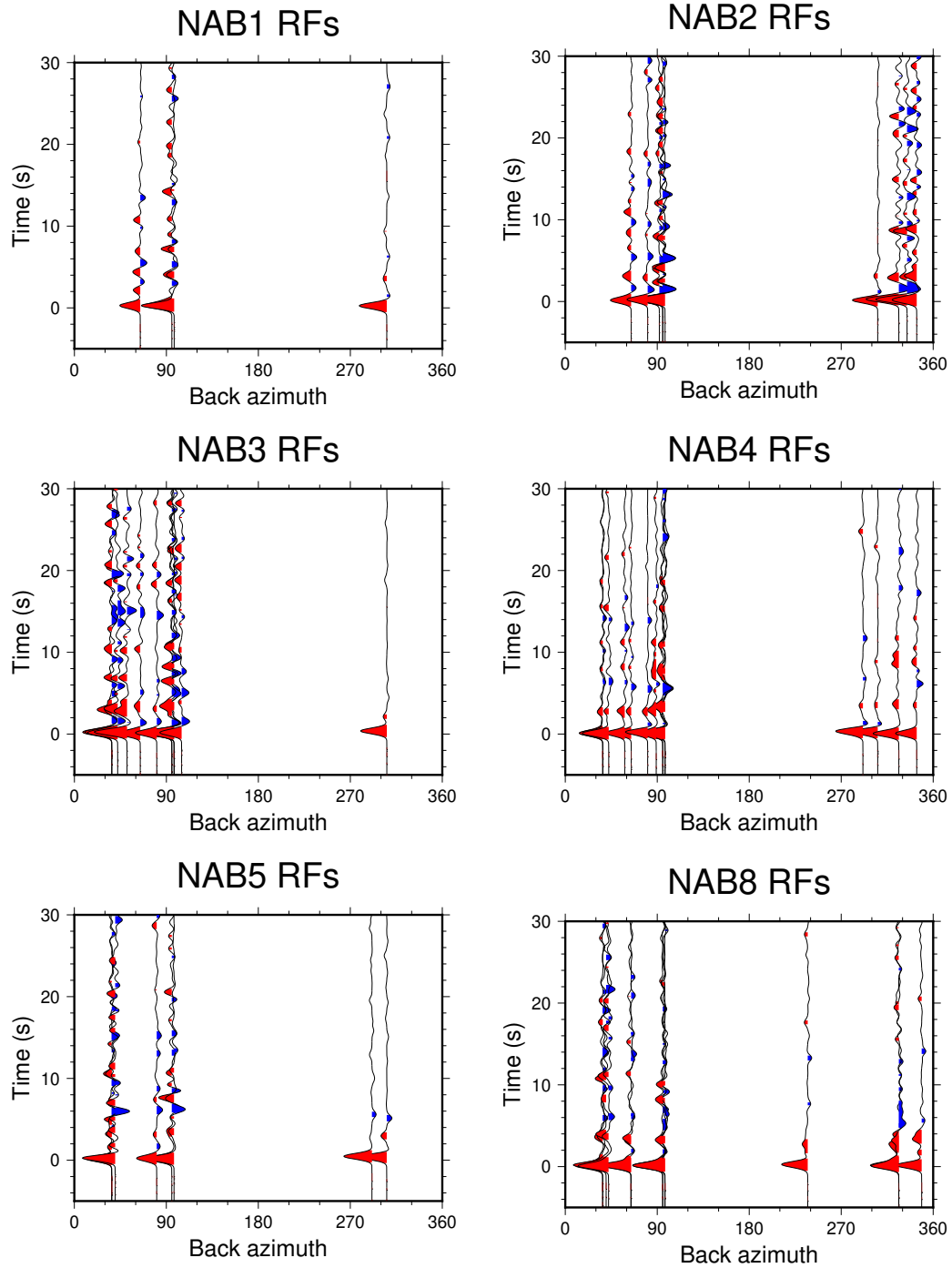


Figure D.1: Composite plot of receiver functions at each Nabro station, plotted against event backazimuth. Positive amplitude arrivals are coloured in red, and negative amplitude arrivals are coloured in blue. See Chapter 3 for more detail on receiver function calculation.

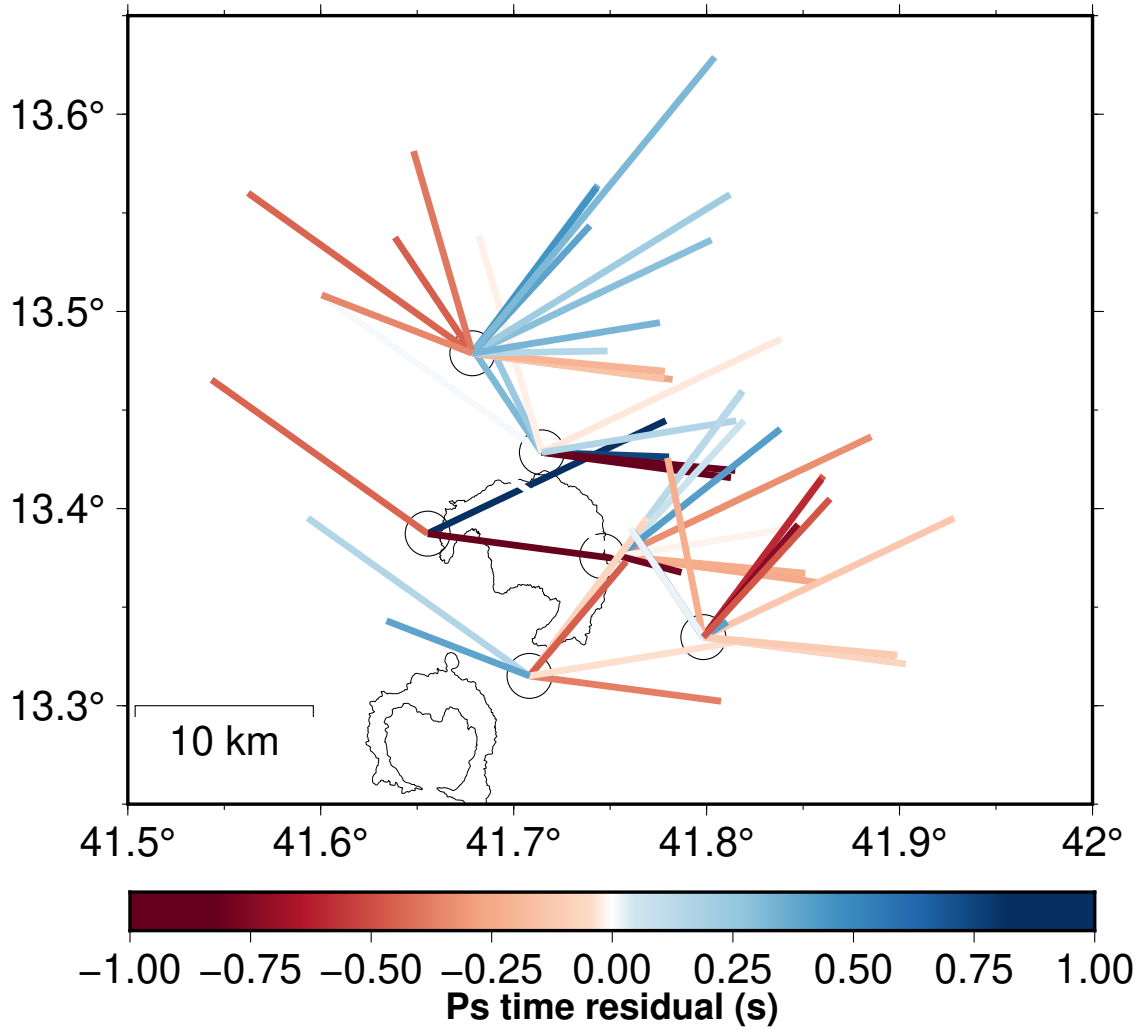


Figure D.2: Plot of differential Ps times around Nabro volcano. The stations from the Nabro Urgency Array network are plotted as black circles. The caldera is outlined in black. Bars are plotted from each station to the pierce point of the P phase, and are coloured by the time residual between the observed and predicted Ps arrival time.

Bibliography

- Afanasyev, A., Blundy, J., Melnik, O., and Sparks, S. (2018). Formation of magmatic brine lenses via focussed fluid-flow beneath volcanoes. *Earth and Planetary Science Letters*, 486, 119–128.
- Ahmed, A., Doubre, C., Leroy, S., Keir, D., Pagli, C., Hammond, J. O., Ayele, A., de Berc, M. B., Grunberg, M., Vergne, J., et al. (2022). Across and along-strike crustal structure variations of the western Afar margin and adjacent plateau: Insights from receiver functions analysis. *Journal of African Earth Sciences*, 192, 104570.
- Airy, G. B. (1855). III. On the computation of the effect of the attraction of mountain-masses, as disturbing the apparent astronomical latitude of stations in geodetic surveys. *Philosophical Transactions of the Royal Society of London*, (145), 101–104.
- Al-Lazki, A., Ebinger, C., Kendall, M., Helffrich, G., Leroy, S., Tiberi, C., Stuart, G., and Al-Toobi, K. (2013). Upper mantle anisotropy of southeast Arabia passive margin [Gulf of Aden northern conjugate margin], Oman. In *Lithosphere Dynamics and Sedimentary Basins: The Arabian Plate and Analogues*, (pp. 429–438). Springer.
- Ali, M. Y., Ismaiel, M., Yusuf, I. M., and Kaviani, A. (2023). Crustal structure and seismic anisotropy of rift basins in Somaliland. *Scientific Reports*, 13(1), 17483.
- Aloisi, M., Cocina, O., Neri, G., Orecchio, B., and Privitera, E. (2002). Seismic tomography of the crust underneath the Etna volcano, Sicily. *Physics of the Earth and Planetary Interiors*, 134(3-4), 139–155.
- Amalokwu, K., Best, A. I., Sothcott, J., Chapman, M., Minshall, T., and Li, X.-Y. (2014). Water saturation effects on elastic wave attenuation in porous rocks with aligned fractures. *Geophysical Journal International*, 197(2), 943–947.
- Angus, D., Kendall, J.-M., Wilson, D., White, D., Sol, S., and Thomson, C. (2009). Stratigraphy of the Archean western Superior Province from P- and S-wave receiver functions: Further evidence for tectonic accretion? *Physics of the Earth and Planetary Interiors*, 177(3-4), 206–216.
- Armitage, J. J., Ferguson, D. J., Goes, S., Hammond, J. O., Calais, E., Rychert, C. A., and Harmon, N. (2015). Upper mantle temperature and the onset of extension and break-up in Afar, Africa. *Earth and Planetary Science Letters*, 418, 78–90.
- Aster, R., and Meyer, R. (1988). Three-dimensional velocity structure and hypocenter distribution in the Campi Flegrei caldera, Italy. *Tectonophysics*, 149(3-4), 195–218.

- Audin, L., Manighetti, I., Tapponnier, P., Métivier, F., Jacques, E., and Huchon, P. (2001). Fault propagation and climatic control of sedimentation on the Ghoubbet rift floor: Insights from the Tadjouraden cruise in the western Gulf of Aden. *Geophysical Journal International*, 144(2), 391–413.
- Ayele, A., Stuart, G., and Kendall, J.-M. (2004). Insights into rifting from shear wave splitting and receiver functions: An example from Ethiopia. *Geophysical Journal International*, 157(1), 354–362.
- Bakker, R. R., Fazio, M., Benson, P. M., Hess, K.-U., and Dingwell, D. B. (2016). The propagation and seismicity of dyke injection, new experimental evidence. *Geophysical Research Letters*, 43(5), 1876–1883.
- Baldrige, W. S., Eyal, Y., Bartov, Y., Steinitz, G., and Eyal, M. (1991). Miocene magmatism of Sinai related to the opening of the Red Sea. *Tectonophysics*, 197(2-4), 181–201.
- Ball, P., White, N., MacLennan, J., and Stephenson, S. (2021). Global influence of mantle temperature and plate thickness on intraplate volcanism. *Nature communications*, 12(1), 1–13.
- Bao, X., Song, X., and Li, J. (2015). High-resolution lithospheric structure beneath Mainland China from ambient noise and earthquake surface-wave tomography. *Earth and Planetary Science Letters*, 417, 132–141.
- Barberi, F., Bonatti, E., Marinelli, G., and Varet, J. (1974). Transverse tectonics during the split of a continent: data from the Afar rift. *Tectonophysics*, 23(1-2), 17–29.
- Barberi, F., Borsi, S., Ferrara, G., Marinelli, G., Santacroce, R., Tazieff, H., and Varet, J. (1972). Evolution of the Danakil Depression (Afar, Ethiopia) in light of radiometric age determinations. *The Journal of Geology*, 80(6), 720–729.
- Barberi, F., Ferrara, G., Santacroce, R., Varet, J., et al. (1975). Structural evolution of the afar triple junction. In *Afar Depression of Ethiopia*", vol. 1, (pp. 38–54). E. Schweizerbart'sche Verlagsbuchhandlung.
- Barberi, F., Santacroce, R., et al. (1980). The Afar Stratoid Series and the magmatic evolution of East African rift system. *Bull. Soc. Geol. Fr*, 22, 891–899.
- Barberi, F., and Varet, J. (1977). Volcanism of Afar: Small-scale plate tectonics implications. *Geological Society of America Bulletin*, 88(9), 1251–1266.
- Barberi, F., and Varet, J. (1978). The Afar rift junction. In *Petrology and Geochemistry of Continental Rifts: Volume One of the Proceedings of the NATO Advanced Study Institute Paleorift Systems with Emphasis on the Permian Oslo Rift, held in Oslo, Norway, July 27–August 5, 1977*, (pp. 55–69). Springer.
- Barbier, E. (2002). Geothermal energy technology and current status: an overview. *Renewable and sustainable energy reviews*, 6(1-2), 3–65.
- Barton, A., and White, R. (1997). Volcanism on the Rockall continental margin. *Journal of the Geological Society*, 154(3), 531–536.

- Bastow, I., Booth, A., Corti, G., Keir, D., Magee, C., Jackson, C. A.-L., Warren, J., Wilkinson, J., and Lascialfari, M. (2018). The development of late-stage continental breakup: Seismic reflection and borehole evidence from the Danakil Depression, Ethiopia. *Tectonics*, 37(9), 2848–2862.
- Bastow, I., and Keir, D. (2011). The protracted development of the continent–ocean transition in Afar. *Nature Geoscience*, 4(4), 248–250.
- Bastow, I., Nyblade, A., Stuart, G., Rooney, T., and Benoit, M. (2008). Upper mantle seismic structure beneath the Ethiopian hot spot: Rifting at the edge of the African low-velocity anomaly. *Geochemistry, Geophysics, Geosystems*, 9(12).
- Bastow, I., Pilidou, S., Kendall, J.-M., and Stuart, G. (2010). Melt-induced seismic anisotropy and magma assisted rifting in Ethiopia: Evidence from surface waves. *Geochemistry, Geophysics, Geosystems*, 11(6).
- Becker, T. W., Conrad, C. P., Schaeffer, A. J., and Lebedev, S. (2014). Origin of azimuthal seismic anisotropy in oceanic plates and mantle. *Earth and Planetary Science Letters*, 401, 236–250.
- Becker, T. W., Lebedev, S., and Long, M. (2012). On the relationship between azimuthal anisotropy from shear wave splitting and surface wave tomography. *Journal of Geophysical Research: Solid Earth*, 117(B1).
- Bellahsen, N., Faccenna, C., Funiciello, F., Daniel, J., and Jolivet, L. (2003). Why did Arabia separate from Africa? insights from 3-D laboratory experiments. *Earth and Planetary Science Letters*, 216(3), 365–381.
- Benoit, M. H., Nyblade, A. A., and VanDecar, J. C. (2006). Upper mantle P-wave speed variations beneath Ethiopia and the origin of the Afar hotspot. *Geology*, 34(5), 329–332.
- Bensen, G., Ritzwoller, M., Barmin, M., Levshin, A. L., Lin, F., Moschetti, M., Shapiro, N., and Yang, Y. (2007). Processing seismic ambient noise data to obtain reliable broad-band surface wave dispersion measurements. *Geophysical journal international*, 169(3), 1239–1260.
- Bensen, G., Ritzwoller, M., and Shapiro, N. M. (2008). Broadband ambient noise surface wave tomography across the United States. *Journal of Geophysical Research: Solid Earth*, 113(B5).
- Berckhemer, H., Baier, B., Bartelsen, H., Behle, A., Burkhardt, H., Gebrande, H., Makris, J., Menzel, H., Miller, H., and Vees, R. (1975). Deep seismic soundings in the Afar region and on the highland of Ethiopia. *Afar depression of Ethiopia*, 1, 89–107.
- Bertelloni, C. L., and Gurnis, M. (1997). Cenozoic subsidence and uplift of continents from time-varying dynamic topography. *Geology*, 25(8), 735–738.
- Bianchi, I., Park, J., Piana Agostinetti, N., and Levin, V. (2010). Mapping seismic anisotropy using harmonic decomposition of receiver functions: An application to Northern Apennines, Italy. *Journal of Geophysical Research: Solid Earth*, 115(B12).

- Biggs, J., Bastow, I., Keir, D., and Lewi, E. (2011). Pulses of deformation reveal frequently recurring shallow magmatic activity beneath the Main Ethiopian Rift. *Geochemistry, Geophysics, Geosystems*, 12(9).
- Bird, P. (1978). Initiation of intracontinental subduction in the Himalaya. *Journal of Geophysical Research: Solid Earth*, 83(B10), 4975–4987.
- Bird, P. (1979). Continental delamination and the Colorado Plateau. *Journal of Geophysical Research: Solid Earth*, 84(B13), 7561–7571.
- Bird, P. (2003). An updated digital model of plate boundaries. *Geochemistry, Geophysics, Geosystems*, 4(3).
- Blundy, J., Afanasyev, A., Tattitch, B., Sparks, S., Melnik, O., Utkin, I., and Rust, A. (2021). The economic potential of metalliferous sub-volcanic brines. *Royal Society Open Science*, 8(6), 202192.
- Bohannon, R. (1989). Style of extensional tectonism during rifting, Red Sea and Gulf of Aden. *Journal of African Earth Sciences (and the Middle East)*, 8(2-4), 589–602.
- Bosworth, W., Huchon, P., and McClay, K. (2005). The Red Sea and Gulf of Aden basins. *Journal of African Earth Sciences*, 43(1-3), 334–378.
- Bowers, G. L. (2002). Detecting high overpressure. *The leading edge*, 21(2), 174–177.
- Boyce, A., Bastow, I., Cottaar, S., Kounoudis, R., Guilloud De Courbeville, J., Caunt, E., and Desai, S. (2021). AFRP20: New P-wavespeed model for the African mantle reveals two whole-mantle plumes below East Africa and neoproterozoic modification of the Tanzania Craton. *Geochemistry, Geophysics, Geosystems*, 22(3), e2020GC0009302.
- Bridges, D. L., Mickus, K., Gao, S. S., Abdelsalam, M. G., and Alemu, A. (2012). Magnetic stripes of a transitional continental rift in Afar. *Geology*, 40(3), 203–206.
- Brikke, N. E. A. (2010). 3D seismic traveltime tomography of the central South Island, New Zealand.
- Brown, F. H., and McDougall, I. (2011). Geochronology of the Turkana depression of northern Kenya and southern Ethiopia. *Evolutionary Anthropology: Issues, News, and Reviews*, 20(6), 217–227.
- Brown, R., Kavanagh, J., Sparks, R., Tait, M., and Field, M. (2007). Mechanically disrupted and chemically weakened zones in segmented dike systems cause vent localization: Evidence from kimberlite volcanic systems. *Geology*, 35(9), 815–818.
- Brun, J.-P. (1999). Narrow rifts versus wide rifts: inferences for the mechanics of rifting from laboratory experiments. *Philosophical Transactions of the Royal Society of London. Series A: Mathematical, Physical and Engineering Sciences*, 357(1753), 695–712.
- Buck, W. R. (2004). Consequences of asthenospheric variability on continental rifting. In *Rheology and deformation of the lithosphere at continental margins*, (pp. 1–30). Columbia University Press.

- Burke, K. (1996). The African plate. *South African Journal of Geology*, 99(4), 341–409.
- Burke, K. (2011). Plate tectonics, the Wilson Cycle, and mantle plumes: geodynamics from the top. *Annual Review of Earth and Planetary Sciences*, 39, 1–29.
- Burke, K., and Dewey, J. F. (1973). Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks. *The Journal of Geology*, 81(4), 406–433.
- Bushenkova, N., Koulakov, I., Senyukov, S., Gordeev, E. I., Huang, H.-H., El Khrepy, S., and Al Arifi, N. (2019). Tomographic images of magma chambers beneath the Avacha and Koryaksky volcanoes in Kamchatka. *Journal of Geophysical Research: Solid Earth*, 124(9), 9694–9713.
- Camp, V. E., Hooper, P. R., Roobol, M. J., and White, D. (1987). The Madinah eruption, Saudi Arabia: magma mixing and simultaneous extrusion of three basaltic chemical types. *Bulletin of volcanology*, 49, 489–508.
- Camp, V. E., and Roobol, M. J. (1992). Upwelling asthenosphere beneath western Arabia and its regional implications. *Journal of Geophysical Research: Solid Earth*, 97(B11), 15255–15271.
- Capaldi, G., Chiesa, S., Conticelli, S., and Manetti, P. (1987). Jabal An Nar: an upper Miocene volcanic centre near Al Mukha (Yemen Arab Republic). *Journal of volcanology and geothermal research*, 31(3-4), 345–351.
- Capaldi, G., Manetti, P., and Piccardo, G. (1983). Preliminary investigations on volcanism of the Sadah region (Yemen Arabic Republic). *Bulletin volcanologique*, 46, 413–427.
- Carbotte, S. M., Marjanović, M., Arnulf, A. F., Nedimović, M. R., Canales, J. P., and Arnoux, G. M. (2021). Stacked magma lenses beneath mid-ocean ridges: Insights from new seismic observations and synthesis with prior geophysical and geologic findings. *Journal of Geophysical Research: Solid Earth*, 126(4), e2020JB021434.
- Celli, N., Lebedev, S., Schaeffer, A., Ravenna, M., and Gaina, C. (2020a). The upper mantle beneath the South Atlantic Ocean, South America and Africa from waveform tomography with massive data sets. *Geophysical Journal International*, 221(1), 178–204.
- Celli, N. L., Lebedev, S., Schaeffer, A. J., Ravenna, M., and Gaina, C. (2020b). The upper mantle beneath the South Atlantic Ocean, South America and Africa from waveform tomography with massive data sets. *Geophysical Journal International*, 221(1), 178–204.
- Cesca, S., Letort, J., Razafindrakoto, H. N., Heimann, S., Rivalta, E., Isken, M. P., Nikkhoo, M., Passarelli, L., Petersen, G. M., Cotton, F., et al. (2020). Drainage of a deep magma reservoir near Mayotte inferred from seismicity and deformation. *Nature Geoscience*, 13(1), 87–93.
- Chambers, E. L., Harmon, N., Keir, D., and Rychert, C. A. (2019). Using ambient noise to image the northern East African Rift. *Geochemistry, Geophysics, Geosystems*, 20(4), 2091–2109.

- Chang, S.-J., Merino, M., Van der Lee, S., Stein, S., and Stein, C. A. (2011). Mantle flow beneath Arabia offset from the opening Red Sea. *Geophysical Research Letters*, 38(4).
- Chen, B., Kaban, M. K., El Khrepy, S., and Al-Arifi, N. (2015). Effective elastic thickness of the Arabian plate: weak shield versus strong platform. *Geophysical Research Letters*, 42(9), 3298–3304.
- Chernet, T., Hart, W. K., Aronson, J. L., and Walter, R. C. (1998). New age constraints on the timing of volcanism and tectonism in the northern Main Ethiopian Rift–southern Afar transition zone (Ethiopia). *Journal of Volcanology and Geothermal Research*, 80(3-4), 267–280.
- Chiarabba, C., and Moretti, M. (2006). An insight into the unrest phenomena at the Campi Flegrei caldera from Vp and Vp/Vs tomography. *Terra Nova*, 18(6), 373–379.
- Chiodini, G., Cioni, R., Guidi, M., Raco, B., and Marini, L. (1998). Soil CO₂ flux measurements in volcanic and geothermal areas. *Applied geochemistry*, 13(5), 543–552.
- Chorowicz, J. (2005). The East African rift system. *Journal of African Earth Sciences*, 43(1-3), 379–410.
- Chouet, B. A., and Matoza, R. S. (2013). A multi-decadal view of seismic methods for detecting precursors of magma movement and eruption. *Journal of Volcanology and Geothermal Research*, 252, 108–175.
- Christensen, N. I. (1996). Poisson's ratio and crustal seismology. *Journal of Geophysical Research: Solid Earth*, 101(B2), 3139–3156.
- Christensen, N. I., and Mooney, W. D. (1995). Seismic velocity structure and composition of the continental crust: A global view. *Journal of Geophysical Research: Solid Earth*, 100(B6), 9761–9788.
- Chu, D., and Gordon, R. G. (1999). Evidence for motion between Nubia and Somalia along the Southwest Indian Ridge. *Nature*, 398(6722), 64–67.
- Civiero, C., Celli, N. L., and Tesauro, M. (2023). Revisiting the geodynamics of the Middle East region from an integrated geophysical perspective. *Journal of Geodynamics*, 158, 102005.
- Civiero, C., Goes, S., Hammond, J. O., Fishwick, S., Ahmed, A., Ayele, A., Doubre, C., Goitom, B., Keir, D., Kendall, J. M., et al. (2016). Small-scale thermal upwellings under the northern East African Rift from S travel time tomography. *Journal of Geophysical Research: Solid Earth*, 121(10), 7395–7408.
- Civiero, C., Hammond, J. O., Goes, S., Fishwick, S., Ahmed, A., Ayele, A., Doubre, C., Goitom, B., Keir, D., Kendall, J.-M., et al. (2015). Multiple mantle upwellings in the transition zone beneath the northern East-African Rift system from relative P-wave travel-time tomography. *Geochemistry, Geophysics, Geosystems*, 16(9), 2949–2968.
- Civiero, C., Lebedev, S., and Celli, N. L. (2022). A complex mantle plume head below East Africa-Arabia shaped by the lithosphere-asthenosphere boundary topography. *Geochemistry, Geophysics, Geosystems*, 23(11), e2022GC010610.

- Clarisse, L., Hurtmans, D., Clerbaux, C., Hadji-Lazaro, J., Ngadi, Y., and Coheur, P.-F. (2012). Retrieval of sulphur dioxide from the infrared atmospheric sounding interferometer (IASI). *Atmospheric Measurement Techniques*, 5(3), 581–594.
- CNR-CNRS (1973). Geology of northern Afar (Ethiopia). *Revue de Géographie Physique et de Géologie Dynamique*, 15, 443–490.
- Cochran, J. R. (1983). A model for development of Red Sea. *Aapg Bulletin*, 67(1), 41–69.
- Coleman, R. G. (1993). Geological evolution of the Red Sea. *Oxford Monographs on Geology and Geophysics*. Oxford University Press, New York, 186.
- Coleman, R. G., Gregory, R., and Brown, G. F. (1983). *Cenozoic volcanic rocks of Saudi Arabia*, vol. 83. US Department of the Interior, Geological Survey.
- Corbeau, J., Rolandone, F., Leroy, S., Al-Lazki, A., Stork, A. L., Keir, D., Stuart, G., Hammond, J. O., Doubre, C., Vergne, J., et al. (2014). Uppermost mantle velocity from Pn tomography in the Gulf of Aden. *Geosphere*, 10(5), 958–968.
- Cornwell, D., Hetényi, G., and Blanchard, T. (2011). Mantle transition zone variations beneath the Ethiopian Rift and Afar: Chemical heterogeneity within a hot mantle? *Geophysical Research Letters*, 38(16).
- Cornwell, D., Maguire, P., England, R., and Stuart, G. (2010). Imaging detailed crustal structure and magmatic intrusion across the Ethiopian Rift using a dense linear broadband array. *Geochemistry, Geophysics, Geosystems*, 11(1).
- Coulié, E., Quidelleur, X., Gillot, P.-Y., Courtillot, V., Lefèvre, J.-C., and Chiesa, S. (2003). Comparative K–Ar and Ar/Ar dating of Ethiopian and Yemenite Oligocene volcanism: implications for timing and duration of the Ethiopian traps. *Earth and Planetary Science Letters*, 206(3–4), 477–492.
- Courtillot, V., Achache, J., Landre, F., Bonhommet, N., Montigny, R., and Féraud, G. (1984). Episodic spreading and rift propagation: new paleomagnetic and geochronologic data from the Afar nascent passive margin. *Journal of Geophysical Research: Solid Earth*, 89(B5), 3315–3333.
- Courtillot, V., Galdeano, A., and Le Mouél, J. (1980). Propagation of an accreting plate boundary: a discussion of new aeromagnetic data in the Gulf of Tadjurah and southern Afar. *Earth and Planetary Science Letters*, 47(1), 144–160.
- Crisp, J. A. (1984). Rates of magma emplacement and volcanic output. *Journal of Volcanology and Geothermal Research*, 20(3–4), 177–211.
- Crowley, J. K., and Zimbelman, D. R. (1997). Mapping hydrothermally altered rocks on Mount Rainier, Washington, with airborne visible/infrared imaging spectrometer (AVIRIS) data. *Geology*, 25(6), 559–562.
- d’Acremont, E., Leroy, S., Beslier, M.-O., Bellahsen, N., Fournier, M., Robin, C., Maia, M., and Gente, P. (2005). Structure and evolution of the eastern Gulf of Aden conjugate margins from seismic reflection data. *Geophysical Journal International*, 160(3), 869–890.

- d'Acremont, E., Leroy, S., Maia, M., Gente, P., and Autin, J. (2010). Volcanism, jump and propagation on the Sheba ridge, eastern Gulf of Aden: segmentation evolution and implications for oceanic accretion processes. *Geophysical Journal International*, 180(2), 535–551.
- Daniels, K. A., Bastow, I., Keir, D., Sparks, R., and Menand, T. (2014). Thermal models of dyke intrusion during development of continent–ocean transition. *Earth and Planetary Science Letters*, 385, 145–153.
- Daoud, M. A., Maury, R. C., Barrat, J.-A., Taylor, R. N., Le Gall, B., Guillou, H., Cotten, J., and Rolet, J. (2010). A LREE-depleted component in the Afar plume: Further evidence from Quaternary Djibouti basalts. *Lithos*, 114(3-4), 327–336.
- Daradich, A., Mitrovica, J. X., Pysklywec, R. N., Willett, S. D., and Forte, A. M. (2003). Mantle flow, dynamic topography, and rift-flank uplift of Arabia. *Geology*, 31(10), 901–904.
- Davison, I., Al-Kadasi, M., Al-Khirbash, S., Al-Subbary, A. K., Baker, J., Blakey, S., Bosence, D., Dart, C., Heaton, R., McClay, K., et al. (1994). Geological evolution of the southeastern Red Sea Rift margin, Republic of Yemen. *Geological Society of America Bulletin*, 106(11), 1474–1493.
- Dawson, P., Chouet, B., and Pitt, A. (2016). Tomographic image of a seismically active volcano: Mammoth Mountain, California. *Journal of Geophysical Research: Solid Earth*, 121(1), 114–133.
- de Kool, M., Rawlinson, N., and Sambridge, M. (2006). A practical grid-based method for tracking multiple refraction and reflection phases in three-dimensional heterogeneous media. *Geophysical Journal International*, 167(1), 253–270.
- DeMets, C., Gordon, R. G., and Argus, D. F. (2010). Geologically current plate motions. *Geophysical journal international*, 181(1), 1–80.
- Dixon, T. H., Ivins, E. R., and Franklin, B. J. (1989). Topographic and volcanic asymmetry around the Red Sea: Constraints on rift models. *Tectonics*, 8(6), 1193–1216.
- Donovan, A., Blundy, J., Oppenheimer, C., and Buisman, I. (2018). The 2011 eruption of Nabro volcano, Eritrea: perspectives on magmatic processes from melt inclusions. *Contributions to Mineralogy and Petrology*, 173(1), 1–23.
- Drury, S., Kelley, S., Berhe, S., Collier, R. L., and Abraha, M. (1994). Structures related to Red Sea evolution in northern Eritrea. *Tectonics*, 13(6), 1371–1380.
- Dueker, K. G., and Sheehan, A. F. (1998). Mantle discontinuity structure beneath the Colorado Rocky Mountains and high plains. *Journal of Geophysical Research: Solid Earth*, 103(B4), 7153–7169.
- Dugda, M. T., and Nyblade, A. A. (2006). New constraints on crustal structure in eastern Afar from the analysis of receiver functions and surface wave dispersion in Djibouti. *Geological Society, London, Special Publications*, 259(1), 239–251.

- Dugda, M. T., Nyblade, A. A., Julia, J., Langston, C. A., Ammon, C. J., and Simiyu, S. (2005). Crustal structure in Ethiopia and Kenya from receiver function analysis: Implications for rift development in eastern Africa. *Journal of Geophysical Research: Solid Earth*, 110(B1).
- Duncan, R. A., and Al-Amri, A. M. (2013). Timing and composition of volcanic activity at Harrat Lunayyir, western Saudi Arabia. *Journal of Volcanology and Geothermal Research*, 260, 103–116.
- Eagles, G., Gloaguen, R., and Ebinger, C. (2002). Kinematics of the Danakil microplate. *Earth and Planetary Science Letters*, 203(2), 607–620.
- Eberhart-Phillips, D., and Reyners, M. (2012). Imaging the Hikurangi Plate interface region, with improved local-earthquake tomography. *Geophysical Journal International*, 190(2), 1221–1242.
- Ebinger, C., Bechtel, T., Forsyth, D., and Bowin, C. (1989a). Effective elastic plate thickness beneath the East African and Afar plateaus and dynamic compensation of the uplifts. *Journal of Geophysical Research: Solid Earth*, 94(B3), 2883–2901.
- Ebinger, C., and Casey, M. (2001). Continental breakup in magmatic provinces: An Ethiopian example. *Geology*, 29(6), 527–530.
- Ebinger, C., and Hayward, N. (1996). Soft plates and hot spots: Views from Afar. *Journal of Geophysical Research: Solid Earth*, 101(B10), 21859–21876.
- Ebinger, C., Yemane, T., Woldegabriel, G., Aronson, J., and Walter, R. (1993). Late Eocene–Recent volcanism and faulting in the southern main Ethiopian rift. *Journal of the Geological Society*, 150(1), 99–108.
- Ebinger, C. J., Deino, A. L., Drake, R., and Tesha, A. (1989b). Chronology of volcanism and rift basin propagation: Rungwe volcanic province, East Africa. *Journal of Geophysical Research: Solid Earth*, 94(B11), 15785–15803.
- Ebinger, C. J., and Sleep, N. (1998). Cenozoic magmatism throughout east Africa resulting from impact of a single plume. *Nature*, 395(6704), 788–791.
- Egloff, F., Rihm, R., Makris, J., Izzeldin, Y., Bobsien, M., Meier, K., Junge, P., Noman, T., and Warsi, W. (1991). Contrasting structural styles of the eastern and western margins of the southern Red Sea: the 1988 SONNE experiment. *Tectonophysics*, 198(2-4), 329–353.
- Elsheikh, A. A. (2019). Seismic anisotropy and mantle flow beneath East Africa and Arabia. *Journal of African Earth Sciences*, 149, 97–108.
- Evans, J. R., Eberhart-Phillips, D., and Thurber, C. (1994). User's manual for SIMULPS12 for imaging Vp and Vp/Vs; a derivative of the "thurber" tomographic inversion simul3 for local earthquakes and explosions. Tech. rep., US Geological Survey,.
- Faccenda, M., and VanderBeek, B. P. (2023). On constraining 3D seismic anisotropy in subduction, mid-ocean-ridge, and plume environments with teleseismic body wave data. *Journal of Geodynamics*, 158, 102003.

- Faccenna, C., Becker, T. W., Auer, L., Billi, A., Boschi, L., Brun, J. P., Capitanio, F. A., Funiciello, F., Horv  th, F., Jolivet, L., et al. (2014). Mantle dynamics in the Mediterranean. *Reviews of Geophysics*, 52(3), 283–332.
- Faccenna, C., Becker, T. W., Jolivet, L., and Keskin, M. (2013). Mantle convection in the Middle East: Reconciling Afar upwelling, Arabia indentation and Aegean trench rollback. *Earth and Planetary Science Letters*, 375, 254–269.
- Fee, D., Matoza, R. S., Gee, K. L., Neilsen, T. B., and Ogden, D. E. (2013). Infrasonic crackle and supersonic jet noise from the eruption of Nabro Volcano, Eritrea. *Geophysical Research Letters*, 40(16), 4199–4203.
- Fee, D., Tan, D., Girona, T., Lyons, J. J., Hotovec-Ellis, A. J., Cannata, A., De Angelis, S., Haney, M. M., Wech, A., and Sciotto, M. (2023). Towards a Generalized Machine Learning Model to Detect and Classify Volcano Seismicity. *AGU23*.
- Ferguson, D. J., MacLennan, J., Bastow, I., Pyle, D., Jones, S., Keir, D., Blundy, J., Plank, T., and Yirgu, G. (2013). Melting during late-stage rifting in Afar is hot and deep. *Nature*, 499(7456), 70–73.
- Field, L., Barnie, T., Blundy, J., Brooker, R. A., Keir, D., Lewi, E., and Saunders, K. (2012). Integrated field, satellite and petrological observations of the November 2010 eruption of Erta Ale. *Bulletin of Volcanology*, 74, 2251–2271.
- Fishwick, S. (2010). Surface wave tomography: imaging of the lithosphere–asthenosphere boundary beneath central and southern Africa? *Lithos*, 120(1-2), 63–73.
- Forte, A. M., Peltier, W. R., Dziewonski, A. M., and Woodward, R. L. (1993). Dynamic surface topography: A new interpretation based upon mantle flow models derived from seismic tomography. *Geophysical Research Letters*, 20(3), 225–228.
- Foulger, G. R., Julian, B., Pitt, A., Hill, D., Malin, P., and Shalev, E. (2003). Three-dimensional crustal structure of Long Valley caldera, California, and evidence for the migration of CO₂ under Mammoth Mountain. *Journal of Geophysical Research: Solid Earth*, 108(B3).
- Fournier, M., Chamot-Rooke, N., Petit, C., Huchon, P., Al-Kathiri, A., Audin, L., Beslier, M.-O., d’Acremont, E., Fabbri, O., Fleury, J.-M., et al. (2010). Arabia-Somalia plate kinematics, evolution of the Aden-Owen-Carlsberg triple junction, and opening of the Gulf of Aden. *Journal of Geophysical Research: Solid Earth*, 115(B4).
- Frassetto, A. M., Zandt, G., Gilbert, H., Owens, T. J., and Jones, C. H. (2011). Structure of the Sierra Nevada from receiver functions and implications for lithospheric foundering. *Geosphere*, 7(4), 898–921.
- Freund, R. (1970). Plate tectonics of the Red Sea and east Africa. *Nature*, 228(5270), 453–453.
- Fromm, M., Kablick III, G., Nedoluha, G., Carboni, E., Grainger, R., Campbell, J., and Lewis, J. (2014). Correcting the record of volcanic stratospheric aerosol impact: Nabro and Sarychev Peak. *Journal of Geophysical Research: Atmospheres*, 119(17), 10–343.

- Furman, T. (2007). Geochemistry of East African Rift basalts: an overview. *Journal of African Earth Sciences*, 48(2-3), 147–160.
- Furman, T., Bryce, J., Rooney, T., Hanan, B., Yirgu, G., and Ayalew, D. (2006a). Heads and tails: 30 million years of the Afar plume. *Geological Society, London, Special Publications*, 259(1), 95–119.
- Furman, T., Kaleta, K. M., Bryce, J. G., and Hanan, B. B. (2006b). Tertiary mafic lavas of Turkana, Kenya: constraints on East African plume structure and the occurrence of high- μ volcanism in Africa. *Journal of Petrology*, 47(6), 1221–1244.
- Gallacher, R., and Bastow, I. (2012). The development of magmatism along the Cameroon Volcanic Line: Evidence from teleseismic receiver functions. *Tectonics*, 31(3).
- Gallacher, R. J. (2016). *Mantle response to continental breakup and elevated potential temperatures beneath the Afar triple junction*. Ph.D. thesis, University of Southampton.
- Gao, S. S., Liu, K. H., and Abdelsalam, M. G. (2010). Seismic anisotropy beneath the Afar Depression and adjacent areas: Implications for mantle flow. *Journal of Geophysical Research: Solid Earth*, 115(B12).
- Garnero, E. J., and McNamara, A. K. (2008). Structure and dynamics of Earth's lower mantle. *science*, 320(5876), 626–628.
- Gashawbeza, E. M., Klemperer, S. L., Nyblade, A. A., Walker, K. T., and Keranen, K. M. (2004). Shear-wave splitting in Ethiopia: Precambrian mantle anisotropy locally modified by Neogene rifting. *Geophysical Research Letters*, 31(18).
- Gass, I., and Mallick, D. (1968). Jebel Khariz: an Upper Miocene strato-volcano of comenditic affinity on the South Arabian coast. *Bulletin Volcanologique*, 32, 33–88.
- Gauntlett, M., Hudson, T., Kendall, J.-M., Rawlinson, N., Blundy, J., Lapins, S., Goitom, B., Hammond, J., Oppenheimer, C., and Ogubazghi, G. (2023). Nabro 3D velocity model produced by the FMTOMO code.
URL <https://doi.org/10.5281/zenodo.7701958>
- George, R., Rogers, N., and Kelley, S. (1998). Earliest magmatism in Ethiopia: evidence for two mantle plumes in one flood basalt province. *Geology*, 26(10), 923–926.
- Ghebreab, W. (1998). Tectonics of the Red Sea region reassessed. *Earth-Science Reviews*, 45(1-2), 1–44.
- Ginzburg, A., Makris, J., Fuchs, K., and Prodehl, C. (1981). The structure of the crust and upper mantle in the Dead Sea rift. *Tectonophysics*, 80(1-4), 109–119.
- Glahn, A., Granet, M., and Group, R. G. T. (1993). Southern Rhine Graben: small-wavelength tomographic study and implications for the dynamic evolution of the graben. *Geophysical Journal International*, 113(2), 399–418.
- Goitom, B. (2017). *The Nabro volcano, tectonic framework and seismic hazard assessment of Eritrea*. Ph.D. thesis, University of Bristol.

- Goitom, B., Oppenheimer, C., Hammond, J. O., Grandin, R., Barnie, T., Donovan, A., Ogubazghi, G., Yohannes, E., Kibrom, G., Kendall, J.-M., et al. (2015). First recorded eruption of Nabro volcano, Eritrea, 2011. *Bulletin of Volcanology*, 77(10), 1–21.
- Greenfield, T., Keir, D., Kendall, J.-M., and Ayele, A. (2019). Seismicity of the Bora-Tullu Moye volcanic field, 2016–2017. *Geochemistry, Geophysics, Geosystems*, 20(2), 548–570.
- Greenfield, T., and White, R. S. (2015). Building Icelandic igneous crust by repeated melt injections. *Journal of Geophysical Research: Solid Earth*, 120(11), 7771–7788.
- Greenfield, T., White, R. S., and Roecker, S. (2016). The magmatic plumbing system of the Askja central volcano, Iceland, as imaged by seismic tomography. *Journal of Geophysical Research: Solid Earth*, 121(10), 7211–7229.
- Hager, B. H., Clayton, R. W., Richards, M. A., Comer, R. P., and Dziewonski, A. M. (1985). Lower mantle heterogeneity, dynamic topography and the geoid. *Nature*, 313(6003), 541–545.
- Hager, B. H., and Richards, M. A. (1989). Long-wavelength variations in Earth's geoid: physical models and dynamical implications. *Philosophical Transactions of the Royal Society of London. Series A, Mathematical and Physical Sciences*, 328(1599), 309–327.
- Halldórsson, S. A., Hilton, D. R., Scarsi, P., Abebe, T., and Hopp, J. (2014). A common mantle plume source beneath the entire East African Rift System revealed by coupled helium-neon systematics. *Geophysical Research Letters*, 41(7), 2304–2311.
- Hamling, I. J., Ayele, A., Bennati, L., Calais, E., Ebinger, C. J., Keir, D., Lewi, E., Wright, T. J., and Yirgu, G. (2009). Geodetic observations of the ongoing Dabbahu rifting episode: new dyke intrusions in 2006 and 2007. *Geophysical Journal International*, 178(2), 989–1003.
- Hamlyn, J., Wright, T., Walters, R., Pagli, C., Sansosti, E., Casu, F., Pepe, S., Edmonds, M., Kilbride, B. M., Keir, D., et al. (2018). What causes subsidence following the 2011 eruption at Nabro (Eritrea)? *Progress in Earth and Planetary Science*, 5(1), 1–15.
- Hamlyn, J. E., Keir, D., Wright, T. J., Neuberg, J. W., Goitom, B., Hammond, J. O., Pagli, C., Oppenheimer, C., Kendall, J.-M., and Grandin, R. (2014). Seismicity and subsidence following the 2011 Nabro eruption, Eritrea: Insights into the plumbing system of an off-rift volcano. *Journal of Geophysical Research: Solid Earth*, 119(11), 8267–8282.
- Hammond, J. (2014). Constraining melt geometries beneath the Afar Depression, Ethiopia from teleseismic receiver functions: The anisotropic H- κ stacking technique. *Geochemistry, Geophysics, Geosystems*, 15(4), 1316–1332.
- Hammond, J., Goitom, B., and Kendall, J.-M. (2014a). Rifting in the Horn of Africa: The Eritrea Seismic Project (June 2011–October 2012). *NERC Geophysical Equipment Facility, Scientific Report 913*.
- Hammond, J., Goitom, B., Kendall, J. M., and Ogubazghi, G. (2011a). Eritrea Seismic Project.
URL https://www.fdsn.org/networks/detail/5H_2011/

- Hammond, J., Goitom, B., Kendall, J. M., and Ogubazghi, G. (2011b). Nabro Urgency Array.
URL https://www.fdsn.org/networks/detail/4H_2011/
- Hammond, J., Kendall, J.-M., Stuart, G., Keir, D., Ebinger, C., Ayele, A., and Belachew, M. (2011c). The nature of the crust beneath the Afar triple junction: Evidence from receiver functions. *Geochemistry, Geophysics, Geosystems*, 12(12).
- Hammond, J., Kendall, M., Stuart, G., Ebinger, C., Bastow, I. D., Keir, D., Ayele, A., Belachew, M., Goitom, B., Ogubazghi, G., et al. (2013). Mantle upwelling and initiation of rift segmentation beneath the Afar Depression. *Geology*, 41(6), 635–638.
- Hammond, J. O., Kendall, J.-M., Wookey, J., Stuart, G., Keir, D., and Ayele, A. (2014b). Differentiating flow, melt, or fossil seismic anisotropy beneath Ethiopia. *Geochemistry, Geophysics, Geosystems*, 15(5), 1878–1894.
- Hansen, S., Schwartz, S., Al-Amri, A., and Rodgers, A. (2006). Combined plate motion and density-driven flow in the asthenosphere beneath Saudi Arabia: Evidence from shear-wave splitting and seismic anisotropy. *Geology*, 34(10), 869–872.
- Hansen, S. E., Nyblade, A. A., and Benoit, M. H. (2012). Mantle structure beneath Africa and Arabia from adaptively parameterized P-wave tomography: Implications for the origin of Cenozoic Afro-Arabian tectonism. *Earth and Planetary Science Letters*, 319, 23–34.
- Hansen, S. E., Rodgers, A. J., Schwartz, S. Y., and Al-Amri, A. M. (2007). Imaging ruptured lithosphere beneath the Red Sea and Arabian Peninsula. *Earth and Planetary Science Letters*, 259(3–4), 256–265.
- Hauksson, E., and Shearer, P. M. (2006). Attenuation models (q_p and q_s) in three dimensions of the southern California crust: Inferred fluid saturation at seismogenic depths. *Journal of Geophysical Research: Solid Earth*, 111(B5).
- Hayward, N., and Ebinger, C. (1996). Variations in the along-axis segmentation of the Afar Rift system. *Tectonics*, 15(2), 244–257.
- Hearn, T. M., and Clayton, R. W. (1986). Lateral velocity variations in southern California. I. Results for the upper crust from Pg waves. *Bulletin of the Seismological Society of America*, 76(2), 495–509.
- Helffrich, G., Wookey, J., and Bastow, I. (2013). *The seismic analysis code: A primer and user's guide*. Cambridge University Press.
- Hill, R. I. (1993). Mantle plumes and continental tectonics. *Lithos*, 30(3–4), 193–206.
- Hofmann, C., Courtillot, V., Feraud, G., Rochette, P., Yirgu, G., Ketefo, E., and Pik, R. (1997). Timing of the Ethiopian flood basalt event and implications for plume birth and global change. *Nature*, 389(6653), 838–841.
- Hofstetter, R., and Beyth, M. (2003). The Afar Depression: interpretation of the 1960–2000 earthquakes. *Geophysical Journal International*, 155(2), 715–732.

- Hoggard, M., Austermann, J., Randel, C., and Stephenson, S. (2021). Observational estimates of dynamic topography through space and time. *Mantle convection and surface expressions*, (pp. 371–411).
- Hoggard, M., Czarnota, K., Richards, F. D., Huston, D. L., Jaques, A. L., and Ghelichkhan, S. (2020). Global distribution of sediment-hosted metals controlled by craton edge stability. *Nature Geoscience*, 13(7), 504–510.
- Holdt, M., White, N., Stephenson, S., and Conway-Jones, B. (2022). Densely sampled global dynamic topographic observations and their significance. *Journal of Geophysical Research: Solid Earth*, 127(7), e2022JB024391.
- Horspool, N., Savage, M., and Bannister, S. (2006). Implications for intraplate volcanism and back-arc deformation in northwestern New Zealand, from joint inversion of receiver functions and surface waves. *Geophysical Journal International*, 166(3), 1466–1483.
- Hosseini, K., Matthews, K. J., Sigloch, K., Shephard, G. E., Domeier, M., and Tsekhmistrenko, M. (2018). SubMachine: Web-based tools for exploring seismic tomography and other models of Earth’s deep interior. *Geochemistry, Geophysics, Geosystems*, 19(5), 1464–1483.
- Hottmann, C., and Johnson, R. (1965). Estimation of formation pressures from log-derived shale properties. *Journal of Petroleum Technology*, 17(06), 717–722.
- Huang, H., Yao, H., and van der Hilst, R. D. (2010). Radial anisotropy in the crust of SE Tibet and SW China from ambient noise interferometry. *Geophysical Research Letters*, 37(21).
- Hudson, T. S., Kendall, J.-M., Pritchard, M. E., Blundy, J. D., and Gottsmann, J. H. (2022). From slab to surface: Earthquake evidence for fluid migration at Uturuncu volcano, Bolivia. *Earth and Planetary Science Letters*, 577, 117268.
- Hurman, G. L., Keir, D., Bull, J. M., McNeill, L. C., Booth, A. D., and Bastow, I. D. (2023). Quantitative Analysis of Faulting in the Danakil Depression Rift of Afar: The Importance of Faulting in the Final Stages of Magma-Rich Rifting. *Tectonics*, 42(6), e2022TC007607.
- Hutchinson, R., and Engels, G. (1972). Tectonic evolution in the southern Red Sea and its possible significance to older rifted continental margins. *Geological Society of America Bulletin*, 83(10), 2989–3002.
- Illsley-Kemp, F., Savage, M. K., Keir, D., Hirschberg, H. P., Bull, J. M., Gernon, T. M., Hammond, J. O., Kendall, J.-M., Ayele, A., and Goitom, B. (2017). Extension and stress during continental breakup: Seismic anisotropy of the crust in Northern Afar. *Earth and Planetary Science Letters*, 477, 41–51.
- Ito, H., DeVilbiss, J., and Nur, A. (1979). Compressional and shear waves in saturated rock during water-steam transition. *Journal of Geophysical Research: Solid Earth*, 84(B9), 4731–4735.
- Izzeldin, A. (1987). Seismic, gravity and magnetic surveys in the central part of the Red Sea: their interpretation and implications for the structure and evolution of the Red Sea. *Tectonophysics*, 143(4), 269–306.

- Janiszewski, H. A., Wagner, L. S., and Roman, D. C. (2020). Aseismic mid-crustal magma reservoir at cleveland volcano imaged through novel receiver function analyses. *Scientific reports*, 10(1), 1–9.
- Jenkins, J., Stephenson, S. N., Martínez-Garzón, P., Bohnhoff, M., and Nurlu, M. (2020). Crustal thickness variation across the Sea of Marmara region, NW Turkey: A reflection of modern and ancient tectonic processes. *Tectonics*, 39(7), e2019TC005986.
- Jolivet, L., and Faccenna, C. (2000). Mediterranean extension and the Africa-Eurasia collision. *Tectonics*, 19(6), 1095–1106.
- Journeau, C., Shapiro, N. M., Seydoux, L., Soubestre, J., Koulakov, I. Y., Jakovlev, A. V., Abkadyrov, I., Gordeev, E. I., Chebrov, D. V., Droznin, D. V., et al. (2022). Seismic tremor reveals active trans-crustal magmatic system beneath kamchatka volcanoes. *Science advances*, 8(5), eabj1571.
- Julia, J., and Nyblade, A. A. (2013). Probing the upper mantle transition zone under Africa with P520s conversions: Implications for temperature and composition. *Earth and Planetary Science Letters*, 368, 151–162.
- Julian, B. R., Pitt, A., and Foulger, G. (1998). Seismic image of a CO₂ reservoir beneath a seismically active volcano. *Geophysical Journal International*, 133(1), F7–F10.
- Kaviani, A., Mahmoodabadi, M., Rümper, G., Pilia, S., Tatar, M., Nilfouroushan, F., Yamini-Fard, F., Moradi, A., and Ali, M. Y. (2021). Mantle-flow diversion beneath the Iranian plateau induced by Zagros’ lithospheric keel. *Scientific reports*, 11(1), 2848.
- Kazmin, V., Shifferaw, A., and Balcha, T. (1978). The Ethiopian basement: stratigraphy and possible manner of evolution. *Geologische Rundschau*, 67, 531–546.
- Kebede, S. (2012). Geothermal exploration and development in Ethiopia: Status and future plan. *Short Course VII on Exploration for Geothermal Resources Organized by UNU-GTP, GDC and KenGen*.
- Keir, D., Bastow, I., Pagli, C., and Chambers, E. L. (2013). The development of extension and magmatism in the Red Sea rift of Afar. *Tectonophysics*, 607, 98–114.
- Keir, D., Bastow, I. D., Whaler, K. A., Daly, E., Cornwell, D. G., and Hautot, S. (2009a). Lower crustal earthquakes near the Ethiopian rift induced by magmatic processes. *Geochemistry, Geophysics, Geosystems*, 10(6).
- Keir, D., Belachew, M., Ebinger, C., Kendall, J.-M., Hammond, J. O., Stuart, G., Ayele, A., and Rowland, J. (2011a). Mapping the evolving strain field during continental breakup from crustal anisotropy in the Afar Depression. *Nature Communications*, 2(1), 285.
- Keir, D., Ebinger, C., Stuart, G., Daly, E., and Ayele, A. (2006). Strain accommodation by magmatism and faulting as rifting proceeds to breakup: Seismicity of the northern Ethiopian rift. *Journal of Geophysical Research: Solid Earth*, 111(B5).

- Keir, D., Hamling, I. J., Ayele, A., Calais, E., Ebinger, C., Wright, T. J., Jacques, E., Mohamed, K., Hammond, J. O., Belachew, M., et al. (2009b). Evidence for focused magmatic accretion at segment centers from lateral dike injections captured beneath the Red Sea rift in Afar. *Geology*, *37*(1), 59–62.
- Keir, D., and Hammond, J. (2009). AFAR0911.
URL https://www.fdsn.org/networks/detail/2H_2009/
- Keir, D., Kendall, J.-M., Ebinger, C., and Stuart, G. (2005). Variations in late syn-rift melt alignment inferred from shear-wave splitting in crustal earthquakes beneath the Ethiopian rift. *Geophysical Research Letters*, *32*(23).
- Keir, D., Pagli, C., Bastow, I. D., and Ayele, A. (2011b). The magma-assisted removal of Arabia in Afar: Evidence from dike injection in the Ethiopian rift captured using InSAR and seismicity. *Tectonics*, *30*(2).
- Kendall, J.-M. (2000). Seismic anisotropy in the boundary layers of the mantle. *Washington DC American Geophysical Union Geophysical Monograph Series*, *117*, 133–159.
- Kendall, J.-M., and Lithgow-Bertelloni, C. (2016). Why is Africa rifting? *Special Publications*, *420*(1), 11–30.
- Kendall, J.-M., and Silver, P. G. (1996). Constraints from seismic anisotropy on the nature of the lowermost mantle. *Nature*, *381*(6581), 409–412.
- Kendall, J.-M., Stuart, G., Ebinger, C., Bastow, I., and Keir, D. (2005). Magma-assisted rifting in Ethiopia. *Nature*, *433*(7022), 146–148.
- Kennett, B., Sambridge, M., and Williamson, P. (1988). Subspace methods for large inverse problems with multiple parameter classes. *Geophysical Journal International*, *94*(2), 237–247.
- Kieffer, B., Arndt, N., Lapierre, H., Bastien, F., Bosch, D., Pecher, A., Yirgu, G., Ayalew, D., Weis, D., Jerram, D. A., et al. (2004). Flood and shield basalts from Ethiopia: magmas from the African superswell. *Journal of Petrology*, *45*(4), 793–834.
- Kiser, E., Palomeras, I., Levander, A., Zelt, C., Harder, S., Schmandt, B., Hansen, S., Creager, K., and Ulberg, C. (2016). Magma reservoirs from the upper crust to the Moho inferred from high-resolution Vp and Vs models beneath Mount St. Helens, Washington State, USA. *Geology*, *44*(6), 411–414.
- Koelemeijer, P., Deuss, A., and Ritsema, J. (2017). Density structure of Earth’s lowermost mantle from Stoneley mode splitting observations. *Nature Communications*, *8*(1), 15241.
- Kogan, L., Fisseha, S., Bendick, R., Reilinger, R., McClusky, S., King, R., and Solomon, T. (2012). Lithospheric strength and strain localization in continental extension from observations of the East African Rift. *Journal of Geophysical Research: Solid Earth*, *117*(B3).

- Korger, E., and Schlindwein, V. (2014). Seismicity and structure of the 85 E volcanic complex at the ultraslow spreading Gakkel Ridge from local earthquake tomography. *Geophysical Journal International*, 196(1), 539–551.
- Korostelev, F., Basuyau, C., Leroy, S., Tiberi, C., Ahmed, A., Stuart, G. W., Keir, D., Rolandone, F., Al Ganad, I., Khanbari, K., et al. (2014). Crustal and upper mantle structure beneath south-western margin of the Arabian Peninsula from teleseismic tomography. *Geochemistry, Geophysics, Geosystems*, 15(7), 2850–2864.
- Koulakov, I., Burov, E., Cloetingh, S., El Khrepy, S., Al-Arifi, N., and Bushenkova, N. (2016). Evidence for anomalous mantle upwelling beneath the Arabian Platform from travel time tomography inversion. *Tectonophysics*, 667, 176–188.
- Koulakov, I., Komzeleva, V., Smirnov, S. Z., and Bortnikova, S. B. (2021). Magma-Fluid Interactions Beneath Akutan Volcano in the Aleutian Arc Based on the Results of Local Earthquake Tomography. *Journal of Geophysical Research: Solid Earth*, 126(3), e2020JB021192.
- Kumar, U., Legendre, C., and Huang, B. (2021). Crustal structure and upper mantle anisotropy of the Afar triple junction. *Earth, Planets and Space*, 73(1), 1–17.
- Lamb, S., Moore, J. D., Perez-Gussinye, M., and Stern, T. (2020). Global whole lithosphere isostasy: implications for surface elevations, structure, strength, and densities of the continental lithosphere. *Geochemistry, Geophysics, Geosystems*, 21(10), e2020GC009150.
- Langston, C. A. (1979). Structure under Mount Rainier, Washington, inferred from teleseismic body waves. *Journal of Geophysical Research: Solid Earth*, 84(B9), 4749–4762.
- Lapins, S. (2021a). *Detecting and characterising seismicity associated with volcanic and magmatic processes through deep learning and the continuous wavelet transform*. Ph.D. thesis, University of Bristol, <http://hdl.handle.net/1983/ea90148c-a1b2-47ae-afad-5dd0a8b5ebbd>.
- Lapins, S. (2021b). Python notebooks to accompany paper 'A Little Data Goes a Long Way: Automating Seismic Phase Arrival Picking at Nabro Volcano with Transfer Learning'. URL <https://doi.org/10.5281/zenodo.4558154>
- Lapins, S. (2022). Nabro volcano event catalogue from Lapins et al., 2021, JGR Solid Earth. URL <https://doi.org/10.5281/zenodo.7398824>
- Lapins, S., Goitom, B., Kendall, J.-M., Werner, M. J., Cashman, K. V., and Hammond, J. O. (2021). A little data goes a long way: Automating seismic phase arrival picking at Nabro volcano with transfer learning. *Journal of Geophysical Research: Solid Earth*, 126(7), e2021JB021910.
- Laughton, A. S. (1966). A discussion concerning the floor of the northwest Indian Ocean-The Gulf of Aden. *Philosophical Transactions of the Royal Society of London. Series A, Mathematical and Physical Sciences*, 259(1099), 150–171.

- Lavayssière, A., Crawford, W. C., Saurel, J.-M., Satriano, C., Feuillet, N., Jacques, E., and Komorowski, J.-C. (2022). A new 1D velocity model and absolute locations image the Mayotte seismo-volcanic region. *Journal of Volcanology and Geothermal Research*, 421, 107440.
- Lavayssière, A., Greenfield, T., Keir, D., Ayele, A., and Kendall, J.-M. (2019). Local seismicity near the actively deforming Corbetti volcano in the Main Ethiopian Rift. *Journal of Volcanology and Geothermal Research*, 381, 227–237.
- Lavayssière, A., Rychert, C., Harmon, N., Keir, D., Hammond, J. O., Kendall, J.-M., Doubre, C., and Leroy, S. (2018). Imaging lithospheric discontinuities beneath the northern East African Rift using S-to-P receiver functions. *Geochemistry, Geophysics, Geosystems*, 19(10), 4048–4062.
- Lees, J. M. (1992). The magma system of Mount St. Helens: non-linear high-resolution P-wave tomography. *Journal of volcanology and geothermal research*, 53(1-4), 103–116.
- Lees, J. M. (2007). Seismic tomography of magmatic systems. *Journal of Volcanology and Geothermal Research*, 167(1-4), 37–56.
- Leroy, S., d’Acremont, E., Tiberi, C., Basuyau, C., Autin, J., Lucazeau, F., and Sloan, H. (2010). Recent off-axis volcanism in the eastern Gulf of Aden: implications for plume–ridge interaction. *Earth and Planetary Science Letters*, 293(1-2), 140–153.
- Leroy, S., Gente, P., Fournier, M., d’Acremont, E., Patriat, P., Beslier, M.-O., Bellahsen, N., Maia, M., Blais, A., Perrot, J., et al. (2004). From rifting to spreading in the eastern Gulf of Aden: a geophysical survey of a young oceanic basin from margin to margin. *Terra nova*, 16(4), 185–192.
- Ligorria, J. P., and Ammon, C. J. (1999). Iterative deconvolution and receiver-function estimation. *Bulletin of the seismological Society of America*, 89(5), 1395–1400.
- Lim, J.-A., Chang, S.-J., Mai, P. M., and Zahran, H. (2020). Asthenospheric flow of plume material beneath Arabia inferred from S wave traveltime tomography. *Journal of Geophysical Research: Solid Earth*, 125(8), e2020JB019668.
- Lin, G. (2013). Seismic investigation of magmatic unrest beneath Mammoth Mountain, California, USA. *Geology*, 41(8), 847–850.
- Lin, G., Amelung, F., Lavallée, Y., and Okubo, P. G. (2014). Seismic evidence for a crustal magma reservoir beneath the upper east rift zone of Kilauea volcano, Hawaii. *Geology*, 42(3), 187–190.
- Lithgow-Bertelloni, C., and Silver, P. G. (1998). Dynamic topography, plate driving forces and the African superswell. *Nature*, 395(6699), 269–272.
- Liu, C., Yao, H., Yang, H.-Y., Shen, W., Fang, H., Hu, S., and Qiao, L. (2019). Direct inversion for three-dimensional shear wave speed azimuthal anisotropy based on surface wave ray tracing: Methodology and application to Yunnan, southwest China. *Journal of Geophysical Research: Solid Earth*, 124(11), 11394–11413.

- Liu, H., and Niu, F. (2012). Estimating crustal seismic anisotropy with a joint analysis of radial and transverse receiver function data. *Geophysical Journal International*, 188(1), 144–164.
- Lomax, A., Virieux, J., Volant, P., and Berge-Thierry, C. (2000). Probabilistic earthquake location in 3D and layered models. In *Advances in seismic event location*, (pp. 101–134). Springer.
- Londoño, J. M., and Sudo, Y. (2003). Velocity structure and a seismic model for Nevado del Ruiz Volcano (Colombia). *Journal of Volcanology and Geothermal Research*, 119(1-4), 61–87.
- Long, M. D., and Becker, T. W. (2010). Mantle dynamics and seismic anisotropy. *Earth and Planetary Science Letters*, 297(3-4), 341–354.
- Lucazeau, F., Leroy, S., Autin, J., Bonneville, A., Goutorbe, B., Watremez, L., D’acremont, E., Düsünür, D., Rolandone, F., Huchon, P., et al. (2009). Post-Rift volcanism and high heat-flow at the ocean-continent transition of the eastern Gulf of Aden. *Terra nova*, 21(4), 285–292.
- Maccaferri, F., Rivalta, E., Keir, D., and Acocella, V. (2014). Off-rift volcanism in rift zones determined by crustal unloading. *Nature Geoscience*, 7(4), 297–300.
- Mackenzie, G., Thybo, H., and Maguire, P. (2005). Crustal velocity structure across the Main Ethiopian Rift: results from two-dimensional wide-angle seismic modelling. *Geophysical Journal International*, 162(3), 994–1006.
- MacQueen, P., Gottsmann, J., Pritchard, M. E., Young, N., Ticona J, F., Ticona, E. T., and Tintaya, R. (2021). Dissecting a Zombie: Joint analysis of density and resistivity models reveals shallow structure and possible sulfide deposition at Uturuncu Volcano, Bolivia. *Frontiers in Earth Science*, (p. 755).
- Maguire, P., Keller, G., Klemperer, S., Mackenzie, G., Keranen, K., Harder, S., O’reilly, B., Thybo, H., Asfaw, L., Khan, M., et al. (2006). Crustal structure of the northern Main Ethiopian Rift from the EAGLE controlled-source survey; a snapshot of incipient lithospheric break-up. *Geological Society, London, Special Publications*, 259(1), 269–292.
- Makris, J., and Ginzburg, A. (1987). The Afar Depression: transition between continental rifting and sea-floor spreading. *Tectonophysics*, 141(1-3), 199–214.
- Mallick, D., Gass, I., Cox, K., De Vries, B., and Tindle, A. G. (1990). Perim Island, a volcanic remnant in the southern entrance to the Red Sea. *Geological Magazine*, 127(4), 309–318.
- Manetti, P., Capaldi, G., Chiesa, S., Civetta, L., Conticelli, S., Gasparon, M., La Volpe, L., and Orsi, G. (1991). Magmatism of the eastern Red Sea margin in the northern part of Yemen from Oligocene to present. *Tectonophysics*, 198(2-4), 181–202.
- Manighetti, I., Tapponnier, P., Courtillot, V., Gallet, Y., Jacques, E., and Gillot, P.-Y. (2001). Strain transfer between disconnected, propagating rifts in Afar. *Journal of Geophysical Research: Solid Earth*, 106(B7), 13613–13665.

- Manighetti, I., Tapponnier, P., Courtillot, V., Gruszow, S., and Gillot, P.-Y. (1997). Propagation of rifting along the Arabia-Somalia plate boundary: The gulfs of Aden and Tadjoura. *Journal of Geophysical Research: Solid Earth*, 102(B2), 2681–2710.
- Manighetti, I., Tapponnier, P., Gillot, P., Jacques, E., Courtillot, V., Armijo, R., Ruegg, J., and King, G. (1998). Propagation of rifting along the Arabia-Somalia plate boundary: Into Afar. *Journal of Geophysical Research: Solid Earth*, 103(B3), 4947–4974.
- Martinez, F., and Cochran, J. R. (1988). Structure and tectonics of the northern Red Sea: catching a continental margin between rifting and drifting. *Tectonophysics*, 150(1-2), 1–31.
- Mattash, M., Pinarelli, L., Vaselli, O., Minissale, A., Jaimes-Viera, M., Al-Kadasi, M., Shawki, M., and Tassi, F. (2014). Geochemical evolution of southern Red Sea and Yemen flood volcanism: evidence for mantle heterogeneity. *Arabian Journal of Geosciences*, 7, 4831–4850.
- McClusky, S., Reilinger, R., Ogubazghi, G., Amleson, A., Healeb, B., Vernant, P., Sholan, J., Fisseha, S., Asfaw, L., Bendick, R., et al. (2010). Kinematics of the southern Red Sea–Afar Triple Junction and implications for plate dynamics. *Geophysical Research Letters*, 37(5).
- McGuire, A. V., and Bohannon, R. G. (1989). Timing of mantle upwelling: evidence for a passive origin for the Red Sea rift. *Journal of Geophysical Research: Solid Earth*, 94(B2), 1677–1682.
- McKenzie, D. (1978). Some remarks on the development of sedimentary basins. *Earth and Planetary science letters*, 40(1), 25–32.
- Mckenzie, D., and Bickle, M. (1988). The volume and composition of melt generated by extension of the lithosphere. *Journal of petrology*, 29(3), 625–679.
- McKenzie, D., Davies, D., and Molnar, P. (1970). Plate tectonics of the Red Sea and East Africa. *Nature*, 226(5242), 243–248.
- McKenzie, D. P., and Morgan, W. (1969). Evolution of triple junctions. *Nature*, 224(5215), 125–133.
- McNab, F., and Ball, P. W. (2023). meltPT: A Python package for basaltic whole-rock thermobarometric analysis with application to Hawai’i. *Volcanica*, 6(1).
- McQuarrie, N., Stock, J., Verdel, C., and Wernicke, B. (2003). Cenozoic evolution of Neotethys and implications for the causes of plate motions. *Geophysical research letters*, 30(20).
- Menzies, M. A., Klemperer, S. L., Ebinger, C. J., and Baker, J. (2002). Characteristics of volcanic rifted margins. *Special Papers-Geological Society of America*, (pp. 1–14).
- Miller, M. S., and Moresi, L. (2018). Mapping the Alaskan Moho. *Seismological Research Letters*, 89(6), 2430–2436.

- Miller, M. S., O'Driscoll, L. J., Porritt, R. W., and Roeske, S. M. (2018). Multiscale crustal architecture of Alaska inferred from P receiver functions. *Lithosphere*, 10(2), 267–278.
- Mohr, P. (1970). The Afar Triple Junction and sea-floor spreading. *Journal of Geophysical Research*, 75(35), 7340–7352.
- Mohr, P. (1978). Afar. *Annual Review of Earth and Planetary Sciences*, 6(1), 145–172.
- Mohr, P. (1983). Ethiopian flood basalt province. *Nature*, 303(5918), 577–584.
- Mohr, P. (1989). Nature of the crust under Afar: new igneous, not thinned continental. *Tectonophysics*, 167(1), 1–11.
- Mohr, P. (1991). Structure of Yemeni Miocene dike swarms and emplacement of coeval granite plutons. *Tectonophysics*, 198(2-4), 203–221.
- Molina, I., Kumagai, H., Le Pennec, J.-L., and Hall, M. (2005). Three-dimensional P-wave velocity structure of Tungurahua Volcano, Ecuador. *Journal of Volcanology and Geothermal Research*, 147(1-2), 144–156.
- Molnar, N. E., Cruden, A. R., and Betts, P. G. (2018). Unzipping continents and the birth of microcontinents. *Geology*, 46(5), 451–454.
- Molnar, P. (2015). Tectonics of continents. In *Plate Tectonics: A Very Short Introduction*. Oxford University Press.
URL <https://doi.org/10.1093/actrade/9780198728269.003.0006>
- Molnar, P., England, P. C., and Jones, C. H. (2015). Mantle dynamics, isostasy, and the support of high terrain. *Journal of Geophysical Research: Solid Earth*, 120(3), 1932–1957.
- Montelli, R., Nolet, G., Dahlen, F., and Masters, G. (2006). A catalogue of deep mantle plumes: New results from finite-frequency tomography. *Geochemistry, Geophysics, Geosystems*, 7(11).
- Morton, W., Black, R., Pilger, A., and Rosler, A. (1975). Afar depression of Ethiopia. In *Afar Depression of Ethiopia, Proceedings of an International Symposium on the Afar Region and Related Rift Problems, Bad Bergzabern. E. Schweizerbart_sche Verlagsbuchhandlung, FR*.
- Moucha, R., and Forte, A. M. (2011). Changes in African topography driven by mantle convection. *Nature Geoscience*, 4(10), 707–712.
- Moufti, M., and Hashad, M. (2005). Volcanic hazards assessment of Saudi Arabian Harrats: geochemical and isotopic studies of selected areas of active Makkah-Madinah-Nafud (MMN) volcanic rocks. *Final project Report (LGP-5-27)*, 679.
- Nelson, W., Furman, T., and Hanan, B. (2008). Sr, Nd, Pb and Hf evidence for two-plume mixing beneath the East African Rift System. *Geochimica et Cosmochimica Acta*, 72, A676.

- Neugebauer, H. J. (1978). Crustal doming and the mechanism of rifting Part I: Rift formation. *Tectonophysics*, 45(2-3), 159–186.
- Nicolas, A., and Christensen, N. I. (1987). Formation of anisotropy in upper mantle peridotites: A review. *Composition, structure and dynamics of the lithosphere-asthenosphere system*, 16, 111–123.
- Nobile, A., Pagli, C., Keir, D., Wright, T. J., Ayele, A., Ruch, J., and Acocella, V. (2012). Dike-fault interaction during the 2004 Dallol intrusion at the northern edge of the Erta Ale Ridge (Afar, Ethiopia). *Geophysical Research Letters*, 39(19).
- Nowacki, A., Wilks, M., Kendall, J.-M., Biggs, J., and Ayele, A. (2018). Characterising hydrothermal fluid pathways beneath Aluto volcano, Main Ethiopian Rift, using shear wave splitting. *Journal of Volcanology and Geothermal Research*, 356, 331–341.
- Nyblade, A. A., Owens, T. J., Gurrola, H., Ritsema, J., and Langston, C. A. (2000). Seismic evidence for a deep upper mantle thermal anomaly beneath east Africa. *Geology*, 28(7), 599–602.
- Nyblade, A. A., and Robinson, S. W. (1994). The African superswell. *Geophysical research letters*, 21(9), 765–768.
- Obrebski, M., Abers, G. A., and Foster, A. (2015). Magmatic arc structure around Mount Rainier, WA, from the joint inversion of receiver functions and surface wave dispersion. *Geochemistry, Geophysics, Geosystems*, 16(1), 178–194.
- Ogden, C., and Bastow, I. (2022). The crustal structure of the Anatolian Plate from receiver functions and implications for the uplift of the central and eastern Anatolian plateaus. *Geophysical Journal International*, 229(2), 1041–1062.
- Ogden, C., Bastow, I. D., Gilligan, A., and Rondenay, S. (2019). A reappraisal of the H- κ stacking technique: implications for global crustal structure. *Geophysical Journal International*, 219(3), 1491–1513.
- Ogubazghi, G., Goitom, B., Kibreab, A., and King, J. G. (2014). Relocation of the 31 March 2011 earthquake in the Nabro Volcanic Range of southern Eritrea. *Eritrean Journal of Science and Engineering*, 1(1), 93–102.
- Oppenheimer, C., Khalidi, L., Gratuze, B., Iverson, N., Lane, C., Vidal, C., Sahle, Y., Blegen, N., Yohannes, E., Donovan, A., et al. (2019). Risk and reward: Explosive eruptions and obsidian lithic resource at Nabro volcano (Eritrea). *Quaternary Science Reviews*, 226, 105995.
- Pagli, C., Wright, T. J., Ebinger, C. J., Yun, S.-H., Cann, J. R., Barnie, T., and Ayele, A. (2012). Shallow axial magma chamber at the slow-spreading Erta Ale Ridge. *Nature Geoscience*, 5(4), 284–288.
- Pallister, J. S., McCausland, W. A., Jónsson, S., Lu, Z., Zahran, H. M., Hadidy, S. E., Aburukbah, A., Stewart, I. C., Lundgren, P. R., White, R. A., et al. (2010). Broad accommodation of rift-related extension recorded by dyke intrusion in Saudi Arabia. *Nature Geoscience*, 3(10), 705–712.

- Park, Y., Nyblade, A. A., Rodgers, A. J., and Al-Amri, A. (2007). Upper mantle structure beneath the Arabian Peninsula and northern Red Sea from teleseismic body wave tomography: Implications for the origin of Cenozoic uplift and volcanism in the Arabian Shield. *Geochemistry, Geophysics, Geosystems*, 8(6).
- Park, Y., Nyblade, A. A., Rodgers, A. J., and Al-Amri, A. (2008). S wave velocity structure of the Arabian Shield upper mantle from Rayleigh wave tomography. *Geochemistry, Geophysics, Geosystems*, 9(7).
- Patane, D., Chiarabba, C., Cocina, O., De Gori, P., Moretti, M., and Boschi, E. (2002). Tomographic images and 3D earthquake locations of the seismic swarm preceding the 2001 Mt. Etna eruption: evidence for a dyke intrusion. *Geophysical Research Letters*, 29(10), 135–1.
- Paulatto, M., Minshull, T., Baptie, B., Dean, S., Hammond, J. O., Henstock, T., Kenedi, C., Kiddie, E., Malin, P., Peirce, C., et al. (2010). Upper crustal structure of an active volcano from refraction/reflection tomography, Montserrat, Lesser Antilles. *Geophysical Journal International*, 180(2), 685–696.
- Paulsen, T. S., and Wilson, T. J. (2010). New criteria for systematic mapping and reliability assessment of monogenetic volcanic vent alignments and elongate volcanic vents for crustal stress analyses. *Tectonophysics*, 482(1-4), 16–28.
- Permana, T., Nishimura, T., Nakahara, H., and Shapiro, N. (2022). Classification of volcanic tremors and earthquakes based on seismic correlation: application at sakurajima volcano, japan. *Geophysical Journal International*, 229(2), 1077–1097.
- Péron-Pinvidic, G., and Manatschal, G. (2009). The final rifting evolution at deep magma-poor passive margins from Iberia-Newfoundland: a new point of view. *International Journal of Earth Sciences*, 98(7), 1581–1597.
- Pik, R., Marty, B., and Hilton, D. (2006). How many mantle plumes in Africa? The geochemical point of view. *Chemical Geology*, 226(3-4), 100–114.
- Pikington, P. (1988). Uses of pressure and temperature data in exploration and now developments in overpressure detection. *Journal of petroleum technology*, 40(05), 543–549.
- Pilia, S., Rawlinson, N., Direen, N., Cummins, P. R., and Balfour, N. (2013). Structural controls on localized intraplate deformation and seismicity in Southern Australia: Insights from local earthquake tomography of the Flinders Ranges. *Journal of Geophysical Research: Solid Earth*, 118(5), 2176–2190.
- Plank, T., and Forsyth, D. (2016). Thermal structure and melting conditions in the mantle beneath the Basin and Range province from seismology and petrology. *Geochemistry, Geophysics, Geosystems*, 17, 1312–1338.
- Pratt, J. H. (1855). I. On the attraction of the Himalaya Mountains, and of the elevated regions beyond them, upon the plumb-line in India. *Philosophical Transactions of the Royal Society of London*, (145), 53–100.

- Priestley, K., and McKenzie, D. (2013). The relationship between shear wave velocity, temperature, attenuation and viscosity in the shallow part of the mantle. *Earth and Planetary Science Letters*, 381, 78–91.
- Pritchard, M., Mather, T., McNutt, S. R., Delgado, F., and Reath, K. (2019). Thoughts on the criteria to determine the origin of volcanic unrest as magmatic or non-magmatic. *Philosophical Transactions of the Royal Society A*, 377(2139), 20180008.
- Pürschel, M., Gloaguen, R., and Stadler, S. (2013). Geothermal activities in the Main Ethiopian Rift: hydrogeochemical characterization of geothermal waters and geothermometry applications (Dofan-Fantale, Gergede-Sodere, Aluto-Langano). *Geothermics*, 47, 1–12.
- Pyle, D. M., and Elliott, J. R. (2006). Quantitative morphology, recent evolution, and future activity of the Kameni Islands volcano, Santorini, Greece. *Geosphere*, 2(5), 253–268.
- Qaysi, S., Liu, K. H., and Gao, S. S. (2018). A database of shear-wave splitting measurements for the Arabian plate. *Seismological Research Letters*, 89(6), 2294–2298.
- Rappisi, F., VanderBeek, B., Faccenda, M., Morelli, A., and Molinari, I. (2022). Slab geometry and upper mantle flow patterns in the Central Mediterranean from 3D anisotropic P-wave tomography. *Journal of Geophysical Research: Solid Earth*, 127(5), e2021JB023488.
- Rawlinson, N., and Kennett, B. (2008). Teleseismic tomography of the upper mantle beneath the southern lachlan orogen, australia. *Physics of the Earth and Planetary Interiors*, 167(1-2), 84–97.
- Rawlinson, N., Reading, A. M., and Kennett, B. L. (2006). Lithospheric structure of Tasmania from a novel form of teleseismic tomography. *Journal of Geophysical Research: Solid Earth*, 111(B2).
- Rawlinson, N., and Sambridge, M. (2004a). Multiple reflection and transmission phases in complex layered media using a multistage fast marching method. *Geophysics*, 69(5), 1338–1350.
- Rawlinson, N., and Sambridge, M. (2004b). Wave front evolution in strongly heterogeneous layered media using the fast marching method. *Geophysical Journal International*, 156(3), 631–647.
- Rawlinson, N., Sambridge, M., et al. (2003). Seismic traveltime tomography of the crust and lithosphere. *Advances in geophysics*, 46, 81–199.
- Rawlinson, N., and Spakman, W. (2016). On the use of sensitivity tests in seismic tomography. *Geophysical Journal International*, 205(2), 1221–1243.
- Rawlinson, N., Tkalčić, H., and Reading, A. M. (2010). Structure of the Tasmanian lithosphere from 3D seismic tomography. *Australian Journal of Earth Sciences*, 57(4), 381–394.

- Rawlinson, N., and Urvoy, M. (2006). Simultaneous inversion of active and passive source datasets for 3-D seismic structure with application to Tasmania. *Geophysical Research Letters*, 33(24).
- Redfield, T., Wheeler, W., and Often, M. (2003). A kinematic model for the development of the Afar Depression and its paleogeographic implications. *Earth and Planetary Science Letters*, 216(3), 383–398.
- Reed, C. A., Almadani, S., Gao, S. S., Elsheikh, A. A., Cherie, S., Abdelsalam, M. G., Thurmond, A. K., and Liu, K. H. (2014). Receiver function constraints on crustal seismic velocities and partial melting beneath the Red Sea rift and adjacent regions, Afar Depression. *Journal of Geophysical Research: Solid Earth*, 119(3), 2138–2152.
- Reilinger, R., and McClusky, S. (2011). Nubia–Arabia–Eurasia plate motions and the dynamics of Mediterranean and Middle East tectonics. *Geophysical Journal International*, 186(3), 971–979.
- Reinsch, T., Dobson, P., Asanuma, H., Huenges, E., Poletto, F., and Sanjuan, B. (2017). Utilizing supercritical geothermal systems: a review of past ventures and ongoing research activities. *Geothermal Energy*, 5(1), 1–25.
- Renne, P. R., WoldeGabriel, G., Hart, W. K., Heiken, G., and White, T. D. (1999). Chronostratigraphy of the Miocene–Pliocene Sagantole Formation, Middle Awash Valley, Afar rift, Ethiopia. *Geological Society of America Bulletin*, 111(6), 869–885.
- Richards, F. D., Hoggard, M. J., White, N., and Ghelichkhan, S. (2020). Quantifying the relationship between short-wavelength dynamic topography and thermomechanical structure of the upper mantle using calibrated parameterization of anelasticity. *Journal of Geophysical Research: Solid Earth*, (p. e2019JB019062).
- Rime, V., Foubert, A., Ruch, J., and Kidane, T. (2023). Tectonostratigraphic evolution and significance of the Afar Depression. *Earth-Science Reviews*, (p. 104519).
- Ritsema, J., Deuss, A., Van Heijst, H., and Woodhouse, J. (2011). S40RTS: a degree-40 shear-velocity model for the mantle from new Rayleigh wave dispersion, teleseismic traveltime and normal-mode splitting function measurements. *Geophysical Journal International*, 184(3), 1223–1236.
- Ritsema, J., Heijst, H. J. v., and Woodhouse, J. H. (1999). Complex shear wave velocity structure imaged beneath Africa and Iceland. *Science*, 286(5446), 1925–1928.
- Ritsema, J., and Van Heijst, H.-J. (2000). Seismic imaging of structural heterogeneity in Earth’s mantle: evidence for large-scale mantle flow. *Science Progress (1933-)*, (pp. 243–259).
- Roman, D. C., and Cashman, K. V. (2006). The origin of volcano-tectonic earthquake swarms. *Geology*, 34(6), 457–460.
- Rooney, T. O. (2017). The Cenozoic magmatism of East-Africa: Part I—Flood basalts and pulsed magmatism. *Lithos*, 286, 264–301.
- Rooney, T. O. (2020a). The Cenozoic magmatism of East Africa: Part II—Rifting of the mobile belt. *Lithos*, 360, 105291.

- Rooney, T. O. (2020b). The Cenozoic magmatism of East Africa: Part III—rifting of the craton. *Lithos*, 360, 105390.
- Rooney, T. O. (2020c). The Cenozoic magmatism of East Africa: part IV—the terminal stages of rifting preserved in the Northern East African Rift System. *Lithos*, 360, 105381.
- Rooney, T. O., Herzberg, C., and Bastow, I. D. (2012). Elevated mantle temperature beneath East Africa. *Geology*, 40(1), 27–30.
- Ross, Z. E., Meier, M.-A., Hauksson, E., and Heaton, T. H. (2018). Generalized seismic phase detection with deep learning—short note. *Bulletin of the Seismological Society of America*, 108(5A), 2894–2901.
- Rudge, J. F., Champion, M. E. S., White, N., McKenzie, D., and Lovell, B. (2008). A plume model of transient diachronous uplift at the Earth's surface. *Earth and Planetary Science Letters*, 267(1–2), 146–160.
- Sambridge, M., and Kennett, B. (1986). A novel method of hypocentre location. *Geophysical Journal International*, 87(2), 679–697.
- Samrock, F., Grayver, A. V., Eysteinnsson, H., and Saar, M. O. (2018). Magnetotelluric image of transcrustal magmatic system beneath the Tulu Moye geothermal prospect in the Ethiopian Rift. *Geophysical Research Letters*, 45(23), 12–847.
- Sanders, C., Ponko, S., Nixon, L., and Schwartz, E. (1995). Seismological evidence for magmatic and hydrothermal structure in Long Valley caldera from local earthquake attenuation and velocity tomography. *Journal of Geophysical Research: Solid Earth*, 100(B5), 8311–8326.
- Sandvol, E., Seber, D., Barazangi, M., Vernon, F., Mellors, R., and Al-Amri, A. (1998). Lithospheric seismic velocity discontinuities beneath the Arabian Shield. *Geophysical Research Letters*, 25(15), 2873–2876.
- Sani, F., Ghinassi, M., Papini, M., Oms, O., and Finotello, A. (2017). Evolution of the northern tip of Afar triangle: Inferences from the Quaternary succession of the Dandiero—Massawa area (Eritrea). *Tectonophysics*, 717, 339–357.
- Schettino, A., Macchiavelli, C., Pierantoni, P. P., Zanoni, D., and Rasul, N. (2016). Recent kinematics of the tectonic plates surrounding the Red Sea and Gulf of Aden. *Geophysical Journal International*, 207(1), 457–480.
- Schouten, H., Klitgord, K. D., and Whitehead, J. A. (1985). Segmentation of mid-ocean ridges. *Nature*, 317(6034), 225–229.
- Sebai, A., Zumbo, V., Féraud, G., Bertrand, H., Hussain, A., Giannerini, G., and Campredon, R. (1991). $^{40}\text{Ar}/^{39}\text{Ar}$ dating of alkaline and tholeiitic magmatism of Saudi Arabia related to the early Red Sea rifting. *Earth and Planetary Science Letters*, 104(2–4), 473–487.
- Sembroni, A., Faccenna, C., Becker, T. W., Molin, P., and Abebe, B. (2016). Long-term, deep-mantle support of the Ethiopia-Yemen Plateau. *Tectonics*, (pp. 1–20).

- Sengör, A. C., and Burke, K. (1978). Relative timing of rifting and volcanism on Earth and its tectonic implications. *Geophysical Research Letters*, 5(6), 419–421.
- Sethian, J. A. (1996). A fast marching level set method for monotonically advancing fronts. *Proceedings of the National Academy of Sciences*, 93(4), 1591–1595.
- Sethian, J. A., and Popovici, A. M. (1999). 3-d traveltimes computation using the fast marching method. *Geophysics*, 64(2), 516–523.
- Shalev, E., Kenedi, C., Malin, P., Voight, V., Miller, V., Hidayat, D., Sparks, R., Minshull, T., Paulatto, M., Brown, L., et al. (2010). Three-dimensional seismic velocity tomography of Montserrat from the SEA-CALIPSO offshore/onshore experiment. *Geophysical Research Letters*, 37(19).
- Shapiro, N. M., Campillo, M., Stehly, L., and Ritzwoller, M. H. (2005). High-resolution surface-wave tomography from ambient seismic noise. *Science*, 307(5715), 1615–1618.
- Shapiro, N. M., Nakata, N., Gualtieri, L., and Fichtner, A. (2019). Applications with surface waves extracted from ambient seismic noise. *Seismic ambient noise*, 218.
- Shaw, J. E., Baker, J. A., Menzies, M. A., Thirlwall, M. F., and Ibrahim, K. M. (2003). Petrogenesis of the largest intraplate volcanic field on the Arabian Plate (Jordan): a mixed lithosphere–asthenosphere source activated by lithospheric extension. *Journal of Petrology*, 44(9), 1657–1679.
- Sichler, B. (1980). La bielle danakile: Un modèle pour l'évolution géodynamique de l'Afar. *Bulletin de la Société Géologique de France*, 22(6), 925–932.
- Silver, P. G., and Chan, W. W. (1991). Shear wave splitting and subcontinental mantle deformation. *Journal of Geophysical Research: Solid Earth*, 96(B10), 16429–16454.
- Silverii, F., Mantiloni, L., Rivalta, E., and Dahm, T. (2023). Lithospheric Sill Intrusions and Present-Day Ground Deformation at Rhenish Massif, Central Europe. *Geophysical Research Letters*, 50(23), e2023GL105824.
- Siniscalchi, A., Tripaldi, S., Neri, M., Giammanco, S., Piscitelli, S., Balasco, M., Behncke, B., Magri, C., Naudet, V., and Rizzo, E. (2010). Insights into fluid circulation across the Pernicana Fault (Mt. Etna, Italy) and implications for flank instability. *Journal of Volcanology and Geothermal Research*, 193(1-2), 137–142.
- Smithsonian (2023). Volcanoes of the World (v. 5.0.2).
URL <https://doi.org/10.5479/si.GVP.VOTW5-2022.5.0>
- Stab, M., Bellahsen, N., Pik, R., Quidelleur, X., Ayalew, D., and Leroy, S. (2016). Modes of rifting in magma-rich settings: Tectono-magmatic evolution of Central Afar. *Tectonics*, 35(1), 2–38.
- Stamps, D. S., Calais, E., Saria, E., Hartnady, C., Nocquet, J.-M., Ebinger, C. J., and Fernandes, R. M. (2008). A kinematic model for the East African Rift. *Geophysical Research Letters*, 35(5).

- Stein, C. A., and Cochran, J. R. (1985). The transition between the Sheba Ridge and Owen Basin: rifting of old oceanic lithosphere. *Geophysical Journal International*, 81(1), 47–74.
- Stein, S., Stein, C. A., Elling, R., Kley, J., Keller, G. R., Wyssession, M., Rooney, T., Frederiksen, A., and Moucha, R. (2018). Insights from North America’s failed Midcontinent Rift into the evolution of continental rifts and passive continental margins. *Tectonophysics*, 744, 403–421.
- Steinmann, R., Seydoux, L., Journeau, C., Shapiro, N. M., and Campillo, M. (2024). Machine learning analysis of seismograms reveals a continuous plumbing system evolution beneath the klyuchevskoy volcano in kamchatka, russia. *Journal of Geophysical Research: Solid Earth*, 129(3), e2023JB027167.
- Stephenson, S. N., Ball, P. W., and Richards, F. D. (2023). Destruction and regrowth of lithospheric mantle beneath large igneous provinces. *Science Advances*, 9(36), eadf6216.
- Stephenson, S. N., Hoggard, M. J., Holdt, M. C., and White, N. (2024). Continental residual topography extracted from global analysis of crustal structure. *forthcoming, Journal of Geophysical Research: Solid Earth*.
- Stephenson, S. N., White, N., Carter, A., Seward, D., Ball, P., and Klöcking, M. (2021). Cenozoic dynamic topography of Madagascar. *Geochemistry, Geophysics, Geosystems*, 22(6), e2020GC009624.
- Stern, R. J., and Johnson, P. (2010). Continental lithosphere of the Arabian Plate: a geologic, petrologic, and geophysical synthesis. *Earth-Science Reviews*, 101(1-2), 29–67.
- Stuart, G., Bastow, I., and Ebinger, C. (2006). Crustal structure of the northern Main Ethiopian Rift from receiver function studies. *Geological Society, London, Special Publications*, 259(1), 253–267.
- Takei, Y. (2002). Effect of pore geometry on VP/VS: From equilibrium geometry to crack. *Journal of Geophysical Research: Solid Earth*, 107(B2), ECV–6.
- Tang, Z., Mai, P. M., Julià, J., and Zahran, H. (2019). Shear velocity structure beneath Saudi Arabia from the joint inversion of P and S wave receiver functions, and Rayleigh wave group velocity dispersion data. *Journal of Geophysical Research: Solid Earth*, 124(5), 4767–4787.
- Tazieff, H., Varet, J., Barberi, F., and Giglia, G. (1972). Tectonic significance of the Afar (or Danakil) depression. *Nature*, 235(5334), 144–147.
- Teanby, N. A., Kendall, J.-M., and Van der Baan, M. (2004). Automation of shear-wave splitting measurements using cluster analysis. *Bulletin of the Seismological Society of America*, 94(2), 453–463.
- Tesfaye, S., Harding, D. J., and Kusky, T. M. (2003). Early continental breakup boundary and migration of the Afar triple junction, ethiopia. *Geological Society of America Bulletin*, 115(9), 1053–1067.

- Thatcher, W. (2007). Microplate model for the present-day deformation of Tibet. *Journal of Geophysical Research: Solid Earth*, 112(B1).
- Theys, N., Campion, R., Clarisse, L., Brenot, H., Gent, J. v., Dils, B., Corradini, S., Merucci, L., Coheur, P.-F., Roozendael, M. V., et al. (2013). Volcanic SO₂ fluxes derived from satellite data: a survey using OMI, GOME-2, IASI and MODIS. *Atmospheric Chemistry and Physics*, 13(12), 5945–5968.
- Thompson, D., Bastow, I., Helffrich, G., Kendall, J., Wookey, J., Snyder, D., and Eaton, D. (2010). Precambrian crustal evolution: seismic constraints from the Canadian Shield. *Earth and Planetary Science Letters*, 297(3-4), 655–666.
- Thompson, D., Hammond, J. O., Kendall, J.-M., Stuart, G., Helffrich, G., Keir, D., Ayele, A., and Goitom, B. (2015). Hydrous upwelling across the mantle transition zone beneath the Afar Triple Junction. *Geochemistry, Geophysics, Geosystems*, 16(3), 834–846.
- Thurber, C. (1993). Local earthquake tomography: velocities and V_p/V_s -Theory. *Seismic tomography: theory and practice*, (pp. 563–583).
- Tiberi, C., Ebinger, C., Ballu, V., Stuart, G., and Oluma, B. (2005). Inverse models of gravity data from the Red Sea-Aden-East African rifts triple junction zone. *Geophysical Journal International*, 163(2), 775–787.
- Toksöz, M. N., Cheng, C., and Timur, A. (1976). Velocities of seismic waves in porous rocks. *Geophysics*, 41(4), 621–645.
- Torsvik, T. H., Smethurst, M. A., Burke, K., and Steinberger, B. (2008). Long term stability in deep mantle structure: Evidence from the ~ 300 Ma Skagerrak-Centered Large Igneous Province (the SCLIP). *Earth and Planetary Science Letters*, 267(3-4), 444–452.
- Tortelli, G., Gioncada, A., Pagli, C., Braschi, E., Gebru, E., and Keir, D. (2022). Constraints on the magma source and rift evolution from geochemistry of the Stratoid flood basalts (Afar, Ethiopia). *Geochemistry, Geophysics, Geosystems*, 23(8), e2022GC010434.
- Turcotte, D. L., and Emerman, S. H. (1983). Mechanisms of active and passive rifting. In *Developments in geotectonics*, vol. 19, (pp. 39–50). Elsevier.
- Ukstins, I. A., Renne, P. R., Wolfenden, E., Baker, J., Ayalew, D., and Menzies, M. (2002). Matching conjugate volcanic rifted margins: ⁴⁰Ar/³⁹Ar chrono-stratigraphy of pre-and syn-rift bimodal flood volcanism in Ethiopia and Yemen. *Earth and Planetary Science Letters*, 198(3-4), 289–306.
- VanderBeek, B. P., Bue, R. L., Rappisi, F., and Faccenda, M. (2023). Imaging upper mantle anisotropy with traveltimes and splitting intensity observations from teleseismic shear waves: insights from tomographic reconstructions of subduction simulations. *Geophysical Journal International*, 235(3), 2640–2670.

- Vanorio, T., Virieux, J., Capuano, P., and Russo, G. (2005). Three-dimensional seismic tomography from P wave and S wave microearthquake travel times and rock physics characterization of the Campi Flegrei Caldera. *Journal of Geophysical Research: Solid Earth*, 110(B3).
- Varet, J. (2017). *Geology of Afar (East Africa)*. Springer.
- Varet, J., and Gasse, F. (1978). Notice de la carte g  ologique de l'Afar du Sud-Est. *ed CNRS, France*.
- Viltres, R., J  nsson, S., Ruch, J., Doubre, C., Reilinger, R., Floyd, M., and Ogubazghi, G. (2020). Kinematics and deformation of the southern Red Sea region from GPS observations. *Geophysical Journal International*, 221(3), 2143–2154.
- Walck, M. C. (1988). Three-dimensional VP/VS variations for the Coso region, California. *Journal of Geophysical Research: Solid Earth*, 93(B3), 2047–2052.
- Wang, Z., and Nur, A. (1986). Effect of temperature on wave velocities in sands and sandstones with heavy hydrocarbons: 56th Ann. *Internat. Mtg., Soc. Expl. Geophys.*
- Ward, K. M., Zandt, G., Beck, S. L., Christensen, D. H., and McFarlin, H. (2014). Seismic imaging of the magmatic underpinnings beneath the Altiplano-Puna volcanic complex from the joint inversion of surface wave dispersion and receiver functions. *Earth and Planetary Science Letters*, 404, 43–53.
- Watanabe, T. (1993). Effects of water and melt on seismic velocities and their application to characterization of seismic reflectors. *Geophysical Research Letters*, 20(24), 2933–2936.
- Wegener, A. (1912). Die entstehung der kontinente. *Geologische Rundschau*, 3(4), 276–292.
- Wessel, P., Luis, J., Uieda, L., Scharroo, R., Wobbe, F., Smith, W. H., and Tian, D. (2019). The generic mapping tools version 6. *Geochemistry, Geophysics, Geosystems*, 20(11), 5556–5564.
- White, R., and McKenzie, D. (1989). Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. *Journal of Geophysical Research: Solid Earth*, 94(B6), 7685–7729.
- White, R. S., Edmonds, M., MacLennan, J., Greenfield, T., and Agustsdottir, T. (2019). Melt movement through the Icelandic crust. *Philosophical Transactions of the Royal Society A*, 377(2139), 20180010.
- Wiar, P., and Oppenheimer, C. (2005). Large magnitude silicic volcanism in north Afar: the Nabro Volcanic Range and Ma'alalta volcano. *Bulletin of Volcanology*, 67(2), 99–115.
- Wilks, M. (2016). *A Seismological Investigation into Tectonic, Magmatic and Hydrothermal Processes at Aluto and Corbetti, Two Restless Volcanoes in the Main Ethiopian Rift*. Ph.D. thesis, University of Bristol.

- Wilks, M., Kendall, J.-M., Nowacki, A., Biggs, J., Wookey, J., Birhanu, Y., Ayele, A., and Bedada, T. (2017). Seismicity associated with magmatism, faulting and hydrothermal circulation at Aluto Volcano, Main Ethiopian Rift. *Journal of Volcanology and Geothermal Research*, 340, 52–67.
- Wilks, M., Rawlinson, N., Kendall, J.-M., Nowacki, A., Biggs, J., Ayele, A., and Wookey, J. (2020). The coupled magmatic and hydrothermal systems of the restless Aluto caldera, Ethiopia.
- Wilson, J. T. (1965). A new class of faults and their bearing on continental drift. *Nature*, 207(4995), 343–347.
- Wilson, J. T., et al. (1966). Did the Atlantic close and then re-open?
- Winkler, K., and Nur, A. (1979). Pore fluids and seismic attenuation in rocks. *Geophysical Research Letters*, 6(1), 1–4.
- Woldegabriel, G., Aronson, J. L., and Walter, R. C. (1990). Geology, geochronology, and rift basin development in the central sector of the Main Ethiopian Rift. *Geological Society of America Bulletin*, 102(4), 439–458.
- Wolfenden, E., Ebinger, C., Yirgu, G., Deino, A., and Ayalew, D. (2004). Evolution of the northern Main Ethiopian rift: birth of a triple junction. *Earth and Planetary Science Letters*, 224(1-2), 213–228.
- Wolfenden, E., Ebinger, C., Yirgu, G., Renne, P. R., and Kelley, S. P. (2005). Evolution of a volcanic rifted margin: Southern Red Sea, Ethiopia. *Geological Society of America Bulletin*, 117(7-8), 846–864.
- Wood, C. A. (1983). Continental rift jumps. In *Developments in Geotectonics*, vol. 19, (pp. 529–540). Elsevier.
- Wookey, J. (2012). Direct probabilistic inversion of shear wave data for seismic anisotropy. *Geophysical Journal International*, 189(2), 1025–1037.
- Wright, T. J., Ebinger, C., Biggs, J., Ayele, A., Yirgu, G., Keir, D., and Stork, A. (2006). Magma-maintained rift segmentation at continental rupture in the 2005 Afar dyking episode. *Nature*, 442(7100), 291–294.
- Wright, T. J., Sigmundsson, F., Pagli, C., Belachew, M., Hamling, I. J., Brandsdóttir, B., Keir, D., Pedersen, R., Ayele, A., Ebinger, C., et al. (2012). Geophysical constraints on the dynamics of spreading centres from rifting episodes on land. *Nature Geoscience*, 5(4), 242–250.
- Wuestefeld, A., Al-Harrasi, O., Verdon, J. P., Wookey, J., and Kendall, J. M. (2010). A strategy for automated analysis of passive microseismic data to image seismic anisotropy and fracture characteristics. *Geophysical Prospecting*, 58(5), 755–773.
- Wüstefeld, A., Bokelmann, G., Zaroli, C., and Barruol, G. (2008). SplitLab: A shear-wave splitting environment in Matlab. *Computers & Geosciences*, 34(5), 515–528.

- Xu, W., Rivalta, E., and Li, X. (2017). Magmatic architecture within a rift segment: Articulate axial magma storage at Erta Ale volcano, Ethiopia. *Earth and Planetary Science Letters*, 476, 79–86.
- Yamauchi, H., and Takei, Y. (2016). Polycrystal anelasticity at near-solidus temperatures. *Journal of Geophysical Research: Solid Earth*, 121(11), 7790–7820.
- Yang, Y., Ritzwoller, M. H., Levshin, A. L., and Shapiro, N. M. (2007). Ambient noise Rayleigh wave tomography across Europe. *Geophysical Journal International*, 168(1), 259–274.
- Yohannes, E. (2012). Geothermal development in Eritrea: A country update. In *Proceedings 4th African Rift Geothermal Conference*.
- Zandt, G., and Ammon, C. J. (1995). Continental crust composition constrained by measurements of crustal Poisson's ratio. *Nature*, 374(6518), 152–154.
- Zelt, B., and Ellis, R. (1999). Receiver-function studies in the Trans-Hudson orogen, Saskatchewan. *Canadian Journal of Earth Sciences*, 36(4), 585–603.
- Zenonos, A., De Siena, L., Widiyantoro, S., and Rawlinson, N. (2019). P and S wave travel time tomography of the SE Asia-Australia collision zone. *Physics of the Earth and Planetary Interiors*, 293, 106267.
- Zhu, L., and Kanamori, H. (2000). Moho depth variation in southern California from teleseismic receiver functions. *Journal of Geophysical Research: Solid Earth*, 105(B2), 2969–2980.
- Zhu, W., and Beroza, G. C. (2019). PhaseNet: a deep-neural-network-based seismic arrival-time picking method. *Geophysical Journal International*, 216(1), 261–273.