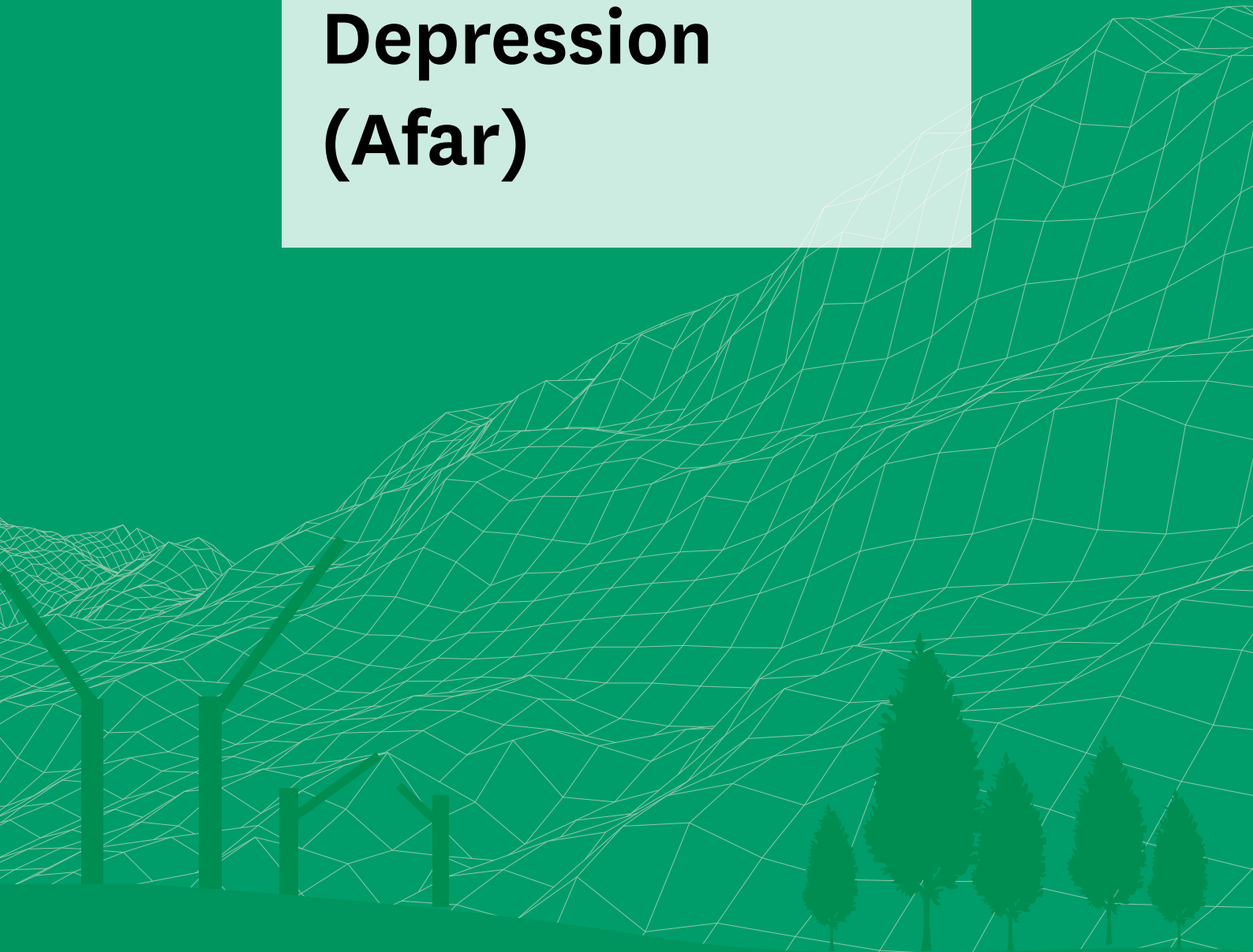


Valentin Rime

Multi-scale basin dynamic study of the Danakil Depression (Afar)



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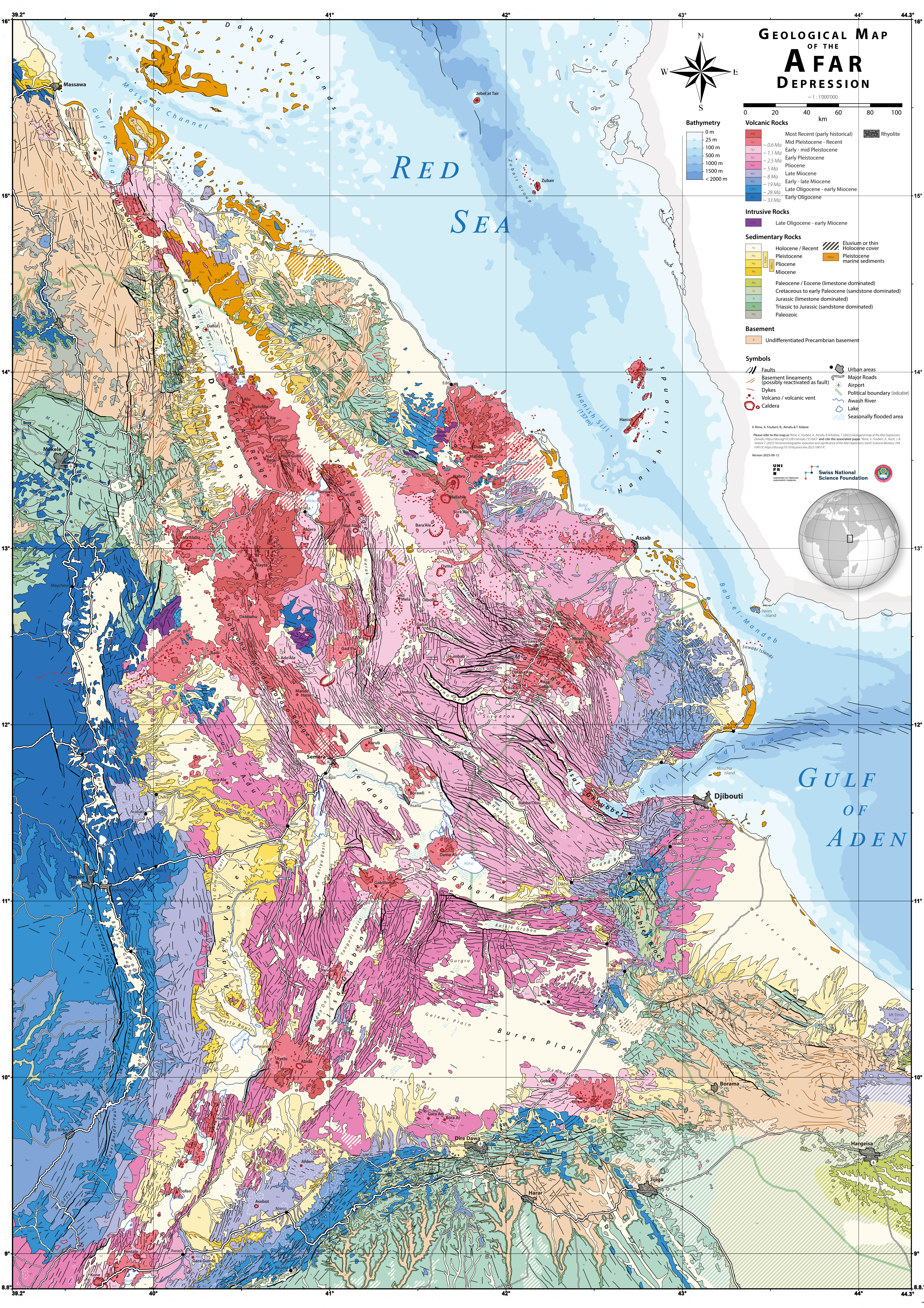


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Valentin RIME

Multi-scale basin dynamic study of the Danakil Depression (Afar)



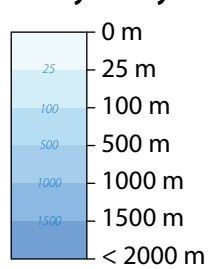


GEOLOGICAL MAP OF THE AFAR DEPRESSION

~ 1 : 1'000'000

0 20 40 60 80 100
km

Bathymetry



Volcanic Rocks

~ 0.6 Ma	Most Recent (early historical)
~ 1.1 Ma	Mid Pleistocene - Recent
~ 2.5 Ma	Early - mid Pleistocene
~ 5 Ma	Early Pleistocene
~ 8 Ma	Pliocene
~ 19 Ma	Late Miocene
~ 28 Ma	Early - late Miocene
~ 33 Ma	Late Oligocene - early Miocene
	Early Oligocene

Intrusive Rocks

	Late Oligocene - early Miocene
--	--------------------------------

Sedimentary Rocks

Hs	Holocene / Recent	Eluvium or thin Holocene cover
Pls	Pleistocene	Pleistocene marine sediments
Pli	Pliocene	
Mi	Miocene	
Pa	Paleocene / Eocene (limestone dominated)	
Cr	Cretaceous to early Paleocene (sandstone dominated)	
Ju	Jurassic (limestone dominated)	
Tri	Triassic to Jurassic (sandstone dominated)	
Pz	Paleozoic	

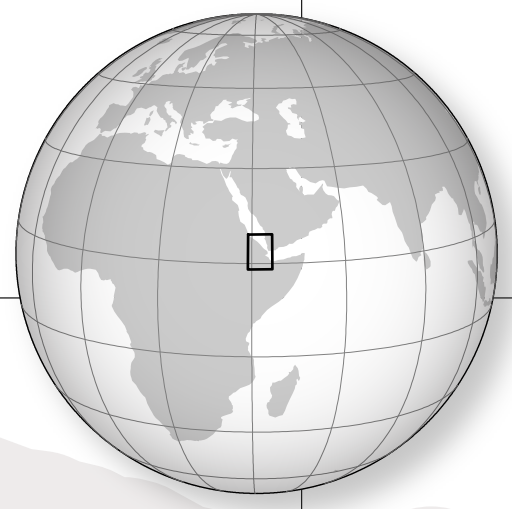
Basement

	Undifferentiated Precambrian basement
--	---------------------------------------

Symbols

///	Faults	•	Urban areas
---	Basement lineaments (possibly reactivated as fault)	✈	Major Roads
---	Dykes	✈	Airport
•	Volcano / volcanic vent	---	Political boundary (indicative)
•	Caldera	---	Awash River
		---	Lake
		---	Seasonally flooded area

V. Rime, A. Foubert, B. Atrushi & T. Kidane
Please refer to this map as Rime, V., Foubert, A., Atrushi, B. & Kidane, T. (2023) Geological map of the Afar Depression, Ethiopia. <https://doi.org/10.5281/zenodo.7515447> and cite the associated paper: Rime, V., Foubert, A., Atrushi, B. & Kidane, T. (2023) Tectonostratigraphic evolution and significance of the Afar Depression. *Earth-Science Reviews*, 244, 104519. <https://doi.org/10.1016/j.earscirev.2023.104519>
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Department of Geosciences
University of Fribourg (Switzerland)

MULTI-SCALE BASIN DYNAMIC STUDY OF THE DANAKIL DEPRESSION (AFAR)

THESIS

presented to the Faculty of Science and Medicine of the University of Fribourg
(Switzerland) in consideration for the award of the academic grade of

Doctor of Philosophy in Earth Sciences

by

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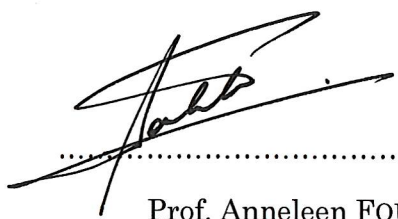
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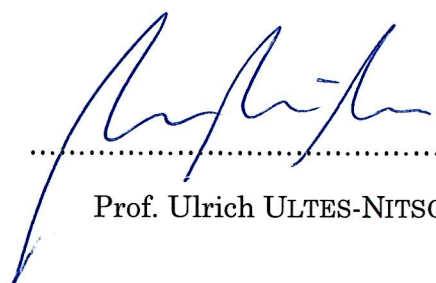
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ABSTRACT

Rifts and passive margins are fundamental features of the structure of Planet Earth. They mark the transition from continents to oceans and, as such, are an essential component of plate tectonics. However, the interactions between tectonics, sedimentology and magmatism controlling rift structure and evolution are poorly known due to the complexity of the factors playing at very different temporal and spatial scales. Because they are the locus of large sediment accumulations, rift basin form detailed archives to reconstruct paleoenvironmental and paleoclimatic conditions and host important natural resources.

The Danakil Depression is part of the Afro-Arabian Rift System and is the only emerged region on Earth currently experiencing the transition from continental rifting to oceanic spreading. As such, it provides a unique natural laboratory to study rifting processes. Despite its indisputable interest, scientific knowledge of this region remains fragmented due to challenging field access. The aim of the present work is to better understand the basin dynamics of the Danakil Depression using a multi-scale and multi-disciplinary approach.

An integrative review of the tectonic, magmatic and sedimentological history of the Afar region and the compilation of new geological maps allows to develop a new model for the evolution of the Afar Depression. This results in the better understanding of the regional rift system

kinematics. In particular, results highlight that rifting is propagating northward in Afar. Results also evidence that the Danakil Depression and Central Afar have a different structure and contrasting geological evolution. They represent two different rift types and not two different stages of evolution of the same type of rift. Differences can be explained by the influence of the Afar hotspot.

The gradual and active transition from continent to ocean is also reflected in the Pleistocene to Holocene environmental conditions of the Danakil Depression. Field studies evidence at least four times flooding of the basin by the Red Sea during the Quaternary. Field data complemented by remote sensing allow to characterize the parameters controlling the marine incursions in the basin. Tectonic uplift of the rift shoulder modulates the height of the gateway between the Red Sea and the Danakil Depression and forms the long-term, first-order control. The short-term, second-order control is eustatic sea-level variations. The tectonic uplift explains why the basin was not flooded during the Holocene. Tentative predictions on the future flooding of the Danakil Depression are discussed.

The last marine incursion occurred ~130 ky ago, resulting in the deposition of several hundred of meters of evaporites during its desiccation stage. The study of a 625 m deep core drilled in the central part of the Danakil Depression allows to

characterize the paleoenvironmental and paleoclimatic conditions of the Depression since the Late Pleistocene. This study furthermore identifies the mechanisms of evaporite formation, providing important clues on the understanding of older enigmatic *salt giant* deposits in Earth history.

Finally, the hypersaline Lake Afdera is a peculiar Recent sedimentary setting in the Danakil Depression. The origin of the Lake is discussed based on bathymetric mapping. Bathymetry shows that the deepest part of the Lake represents the lowest known elevation of the Danakil Depression. This is explained by the interaction of faults of different orientations representing the intersection of the rift axis with two transfer zones.

Overall, the present study allows the better understanding of the evolution, the chronology, the structure and the paleoenvironmental and paleoclimatic conditions of the Danakil and Afar Depression, as well as some of the mechanisms controlling these dynamic and complex systems. This regional study gives new insights on the understanding of rifting processes shaping our Planet since more than two billion of years. The results of this thesis will also have an impact on the exploration, management and exploitation of natural resources in this region, such as groundwater but also geothermal energy, salt, potash, and other mineral resources. Finally, results may contribute to highlight and protect the unique and fascinating geology of the Danakil Depression.

RÉSUMÉ

Les rifts et les marges passives sont des éléments fondamentaux de la structure de notre planète. Ils marquent la transition entre continents et océans et sont ainsi une composante essentielle de la tectonique des plaques. Cependant, les interactions entre la tectonique, la sédimentologie et le magmatisme qui contrôlent la structure et l'évolution d'un rift sont mal comprises en raison de la complexité des facteurs qui l'influencent à des échelles de temps et d'espace très variables. Les rifts accumulent de grandes quantités de sédiments qui sont autant d'enregistrement paleoenvironnementaux et paléoclimatiques précieux. Ils sont également une source importante de ressources naturelles.

La Dépression de Danakil fait partie du Système de Rift Afro-Arabique et est la seule région émergée au monde qui soit en train de passer d'un rift continental à une croute océanique. Il représente donc un formidable laboratoire à ciel ouvert pour étudier ces processus de rifting. Malgré son intérêt indéniable, les connaissances scientifiques de cette région restent parcellaires en raison des difficultés de conduire des études de terrain. Le but de cette thèse est de mieux comprendre la dynamique du bassin de Danakil grâce à une approche multidisciplinaire à différentes échelles.

Une revue holistique de l'histoire tectonique, magmatique et sédimentaire de la région de l'Afar, ainsi que la compilation de nouvelles cartes géologiques a permis de développer un

nouveau modèle d'évolution de la Dépression de l'Afar. Ces résultats apportent une meilleure compréhension de la cinématique régionale du rift. En particulier, ils montrent qu'en Afar, le rift se propage vers le nord et non vers le sud. Ils montrent aussi que la Dépression de Danakil et l'Afar Central ont une structure et une histoire géologique différente. Ces deux régions représentent deux types de rift distincts et pas deux étapes de l'évolution d'un même type de rift. Ces différences peuvent être expliquées par l'influence du hotspot de l'Afar.

La transition graduelle d'un continent à un océan se manifeste également dans les conditions environnementales de la Dépression de Danakil durant le Pléistocène et l'Holocène. Des explorations de terrain ont montrés que la Mer Rouge a inondé ce bassin au moins quatre fois durant le Quaternaire. Des études de terrain et par télédétection permettent de caractériser les paramètres qui contrôlent ces incursion marines. Les variations eustatiques du niveau marin sont le mécanisme dominant mais les mouvements tectoniques de surrection des épaules du rift modulent l'élévation du passage entre la Mer Rouge et la Dépression de Danakil. Ce phénomène de surrection explique pourquoi la dépression n'est pas inondée de nos jours. Des prédictions sur l'inondation future de la dépression sont également discutées.

La dernière incursion marine a eu lieu il y a environ 130 ka et a laissé derrière

elle plusieurs centaines de mètres d'évaporites durant son assèchement. L'étude d'une carotte de 625 m forée dans la partie centrale du bassin a permis de caractériser les conditions paléoenvironnementales et paleoclimatiques de la Dépression depuis le Pleistocène Supérieur. Cette étude a également permis d'identifier les mécanismes de formation de ces évaporites et contribue à une meilleure compréhension des énigmatiques *géants de sels* qui ont ponctué l'histoire géologique de notre planète.

Enfin, le lac salé d'Afdera est un exemple particulier d'environnement sédimentaire récent de la Dépression de Danakil. L'origine de ce lac est discutée basé sur une nouvelle carte bathymétrique du lac. Elle montre que la partie la plus profonde du lac correspond à l'élévation connue la plus basse dans la Dépression de Danakil. Cette particularité est expliquée par l'interaction de failles de différentes orientations qui représentent l'intersection de l'axe du rift avec deux *transfer zones*.

Ce travail permet de mieux comprendre l'évolution, la chronologie, la structure et les conditions paléoenvironnementales et paleoclimatiques des Dépression de Danakil et de l'Afar, mais aussi certains des mécanismes qui contrôlent ces systèmes complexes. Ils donnent des indications utiles à la compréhension des processus de rifting qui ont modelé notre planète depuis plus de deux milliards d'années. Les résultats de cette thèse peuvent également bénéficier à la population locale, leur permettant de mieux comprendre et gérer les ressources naturelles de leur terre, en premier lieu l'eau, mais aussi la géothermie, le sel, la

potasse et d'autre ressources minérales. Finalement, ces résultats peuvent contribuer à la mise en valeur et la protection de l'exceptionnelle géologie de la Dépression de Danakil.

ZUSAMMENFASSUNG

Riftzonen und passive Kontinentalränder sind grundlegende Merkmale der Struktur unseres Planeten. Sie markieren den Übergang von Kontinenten zu Ozeanen und sind darum ein wesentlicher Bestandteil der Plattentektonik. Allerdings sind die Wechselwirkungen zwischen Tektonik, Sedimentologie und Magmatismus, die die Struktur und Entwicklung von Riftzonen steuern, aufgrund der Komplexität der Faktoren, kaum bekannt. Diese Faktoren wirken auf sehr unterschiedlichen zeitlichen und räumlichen Massstäben. Riftbecken akkumulieren grosse Mengen an Sedimenten, die detaillierte Archive zur Rekonstruktion der paläoökologischen und paläoklimatischen Bedingungen darstellen. Sie sind außerdem eine wichtige Quelle für natürliche Ressourcen.

Die Danakil-Senke ist Teil des Afro-Arabischen Grabensystems und die einzige nicht unter Wasser stehende Region der Erde, die sich im Übergang von einem Kontinentalrift zu einer ozeanischen Kruste befindet. Sie bietet daher ein einzigartiges natürliches Labor zur Untersuchung von Rifting-Prozessen. Trotz ihres unbestreitbaren Interesses sind die wissenschaftlichen Kenntnisse über diese Region, aufgrund der Schwierigkeiten bei der Durchführung von Feldstudien, lückenhaft. Das Ziel dieser Dissertation ist es, die Dynamik des Danakil-Beckens durch einen multidisziplinären Ansatz auf

verschiedenen Massstäben besser zu verstehen.

Ein integrativer Überblick über die tektonische, magmatische und sedimentäre Geschichte der Afar-Region und die Zusammenstellung neuer geologischer Karten ermöglicht die Entwicklung eines neuen Modells für die Entwicklung des Afar-Rifts. Diese Ergebnisse tragen zu einem besseren Verständnis der regionalen Kinematik des Rifts bei. Insbesondere zeigen die Ergebnisse, dass sich das Rifting in der Afar-Region nach Norden und nicht nach Süden ausbreitet. Die Ergebnisse belegen auch, dass die Danakil-Senke und Zentral-Afar eine unterschiedliche Struktur und eine gegensätzliche geologische Entwicklung aufweisen. Diese beiden Regionen repräsentieren zwei unterschiedliche Rifttypen und nicht zwei verschiedene Entwicklungsstadien desselben Rifttyps. Die Unterschiede lassen sich durch den Einfluss des Afar-Hotspots erklären.

Der allmähliche Übergang vom Kontinent zum Ozean zeigt sich in den Umweltbedingungen der Danakil-Senke während des Pleistozäns und des Holozäns. Feldstudien zeigen im Quartär mindestens vier Überflutungen des Beckens durch das Rote Meer. Feldstudien und Fernerkundung ermöglichen es, die Parameter zu charakterisieren, die die marinen Überflutungen des Beckens steuern. Eustatische Meeresspiegelschwankungen sind der dominierende Mechanismus,

aber die tektonische Hebung der Riftschultern moduliert die Höhe des Durchgangs zwischen dem Roten Meer und der Danakil-Senke. Die tektonische Hebung erklärt, warum das Becken während des Holozäns nicht überflutet wurde. Vorhersagen über die zukünftige Überflutung der Danakil-Senke werden ebenfalls diskutiert.

Der letzte Meereseinbruch fand vor etwa 130'000 Jahren statt und hinterliess während seiner Austrocknung mehrere hundert Meter Evaporite. Die Untersuchung eines 625 m tiefen Bohrkerns, der im zentralen Teil der Danakil-Senke gebohrt wurde, ermöglicht die Charakterisierung der paläoökologischen und paläoklimatischen Bedingungen der Senke seit dem Spätpleistozän. Darüber hinaus werden in dieser Studie die Mechanismen der Evaporitbildung identifiziert. Sie liefern wichtige Hinweise über das Verständnis älterer rätselhafter *Salzgiganten*.

Schliesslich stellt der Afdera-Salzsee ein besonderes Beispiel eines rezenten Sedimentbeckens in der Danakil-Senke dar. Der Ursprung des Sees wird anhand von bathymetrischen Kartierungen diskutiert. Die Bathymetrie zeigt, dass der tiefste Teil des Sees die niedrigste bekannte Höhe in der Danakil-Senke darstellt. Diese Besonderheit wird durch die Interaktion von Verwerfungen unterschiedlicher Ausrichtung erklärt, die den Schnittpunkt der Riftachse mit zwei Transferzonen darstellen.

Insgesamt ermöglicht die vorliegende Studie ein besseres Verständnis der Entwicklung, der Chronologie, der Struktur und der paläoökologischen und paläoklimatischen Bedingungen der Danakil- und Afar-Senke sowie einiger der Mechanismen, die diese dynamischen

und komplexen Systeme steuern. Diese regionale Studie gibt neue Einblicke in das Verständnis der Rifting-Prozesse, die unseren Planeten seit mehr als zwei Milliarden Jahren prägen. Die Ergebnisse dieser Dissertation können auch der lokalen Bevölkerung zugutekommen, indem sie ihr helfen, die natürlichen Ressourcen ihres Landes besser zu verstehen und zu verwalten, in erster Linie Wasser, aber auch Geothermie, Salz, Kalisalz, und andere Mineralressourcen. Schliesslich können die Ergebnisse dazu beitragen, die einzigartige und faszinierende Geologie der Danakil-Senke hervorzuheben und zu schützen.

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I INTRODUCTION

Rifts and passive margins are fundamental and ubiquitous features of the structure of our Planet. They are elemental parts of plate tectonics and of the Wilson cycle (Wilson, 1966; Burke, 1978; Şengör & Natal'in, 2001). Wilson cycles shape our planet probably since more than two billions of years (Korenaga, 2013) and allowed life to develop on Earth (e.g. Parnell, 2004). Wilson cycles and rifting processes are believed to have profound and long term influence on climate (e.g. Fischer, 1984; Worsley *et al.*, 1984; Nance *et al.*, 2014). Rifts form important sedimentary basins. As such, their sedimentary records provide unique paleoenvironmental and paleoclimatic archives that can help better understand past climate at a relatively high spatial and temporal resolution. Rift sediments are furthermore hosting important natural resources, including hydrocarbons, salts, potash but also offering CO₂ sequestration opportunities and geothermal energy potential. The dynamic nature of rifts also comes with natural hazards such as earthquakes, volcanic eruptions and tsunamis. The study of which can help in natural hazard management and assessment. Eventually, rifts, passive margins and their sediments get trapped into orogenic belts forming the building bricks of continental crust.

However, knowledge on the tectonic structure and the sedimentary architecture of rifts as well as their evolution and the factors controlling them remains fragmented. In particular, many

questions remain on the coupling and feedback between tectonics, magmatism (including plume) and sedimentology (Hill, 1991; Storey, 1995; Courtillot *et al.*, 1999; Paton *et al.*, 2017; Pérez-Gussinyé *et al.*, 2020; Sapin *et al.*, 2021; Chenin *et al.*, 2022). The interaction between different rifts and rift segments, linkage mechanisms, rift jump and failed rifts are complex (Müller *et al.*, 2001; Pagli *et al.*, 2014; Molnar *et al.*, 2019; Zwaan & Schreurs, 2020; Neuharth *et al.*, 2021; La Rosa *et al.*, 2022) and, as such, the basin dynamics, the interaction between sedimentary basins, the flooding mechanisms and the factors controlling deposition of a wide variety of sediments (in particular evaporites) in incipient and restricted ocean basin remain poorly understood (Orszag-Sperber *et al.*, 1998; Jackson *et al.*, 2000; Bryant *et al.*, 2012; Hudec *et al.*, 2013; Strozyk *et al.*, 2017; Kukla *et al.*, 2018; Rodriguez *et al.*, 2018; Mitchell *et al.*, 2021; Decalf & Heyn, 2022). These unknowns are partly explained by the inaccessibility of late syn-rift sediments, often situated offshore and buried under kilometres of post-rift sediments. Datings in these older rift systems also feature large uncertainties, preventing a good understanding of the rates and velocities of geological processes. The shortcomings in understanding rift systems are also explained by the complexity of these systems, where a multitude of controlling and interconnected factors are influencing each other at a wide variety of temporal and spatial scales.

The Afar Depression is probably the best location on Earth to study these rift processes. It is the only onshore region nearing continental breakup. It furthermore experiences the unique influence of a hotspot. The Danakil Depression forms the northernmost part of the Afar Depression. It probably represents the part of Afar that is the closest to ocean spreading and, as such, represents an advanced stage of rifting (Hayward & Ebinger, 1996; Bastow & Keir, 2011; Keir *et al.*, 2013). The gradual transition from continent to ocean is reflected in the sediments of the Danakil Depression. Coralgall reefs evidenced several flooding phases of the basin by the Red Sea during the Quaternary (Atnafu *et al.*, 2015; Jaramillo-Vogel *et al.*, 2019). These floodings and subsequent desiccations filled the basin with important evaporite deposits (Holwerda & Hutchinson, 1968). The young age of these geological processes allows a high temporal resolution with very precise datings and a good understanding of external factors like eustatic sea level changes. Its emerged situation and the nearly complete absence of vegetation also allows a very high spatial resolution as outcrops are easier to study by remote sensing or fieldwork. The area therefore represents a unique opportunity to study the interactions between tectonics, magmatism, climate, hydrology, biosphere and sedimentation in particular for the transition phase between continental and marine conditions and for *salt giant* formation.

The Afar Depression also presents an important sedimentary and paleoclimatic archive in a region where only few records were studied (Gasse & Street, 1978; Gillespie *et al.*, 1983; Gasse & Fontes, 1989; Hailemichael *et al.*, 2002; Fleitmann

et al., 2011; Tierney *et al.*, 2017; Lamb *et al.*, 2018; Schaebitz *et al.*, 2021; Foerster *et al.*, 2022). The study of paleoclimatic and paleoenvironmental conditions is particularly important in this area where some of our oldest ancestors have been discovered, in particular the famous Lucy (Johanson *et al.*, 1982) and the oldest known *Homo* (Villmoare *et al.*, 2015), among others. The paleoenvironment and paleoclimate of the Danakil Depression itself has only been briefly investigated for the Holocene in Lake Afdera (Gasse, 1974) and for the Early to Middle Pleistocene in the Dandiero Basin (Abbate *et al.*, 2004; Ghinassi *et al.*, 2009) but no records have been published for most of the Mid-Pleistocene to Late Pleistocene.

The local population of the Danakil Depression is exploiting the geology of the basin since centuries with the mining and trade of salt (Munzinger, 1869). Since the XXth century, this activity expanded to the exploitation of potash minerals mainly used as fertilizers (Holwerda & Hutchinson, 1968; Warren, 2016). The knowledge on these resources is, however, fragmented. The local population is also in need of freshwater and energy resources, as well as impacted often by natural hazards such as earthquakes and volcanic eruptions (Gouin, 1979; Battistelli *et al.*, 2002; Goitom *et al.*, 2015; Gardo & Varet, 2018; Varet, 2018; Varet & Gardo, 2020; Rizzello *et al.*, 2021; Ruch *et al.*, 2021; Hearn, 2022). The better understanding of local geology and hydrogeology would certainly be beneficial for the local population. Furthermore, scientific work may provide material for the local communities to highlight and valorise the unique geology of the Danakil Depression with the development of local and sustainable geotourism, but also to protect these exceptional geotopes.

Despite the high scientific interest, few field studies have been performed in the Danakil Depression. The exploration of the Danakil Depression is challenging. Dallol holds the world record for the highest annual mean temperature (Cerveny *et al.*, 2007) and the area is also exceptionally dry, lacking freshwater sources. The region was furthermore often not accessible due to geopolitical conflicts.

Munzinger (1869) was the first explorer who described the area. The expeditions of 1875, 1881 and 1884 led by Munzinger, Giulietti and Bianchi were unsuccessful as all the explorers were killed (Nesbitt, 1935). The next generation of explorers was more successful and Vinassa De Regny in 1919, Nesbitt in 1928 and Franchetti in 1929 managed to cross the Danakil desert and collect geographical and geological data (Lupi, 2009). Detailed and extensive geological field explorations were conducted by French, Italian and German teams in the late 60s' (Bannert *et al.*, 1970; C.N.R. - C.N.R.S. Afar Team, 1973). Since the 70s', however, safety concerns in the area interrupted scientific expeditions. Since the 2010s', the safety situation improved again and scientific expeditions in the Danakil Depression could resume.

This study frames into the SNF funded project SERENA (SEdimentary REcord of the Northern Afar, 200021_163114) which aims at better understanding the flooding history and the sedimentological processes in the Danakil Depression. The SERENA project initiated when researchers from Addis Ababa University contacted geologists from the University of Fribourg (Switzerland) and the University of Ghent (Belgium) to analyse the origin and significance of fossils they discovered in the field. This first contact led to an official collaboration between the

universities of Fribourg and Addis Ababa. The first joint field expedition took place in 2013 (Atnafu *et al.*, 2015). It was followed by expeditions in 2015, 2016, 2017, 2019 and 2020, and complemented by travels to Addis Ababa in 2018 and 2021.

To answer some of the major questions highlighted above related to global, regional or local rift evolution, paleoenvironment, paleoclimate and resources, the major aim of this study is to better understand the basin dynamics of the Danakil Depression. In particular, it is aiming a better understanding of its structure, its evolution, the rates of geological processes and the factors controlling them using a multi-scale and interdisciplinary approach.

The aim of **Chapter II: Tectonostratigraphic evolution and significance of the Afar Depression**, as well as **Appendix A: Geological map of the Afar Depression** (Rime *et al.*, 2022a) and **Appendix B: Geological map of the southern Red Sea and western Gulf of Aden region** (Rime *et al.*, 2022b) is to analyse and understand the complex rifting history of the whole Afro-Arabian Rift System in order to better understand rifting mechanisms and the interactions between tectonics, sedimentation and magmatism, but also to understand the framework of the more local and detailed studies focussing on specific regions in the Afar Rift and the Danakil Depression. This study also indirectly results into **Appendix C: Oceanic isostasy as a trigger for the rift-to-drift transition: COMMENT** (Rime & Foubert, 2022).

Chapter III: The flooding of a young rift basin: past and future of the Danakil Depression, Afar is aiming

at the better understanding of mechanisms of opening and closure of the Danakil Depression to grasp the factors controlling the sedimentary architecture of rifts. It builds on the observation that coral reef terraces are not systematically situated near present-day sea level and that tectonic movements probably played a role in the opening and closure of the Danakil basin (Atnafu *et al.*, 2015; Perrochet, 2017; Jaramillo-Vogel *et al.*, 2019). We analyse the elevation of coralgall reef terraces and marine deposits to quantify vertical tectonic movement that controlled the connexion between the Red Sea and the Danakil Depression. We furthermore extrapolate our results to predict the future floodings of the Depression.

Chapter IV: Origin of Late Pleistocene to Holocene evaporite in the Danakil Depression: a modern analogue for understanding salt giants in Earth history aims at the better characterization of the sedimentary architecture of evaporite basins, as well as the paleoenvironmental and paleoclimatic history of the Danakil Depression. Sedimentological studies conducted on outcrops of the margins of the Depression by the SERENA team (Jaramillo-Vogel *et al.*, 2019; Endeshaw *et al.*, in prep.; Foubert *et al.*, in prep.; Negga *et al.*, in prep.) shed new light on the sedimentary architecture of the basin margins during alternating flooding and desiccation cycles, and allowed unravelling their paleoenvironmental and palaeoclimatological record. To have a basin-wide overview of the sedimentary architecture, studying the sediments in the subsurface of the central part of the basin is necessary. Moreover, basinal sediments provide a more continuous paleoenvironmental and

palaeoclimatological record compared to the punctuated nature of marginal sediments. A 625 m long core drilled in the central part of the basin is therefore studied. Alongside paleoenvironmental and paleoclimatic data, this study brings clues on the mechanisms of formation of large evaporite deposits.

Finally, the goal of **Chapter V: Novel Bathymetry of Lake Afdera Reveals Fault Structures and Volcano-Tectonic Features of an Incipient Transform Zone (Afar, Ethiopia)** (Schaegis *et al.*, 2021) is to understand the structure, origin and development of a Recent hypersaline lake in the Danakil Depression: Lake Afdera. This lake is a crucial resource for the local communities but is also probably a key element for the understanding of water circulation in the Depression.

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II TECTONOSTRATIGRAPHIC EVOLUTION AND SIGNIFICANCE OF THE AFAR DEPRESSION

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ABSTRACT

The Afar Depression is a unique place on Earth where active rift processes can be directly observed. It is believed to be close to continental breakup. The Afar hotspot has a strong influence on the geology of the Depression. Despite the strong geological interest in the region, difficult field access slowed scientific discoveries. During the last two decades, new projects and studies resulted in a better characterization of the region. New field data and global advances in understanding rift processes call for an integrative and holistic review of the tectonostratigraphic evolution of the Afar Depression. This study compiles new geological maps and reviews the stratigraphy and the geological history of the Afar Depression and the Afro-Arabian Rift System. A new kinematic evolution model and integrative paleogeological maps are proposed. Results show that geological events are diachronous throughout the region. We consider the Afar Rift to be distinct from the Red Sea Rift, both being separated by the Arrata Microplate. The Afar Rift is propagating northwards and forms a relay structure with the Red Sea Rift, linked to the counter-clockwise rotation of the Danakil Block since the Mid- to Late Miocene. The Afar Depression can be segmented into two distinct domains, Central Afar and the Danakil Depression. Central Afar experienced significant extension, protracted and extensive magmatism and magma-compensated thinning. It is believed to be strongly influenced by the Afar hotspot and bears similarities with oceanic plateaus. As such, Central Afar cannot be considered as a conventional rift system. In comparison, the Danakil Depression is younger and went through less extension and less magmatic activity until Recent (~0.6 Ma) times. The Danakil Depression probably represents the equivalent of a magma-rich margin. The absence of magma-compensated thinning allowed the development of an evolved stage of continental breakup. The tectonostratigraphic evolution of the Afar Depression with distinct rifting styles shows the complexity of continental break-up and forms an excellent analogue for understanding the development of passive margins and oceanic plateaus world-wide.

II.1 INTRODUCTION

The Afar Depression is the only emerged region on Earth that is close to continental breakup. It furthermore experiences the unique influence of a hotspot. As such, the Afar Depression is a unique place to study ongoing rift mechanisms that shaped planet Earth.

The Afar Depression has triggered geological interest since the 19th century (Blanford, 1869; Munzinger, 1869). The geological significance of the region was heavily debated during the rise of the continental drift theory and the understanding of plate tectonics (e.g. Wegener, 1929; Teilhard de Chardin, 1930; Mohr, 1967; McKenzie et al., 1970; Tazieff et al., 1972). The kinematic reconstruction of the Afar region has been at the core of this discussion for nearly a century. Wegener (1929) argued for continental drift based on the fitting between the African and Arabian continents when removing Afar. Conversely, Mohr (1967) denied this theory based on the presence of pre-rift blocks in the Afar Depression. It remains a spectacular natural laboratory for the study of tectonic processes, such as the transition from rifting to drifting, crustal structure, strain accommodation, the evolution of tectonic styles, mechanical extension vs magma assisted rifting, seismic activity, fault nucleation, propagation and interactions, and rift linkage, amongst others (e.g. Rowland et al., 2007; Bastow & Keir, 2011; Keir et al., 2013; Pagli et al., 2014, 2019; Stab et al., 2016; Doubre et al., 2017; Bastow et al., 2018; Illsley-Kemp et al., 2018; Zwaan et al., 2020b; Moore et al., 2021; Ruch et al., 2021; Sangha, 2021; Ahmed et al., 2022; La Rosa et al., 2022). It is also an essential site for studies related to the

understanding of mantle dynamics (e.g. Chang & Van der Lee, 2011; Hansen et al., 2012; Faccenna et al., 2013; Hammond et al., 2013b; Armitage et al., 2015; Civiero et al., 2015; Chambers et al., 2019). The region constitutes the second most recent Large Igneous Province (after Yellowstone) with ~30 Ma old Ethiopian flood basalts (Courtillet *et al.*, 2003), but is also characterized by a Recent volcanic cover, leading to numerous studies on petrology, volcanology and volcano-tectonics (e.g. Barberi et al., 1972b; Barberi & Varet, 1975; Mohr & Zanettin, 1988; Kieffer et al., 2004; Wright et al., 2006; Ayele et al., 2007; Acocella, 2010; Ebinger et al., 2010; Pagli et al., 2012; Ferguson et al., 2013b; Rooney et al., 2013; Barnie et al., 2016; Hagos et al., 2016a; Medynski et al., 2016; Rooney, 2017, 2020a; Xu et al., 2017; Tortelli et al., 2022). While dominated by magmatic products, the sedimentary architecture and infill of the Afar Depression are particular. The Danakil Depression witnessed episodic marine flooding that led to the formation of marine sediments as well as thick evaporite deposits, possibly forming a modern analogue for salt giants (Holwerda & Hutchinson, 1968; Atnafu et al., 2015; Jaramillo-Vogel et al., 2019, 2023). The interaction of sediments and magmatism formed the unique Dallol salt volcano that is believed to be a Mars-analogue for the study of extremophiles (Cavalazzi et al., 2019; López-García et al., 2020; Belilla et al., 2021; Sanz et al., 2021; Otálora et al., 2022). The Pliocene to Pleistocene sediments in the Afar Depression yielded crucial Hominine discoveries (Johanson et al., 1982; Abbate et al., 1998; Pan et al., 2002; Alemseged et al., 2006; Villmoare et al., 2015). During Late Pleistocene times, the region provided a critical corridor for the early

waves of *Homo sapiens* out-of-Africa migration (Beyin, 2006, 2021; Abbate & Sagri, 2012; López et al., 2015; Tierney et al., 2017).

Due to challenging environmental conditions, complex logistics and geopolitical instability, few field studies have been performed considering the significant scientific interest. The emergence of new projects and field data and advances in understanding rifts and passive margin development during the last two decades call for an integrative and holistic review and reassessment of the current knowledge on the tectonostratigraphic evolution of the Afar Depression. This holistic approach implies the consideration of the entire Afar Depression and its margins, as well as the Afro-Arabian Rift System. It requires not only the study of the dominant volcanic rocks, but also sediments (including pre-rift and syn-rift sediments) and the basement. Likewise, the structural analysis includes pre-rift and Precambrian structures that might have influenced the rifting processes.

The core of this work is the compilation of more than 750 datings and more than 120 geological maps, along with other published geological data, new field mapping and remote sensing analysis. All data are integrated, resulting in a new geological map of the Afar Depression (Rime *et al.*, 2022a). It is the first detailed (1:1'000'000) map comprising the complete Afar Depression and its margins. The map combines basement, sedimentary, intrusive and volcanic units, including pre-rift units. Lithological formations are homogenized and age constraints are improved. This map, combined with revised maps of the surrounding regions (section 3) and a review of the geological history of the

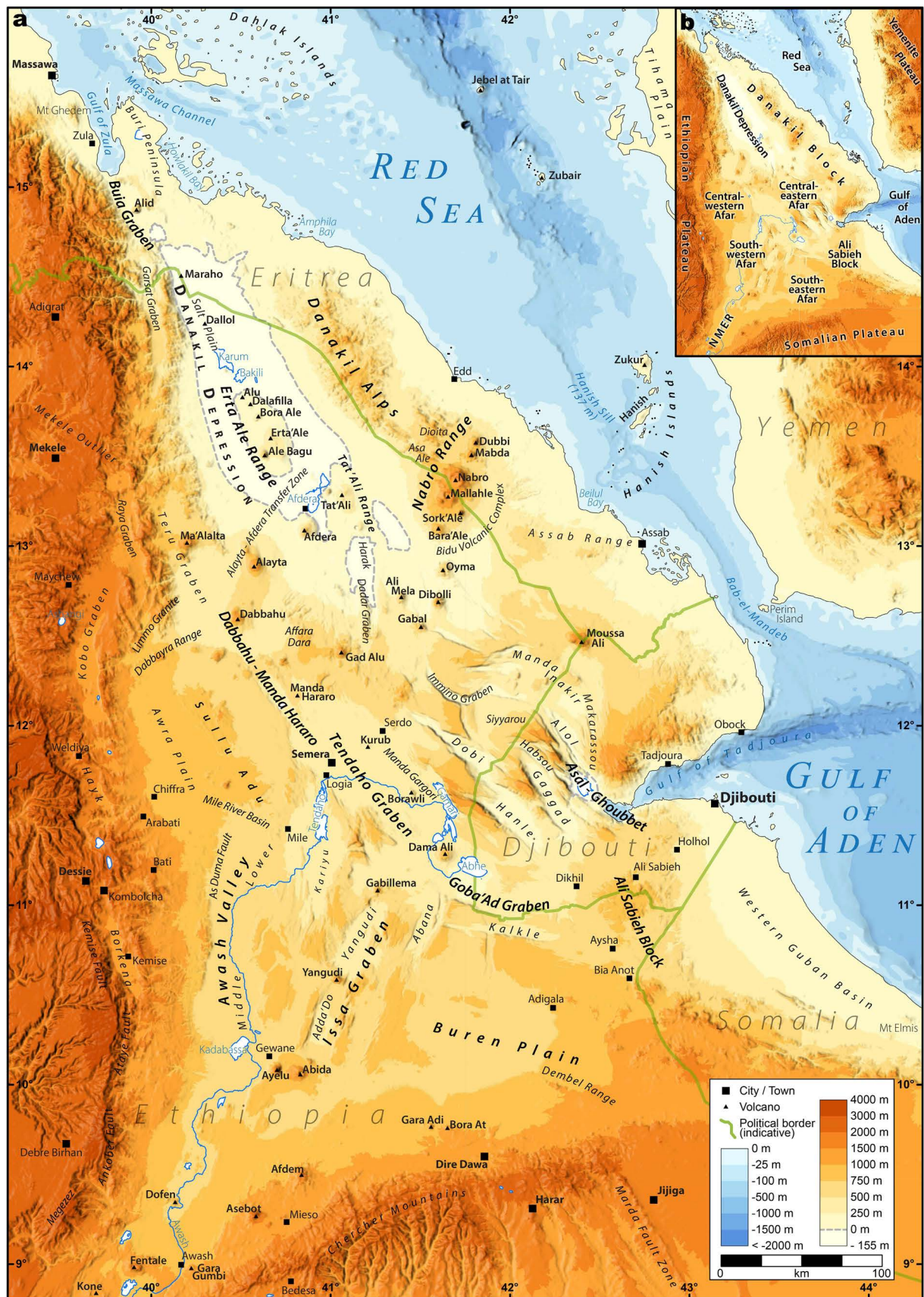
Afar region (section 4), allows for (1) reevaluation of the tectonic structure of the Depression (section 5), (2) reassessment of the tectonostratigraphic evolution of area with a new kinematic model (section 6), and (3) discussion of concepts and mechanisms of rift evolution in the Afar Depression (section 7).

This revised tectonostratigraphic review will form the base for further detailed studies in the fields of volcanology, structural geology, tectonics, petrology, sedimentology, geo(micro)biology, hominid evolution and human dispersal, and the exploration of the limits of life in extreme settings. The map of the Afar Depression will help local communities in risk assessment of natural hazards and the exploitation of natural resources such as water, salt, and geothermal energy, amongst others (Battistelli et al., 2002; Gardo & Varet, 2018; Varet, 2018; Varet & Gardo, 2020; Rizzello et al., 2021; Schaegis et al., 2021; Hearn, 2022; Pasquet et al., 2022).

The full review of the magmatic petrology and geochemistry of the Afar Depression is out of the scope of this study. For more information on this topic, the reader is referred to the book of (Varet, 2018), the reviews of (Rooney, 2017, 2020b; c; a) and references therein.

II.2 GEOLOGICAL SETTING

The *Afar Depression* (also called *Afar Triangle*) is a low-lying triangular region of approximately 600 km by 300 km covering about 105'000 km² and extending in Ethiopia, Djibouti and Eritrea (fig. 1). It is bordered in the west by the Ethiopian plateau, in the south by the Somalian plateau and in the east by the Danakil Block. In the south-west, it



◀ **Figure 1: Topographic map of the Afar Depression.** Topography data are from SRTMGL1 and GEBCO. **a)** The different regions, basins, structures, volcanoes (block triangles), cities (black squares), lakes and river discussed in the text are shown. Green lines represent political boundaries (Openstreetmap, retrieved on 06.01.2020). Multiple names and spelling are common in the region. The most commonly used names in the scientific literature were favoured. The list of sources for the nomenclature is presented in section Table 1. More local names are presented in Rime *et al.* (2022a). **b)** Names of the main regions used in the text and in fig. 7.

opens up towards the Northern Main Ethiopian Rift (NMER). While the elevation of the Depression is situated between – 154 m (lake Asal) and approximately 1000 m, the plateaus reach more than 3000 m. The broader *Afar region*¹ term used in this paper encompasses the Afar Depression, the Danakil Block, the southern Red Sea, the western Gulf of Aden and the surrounding margins of the Ethiopian, Somalian and Yemenite plateaus.

The Afar Depression is situated at the border between the Nubian, Arabian and Somalian plates, forming a triple junction (McKenzie *et al.*, 1970; McClusky *et al.*, 2003; Reilinger *et al.*, 2006; Reilinger & McClusky, 2011; Viltres *et al.*, 2022, fig. 2). The present-day plate velocities show that the movement of the Arabian plate relative to the Nubian and Somalian plates is significantly larger (~20 mm/yr in the southern Red Sea) than the movement between the Nubian and the Somalian plates (~5 mm/yr in northern MER) (e.g. McClusky *et al.*, 2010; Stamps *et al.*, 2021). The Arabian plate has constantly been moving northward relative to the Eurasian plate since 56 Ma due to the closure of the Neotethys (McQuarrie *et al.*, 2003). The Bitlis and Zagros sutures (fig. 2) mark the closure of this former ocean at least 10 Ma ago (McQuarrie *et al.*, 2003; Darin & Umhoefer, 2022). At ~25 Ma, the collision

in the Mediterranean was probably locked while the Makran was still actively subducting. This caused a significant slowing of the northward relative movement of the Nubian plate and led to the opening of the Red Sea and Gulf of Aden (Jolivet & Faccenna, 2000; Bellahsen *et al.*, 2003; McQuarrie *et al.*, 2003; ArRajehi *et al.*, 2010; Reilinger & McClusky, 2011; Reilinger *et al.*, 2015). This collision also caused the westward escape of the Anatolian plate (e.g. Reilinger *et al.*, 2006; fig. 2). The East African Rift System (EARS) experienced localized extension since late Eocene, but widespread rifting started to propagate since the Mid to Late Miocene and separated the Somalian plate from the Nubian plate (Michon *et al.*, 2022).

The opening of the Afro-Arabian rift system was preceded by the emplacement of the Ethiopian flood basalts. The first basalts emplacement took place in the Eocene (since 45 Ma) in Southern Ethiopia and Northern Kenya (fig. 3, see also Rooney, 2017). The main eruptive phase occurred during a short time period around 30 Ma in Ethiopia and Yemen, forming the classical Traps or flood basalts (Hofmann *et al.*, 1997; Ukstins *et al.*, 2002; Coulié *et al.*, 2003; Eid *et al.*, 2021).

This Large Igneous Province (LIP) is related to the African Superplume or

¹ Which should not be confused with the Afar Region (with capital R) representing a political division of Ethiopia.

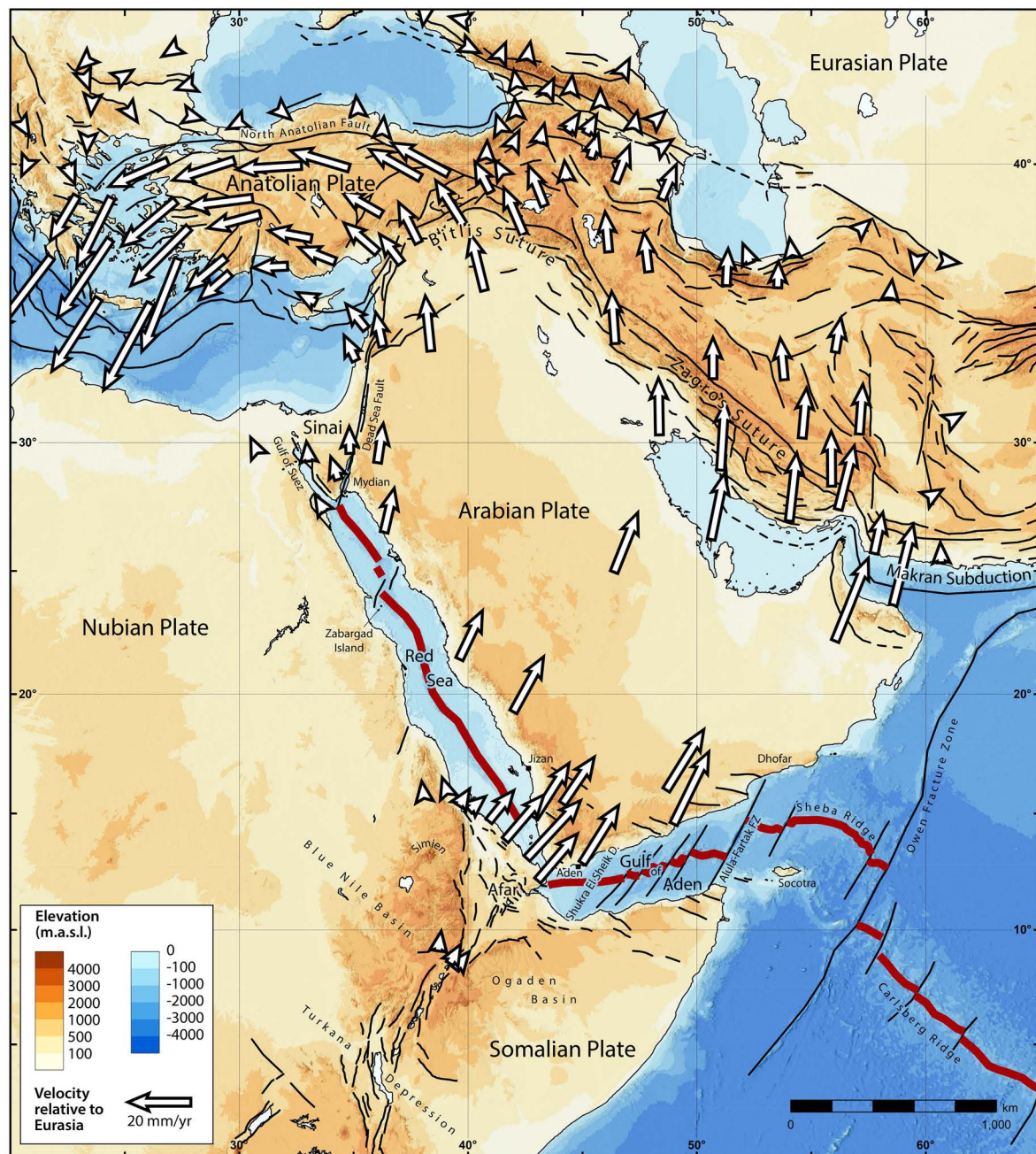


Figure 2: Geodynamic and topographic map of the Middle East. The white arrows represent GPS velocities relative to stable Eurasia. Data from Reilinger et al. (2006) and Reilinger & McClusky (2011). Black lines indicate major faults and thrust, thick red lines indicate oceanic ridge (after Aghanabati, 1993; Barrier et al., 2004; Pubelier, 2008; Fournier et al., 2010; Ali & Watts, 2016; Thiélemont, 2016; Aksu et al., 2021; Augustin et al., 2021; Rime et al., 2022a; b)

African Large Low Shear Velocity Provinces (LLSVP) rooted at the core-mantle boundary (Bastow et al., 2005; Ritsema et al., 2011; Hansen et al., 2012; Cottaar & Lekic, 2016; Rooney, 2017). While some authors argued for a minimal plume influence in Recent times in the

Afar Depression (Rychert et al., 2012; Hammond et al., 2013b), others associated it with an elevated mantle temperature (Bastow & Keir, 2011; Rooney et al., 2012; Ferguson et al., 2013b; Armitage et al., 2015) and resulting in dynamic topography,

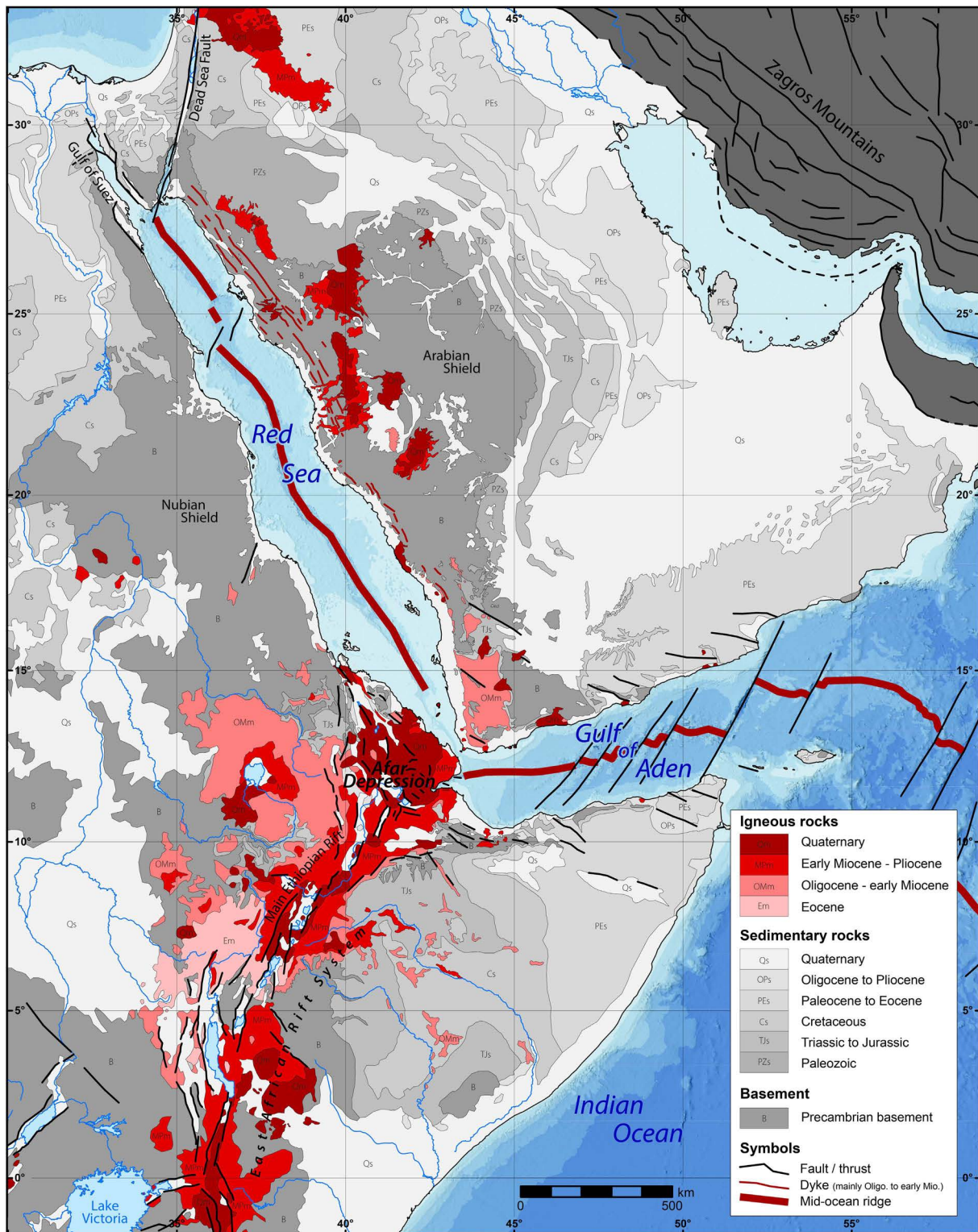
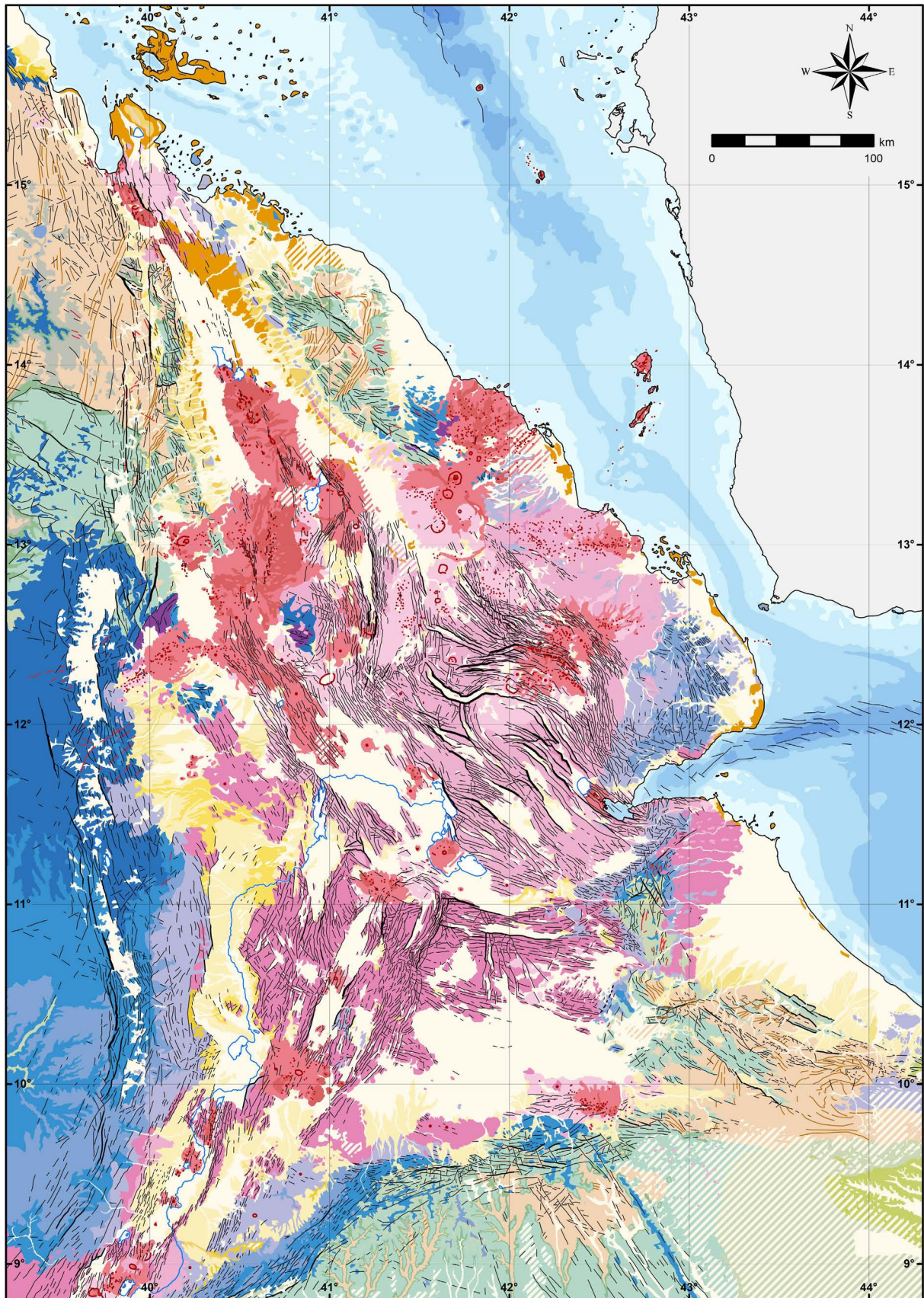


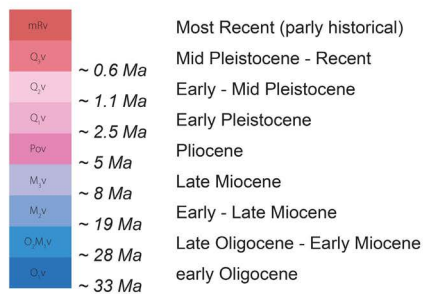
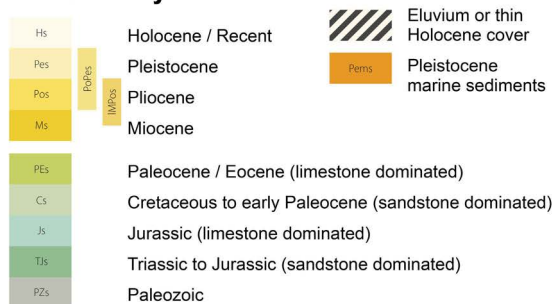
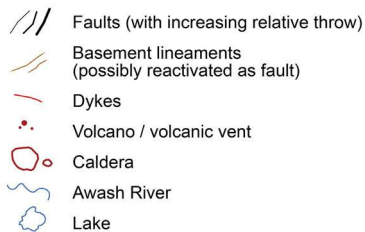
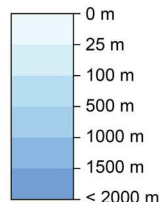
Figure 3: Geological map of the Afro-Arabian Rift System. This map particularly shows the extent and chronology of Cenozoic volcanic rocks. References listed in table 3.

explaining the high elevations around the Afar region (Lithgow-Bertelloni & Silver, 1998; Moucha & Forte, 2011).

There is a long-standing and unsolved controversy about the number and

configuration of plumes in the upper mantle. While some authors argue that the Ethiopian flood basalts were formed by one single plume/superplume



Volcanic Rocks**Intrusive Rocks****Sedimentary Rocks****Basement****Symbols****Bathymetry**

◀ **Figure 4: Geological map of the Afar Depression.** The full-scale version of this map is available in Rime *et al.* (2022a). References listed in table 1.

(Ebinger & Sleep, 1998; Hansen *et al.*, 2012; Hassan *et al.*, 2020), others argue for distinct Kenyan and Afar plumes or several different heads rooted in the African Superplume (George *et al.*, 1998; Rogers *et al.*, 2000; Lin *et al.*, 2005; Furman *et al.*, 2006; Montelli *et al.*, 2006; Chang & Van der Lee, 2011; Halldórsson *et al.*, 2014; Civiero *et al.*, 2015, 2022; Chang *et al.*, 2020).

The origin of the driving force causing the extension of the Afro-Arabian Rift System is often debated (Aldaajani *et al.*, 2021; Michon *et al.*, 2022). For the East African Rift System, plume and mantle effects seem to provide the most significant contribution, mainly as gravitational potential energy (Stamps *et al.*, 2014, 2015; Kendall & Lithgow-Bertelloni, 2016; Michon *et al.*, 2022). For the Red Sea and the Gulf of Aden, the individual contributions of mantle effects and slab-pull linked to the subduction of the Neotethys remain unclear (Bellahsen *et al.*, 2003; Bosworth *et al.*, 2005; Mouthereau *et al.*, 2012; Faccenna *et al.*, 2013; Petrunin *et al.*, 2020). While slab-pull forces were probably important during the first phase of rifting, they significantly reduced after the Arabian-Nubian lithosphere by plume activity played a significant role in the localization of the Afro-Arabian Rift System (Bellahsen *et al.*, 2003; Bosworth *et al.*, 2005; Reilinger *et al.*, 2015; Koptev *et al.*, 2018b; a; Khalil *et al.*, 2020; Guan *et al.*, 2021; Brune *et al.*, 2023).

II.3 GEOLOGICAL MAPPING

II.3.1 Geological map of the Afar Depression

One of the first geological maps of the Afar region was the *Sketch of the Geological Formations of Abyssinia* of Rüppel (1834). However, until the 70s', no detailed geological map existed. During the 70s', high-quality maps were published after field exploration and access to new satellite imagery. Brinckmann & Kürsten (1970) published detailed 1:250'000 maps of the Danakil

Depression. Barberi et al. (1971) and Varet et al. (1975) published 1:500'000 maps covering almost entirely the Afar Depression but excluding the margins and pre-rift rocks. The Geological Survey of Ethiopia (1973) published the first 1:2'000'000 geological map of Ethiopia (including present-day Eritrea). Until today, the only geological map covering the complete Afar Depression and its margins are the 1:2'000'000 maps of Kronberg et al. (1974a; b) (accompanying paper: Kronberg et al., 1975). Small-scale maps were since then published in single articles (e.g. Beyene & Abdelsalam, 2005; Stab et al., 2016; Ayalew et al., 2019) and as part of regional maps (e.g. Thiélemont, 2016).

The new geological map of the Afar Depression (Rime et al., 2022a, simplified version in fig. 4) covers the Danakil Depression and adjacent regions from 8.8°N to 16 °N and from 39.2°E to 44.3°E at a scale of 1:1'000'000. It includes the northernmost part of the Main Ethiopian Rift, the less-studied Southern Afar Depression and Somalian margins, as well as the rift margins. To keep the scale of 1:1'000'000, certain structures were simplified, and some rare outcrops of important geological significance were upscaled to remain visible.

The map was compiled in Esri's ArcGIS Pro. All GIS data and files, as well as the dating database are available through the Zenodo open-access database (Rime et al., 2022a). An uncertainty map and the descriptive uncertainty of each mapped region are also available. The attributes of the single mapped polygons further contain source information and comments on the reliability. Finally, on the full-sized geological map, unit codes followed by “?” indicate higher uncertainty levels (e.g. contradictory

interpretation in the literature, poor age constraint, extrapolation over long distances, unclear remote sensing data, missing coordinates or vague descriptions).

II.3.1.1 *Material*

More than 120 published geological maps of different sizes, scales, precision and reliability were considered for the mapping (see Table 1 and Rime et al., 2022a). The mapped units across the different maps are mostly lithostratigraphic. Some maps also follow a chronostratigraphic approach and/or provide information on geochronology.

We compiled an extensive database of published datings (Rime et al., 2022a, see also suppl. mat. 1). This database contains 773 datapoints extracted from 75 publications. Most datings have been performed on volcanic material. 63% of the samples were dated using the K/Ar technique, 23% using Ar/Ar, 8% by unknown dating techniques, and the last 6% include a variety of techniques like (U-Th-Sm)/He, ³⁶Cl, ¹⁴C, Fission track and ³He. Some techniques are known to be less reliable, particularly, K/Ar datings compared to Ar/Ar datings (Kidane et al., 2003; Lahitte et al., 2003b). The higher uncertainty related to the technique used and the published uncertainty range for each datapoint were weighted during the mapping process (see discussion on reliability in Rime et al., 2022a). The precision and reliability of the localisation were also evaluated and rated for each datapoint. When possible, the position of the datapoints was reviewed based on the sampling description or maps presented in the publication. Several (>40) errors in coordinates and/or inconsistencies were

identified, corrected and indicated in the database.

Outcrop descriptions, cross-sections and sedimentological descriptions were also used for the mapping (Table 1).

The map is complemented by lithological and structural field observations conducted during six field trips (2013, 2015, 2016, 2017, 2019, 2020) led by teams of the universities of Addis Ababa (Ethiopia) and Fribourg

(Switzerland) within the framework of the SNF-funded project SERENA ‘Sedimentary Record of the Northern Afar’ (200021_163114). Observations focussed on the Danakil Depression and its western margin, including western Harak graben, the basins N, NE and E of the Tat’Ali Range, the Erta Ale Range, the basin W of the Erta Ale Range, the Salt plain and its western margin, several transects between the salt plain and the Garsat Graben and the road

	Geological maps	Datings	Other
Gulf of Zula, Danakil Depression and Danakil Alps	Abbate et al. (2004), Antonielli et al. (2009), Barberi et al. (1971), Barberi and Varet (1972), Brinckmann and Kürsten (1970), Buffler et al. (2010), Civetta et al. (1975a), De Fino et al. (1978), Drury et al. (1994), Duffield et al. (1997), Frazier (1970), Gasse (1974), Geological Survey of Ethiopia (1978a, 1971), Ghebreab (1999, 1998), Ghebreab and Talbot (2000), Hagos et al. (2016b), Le Gall et al. (2018), López-García et al. (2020), Lowenstern et al. (2006), Mège et al. (2021), Sagri et al. (1998), Sani et al. (2017), Teklay et al. (2005), Tortelli et al. (2021), Watts et al. (2020)	Barberi et al. (1975, 1972a), Brinckmann and Kürsten (1971), Buffler et al. (2010), Civetta et al. (1975a), Drury et al. (1994), Duffield et al. (1997), Frazier (1970), Le Gall et al. (2018), Lowenstern et al. (2006), Shoshani et al. (2006)	Bannert et al. (1970), Barberi et al. (1970), Barberi and Varet (1970), Brinckmann and Kürsten (1971), Bruggemann et al. (2004), Buffler et al. (2010), Bunter et al. (1998), Duffield et al. (1997), Frazier (1970), Holwerda and Hutchinson (1968), Munzinger (1869), Park (2010), Varet (2018), Walter et al. (2000)
Ethiopian plateau	Ayalew et al. (2021), Geological Survey of Ethiopia (2011, 2010b, 2009a, 2009b, 1978a, 1971), Hagos et al. (2020), Sembroni et al. (2017), Tesfamichael et al. (2010)	Chernet et al. (1998), Coulié et al. (2003), Drury et al. (1994), Eid et al. (2021), Feyissa et al. (2017), Hofmann et al. (1997), Kieffer et al. (2004), Mohr and Zanettin (1988), Morton et al. (1979), Rooney et al. (2013), Stab et al. (2016), Ukstins et al. (2002), Wolfenden et al. (2005)	Mohr and Zanettin (1988)
Ethiopian central-eastern Afar (including here Dabbahu – Manda Hararo and Tendaho)	Ayalew et al. (2019), Field et al. (2013), Gasse et al. (1987, 1980, 1974), Geological Survey of Ethiopia (2014, 2013a), Kidane et al. (2003), Lahitte et al. (2003b, 2003a), Medynski et al. (2016, 2015), Stab et al. (2016), Varet et al. (1975)	Barberi et al. (1975, 1972a), Brinckmann and Kürsten (1971), Civetta et al. (1975a, 1975b), Ferguson et al. (2013a), Field et al. (2013), Kidane et al. (2003, 1999), Lahitte et al. (2003b), Medynski et al. (2015)	Global Volcanism Program (2013), Varet (2018)

Table 1: Reference for the geological map of the Afar Depression.

	Geological maps	Datings	Other
Western Afar	Adamson and Williams (1987), Alene et al. (2017), Asfaw et al. (2002), Black et al. (1972), Chorowicz et al. (1999), Christiansen et al. (1975), Clark et al. (2003), De Heinzelin et al. (1999), DiMaggio et al. (2015b, 2009, 2015a), Garello (2019), Geological Survey of Ethiopia (2014, 2011, 2010b, 2009a), Geraads et al. (2021), Kalb (1995), Kalb et al. (1982), Levin et al. (2004), Mège and Korme (2004), Mohr (1983), Mohr and Zanettin (1988), Niespolo et al. (2021), Quade et al. (2008, 2004), Semaw et al. (2005), Stab et al. (2016), Taieb et al. (1976), Tiercelin (1986), Varet et al. (1975), Wolfenden et al. (2005), Wynn et al. (2008), Zwaan et al. (2020a, 2020b)	Alemseged et al. (2020), Aronson et al. (1977), Ayalew et al. (2019), Barberi et al. (1975, 1972a), Black et al. (1972), Braun et al. (2019), Clark et al. (1994), Coulié (2001), Coulié et al. (2003), Deino et al. (2010), DiMaggio et al. (2015b, 2015a), Haile-Selassie et al. (2015), Kidane et al. (2003), Kleinsasser et al. (2008), Kunz et al. (1975), Lahitte et al. (2003b), Niespolo et al. (2021), Renne et al. (1999), Rooney et al. (2013), Saylor et al. (2019), Stab et al. (2016), Ukstins et al. (2002), White et al. (1993), WoldeGabriel et al. (2013, 2001), Wolfenden et al. (2005)	Black et al. (1972), Campisano (2012), Clark et al. (1984), Hall et al. (1984), Justin-Visentin and Zanettin (1974), Kleinsasser et al. (2008), Rowan et al. (2022), Varet (2018), WoldeGabriel et al. (2013)
NMER	Abebe et al. (2005), Benito-Calvo et al. (2014), Chernet et al. (1998), Feyissa et al. (2017), Geological Survey of Ethiopia (2009b, 2007, 1985, 1978b), Hujer et al. (2015), Juch (1975), Morbidelli et al. (1975), Popp and Scholger (2017), Sickenberg and Schönfeld (1975), Siegburg et al. (2018), Williams et al. (2004), WoldeGabriel et al. (1992b), Wolfenden et al. (2004)	Benito-Calvo et al. (2014), Chernet et al. (1998), Feyissa et al. (2017), Hujer et al. (2015), Kazmin et al. (1980), Kidane et al. (2010, 2006), Kunz et al. (1975), Lahitte et al. (2003b), Morbidelli et al. (1975), Morton et al. (1979), Siegburg et al. (2018), Tiercelin et al. (1979), Ukstins et al. (2002), Vogel et al. (2006), Williams et al. (2004), WoldeGabriel et al. (1992b, 1992a), Wolfenden et al. (2005, 2004)	Rooney (2020b), Tiercelin et al. (1979)
South-eastern Afar and Somali plateau	Black et al. (1975), Geological Survey of Ethiopia (2015, 2014, 2010a, 2007, 1985), Juch (1975), Mège et al. (2016), Varet et al. (1975)	Audin et al. (2004), Barberi et al. (1975), Berhe (1986), Black et al. (1975), Chernet et al. (1998), Kunz et al. (1975), Mège et al. (2016), WoldeGabriel et al. (1992a)	Varet (2018)
Somalia	Ali (2015, 2009), Ali et al. (2018), Ali and Watts (2013), Geological Survey Somali Democratic Republic (1971a, 1971b, 1970a, 1957)		
Djibouti	Arthaud et al. (1980), Audin et al. (2004), Black et al. (1975), Boucarut et al. (1985), Chessex et al. (1975), Daoud et al. (2011), Le Gall et al. (2015, 2010), Manighetti et al. (1998, 1997), Muller and Boucarut (1975)	Audin et al. (2004), Barberi et al. (1975, 1972a), Chessex et al. (1975), Civetta et al. (1975a, 1975b), Courtillot et al. (1984), Daoud et al. (2011), Lahitte et al. (2003b), Le Gall et al. (2010), Manighetti et al. (1998), Piguet and Vellutini (1991), Zumbo et al. (1995b)	Gasse et al. (1986, 1980), Varet (2018)

	Geological maps	Datings	Other
Southern Danakil Block (excl. Djibouti)	Barberi et al. (1971), Civetta et al. (1975a, 1975b), De Fino et al. (1973), Ganci et al. (2020), Wiart and Oppenheimer (2005)	Barberi et al. (1975), Brinckmann and Kürsten (1971), Civetta et al. (1975a, 1975b, 1974), Piguët and Vellutini (1991)	Civetta et al. (1974), De Fino et al. (1973), Varet (2018), Wiart and Oppenheimer (2000), Wiart et al. (2000)
Red Sea and Gulf of Aden (incl. islands)	Carbone et al. (1998), Dauteuil et al. (2001), Frazier (1970), Gass et al. (1973), GEBCO Compilation Group (2020), Lambeck et al. (2011), Le Gall et al. (2015), Mallick et al. (1990), Purkis et al. (2012), Xu et al. (2015)	Mallick et al. (1990)	Al-Mikhlaifi et al. (2018), Angelucci et al. (1975), Conforto et al. (1976), Frazier (1970)
Seasonally flooded areas	2019/2020/2021 Landsat 8 images (Earth Resources Observation and Science (EROS) Center, 2018a)		
Geographic features	Rivers, lakes, roads, coastlines and political boundaries from Openstreetmap (https://www.openstreetmap.org , 06.01.2020). River, lakes, roads and coastlines are manually modified. Data for insert globe from Natural Earth (naturalearthdata.com).		
Names	Ali (2015), Angelucci et al. (1975), Ayalew et al. (2019), Ayele et al. (2009), Black et al. (1972), Campisano (2012), Chernet et al. (1998), Christiansen et al. (1975), Clark et al. (2003), Frazier (1970), Geological Survey of Ethiopia (2015, 2014, 2013b, 2011, 2010a, 2007, 1985, 1978b), Global Volcanism Program (2013), Kidane et al. (2003), Lahitte et al. (2003b, 2003a), Le Gall et al. (2015), Manighetti et al. (2001), Munzinger (1869), Nanis and Aly (2020), Stab et al. (2016), Varet (2018), Varet et al. (1975), Wiart and Oppenheimer (2005), Williams (2016), Wolfenden et al. (2005), Wynn et al. (2008), Zwaan et al. (2020b) and the EthioGIS MapServer Ethiopia (www.ethiogis-mapserver.org). Multiple names and spelling are common in the region. We favoured the most commonly used names in the scientific literature. Names with comments are also featured as a shapefile of Rime et al. (2022a).		

Table 1 (continued)

outcrops between Afdera-Mekele and Dallol-Mekele.

Remote sensing data were used to complement, improve and interpolate published data and field observations. Multispectral satellite images from Landsat 8 OLI and TIRS Level-1 data (Earth Resources Observation and Science (EROS) Center, 2018a) were used for the complete study area. The bands 7-6-5 were used to enhance geological contrasts. Other commercial orthophoto providers (Google Earth, Bing Maps, Esri) were used for higher-resolution images. The digital elevation models

(DEM) used were the SRTMGL1 (Earth Resources Observation and Science (EROS) Center, 2018b) for the complete study area and the ALOS PALSAR RTC (<https://doi.org/10.5067/Z97HFCNKR6V>) for the Danakil Depression.

II.3.1.2 Lithological units

Different stratigraphic classifications have been proposed in the literature for the Afar Depression and its margins since the first lithostratigraphy was presented by Blanford (1869) for the Ethiopian Plateau. suppl. mat. 2 compiles the most commonly used formal and informal

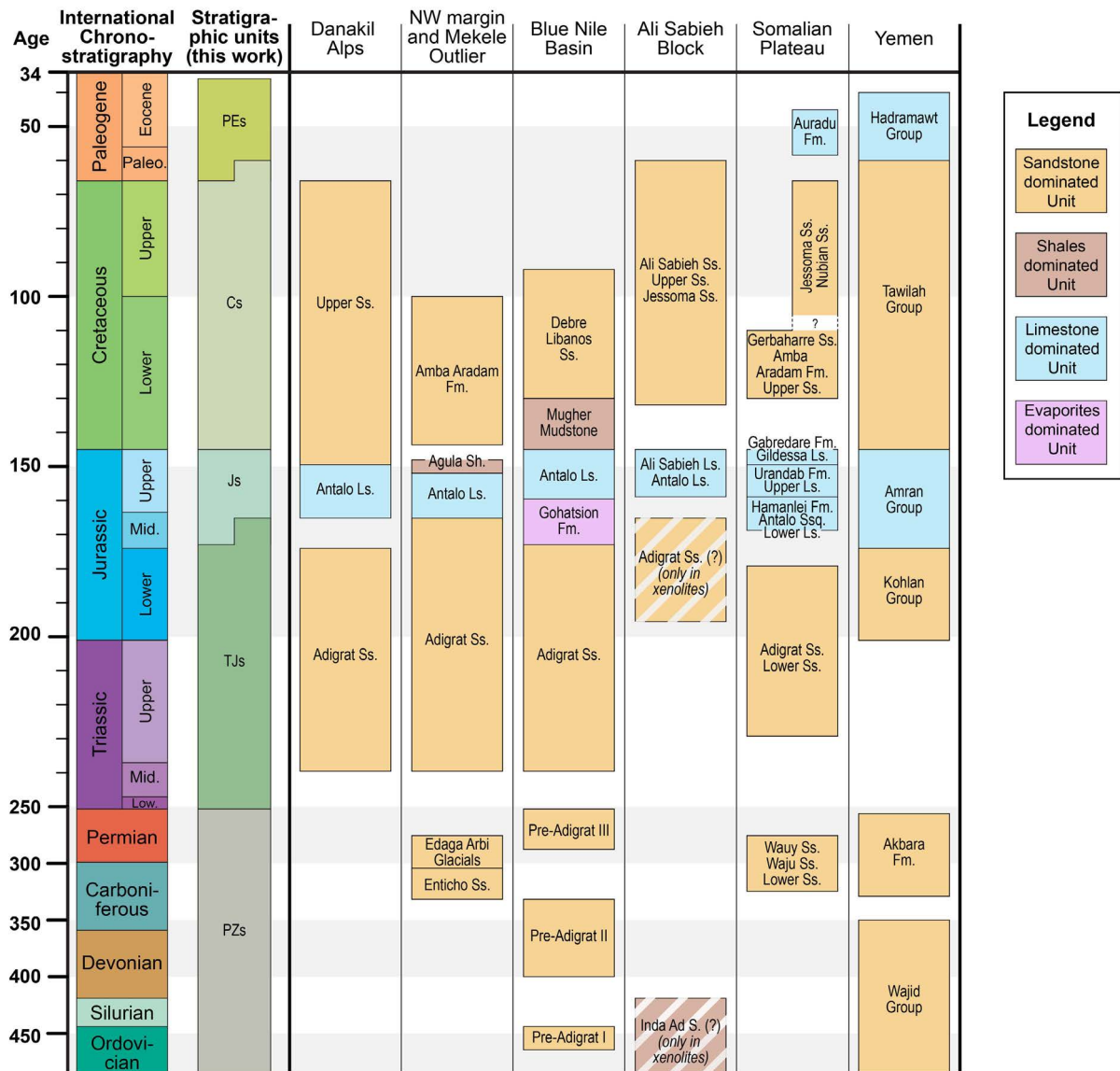


Figure 5: Pre-rift lithological units of the margins of the Afar Depression. Compilation of the most common lithological units. Note that the chronostratigraphy is often poorly constrained. The reader is referred to the original publications for detailed information. Abbreviations: Fm.= Formation; Gr.=Group; Ls. = Limestone; S.= Series; Sh.= Shales; Ss.= Sandstone; Ssq.=Supersequence. Sources: Danakil Alps: Brinckmann & Kürsten (1971) and Sagri et al. (1998); NW margin and Mekele Outlier: Beyth (1972), Kidane et al. (2013), Sembroni et al. (2017) and Hagos et al. (2020); Blue Nile basin : Chernet *et al.* (2020) ; Ali Sabieh Block : Chessex et al. (1975), Gasse et al. (1986), Le Gall et al. (2015) and Kidane (2016) ; Somalian plateau : Geological Survey Somaliland Protectorate (1956), Geological Survey of Ethiopia (1985, 1996, 2007, 2010a), Bosellini et al. (2001) and Ali & Watts (2016) ; Yemen : Grolier & Overstreet (1978), van der Gun & Ahmed (1995) and As-Saruri et al. (2010).

lithostratigraphic classifications in the area for the Paleogene, Neogene and Quaternary based on magmatic and sedimentary units. Fig. 5 compiles the lithostratigraphy from the Ordovician to the Paleogene. As noted by Mohr & Zanettin (1988) and Rooney (2017, 2020a), lithological units and Formation

names used in Afar are based on petrological, morphological or genetic relationships, and distinct Formations sometimes have overlapping ages. Conversely, identical terms are used for rocks of significantly different ages (e.g. Stratoid, Termaber Fm. on suppl. mat. 2). Also, most studies focus either on

magmatic stratigraphy or sedimentary stratigraphy but an integrated stratigraphic approach is rarely proposed.

Considering the principal aim of this study is the overall understanding of the tectonostratigraphic evolution of the study area, the authors took an unconventional chronostratigraphic approach while – where possible – respecting the historically defined lithostratigraphic units. As such, the map is the first of its kind for the Afar Depression and integrates volcanic, magmatic and sedimentary rock units. The authors are aware of the limits of this combined chrono-lithostratigraphic approach. Still, they are convinced that this is the only way to understand better rift evolution at a regional scale.

II.3.1.2.1 Volcanic stratigraphy

The volcanic units are defined mainly based on their age and not necessarily on historical defined Formation names. The units represent lavas erupted synchronously but often have very different petrology, chemistry or origin. Conversely, one former defined single, homogeneous lithostratigraphic unit might exceptionally be mapped in this study by two separate units due to significant age differences at the bottom and top of the succession. Nevertheless, most of the volcanic stratigraphic units defined in this study follow the lithostratigraphic consensus presented in recent literature. The encountered lithologies of volcanic rocks cover various compositions from basalts to rhyolites (e.g. Barberi et al., 1970; Varet et al., 1975; Rooney, 2020a). On the full-scale map (Rime *et al.*, 2022a), rhyolite outcrops are indicated with a pattern. Because of significant contradictions

among different sources, ignimbrites were not included in this category.

The following volcanic stratigraphic units are defined:

- *mRv*: most Recent volcanics. This unit is not defined solely by age but corresponds to the last eruptions identified and mapped easily due to their very dark color. It considers historical eruptions, including the very Recent Erta Ale or Nabro eruptions (Goitom et al., 2015; Global Volcanism Program, 2018, 2019).
- *Q_{3v}*: Mid-Pleistocene to Recent volcanics younger than ~ 0.6 Ma. This unit is comparable to the Recent formations of Kidane et al. (2003), Stab et al. (2016) and Varet (2018), or the *Axial Series* of Zumbo et al., (1995b) and Rooney (2020a).
- *Q_{2v}*: Early – Mid-Pleistocene volcanics, with ages between ~ 0.6 and ~ 1.1 Ma. This unit is comparable to the *Gulf Basalts* of Kidane et al., (2003; Stab et al. (2016) or the upper part of the *Stratoid series* of Varet et al. (1975).
- *Q_{1v}*: Early Pleistocene volcanics between ~ 1.1 and ~ 2.5 Ma old. This unit is comparable to the upper *Stratoid Series* of Kidane et al. (2003).
- *Pov*: Pliocene volcanics between ~ 2.5 and ~ 5 Ma old. This unit includes the lower *Stratoid Series* of Kidane et al. (2003) and Rooney, (2020a), most of the *Sullu Adu Series* of Stab et al. (2016), the *Bachi Fm.* of Wolfenden et al. (2004), and the upper part of the *Somali* and *Dalha Basalts* of Le Gall et al. (2015).
- *M_{3v}*: Late Miocene volcanics between ~ 5 and ~ 8 Ma old. This

unit is shorter but comparable to most of the *Dahla Series* defined in the literature (Varet et al., 1975; Kidane et al., 2003; Wolfenden et al., 2005; Le Gall et al., 2015).

- *M_{2v}*: Early to Late Miocene volcanics between ~ 19 and ~ 8 Ma old. This unit covers an extended period but a relatively small extent (suppl. mat. 1c) and includes the *Mabla Series*.
- *O₂M_{1v}*: late Oligocene to Early Miocene volcanics between ~ 28 and ~ 19 Ma old. It corresponds to the upper part of the *Traps series* (holding a wide variety of names across the region) and includes the *Ali Sabieh basalts* (Le Gall et al., 2015) and the Ogaden dyke swarm (Mège et al., 2016).
- *O_{1v}*: early Oligocene volcanics between ~ 28 and ~ 34 Ma old corresponding to the lower part of the *Traps series*.

II.3.1.2.2 Intrusive rocks

Mapped Cenozoic intrusive rocks in the study area are relatively rare. They are mainly formed by granites and some dolerite (Bannert et al., 1970; Black et al., 1972; Barberi et al., 1975; Civetta et al., 1975a; De Fino et al., 1978; Geological Survey of Ethiopia, 1978a). The major intrusions are the Limmo, Affara Dara, Dioita and Asa Ale intrusives (Black et al., 1972; Barberi et al., 1975; Civetta et al., 1975a; De Fino et al., 1978). Because they all have similar age, a single unit was defined:

- *O₂M_{1i}*: late Oligocene to Early Miocene intrusive magmatic bodies between ~ 28 and ~ 19 Ma old. They include the Limmo, Affara Dara, Dioita and Asa Ale intrusives (Black et al., 1972;

Barberi et al., 1975; Civetta et al., 1975a; De Fino et al., 1978).

II.3.1.2.3 Sedimentary stratigraphy

Stratigraphic units for the syn-rift sedimentary rock are defined following the International Chronostratigraphic Chart. Except for the *Pems* unit, these units are chiefly composed of siliciclastic sediments ranging from very coarse conglomerates to mudstones, but lacustrine carbonates and evaporites are also present.

The following syn-rift stratigraphic units are defined:

- *Hs*: Holocene sediments covering mainly the basin floor and the Red Sea and Gulf of Aden coasts.
- *Pes*: Pleistocene sediments mainly present at the margin of the Danakil Depression, “Terraces gravel” of Brinckmann & Kürsten (1971), in the lower and middle Awash valley, forming the *Busudima Fm.* (Quade et al., 2008), in some graben structures of central-eastern Afar (Gasse & Street, 1978) and in the NMER (WoldeGabriel et al., 1992b).
- *Pems*: Pleistocene marine sediments. This unit is distinguished from the other Pleistocene deposits due to its marine nature and lithology being significant to understand the basin dynamics of the Danakil Depression.
- *Pos*: Pliocene sediments. They are mainly present in the Awash valley as clastic sediments forming the *Sagantole* and *Hadar Formations* (Renne et al., 1999; Quade et al., 2008) and in the eastern Goba’Ad basin (Gasse et al., 1987).
- *PoPes*: Pliocene to Pleistocene sediments. This unit is mainly

present in the Awra plain and Guban basin.

- *Ms*: Miocene sediments. They are present as the *Chorora* and *Adu-Asa Fm.* in the NMER (Sickenberg & Schönfeld, 1975; WoldeGabriel et al., 2001) and as the *Dogali Fm.* near Massawa (Drury et al., 1994; Sagri et al., 1998).
- *LMPos*: Late Miocene to Pliocene sediments. This unit is almost exclusively used for the Red Bed Series around and at the margins of the Danakil Depression.

Pre-rift lithostratigraphic units are well defined in former studies with a homogenized nomenclature (fig. 5). However, timing is poorly constrained. Based on the recurrent, almost synchronous pattern of facies associations, the stratigraphic units defined in this study mostly conform to the former lithostratigraphic units.

The following pre-rift stratigraphic units are defined (see legend of fig. 5 for sources) :

- *PEs*: Paleocene to Eocene sediments dominated by marine limestone. They are only present in the SE part of the map forming the *Auradu Fm.*
- *Cs*: Cretaceous sediments dominated by sandstones. The upper part of this unit (*Jessoma sandstones*) might, however, be early Paleocene in age (Grolier & Overstreet, 1978; van der Gun & Ahmed, 1995; Kidane, 2016). Note that the *Mugher Mudstones* of the Blue Nile basin are incorporated in this *Cs* unit.
- *Js*: Jurassic sediments dominated by marine limestones (often named *Antalo limestones*). This unit includes the *Agula Shales* of the Mekele outlier and the

evaporitic *Gohatsion Fm.* of the Blue Nile basin.

- *TJs*: Triassic to Jurassic sediments dominated by sandstones (often named *Adigrat sandstones*)
- *PZs*: Paleozoic sediments dominated by continental tillites, sandstones and shales.

Finally, the *B* unit represents the undifferentiated Precambrian basement, formed by various plutonic rocks, metasediments and metavolcanic rocks.

Often, thin (a few meters or less) Holocene sedimentary cover (aeolian or lacustrine) or, in the case of the Somalian plateau, eluvium cover is present on top of older units. As it only concerns a very thin layer, they are not mapped as distinct 'Holocene' units (*Hs*) but indicated with hatched symbols.

II.3.1.3 Structural mapping

In addition to the newly defined stratigraphic units, faults, basement lineaments, dykes, volcanoes/volcanic vents and calderas are mapped. Faults are classified into three categories reflecting the relative throw of the fault. The authors are aware that fault mapping mainly based on remote sensing has inherent limitations related to erosion, vegetation cover, lithology contrast, topography and fault orientation. The density of mapped faults is thus not necessarily reflecting the actual fault density, and the classification by throw is only relative at a local scale. Also, differentiation between faults and dykes based on DEM data is not always evident when vegetation covers the area. The basement lineaments represent any visual lineation identified in the basement that cannot be attributed to a fault. They were

identified based on colour/lithology contrasts and/or lineaments visible on the DEM.

II.3.2 Other maps

II.3.2.1 *Geological map of the Southern Red Sea & Western Gulf of Aden region*

The *Geological map of the Southern Red Sea & Western Gulf of Aden region* (fig. 6 and Rime *et al.*, 2022b) was compiled to serve as a base for the kinematic reconstruction. The mapping was performed at an approximate scale of 1:3'000'000 and using a similar chronostratigraphic-lithostratigraphic approach as described for the map of the Afar Depression. Due to the lower resolution of mapping, only the following stratigraphic units are mapped :

- *Qv*: Quaternary volcanic rocks. Includes *mRv*, *Q_{3v}*, *Q_{2v}* and *Q_{1v}* units of the geological map of the Afar Depression.
- *MPv*: Early Miocene to Pliocene volcanic rocks. Includes *M_{2v}* and *M_{3v}* units of the geological map of the Afar Depression.
- *OMv*: Oligocene to Early Miocene volcanic rocks. Includes *O_{1v}* and *O₂M_{1v}* units of the geological map of the Afar Depression.
- *OMi*: Oligocene to Early Miocene intrusive rocks. Similar to the *O₂M_{1i}* units of the geological map of the Afar Depression.
- *Qs*: Quaternary sediments. Includes *Hs* and *Pes* units of the geological map of the Afar Depression.
- *OPs*: Oligocene to Pliocene sediments. Includes *Pos*, *PoPes*, *lMPos* and *Ms* units of the geological map of the Afar

Depression and some older sediments (up to Oligocene) on the coasts of the Gulf of Aden (e.g. USGS & Arabian American Oil Company, 1963; Leroy *et al.*, 2012; Ali, 2015).

- *PEs*: Paleocene to Eocene sediments dominated by marine limestones and including the Auradu Fm. on the Somalian plateau and the Hadramawt Group on the Yemenite plateau (see fig. 5).
- *TJs*: Triassic to Jurassic sediments, including the sandstone-dominated *TJs* and the limestone-dominated *Js* units of the geological map of the Afar Depression.
- *PZs*: Paleozoic sediments similar to the geological map of the Afar Depression.
- *B*: Undifferentiated Precambrian basement.

Sources and reference maps used are summarized in Table 2. Remote sensing data (Landsat imaging and SRTMGL1 DEM) were used to discriminate between the interpretations of different authors, to compensate lack of data and to improve the mapping of faults, dykes and basement lineaments. All data are available in Rime *et al.* (2022b).

II.3.2.2 *Geological map of the Afro-Arabian rift system and basement map of the Red Sea*

The geological map of the Afro-Arabian Rift System (fig. 3) was mapped at an approximate scale of 1:17'000'000 and mainly summarizes the Cenozoic magmatic units. Sedimentary units are the same as the *Geological map of the Southern Red Sea & Western Gulf of Aden*

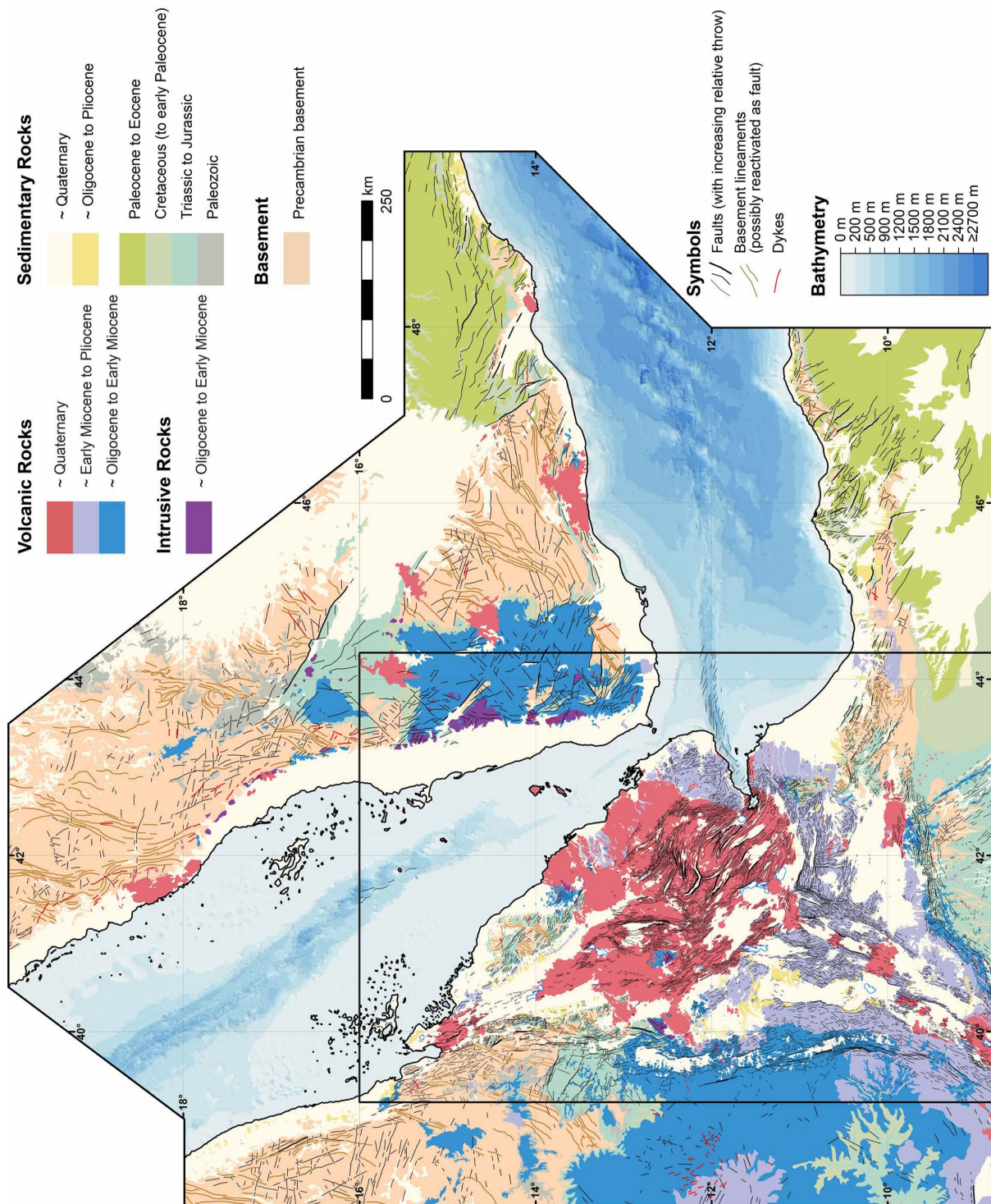


Figure 6: Geological map of the Southern Red Sea & Western Gulf of Aden Region. The full-scale version of this map is available in Rime *et al.* (2022b). References are listed in table 2.

Afar Depression	Geological map of the Afar Depression (Rime et al., 2022a, see Table II.1 for references)
Ethiopia	Abebe et al. (2005), Chernet et al. (2020, 1998), Chorowicz et al. (1998), Coulié et al. (2003), Drury et al. (1994), Geological Survey of Ethiopia (1996, 1978a, 1978b, 1971, 2011, 2010b, 2009b, 2009a), Hagos et al. (2017), Hofmann et al. (1997), Kieffer et al. (2004), Mège and Korme (2004), Mohr and Zanettin (1988), Morton et al. (1979), Rooney (2020b, 2017), Sembroni et al. (2017), Teklay et al. (2005)
Somalia	Ali (2009, 2015), Ali et al. (2018), Billi (2022), Geological Survey Somali Democratic Republic (1970b, 1970c, 1970d, 1970e, 1971c), Geological Survey Somaliland Protectorate (1954, 1956, 1959c, 1959a, 1959b)
Yemen and southern Saudi Arabia	Baker et al. (1996), Brown et al. (1989), Capaldi et al. (1987a, 1987b), Davison et al. (1998, 1994), Fairer (1985), Garzanti et al. (2001), Grolier and Overstreet (1978), Kruck (1980), Kruck et al. (1984), Kruck (1984), Mallick et al. (1990), Prinz (1984), Stern and Johnson (2019), USGS (1963), van der Gun (1995), Windley et al. (1996)
Geographic features	Lakes, coastlines and borders from Openstreetmap (https://www.openstreetmap.org , 06.01.2020) and Natural Earth (naturalearthdata.com). Lakes and coastlines are manually modified. Data for insert globe from Natural Earth.

Table 2: References for the geological map of the southern Red Sea & western Gulf of Aden region (fig. 6).

Geological map of the Afro-Arabian Rift System (fig. 3)	Abbate et al. (2015), Abebe et al. (2005), Aghanabati (1993), Ali and Lee (2019), Ali and Watts (2016), Altherr et al. (2019), Augustin et al. (2021), Barrier et al. (2004), Berhe et al. (1987), Bosworth and Stockli (2016), Brown et al. (1989), Camp et al. (1992, 1991), Camp and Roobol (1989), Chernet et al. (2020), Dawit (2016), Duncan and Al-Amri (2013), Egyptian Geological Survey (1981), Fantozzi and Ali Kassim (2002), Fournier et al. (2010), GEBCO Compilation Group (2020), Geological Survey of Ethiopia (1996), Kenea et al. (2001), Lucassen et al. (2008), Mège et al. (2016), Ministry of mines and energy (1985), Ministry of Petroleum and Mining (2004), Moufti et al. (2012), Mutebi et al. (2019), Pollastro et al. (1999), Pubelier (2008), Rime et al. (2022a, 2022b), Rooney (2020b, 2017), Scoon (2020), Sebai et al. (1991), Sneh et al. (1997a, 1997b, 1997c, 1997d), Stern and Johnson (2019), Survey of Kenya (1962), Thiélemont (2016), Trifonov et al. (2011), USGS (1963), van der Gun (1995), Vrbka et al. (2008), Zawacki et al. (2022) and Natural Earth.
Simplified basement map of the Red Sea and surrounding regions (suppl. mat. 3)	Geological map of the Afro-Arabian Rift System (references above), Brown et al. (1989), Flowerdew et al. (2013), Fritz et al. (2013), Johnson (2021), Johnson et al. (2011), Osman and Fowler (2021), Quick (1991), Rime et al. (2022b), Stern and Johnson (2019), Windley et al. (1996) and Esri orthophoto.

Table 3: References for other maps

region. Volcanic and intrusive rocks are merged as “igneous rocks”. Age divisions are identical except for adding a new unit showing the Eocene lavas in the southern part of the Main Ethiopian Rift. Sources and reference maps used are summarized

in Table 3. Another version of this map focused on basement structures was produced (suppl. mat. 3).

II.4 GEOLOGICAL HISTORY OF THE AFAR DEPRESSION AND AFRO-ARABIAN RIFT SYSTEM

In order to achieve a comprehensive understanding of the Afar Depression, the tectonic, magmatic and sedimentological history of different regions are first reviewed and discussed individually in this section. A chronological chart (fig. 7) summarizes these events. As the Afar Depression is part of and genetically linked with the whole Afro-Arabian Rift System, the latter is also included in this section. The division of the regions discussed in this section is indicated in fig. 1b while local names are presented in fig. 1a and 2. First, the pre-rift history is shortly discussed, followed by the detailed Cenozoic evolution of the Gulf of Aden, Red Sea, Ethiopian Plateau, Yemenite Plateau, Danakil Block, Ali Sabieh Block, NMER, south-western Afar, central-western Afar, central-eastern Afar and the Danakil Depression. A shorter, summarized, interpretative and chronological description of the tectonostratigraphic evolution of the Afar region is presented in section 6.2.

II.4.1 Pre-rift history

The basement of the Afar region is part of the Arabian-Nubian Shield and formed during the Neoproterozoic as part of a supercontinent cycle including the breakup of Rodinia and the East African Orogen. The East African Orogen comprises the accretion of several island-arc terranes, the formation of sedimentary and volcano-sedimentary basins, as well as a complex history of intrusion, volcanism, metamorphism and

faulting. By the end of the Ediacaran, the eroded and weathered Arabian-Nubian Shield formed a vast low-relief peneplain (Johnson et al., 2011; Johnson, 2021).

During Paleozoic and Mesozoic times, eastern Africa was influenced by two main tectonic phases (Frizon de Lamotte *et al.*, 2015; Macgregor, 2018). The first phase (Karoo phase) lasted from Permian to Triassic times and mainly affected the southern part of eastern Africa, as well as the southern Ogaden basin (Worku & Astin, 1992; Hunegnaw et al., 1998; Shawel et al., 2022), characterized by mainly NE-SW oriented structures. The “late Karoo” phase of rifting during the Lower Jurassic led to the separation of Madagascar and India. The second phase of rifting happened in the Upper Jurassic and Lower Cretaceous, forming multiple NW-SE oriented basins, including the Mekele Basin (Beyth, 1972; Sembroni *et al.*, 2017), the Blue Nile Basin (Gani et al., 2009; Chernet et al., 2020), the Somaliland basins (Ali & Watts, 2016; Ali & Lee, 2019) and the southern Yemenites basins (Redfern & Jones, 1995; As-Saruri et al., 2010). As these structures were also active during the Cenozoic, it is sometimes unclear when faulting occurred (Macgregor, 2018; Hagos et al., 2020).

In the Afar region, the first sediments were deposited during Paleozoic times. They consist of terrestrial sediments, often with glacial and peri-glacial affinities (e.g. Enticho Sandstone, Edaga Arbi Glacials, fig. 5). The Triassic to Lower/Middle Jurassic is dominated by widespread terrigenous conditions with the deposition of the Adigrat sandstones. During the Middle and Upper Jurassic, a major marine transgression synchronous with the initiation of the second rifting phase led to the deposition of marine

Broad stretching
Thinning or major tectonic event
Magma-accomodated extension
Other

- Seafloor spreading with well-defined mag. anomalies
- Seafloor spreading postulated
- Magmatism spread over wide area
- Magmatism confined in graben
- Other magmatic activity
- Plutonism or dyking

Continental conditions
Open marine conditions
Evaporitic conditions

◀ **Figure 7: Chronological event chart of the Afar and surrounding regions.** This figure summarizes the geological evolution of different regions discussed in section 4. The abbreviated regions are, as defined in fig. 1, 2 and section 4: the eastern (East.), central (Cent.), western (West.) Gulf of Aden and the Asal-Ghoubbet (As-G.) graben; the southern (South.), central (Centr.) and northern (North.) Red Sea, along with the Dead Sea fault (D); the northern (N.), central (C.) and southern (S.) Ethiopian Plateau; the Yemenite Plateau (Yem P.); the northern (N) and southern (S) Danakil Block, as well as its rotation (rot); the Ali Sabieh Block segmented in its central and eastern part (CE) and its north-western part (NW); the south-eastern Afar with its plateau/escarpment (P./E.) and its basin; the Northern Main Ethiopian Rift (NMER) with its south-eastern margin (SE-m), its north-western margin (NW) and the Wonji segments; the South-Western Afar with its plateau/escarpment (P/E), the Awash valley and the horst east of it (Aw/H.), and the Issa Graben; the central-western Afar with its escarpment (Esc.), the Awra plain, Sullu Adu + Affara Dara + Dabbayra + Limno (S/A) and the Dabbahu/ Manda Hararo rift segment (DMH); the central-eastern Afar (C-E Afar); the northern Afar with the Danakil Basin (Dana.) and the Buia Graben.

sediments in the Afar region (e.g. Antalo Limestone, fig. 5, Beyth, 1972, 1991; Gasse et al., 1986; Sagri et al., 1998; Bosellini et al., 2001; Guiraud et al., 2005; Abbate et al., 2015; Sembroni et al., 2017; Chernet et al., 2020). During the Cretaceous, a large regression resulted again in terrestrial conditions around the Afar Depression while marine deposition continued in the southern Ogaden basin (Beyth, 1991; Guiraud et al., 2005; Abbate et al., 2015; Ali, 2015). One last marine flooding in the Ogaden and Yemen basin occurred from the Paleocene till the Eocene (Auradu Fm. and Hadramawt Group on fig. 5, Guiraud et al., 2005; As-Saruri et al., 2010; Abbate et al., 2015; Ali, 2015).

II.4.2 Gulf of Aden

The Gulf of Aden, situated east of the Afar Depression, is segmented into four zones (fig. 1, 2). The Alula-Fartak Fault Zone separates the eastern and central Gulf of Aden, while the central and western Gulf of Aden are separated by the Shukra El Sheik Discontinuity. We included the Asal-Ghoubbet graben in the Gulf of Aden region as it constitutes its eastern tectonic prolongation. Note that more in-depth reviews of the geological history of the Gulf of Aden and the Red Sea are available (Bosworth et al., 2005;

Leroy et al., 2012; Stockli & Bosworth, 2019).

Leroy et al. (2012) and references therein show that the first syn-rift sediments of the eastern Gulf of Aden in the region of Dhofar (Oman) and Socotra Island (Arabian Sea) are dated to late Priabonian/early Rupelian, and dominated by shallow marine facies. The first tectonic activity is already recorded at the end of the mid-Eocene. Major faults are registered at the end of the Oligocene and a significant unconformity during the Burdigalian, interpreted to represent the onset of the break-up of the continental crust. Post-rift sedimentation has been recorded since Langhian times. These timings are consistent with the initiation of seafloor spreading before 17.5 Ma in the same area (Fournier *et al.*, 2010; Ogg, 2020).

The age for the initiation of clastic syn-rift sedimentation at the Yemenite margin in the central Gulf of Aden is believed to have initiated during the lower Oligocene (Watchorn et al., 1998; Leroy et al., 2012; Ali & Watts, 2016; Nonn et al., 2019), coherent with ages of ~32 Ma for rift initiation on the Somalian margin (Fantozzi & Ali Kassim, 2002; Ali & Watts, 2013). The first marine sediments of the central Gulf of Aden are dated to ~29.5 -27 Ma (Hughes *et al.*,

1991; Speijer *et al.*, 2020). On the Yemenite coasts, the facies changes from continental to marine in the Aquitanian/Burdigalian (Leroy *et al.*, 2012). Syn- to post-rift transition probably happened between 21 and 17 Ma (Watchorn *et al.*, 1998). The formation of SDR (Seaward Dipping Reflectors) is described between 21 Ma and 18 Ma, coinciding with significant thinning of the crust (Nonn *et al.*, 2019). In the westernmost part of this zone, near the Shukra El Sheik Discontinuity, these SDRs are better developed over longer time periods (~30 to 16 Ma), probably due to the more significant influence of the Afar plume (Tard *et al.*, 1991; Leroy *et al.*, 2012). Seafloor spreading starts before 16.3 Ma (Fournier *et al.*, 2010; Ogg, 2020).

West of the Shukra El Sheik Discontinuity, the nature of the crust changes (e.g. Manighetti *et al.*, 1997; Hébert *et al.*, 2001), see also Bouguer anomaly, suppl. mat. 4). The crust ranges from oceanic to the east to continental with an axial valley in the west (Hébert *et al.*, 2001). Few data exist on the rift history before seafloor spreading. SDRs are mentioned between ~30 and 16 Ma offshore Aden (Tard *et al.*, 1991; Leroy *et al.*, 2012). In the Somalian Guban basins, Ali, (2015) mentions extensive basaltic lavas of potential Miocene age. The marine connection between the Gulf of Aden and the Red Sea was established during the late Oligocene (Hughes *et al.*, 1991). Seafloor spreading in the eastern part began between 9 and 6 Ma (Leroy *et al.*, 2012) or 9 and 11.6 Ma (Fournier *et al.*, 2010; Ogg, 2020). Further west, oceanic spreading is believed to have started around 3 Ma ago (Manighetti *et al.*, 1997; Fournier *et al.*, 2010; Leroy *et al.*, 2012; Ogg, 2020). However, all

authors do not accept the oceanic nature of the spreading in this westernmost part of the Gulf (Hébert *et al.*, 2001). In the Gulf of Tadjoura, Daoud *et al.* (2011) report the initiation of faulting to be around 1.7 Ma, coherent with the interpretation of Manighetti *et al.* (1997).

Manighetti *et al.* (1998, 2001) mention the onset of diffuse rifting of the Asal-Ghoubbet rift at ~ 0.9 Ma with the establishment of a fully developed graben by ~0.3 Ma. This rift segment possibly propagated towards Manda Inakir around 0.2 Ma ago. Extension in Asal-Ghoubbet is now partially accommodated by magmatic intrusions (Cattin *et al.*, 2005).

Marine flooding of the Ghoubbet basin occurred around 9 ka ago, but it is plausible that it was also connected to the Gulf of Aden during previous interglacial(s) (Audin *et al.*, 2001). Lake Assal is now evaporitic but was lacustrine until 5-6 kyr ago (Gasse & Fontes, 1989).

II.4.3 Red Sea

II.4.3.1 Tectonic and magmatic history

There is a long-standing debate about the nature of the crust in the Red Sea and the position of the Ocean-Continent Transition (Girdler & Styles, 1974; Cochran, 1983; Bonatti, 1985; Girdler, 1985; Girdler & Underwood, 1985; Bohannon, 1986; Bonatti & Seyler, 1987; Girdler & Southren, 1987; Izzeldin, 1987; Hall, 1989; Egloff *et al.*, 1991; Makris *et al.*, 1991; Sultan, 1992; Sultan *et al.*, 1993; Rihm & Henke, 1998; Almalki *et al.*, 2014, 2015; Augustin *et al.*, 2014, 2016, 2021; Mitchell & Park, 2014; Bonatti *et al.*, 2015; Schettino *et al.*, 2016;

Ligi et al., 2018; Shi et al., 2018; Stockli & Bosworth, 2019; Le Magoarou et al., 2021; Issachar et al., 2022; Okwokwo et al., 2022). While all studies agree on the oceanic nature of the central trough in the central and southern Red Sea, the nature of the crust in the northern Red Sea and the one underlying the platforms in central and southern Red Sea is under debate. Most models acknowledge the presence of a thin crust (~20-8 km, Mooney et al., 1985; Egloff et al., 1991; Almalki et al., 2014) in these areas, thicker however than normal oceanic crust. The presence of significant magmatic material is testified by magnetic, gravimetry, as well as seismic refraction and reflection surveys in the central and southern Red Sea (Mooney et al., 1985; Izzeldin, 1987; Hall, 1989; Bohannon & Eittreim, 1991; Almalki et al., 2014; Mitchell & Park, 2014). It is, however, debated whether these geophysical data indicate heavily intruded continental crust or oceanic crust (Izzeldin, 1987; Hall, 1989). Overall, the distinction between heavily intruded continental crust and thickened oceanic crust is subtle and has only small implications for tectonic reconstructions. Both imply that significant crustal material was accreted during the rifting process.

On the southern margin of the Red Sea (south of approx. 15.5°N), significant volcanism was recorded between 31 and 18 Ma (Huchon et al., 1991; Davison et al., 1994), together with the emplacement of plutons between 20 and 26 Ma (Capaldi *et al.*, 1987b). Renewed phases of post-tectonic volcanic activity occurred at ~10 Ma at the foot of the escarpment (Jabal an Nar) and on Perim Island (Capaldi et al., 1987a; Mallick et al., 1990; Huchon et al., 1991). Huchon et al. (1991) indicate

extension starting at least 25 Ma ago and ending before 10 Ma, with a major pulse of tectonism between 18 and 10 Ma. In the same area, Menzies et al. (1992) show cooling ages of 16 Ma, and Davison et al. (1994) indicate exhumation since 20±5 Ma.

On the Bab-el-Mandeb platform, Mitchell et al. (1992) mention flat beds with only minor subsidence and tectonic influence in the Plio-Pleistocene sequence. These observations, along with the Late Pleistocene tectonic quiescence (Faure et al., 1980; Al-Mikhlafl et al., 2018) and the absence of volcanic material younger than the Mid-Miocene (Capaldi *et al.*, 1987a; Mallick *et al.*, 1990) indicate that tectonic activity in the area ceased after the Mid- to Late-Miocene. North of Bab-el-Mandeb, seismic and recent volcanic activity is recorded, forming the Hanish, Zubair and Jebel at Tair islands (Xu et al., 2015; Ruch et al., 2021).

Further north, on the SW Saudi Arabian margin (central Red Sea on fig. 7), Bohannon (1986), Bohannon et al. (1989) and Bohannon & Eittreim (1991) show faulting activity between 29 and 23 Ma and an intense period of extension around 25 Ma, just before significant dikeing between 24 and 21 Ma and the emplacement of plutons between 21 and 22 Ma (Sebai *et al.*, 1991). Magmatic extension probably dominated afterwards, while the uplift of the margins began at ~20 Ma and accelerated since 13.8 Ma (Bohannon *et al.*, 1989). On the Eritrean coast, Abbate et al., (2002), Ghebreab et al. (2002) and Balestrieri et al., 2005) similarly report denudation between 23 and 15 Ma. Omar & Steckler (1995) interpret a further uplift phase around 34 Ma in SW Saudi Arabia.

In the northern Red Sea, rifting style and chronology differ. While Makris & Rihm (1991), Rihm & Henke (1998) and Augustin et al. (2021) consider seafloor spreading along the whole Red Sea, many authors consider that the northern Red Sea has not yet attained this state or only partially, as testified by geophysical studies and the presence of Precambrian basement and granitic rocks on Zabargad Island and in offshore wells (Bonatti et al., 1981; Bonatti & Seyler, 1987; Bosworth et al., 1993, 2005; Cochran, 2005; Almalki et al., 2014, 2016; Mitchell & Park, 2014; Stockli & Bosworth, 2019; El Khrepy et al., 2021; Le Magoarou et al., 2021; Saada et al., 2021; Ali et al., 2022). This part of the Red Sea probably represents intruded and hyperextended continental crust with a thickness reduced to < 14 km (Le Magoarou et al., 2021; Ali et al., 2022).

At the NW Saudi Arabian margin, no significant extension or tilting occurred before 23 Ma. 23 Ma also corresponds to a significant dyking event (Bosworth & Stockli, 2016; Stockli & Bosworth, 2019) and the first syn-rift sediments in Egypt (Bosworth *et al.*, 2005). In Saudi Arabia, thermochronometry showed the initiation of extension at ~23 Ma with a second rift pulse at ~15 Ma (Szymanski et al., 2016; Stockli & Bosworth, 2019). Stockli & Bosworth (2019) provide a comprehensive review of timing.

The oldest measurable anomalies, identified in the central Red Sea (between roughly 16° N and 19° N), are C3n.2n (4.6 -4.8 Ma) according to Schettino et al. (2016) and 3A (6.3 -7.2 Ma) according to Izzeldin (1987) (age updated after Ogg, 2020). Other authors estimate the initiation of seafloor spreading around 10 Ma (Okwoko *et al.*, 2022), between 10-12 Ma (Izzeldin, 1987;

Rihm & Henke, 1998), at ~13 Ma (Le Pichon & Gaulier, 1988; Augustin et al., 2021) and ~14 Ma (Makris & Rihm, 1991). Augustin et al. (2016) suggest younging of the initiation of seafloor spreading northward, from 12 Ma at 20°N to 8-9 Ma at 23°N.

Mid-Miocene ages are also coherent with the ~14-12 Ma major structural event recorded in the southern Gulf of Suez and Midyan (Bosworth *et al.*, 2005), interpreted to represent the abandonment of the Gulf of Suez rift with the strain being accommodated by the Dead Sea Fault system (or Aqaba-Levant fault). This reorganisation is thought to be caused by the very thick and strong lithosphere of the Eastern Mediterranean hindering the northward propagation of the Gulf of Suez, while the Levant area provided a weaker lithosphere (Steckler & ten Brink, 1986; El-Sharkawy et al., 2021). Based on data from the Gulf of Suez, Bosworth et al. (1998, 2005) estimate that this ~14 Ma reorganisation caused a change in extension direction from rift normal to rift oblique (parallel to the Dead Sea Fault). It also corresponds to a significant increase in the Arabia – Nubia relative motion around 11-13 Ma ago (McQuarrie et al., 2003; ArRajehi et al., 2010; Reilinger et al., 2015) and to the closure of the Neotethys ocean (McQuarrie *et al.*, 2003). Motion along the Dead Sea fault is thought to have initiated before this ~14 Ma reorganization, around 20 Ma ago (Garfunkel, 2014; Nuriel et al., 2017; Morag et al., 2019; Craddock et al., 2021). The sinistral transform movement of the fault accommodates 100-110 km of displacement (Quennell, 1958; see also Garfunkel, 2014 and references therein).

II.4.3.2 Sedimentary environments

According to Hughes *et al.* (1991), Crossley *et al.* (1992) and Hughes & Beydoun (1992), the marine connection between the Gulf of Aden and the Red Sea was established during the late Oligocene (NP25, i.e. ~27 to 23 Ma, Speijer *et al.*, 2020), when adverse marine conditions are observed offshore Eritrea (Thio-1 well), but no indication of syn-rift sediments are reported further north or offshore Yemen. The configuration of this potential connection between the Gulf of Aden and the Red Sea is puzzling as it is not recorded in the sediments of the Tihama Plain (Hakimi *et al.*, 2021, 2022) and some authors (Smith & Santamarina, 2022) suggest that this initial flooding occurred from the north. In the Gulf of Suez and the northern Red Sea, the first syn-rift sediments are reported during the late Oligocene - Early Miocene (Evans, 1988; Montenat *et al.*, 1998; Plaziat *et al.*, 1998; Hughes *et al.*, 1999; Bosworth *et al.*, 2005), with Early Miocene marine waters probably coming from the Mediterranean Sea (Hughes *et al.*, 1991; Mitchell *et al.*, 1992; Orszag-Sperber *et al.*, 1998; Mandur, 2009). During the Early Miocene, open-marine conditions were reached offshore Eritrea and marine, but adverse conditions, possibly due to high clastic inputs, are reported offshore Sudan and Egypt (Hughes & Beydoun, 1992). During the Early or Mid-Miocene, the appearance of marine conditions is also reported below the Tihama Plain in Yemen (As-Saruri *et al.*, 2010; Hakimi *et al.*, 2016, 2021). Near Massawa (Eritrea), the Early Miocene lower Dogali Fm. show continental deposits (Drury *et al.*, 1994; Sagri *et al.*, 1998). The overlying Mid to Late Miocene upper Dogali Fm. is characterized by coastal conditions (Sagri *et al.*, 1998).

During Langhian to Serravallian, the basin became underfilled and evaporitic conditions prevailed throughout the Red Sea, while deep marine conditions were maintained in the Gulf of Aden, clearly indicating a decoupling of both basins (Savoyat *et al.*, 1989; Hughes & Beydoun, 1992; Mitchell *et al.*, 1992, 2021; Orszag-Sperber *et al.*, 1998; Plaziat *et al.*, 1998; Smith & Santamarina, 2022). On the Bab-el-Mandeb platform, Mitchell *et al.* (1992) mention the absence of salt, possibly indicating emersion of the sill.

Following this evaporitic crisis, the re-establishment of open-marine conditions occurred in the Pliocene, establishing shallow to moderately-deep (except for the central trough) marine conditions in the Red Sea (Stoffers & Ross, 1974; Savoyat *et al.*, 1989; Crossley *et al.*, 1992; Hughes & Beydoun, 1992; Mitchell *et al.*, 1992; As-Saruri *et al.*, 2010; Hakimi *et al.*, 2016, 2021).

II.4.4 Ethiopian Plateau

The Ethiopian plateau represents the western margin of the Afar Depression (fig. 1b). In the north, it merges with the Red Sea's western margin. The northern part of the Ethiopian plateau (N of approx. 13°N) is characterized by the sparse Cenozoic volcanic cover. In Eritrea, the main volcanic event is coeval with the emplacement of the Trap Series reported for the whole region and having early Oligocene ages (32.9-28.0; Drury *et al.*, 1994; Zanettin *et al.*, 2006). The second phase of volcanism is reported in the late Oligocene/Early Miocene (datings between 24.6 and 22.1 Ma; Zanettin *et al.*, 2006). In Ethiopia, datings in the Adigrat region (Hofmann *et al.*, 1997) and in the Mekele area similarly indicate the main phase of volcanic activity during the early

Oligocene, possibly extending from 33 to 26 Ma (Justin-Visentin, 1974; Sembroni *et al.*, 2017). A Miocene phase of volcanism (19-15 Ma) formed the Adwa Phonolite and trachyte, as well as volcanic rocks found along the coast (Drury *et al.*, 1994; Natali *et al.*, 2013; Hagos *et al.*, 2017), being synchronous with a late pulse of activity of the Simien Shield volcano (Kieffer *et al.*, 2004). Hagos *et al.* (2020) report some fault reactivation in the Mekele basin after the emplacement of the flood basalts.

In the central part of the Ethiopian plateau (approx. between 11 and 13°N), datings are scarce but Hofmann *et al.* (1997), Ukstins *et al.* (2002), Kieffer *et al.* (2004), Ayalew *et al.* (2021) and Eid *et al.* (2021) indicate ages ranging from 31.2 to 26.7 Ma. In contrast to the northern highlands, these Oligocene lavas almost entirely cover pre-Cenozoic rocks.

The southern part of the Ethiopian plateau (approximately between 11 and 9°N) similarly shows a wide magmatic cover of the pre-Cenozoic rocks, but their ages are younger than in the north. South of approx. 11° N, no early Oligocene lava is present (Morton *et al.*, 1979; Chernet *et al.*, 1998; Ukstins *et al.*, 2002; Coulié *et al.*, 2003; Kieffer *et al.*, 2004; Wolfenden *et al.*, 2004, 2005; Feyissa *et al.*, 2017), the oldest being 26.7±1 Ma (Feyissa *et al.*, 2017). Coulié *et al.* (2003) even measured a 20.7±0.3 Ma flow directly overlying the Mesozoic sandstones. The bulk of the datings indicate ages between 26.7 and 19.7 Ma, and the uppermost plateau lavas are as young as 12-10 Ma (Chernet *et al.*, 1998; Ukstins *et al.*, 2002; Wolfenden *et al.*, 2004).

II.4.5 Yemenite plateau

The Yemenite plateau forms the eastern margin of the southern Red Sea and the northern margin of the Gulf of Aden (fig. 1b). Dating of lavas on this plateau indicate that the eruption of the traps was a very brief event lasting from ~30.9 to ~26.5 Ma, (Zumbo *et al.*, 1995a; Baker *et al.*, 1996; Riisager *et al.*, 2005; Ukstins Peate *et al.*, 2005; Thines *et al.*, 2021). Recent publications indicate that the volcanic phase was even shorter, ending around 27.7 Ma. This contrasts with the Ethiopian plateau, where volcanism continued during the late Oligocene and the Miocene.

II.4.6 Danakil Block

The Danakil Block refers to the broad area of relatively more elevated topography and Bouguer gravity anomaly situated between the Gulf of Zula and the Gulf of Tadjoura (fig. 1, suppl. mat. 4). The geological history of the Danakil Block is poorly constrained due to the lack of detailed field studies since the 70s. We segment the Danakil Block into two parts: the northern part (approx. north of 13.2°N) exposes mainly Precambrian basement, Mesozoic sediments and the massive Nabro volcanic range. The southern part is almost entirely covered by Neogene and Quaternary volcanics. This limit also marks a possible tectonic block boundary (Viltres *et al.*, 2020).

On the northern Danakil Block, Frazier (1970) reports two datings on volcanic rocks in the Howlakil Bay and N of the Amphila Bay at 15 and 23 Ma, respectively. Similarly, Brinckmann & Kürsten (1971) report the presence of isolated igneous rocks of ages ranging from ~25 to ~16 Ma in the Danakil Alps.

In the southern Danakil Alps, the Dioita Massif is covered by 27 to 23 Ma lavas (Brinckmann & Kürsten, 1971; Civetta *et al.*, 1975a). Two ~21-23 Ma old granitic plutons are described: the Dioita granite and the Asa Ale granite (Brinckmann & Kürsten, 1971; Barberi *et al.*, 1975; Civetta *et al.*, 1975a). They are overlain by ~19 Ma old rhyolites and a few Recent lava flows (Civetta *et al.*, 1975a; De Fino *et al.*, 1978).

The relative chronology of the Nabro volcanic range is well-defined (Wiart *et al.*, 2000; Wiart & Oppenheimer, 2005) with the latest eruption in 2011 (e.g. Goitom *et al.*, 2015), but the oldest lavas have not been dated. A maximal Pleistocene age is generally agreed upon (Varet *et al.*, 1975; Wiart & Oppenheimer, 2005). The lavas situated SE of the Nabro range have ages between 8 and 6 Ma (Brinckmann & Kürsten, 1971; Civetta *et al.*, 1975b).

On the southern Danakil Block, the age of the lavas covering most of the Danakil Block between 12.4°N and 13.2°N is poorly constrained. Datings from Barberi *et al.* (1975), Civetta *et al.* (1975b) and Zumbo *et al.* (1995b) indicate Mid-Late Miocene to Recent ages. The youngest lavas are found in the Assab and Gufa volcanic ranges (De Fino *et al.*, 1973; Civetta *et al.*, 1975b). The oldest rocks exposed on the Djiboutian part of the Danakil Block are the Mabla Rhyolites (and associated basalts, sometimes called Andolei basalts). Zumbo *et al.* (1995b) dated this formation between 15.4 and 11.8 Ma. Older K/Ar datings of Barberi *et al.*, (1975) indicated a minimum age of 9.5 Ma. Extrapolation of the same units found on the Ali Sabieh Block suggests that volcanic activity began significantly earlier (Oligocene), but the absence of older outcrops on the

Danakil Block does not allow to confirm this hypothesis. The overlying Dalha basalts in this part of Djibouti yield ages of 6.5 and 6.7 Ma (Barberi *et al.*, 1975; Zumbo *et al.*, 1995b). Comparison with the same unit in Ali Sabieh indicates that this unit might expand from 8.6 to 3.8 Ma (Barberi *et al.*, 1975; Chessex *et al.*, 1975; Zumbo *et al.*, 1995b; Audin *et al.*, 2001; Daoud *et al.*, 2011). These units are overlain in the NW by the Quaternary stratoid basalts (Zumbo *et al.*, 1995b) and near the shore in the SSE by Gulf Basalts with ages ranging from 3.5 to 1 Ma (Barberi *et al.*, 1975; Daoud *et al.*, 2011).

No complete interpretation of the tectonic history of the Danakil Block is available in the literature, but a chronology can nevertheless be proposed. The similarities of the volcanic and plutonic events of the northern Danakil Block and the Yemen coast (fig. 7) suggest a similar tectonic history related to the opening of this part of the Red Sea, i.e. starting before 25 Ma and ending before 10 Ma (see section 4.3.1). The observation that the ~19 Ma rocks of the Dioita Massif are heavily faulted, while the ~7 Ma rocks situated south of the Nabro range are not affected by tectonism (Barberi *et al.*, 1975) strengthens the hypothesis of a cessation of tectonic activity in the Mid-Miocene. Also, C.N.R. - C.N.R.S. Afar Team (1973) mention that Jurassic sediments are tilted in another direction compared to the Red Bed Series, implying deformation before deposition of the latter (~8.5 Ma, see section 4.13.2). The northern Danakil Block experienced a second phase of deformation on its westernmost margin linked to the opening of the Danakil Depression. This phase is discussed in section 4.13.2.

The detailed tectonic history of the southern Danakil Block is unknown

before the Mid-Miocene, but Makris *et al.* (1975), Makris & Ginzburg (1987) and Redfield *et al.* (2003) showed that the crust of the Danakil Block is thinner than the plateaus and thus experienced significant extension. It is reasonable to assume that the early tectonic history of the southern Danakil Block is similar to the one of the northern part of the Danakil Block, Yemen and Ali Sabieh Block, i.e. extension related to the opening of the Red Sea from late Oligocene to Mid Miocene. Surface geology shows that along the Red Sea coast, the Mabla Series is heavily faulted and tilted with an orientation parallel to the Red Sea trend (NNW-SSE) (Barberi *et al.*, 1975). The Dahla/Dalha basalts lie unconformably on them and are only slightly tilted and faulted (Barberi *et al.*, 1975; Chauvet *et al.*, 2023), a further indication of the cessation of tectonism in the Bab-el-Mandeb area during the Late Miocene (see also section 4.3.1). In contrast, on the SW margin of the block (Makarassou region), thickness changes indicate continuous tectonic activity during the eruption of the Dahla basalts (Le Gall *et al.*, 2011; Geoffroy *et al.*, 2014; Chauvet *et al.*, 2023). The overlying Plio-Quaternary lavas similarly show very little faulting except in Moussa Ali, Makarassou fault system and on the coasts of the Gulf of Tadjoura. The faults in the latter zone are parallel to the ones found on the floor of the Gulf of Tadjoura (Dauteuil *et al.*, 2001), and we interpret it as being caused by the Quaternary opening of the Gulf (Manighetti *et al.*, 1997; Daoud *et al.*, 2011; see also section 4.2).

Paleomagnetic data of Manighetti *et al.* (2001) show that the Danakil Block rotated 10.7° since ~ 8 Ma. Assuming a constant rotation rate and a total rotation

of 30° , they estimated the initiation of rotation around 20 Ma ago, but several other authors agree on a significantly younger age, i.e. at least at ~ 11 Ma (Eagles *et al.*, 2002), 9.3 ± 4 Ma (McClusky *et al.*, 2010) and 11 ± 2 Ma (Reilinger & McClusky, 2011).

II.4.7 Ali Sabieh Block

The Ali Sabieh Block (or Aisha Block) is an elevated area situated at the border between Ethiopia, Djibouti and Somalia (fig. 1), covered by Mesozoic sediments attesting the continental nature of the block (e.g. Chessex *et al.*, 1975; Geological Survey of Ethiopia, 2015; Le Gall *et al.*, 2015). It was intruded and covered by trap basalts (19 – 28 Ma, alternatively named Ali Sabieh basalts, Adolei/Andolei basalt or Galile Fm., Barberi *et al.*, 1975; Black *et al.*, 1975; Chessex *et al.*, 1975; Zumbo *et al.*, 1995b; Audin *et al.*, 2004; Le Gall *et al.*, 2010, 2015). The overlying Mabla series have ages ranging between 19 and 13 Ma (Barberi *et al.*, 1975; Black *et al.*, 1975; Chessex *et al.*, 1975; Zumbo *et al.*, 1995b) of which many dykes intrude the Ali Sabieh basalts (Chessex *et al.*, 1975; Zumbo *et al.*, 1995b). The Dahla series lies unconformably on the Mabla Series and mainly yields ages between 8 and 4.5 Ma (Barberi *et al.*, 1975; Black *et al.*, 1975; Chessex *et al.*, 1975; Audin *et al.*, 2004) while some authors consider rocks as recent as 3.5 Ma being part of this unit (Zumbo *et al.*, 1995b; Le Gall *et al.*, 2015). The rocks dated between 3.5 and 4.5 Ma were all sampled on the eastern side of the Block and classified by other authors as part of the Somali Basalts. This latter unit was dated between 7.2 and 3.6 Ma (Daoud *et al.*, 2011) and is sometimes topped by the Pleistocene Goumarre Basalts (~ 2.5 -1.7 Ma; Daoud *et al.*, 2011). Near the Gulf of

Tadjoura, Stratoid and Gulf basalts with ages ranging from ~2.8 to 0.8 Ma (Courtillet et al., 1984; Audin et al., 2004; Daoud et al., 2011) represent the last volcanic events of the Ali Sabieh Block. On the SW side of the Block, the Dahla Series is covered by Stratoid lava of unknown age but probably being Pliocene based on the extrapolation of this unit in south-eastern Afar (see section 4.8).

South of the Ali Sabieh Block *sensu stricto*, the geological history of the region South of Aisha and E of Adigala is similar to the one of the Block itself (Black et al., 1975; Audin et al., 2004; Geological Survey of Ethiopia, 2015).

Towards the coast of the Gulf of Tadjoura, the Ali Sabieh Block is covered by the previously described Dahla unit and by Quaternary Stratoid and Gulf basalts with ages ranging from 2.5 to 0.8 Ma (Barberi et al., 1975; Courtillet et al., 1984; Audin et al., 2004; Daoud et al., 2011).

From a tectonic point of view, Chessex *et al.* (1975) mention a first faulting event before the deposition of the Ali Sabieh lavas. The second stage of deformation occurs after the emplacement of the Ali Sabieh lavas, between 16 and 22 Ma, according to Chessex *et al.* (1975), or between 15 and 20 Ma, according to Black *et al.* (1975). The next faulting event is reported between the emplacement of the Mabla and Dahla Series. After the emplacement of the Dahla Series, renewed faulting activity is described (Chessex et al., 1975; Muller & Boucarut, 1975; Arthaud et al., 1980; Daoud et al., 2011), followed by the faulting of the Stratoid and Gulf basalts in the Quaternary. This faulting is related to the opening of the Gulf of Tadjoura and the Asal – Ghoubbet graben, where most

of the present-day tectonic activity is concentrated (Manighetti et al., 1997; Audin et al., 2004; Daoud et al., 2011).

Paleomagnetic studies (Kidane, 2016) showed a 30° clockwise rotation of the Ali Sabieh Block sometime between 60 and 8 Ma. Audin *et al.* (2004) show a clockwise rotation of the NW margin of the Ali Sabieh Block between 8 and 4 Ma and propose that the movement between the Ali Sabieh and Danakil Block was accommodated by the strike-slip activity of the NE-SW Holhol fault between ~4 and 1 Ma before the strain was transferred in the present-day Gulf of Tadjoura.

On the eastern side of the Block, the absence of faulting indicates tectonic stability, at least since Late Miocene times (Barberi et al., 1975; Audin et al., 2004).

II.4.8 South-eastern Afar

The south-eastern Afar is situated between the Goba'Ad graben, the Issa graben and the Somalian Plateau (fig. 1). The Somalian Plateau differs from the Ethiopian Plateau with a smaller volcanic coverage, almost exclusively confined to the escarpment. The first activity recorded in south-eastern Afar is the formation of the Ogaden Dyke Swarm at 22-25 Ma in the Marda Fault Zone, postulated to be a Precambrian reactivated structure (Black et al., 1974; Purcell, 1976; Mège et al., 2016). More than 200 km further south in Ogaden, the continuation of the same dyke swarm indicates older ages, reaching 30 Ma (Mège *et al.*, 2016). Between 24 and 22 Ma, the Alaji basalts erupted at the present-day escarpment and on the nearby plateau (Kunz et al., 1975; Chernet et al., 1998). Mid-Miocene lavas

(~13 Ma, Berhe, 1986) are also present on the plateau.

At the foot of the escarpment, lava flows of Late Miocene ages (6.8-7.4 Ma) have been reported (Berhe, 1986). They are topped by slightly younger Stratoid basalts with ages ranging from 4.5 to 5.1 Ma (Berhe, 1986; WoldeGabriel *et al.*, 1992a; Chernet *et al.*, 1998).

It is unclear when stretching of the escarpment took place (Black *et al.*, 1975), mainly because it does not expose much syn-rift volcanism. The late Oligocene/Early Miocene traps lie conformably above the Mesozoic sediments (Juch, 1975; Geological Survey of Ethiopia, 2010a). Late Miocene lavas are not outcropping enough to be conclusive, but Pliocene lavas have only undergone limited faulting, mainly towards the N and the Buren plain or towards the W and the NMER, which we interpret as evidence for post-tectonic magmatism, consistently with the results of (Tortelli *et al.*, 2022) who show that these lavas were probably emplaced on an already thinned lithosphere. The main stretching phase could thus have happened anytime between ~22 and ~4.5 Ma.

The Dembel Volcanic Range in the easternmost part of the region has experienced volcanic activity since 3.35 Ma ago, which continued until Recent times, along with some faulting activity (Black *et al.*, 1975; Audin *et al.*, 2004). However, data from Doubre *et al.* (2017), Moore *et al.* (2021) and Sangha *et al.* (2022) show that this region currently hardly accommodates any extension. The Buren Plain is mainly covered by Holocene/Recent sediments, suggesting active subsidence.

The history of the horst between the Buren Plain and the Goba'Ad graben has never been investigated in detail. The oldest rocks are dated at 6.3 Ma (Black *et al.*, 1975) or 4.3 Ma (Barberi *et al.*, 1975). The youngest is 2.4 Ma (Barberi *et al.*, 1975), but Varet *et al.* (1975), Geological Survey of Ethiopia (2015) and our mapping results indicate the presence of younger, undated lavas, hypothetically assigned to Early/Middle Pleistocene. The orientation of feeder dykes (coinciding with magnetic lineations (Purcell, 2018) in the area indicates a SSW-NNE extension during the Pliocene (Christiansen *et al.*, 1975). This would fit the hypothesis of Audin *et al.* (2004), Wolfenden *et al.* (2004) and Hammond *et al.* (2011) that this area was an active rift segment during the Mio-Pliocene.

II.4.9 NMER

The Northern Main Ethiopian Rift (NMER) forms a rift valley between the Somalian and Ethiopian Plateaus, south of approximately 10°N (fig. 1). On the eastern side of the NMER, the geological history of the plateau corresponds to the geological evolution of South-Eastern Afar with Trap basalts between 24 and 22 Ma (Alaji Group). Few datings are reported between 20 and 16 Ma on the present-day escarpment, but datings cluster mainly around 13 Ma (on the plateau and the escarpment) and between 9 and 11 Ma (at the base of the escarpment), indicating renewed magmatic activity in the Mid to Late Miocene (Kunz *et al.*, 1975; Morbidelli *et al.*, 1975; Berhe, 1986; Chernet *et al.*, 1998) (Tarmaber Fm, Anchar Basalts and Arba Guracha Silicics acc. to Geological Survey of Ethiopia, 1985 and Chernet *et al.*, 1998). Intercalated between this volcanic material, the diatomite-

dominated Chorora Fm indicates the presence of a lacustrine basin at ca. 10 Ma. This basin was localized between the future Afdem, Asebot and Gara Gumbi volcanoes and the foot of the escarpment (Sickenberg & Schönfeld, 1975; Tiercelin *et al.*, 1979; Geraads *et al.*, 2002). This timing corresponds to the initiation of subsidence according to (Chernet *et al.*, 1998). NE of Mieso, Juch (1975) reports the first angular unconformity in the Arba Guracha Silicics, which are slightly older (~10-14 Ma; Chernet *et al.*, 1998). Berhe (1986) also reports the first rifting at 14 Ma with the formation of this unit. SE of Mieso, this unit is conformable (Juch, 1975).

The large volcanic edifices of Asebot and Gara Gumbi have been dated between 5 and 6.85 Ma, and the Afdem volcano is assumed to have a similar age (Kunz *et al.*, 1975; Geological Survey of Ethiopia, 1985; WoldeGabriel *et al.*, 1992a; Chernet *et al.*, 1998; Feyissa *et al.*, 2017). They show little faulting except on their W side towards the NMER and were thus probably emplaced after the first stretching phase. (Juch, 1975) reports unconformities in the lowermost part of this Late Miocene unit which we interpret as a second phase of stretching restricted to the rift floor. At the NW foot of the volcanoes, the rift is floored by Mio-Pliocene volcanic products and sediments (~5.6 Ma to ~3.8 Ma) covered near Awash by mostly undeformed Early Pleistocene lavas (WoldeGabriel, 1987; Chernet *et al.*, 1998; Kidane *et al.*, 2010; Hujer *et al.*, 2015). This second tectonic event therefore probably lasted from the Late Miocene to the Pliocene.

On the NW margin of the NMER, Wolfenden *et al.* (2004) similarly note the first unconformity at ~10.6 Ma, linked with significant magmatic production

until ~6.6 Ma (Kessem Fm.) (Chernet *et al.*, 1998; Ukstins *et al.*, 2002; Wolfenden *et al.*, 2004) and a regional southeastward basinal tilt (Wolfenden *et al.*, 2004). Deformation before 11 Ma cannot be excluded but must have been minor. Slightly further N along the Ankober fault, Wolfenden *et al.* (2005) note the first tectonic activity in the ~11 Ma Aliyu Amba Fm. A significant hiatus with a major unconformity is recorded between ~6.6 and ~3.5 Ma and is interpreted to be diachronous with strain migrating towards the centre of the rift. Pliocene (~3.6 – 2.5 Ma, Ukstins *et al.*, 2002; Wolfenden *et al.*, 2004) magmatic products are found in the lower part of the escarpment (Balchi Fm.).

The *Wonji segments* or the *Wonji fault belt* are narrow belts of intense recent faulting in the MER (Mohr, 1967). In the study area, they include the Kone, Fentale, Dofen and Hertali volcanoes. Their age range from ~1.8 Ma to very Recent (Morton *et al.*, 1979; Kazmin *et al.*, 1980; Chernet *et al.*, 1998; Williams *et al.*, 2004; Vogel *et al.*, 2006; Feyissa *et al.*, 2017; Siegburg *et al.*, 2018). Deformation was focused in the Wonji belt since ~2 Ma (Wolfenden *et al.*, 2004; Corti, 2009). Deformation is accommodated by both brittle and magmatic processes (Kurz *et al.*, 2007). One Lower Pleistocene fluvial and lacustrine clastic unit is reported near the Dofen volcanic centre by WoldeGabriel *et al.* (1992b). Sedimentation was probably nearly continuous since then.

II.4.10 South-western Afar

The south-western Afar comprise the area between the Issa Graben and the Ethiopian Plateau (fig. 1). The geological history of the south-western Afar margin

has been well studied, particularly by Wolfenden *et al.* (2005). Early Oligocene flood basalts are reported in the northern part of the region (Ukstins *et al.*, 2002; Coulié *et al.*, 2003; Rooney *et al.*, 2013), but, as discussed in section 4.4, the traps outcropping in the southern part are younger.

The first stage of deformation initiated between 29 and 26 Ma along border faults and lasted until ~20 Ma (Wolfenden *et al.*, 2005). Deformation was accompanied by significant acidic volcanism (Ukstins *et al.*, 2002; Wolfenden *et al.*, 2005; Rooney *et al.*, 2013; Feyissa *et al.*, 2017). In the southern part of the region (9.95°N, transect E-E' of Wolfenden *et al.*, 2005), the first tectonic event is recorded along the Ataye border fault before the deposition of the ~25 Ma old Brikitu Fm. This strongly contrasts with the initiation of tectonic activity ~12 Ma ago in the NMER, only 40 km south (see Wolfenden *et al.*, 2005 and section 4.9).

At ca. 16 Ma, the locus of extension shifted towards the east (Wolfenden *et al.*, 2005), associated with basaltic and intermediate volcanism (~16-14 Ma; Ukstins *et al.*, 2002; Wolfenden *et al.*, 2005).

Around 7 Ma, the strain again migrated eastward towards the foot of the escarpment, corresponding to the 3rd stage of deformation (Wolfenden *et al.*, 2005) and accompanied by volcanic products restricted to the foot of the escarpment (~7-3 Ma; Barberi *et al.*, 1975; Kunz *et al.*, 1975; Coulié, 2001; Wolfenden *et al.*, 2005; Kleinsasser *et al.*, 2008; Feyissa *et al.*, 2017). The absence of large offset faults in this unit is interpreted by (Wolfenden *et al.*, 2005) as an indication that strain was mainly accommodated by dyke injection. At the

SE base of the escarpment, WoldeGabriel *et al.* (2001, 2013) report fluvio-lacustrine clastic sedimentary units with ages from 5.77 to 3.8 Ma.

The oldest deposits of the Awash valley are found in the middle Awash (ca. 10.4°N) with ages ranging from 5.5 to 4 Ma (Renne *et al.*, 1999), and in the ~5.2 – 3.9 Ma fluvio-lacustrine Sagantole Formation formed in an eastward-dipping half-graben localized in the present-day Lower Awash valley (ca. 11.1°N, Quade *et al.*, 2008).

The fluvio-lacustrine Pliocene (3.9 – 2.9 Ma, Quade *et al.*, 2008) Hadar Formation in the lower Awash probably formed in response to extension parallel to the Red Sea Rift System, as testified by the north-northeast depositional direction observed in this unit (Wynn *et al.*, 2008; Campisano, 2012; DiMaggio *et al.*, 2015a), indicating that the region was not yet dominated by MER tectonics. Few studies, however, propose another interpretation with a N-S-oriented basin (see Campisano, 2012 and references therein). Deposits of comparable age (3.39-4.26 Ma, White *et al.*, 1993) in the middle Awash valley are believed to belong to a distinct basin (Wynn *et al.*, 2008). The deposition of the fluvatile Busidima Formation (2.7 – 0.15 Ma, Quade *et al.*, 2008) marks a significant change in the dynamic of the basin, the activation of the N-S As Duma fault and the formation of a marginal graben (Quade *et al.*, 2008; Wynn *et al.*, 2008). Simultaneously, sedimentation began in the axial Adda'Do and Yangudi Basin (Issa Graben), according to Wynn *et al.* (2008). Sedimentation in the lower Awash valley ended approx. 0.15 Ma ago, probably related to base-level drop (Quade *et al.*, 2008) and the formation of the Kariyu Basin (Kalb, 1995).

Magmatic rocks situated on the horst east of the Awash basin were dated between 4 and 2.3 Ma (Barberi *et al.*, 1975; Kunz *et al.*, 1975; Lahitte *et al.*, 2003b) and were emplaced in a thinned lithosphere (Tortelli *et al.*, 2022). One dating of 7.1 Ma is presented by Barberi *et al.* (1975) at the base of the eastern fault scarp of the Issa graben, suggesting that Late Miocene rocks are buried below the outcropping Pliocene lavas.

The age of the Issa graben is probably Quaternary, as proposed by Varet *et al.* (1975) and Wynn *et al.* (2008), and possibly related to the reorganisation of the Awash valley at ~2.8 Ma. This is comparable in age with the tectonic activity of the Wonji segment of the NMER (see section 4.9). No volcanic material in this graben has been dated. Only a few flows on the Abida and Ayelu volcanoes yield ages younger than 0.8 Ma (Kunz *et al.*, 1975; Lahitte *et al.*, 2003b; Rees *et al.*, 2023).

II.4.11 Central-western Afar

The central-western Afar comprises the area between the Affara Dara and the Ethiopian Plateau (fig. 1). At the margin, the trap series were dated between 31.2 Ma and 28.5 Ma (Hofmann *et al.*, 1997; Ukstins *et al.*, 2002; Kieffer *et al.*, 2004; Ayalew *et al.*, 2019; Eid *et al.*, 2021). Datings between ~23 and ~20 Ma are reported by Stab *et al.* (2016) and Ayalew *et al.* (2019) in the lavas of the Chifra series. The top of this formation was dated at 7 Ma (Stab *et al.*, 2016). It is unclear whether low-volume but continuous volcanism existed between these two Miocene age clusters. No thickness variations, unconformities or growth strata were observed in the Trap and Chifra series, but it is postulated that limited extension without block tilting

formed marginal grabens (Stab *et al.*, 2016). The Chifra series is unconformably overlain by the flat-lying Arabati Rhyolite, probably ~6.8-5.5 Ma old (Coulié, 2001; Ayalew *et al.*, 2019). This indicates that significant tectonic activity only occurred during the Late Miocene. Ayalew *et al.* (2019) measured WSW - ENE dykes of 24.3 – 17.4 Ma old in the same region.

At the foot of the escarpment, the Awra plain is covered by poorly studied sediments. Erosion features indicate that most of those sediments are not Recent. Studies lack in this basin except along the Mille River Basin, where sediments were dated as Pliocene (3.4 to 3.8 Ma, Deino *et al.*, 2010; Haile-Selassie *et al.*, 2015; Saylor *et al.*, 2019). Extrapolation to the whole basin is difficult but remains plausible. Volcanoes in the northern part of the basin were dated at 3.5 and 1.6 Ma (Coulié, 2001; Stab *et al.*, 2016). Very recent volcanic cones are visible in the centre of the basin. Faults are rarely identified, and the relative scarcity of Recent sediments could indicate that the basin is not actively subsiding.

Whereas the escarpment seems to lack deformation during the early rifting phase, the Sullu Adu area shows a different evolution. In the northern part, Coulié (2001) and Stab *et al.* (2016) describe an angular unconformity between the Early Oligocene (30.6 to 29.7 Ma) lavas and the Miocene (25.4 to 8.6 Ma) Finto rhyolites. Stab *et al.* (2016) mentions the base of this unit to be 23.5 Ma old but Coulié (2001) shows an age of 25.4 Ma, probably in the same unit. It is unclear whether this long-lasting unit represents continuous deposition or whether it corresponds to two eruptive phases. In each case, the unit lacks any internal deformation or unconformity,

evidencing tectonic quiescence during most of the Miocene. Renewed tectonic activity between this series and the Sullu Adu Series (i.e. between 8.6 and 5.4 Ma) formed a second major angular unconformity (Stab *et al.*, 2016). The Sullu Adu series (5.4 – 3.9 Ma) are horizontal but cut by faults. In the southern part of the Sullu Adu horst (region of Mille), these lavas yield slightly younger ages (3.5 -2.7 Ma, Kunz *et al.*, 1975; Coulié, 2001; Kidane *et al.*, 2003; Stab *et al.*, 2016). Barberi *et al.* (1975) report an age of 7.4 Ma for a tilted block covered by the Sullu Adu series.

NNW of the Sullu Adu horst is the Recent Dabbayra volcanic range. The timing of the initiation of volcanic activity (in the eastern part of the area) is unknown, but extrapolation from datings of Coulié (2001) and Stab *et al.* (2016) in the Awra plain points toward a possible Plio-Pleistocene age². Further to the NNW, the 23 Ma Limmo Granite forms one of the few outcropping plutons of the Afar Depression, intruding in 23 – 26 Ma porphyritic and extrusive rocks (Black *et al.*, 1972).

East of the Dabbahu – Manda Hararo volcanic range, the Affara Dara is formed by another granitic body with intruding trap basalts at ~25 Ma (Barberi *et al.*, 1972a; C.N.R - C.N.R.S. Afar Team, 1973; Varet, 2018). This body is covered by Early-Middle Pleistocene lavas (Barberi *et al.*, 1975). This massif was most probably linked to the Sullu Adu horst before being separated by the proto-Dabbahu- Manda Hararo range.

The Dabbahu-Manda Hararo volcanic range is formed by lavas younger than 0.3 Ma (Barberi *et al.*, 1972a; Lahitte *et al.*,

2003b; Ferguson *et al.*, 2013a; Field *et al.*, 2013; Medynski *et al.*, 2015). During this timeframe, the range formed the locus of extension, with most of the strain being accommodated by dyke intrusion rather than faulting (Ebinger *et al.*, 2010; Ferguson *et al.*, 2013a), as exemplified by the 2005-2010 event (e.g. Wright *et al.*, 2006; Ayele *et al.*, 2009; Grandin *et al.*, 2009, 2010; Barnie *et al.*, 2016). This area shows low crustal contamination, high partial melting, shallow Moho, linear magnetic anomalies, an extensive and long-lasting magma reservoir and is believed to be heavily intruded, representing an almost oceanic crust, close to continental breakup (Ayele *et al.*, 2009; Hammond *et al.*, 2011; Bridges *et al.*, 2012; Desissa *et al.*, 2013; Didana *et al.*, 2014; Reed *et al.*, 2014; Johnson *et al.*, 2016; Lewi *et al.*, 2016; Ayalew *et al.*, 2019; Ahmed *et al.*, 2022). The margins of the Dabbahu-Manda Hararo volcanic range are formed by Pleistocene lavas (2.3-0.6 Ma, Barberi *et al.*, 1975; Kunz *et al.*, 1975; Kidane *et al.*, 2003; Lahitte *et al.*, 2003b).

II.4.12 Central-eastern Afar

The central-eastern Afar is situated between the Tendaho and Goba'Ad grabens and the Danakil Block (fig. 1). The central-eastern Afar is almost exclusively covered by rocks of Quaternary age, hiding part of its geological history. A few late Pliocene K/Ar datings were obtained near the Assal Rift and two on the Gamarri fault scarp (Barberi *et al.*, 1975). Dating of the same scarp by Ar/Ar techniques indicates that previous datings were probably imprecise (Ar/Ar ages ~2 Ma, Kidane *et al.*, 1999, 2003; Lahitte *et al.*, 2003b; Ahn

² If you read this, I pay you a beer. I thought nobody would read the entire text.

et al., 2016). Other datings (> 130 datapoints) indicate ages of 2.3 Ma or younger, assigned to the upper Stratoid series (Brinckmann & Kürsten, 1971; Barberi et al., 1975; Civetta et al., 1975b; a; Courtillot et al., 1984; Zumbo et al., 1995b; Manighetti et al., 1998; Kidane et al., 1999, 2003; Lahitte et al., 2003b). Besides the few pre-Quaternary datings discussed above, most of the oldest datings are found between the Tendaho / Goba'Ad graben and the Asal – Ghoubbet graben. The age of the lavas tends to be young towards both of these active rift segments and Manda Inakir. The Recent (< ~0.6 Ma) lavas are restricted to four areas: the Tendaho/Goba'Ad graben, the Dadar graben, Manda Inakir and Asal-Ghoubbet (the latter discussed in section 4.2 above). Except for Manda Inakir, Recent lavas are confined to active rift segments. Stab *et al.* (2016) named this phase the “localizing phase”.

The central-eastern Afar is dissected by numerous grabens. They are mainly filled by Pleistocene-Recent sediments (e.g. Gasse & Street, 1978; Gasse et al., 1980). The early tectonic history of the area is mainly unknown due to the ubiquitous Quaternary cover. However, data from Tortelli *et al.* (2022) suggest that the lithosphere was still relatively thick during the emplacement of the Early Pleistocene *Upper Stratoid Fm.*, indicating a late thinning. The Tendaho graben is filled in some parts by more than 1000-1300 m of sediments intercalated with basaltic lava flows (Battistelli *et al.*, 2002). Formation of this graben has been postulated to have started 1.8 Ma ago (Acton et al., 1991; Acocella et al., 2008). However, this value is only constrained by the average age of the Stratoid basalts. We instead consider the graben to be initiated around 1.1 to

0.6 Ma ago, as argued by Lahitte *et al.* (2003b), coherently with petrological indications of rift focusing since 1.1 Ma (Tortelli *et al.*, 2022). According to Acocella *et al.* (2008), the rifting activity decreased in the Tendaho Graben since ~0.2 Ma, in favour of Manda Hararo. However, lakeshore markers indicate still active subsidence during the Holocene (Mologni *et al.*, 2021). The eastern part of the Goba'Ad graben comprises older sediments, up to Pliocene, indicating an earlier formation (fig. 4, Gasse *et al.*, 1987). Pagli et al. (2014), Doubre et al. (2017), Moore et al. (2021) and Sangha et al. (2022) show that the Goba'Ad graben hardly accommodates any shortening nowadays.

II.4.13 Danakil Depression

The Danakil Depression *sensu lato* forms an area of lower elevation between the Gulf of Zula, the Ethiopian Plateau, the Affara Dara and the Danakil Block (fig. 1). We segment it here in the Afar-Danakil transition zone, the Danakil Depression *sensu stricto* and the Buia graben.

II.4.13.1 Afar-Danakil transition zone

The *Afar-Danakil transition zone* refers in this study to the area situated between 12.8°N and 13.4°N (fig. 1). This zone marks important changes in morphology, magmatism, tectonics and sedimentology between central Afar south of it and the Danakil Depression *sensu stricto* situated north of it (see also section 7.1). This area is characterized by vast zones covered by Recent (< ~0.6 Ma) lavas of the Ma'Alalta, Alayta and Tat'Ali massifs. Older lavas of unknown age are present on Afdera volcano. Barberi *et al.* (1975) published an age of 3.1 Ma at the

foot of the volcano, but this age is questionable. Varet *et al.* (1975) map this volcano as belonging to the upper stratoid series, corresponding to our Q_{2v} unit. Tectonically, the area is dominated by the Alayta – Afdera Transfer Zone, which links the Dabbahu – Manda Hararo rift segment to the Tat’Ali and Erta Ale rift segments (Schaegis *et al.*, 2021). This transfer zone similarly affects Early Pleistocene to Recent lavas.

The Teru Graben and the Ma’Alalta volcano form the northern prolongation of the Dabbahu-Manda Hararo range. The Recent (<0.6 Ma, Barberi *et al.*, 1972a; Tortelli *et al.*, 2021) lavas of the Ma’Alalta volcano directly cover Mesozoic sediments towards the E, indicating a possibly relatively recent age for the formation of this rift segment. Local seismicity and the very recent filling of the Teru graben, which is possibly cut by active faults (12.827N, 40.317E), indicate that this area is still active.

II.4.13.2 Danakil Depression sensu stricto

The Danakil Depression *sensu stricto* comprises the topographic depression between ~13.3°N and 14.65°N (fig. 1). Its geological history remains debated. Magmatic material at the eastern margins is dated by Brinckmann & Kürsten (1971). However, dating results have large uncertainties (sometimes 200% error ranges). Subvolcanic samples from plugs and dykes at the western edge of the Danakil Alps showed results ranging from ~26 to ~16 Ma, probably corresponding to the same magmatic events studied in the southern Danakil Alps by Civetta *et al.* (1975a) (section 4.6). The rare Syenite and Dolerite sills mapped by the Geological Survey of

Ethiopia (1978a) on the western margin probably have similar ages.

The basin itself is filled by the syn-rift Red Bed Series (also called Danakil Fm. by Brinckmann & Kürsten, 1971 or Polychromatic Fm. by C.N.R - C.N.R.S. Afar Team, 1973 and Varet, 2018) and unconformably overlay the basement and the Mesozoic cover (Brinckmann & Kürsten, 1971; Beyth, 1973, 1978; Le Gall *et al.*, 2018). This unit is formed by alluvial and lacustrine sediments dominated by conglomerates, sandstones and siltstones with some basaltic intercalations (Brinckmann & Kürsten, 1971; Beyth, 1973, 1978; Le Gall *et al.*, 2018). On the eastern margin of the Depression, Brinckmann & Kürsten (1971) dated lava flows embedded within the Red Bed Series. The datings with reasonable uncertainty range from 8.5 to 6 Ma, with an exception being 3.4 Ma. These values are comparable to the datings on the western margin ranging from 5.7 to 6.2 Ma (Brinckmann & Kürsten, 1971; Le Gall *et al.*, 2018) and to Early Pliocene datings reported by (Park, 2010) on the eastern margin.

Continental sedimentation continued until present-day times. Continental conditions were interrupted by Middle to Late Pleistocene marine incursions identified as the Zariga Fm by Brinckmann & Kürsten (1971) and composed of fringing corallgal reefs covered by gypsum. At least four marine incursions were observed based on field observations, of which the two most recent ones were dated as MIS 5e and MIS 7 (i.e. ~0.125 and ~0.2 Ma, respectively, Jaramillo-Vogel *et al.*, 2019, Foubert *et al.*, in prep., Negga *et al.*, in prep., chapter III, this volume). Desiccation after the marine incursions resulted in the deposition of thick

evaporite units (Holwerda & Hutchinson, 1968; Bastow et al., 2018, chapter IV, this volume) recycled by continental water forming a large salt pan that is still active today. The sediment thickness below the salt plain is probably more than 2 km (Behle et al., 1975; Makris & Ginzburg, 1987).

The Erta Ale volcanic range is situated on the rift axis in the southern Danakil Depression. It is formed by Recent lava and is still active nowadays (Burgi et al., 2002; Pagli et al., 2012; Xu et al., 2017, 2020; Global Volcanism Program, 2021). Barberi *et al.* (1972a) propose two datings of the lava deposits, resp. 0.35 ± 0.14 and 1.2 ± 0.5 Ma. However, as mentioned by Watts *et al.* (2020), it is challenging to date very young material, and most authors consider the Erta Ale lavas as Recent, i.e. probably younger than 0.6 Ma.

Rifting in the Danakil Depression has traditionally been postulated to initiate in the Early Miocene (~23 - 25 Ma), such as proposed by Barberi *et al.* (1972a), C.N.R - C.N.R.S. Afar Team (1973) and Le Gall *et al.* (2018) based on datings of magmatic material published by Brinckmann & Kürsten (1971). However, we challenge this hypothesis. As discussed above, all Early to Mid-Miocene datings were sampled in plugs and dykes of the Danakil Alps and not in the syn-rift Red Bed Series as described by Brinckmann & Kürsten (1971) and shown on their maps (Brinckmann & Kürsten, 1970). Only one single dating (sample 438, 22 ± 2 Ma, presumably taken at 13.802N, 40.984E according to their geological map) cannot unambiguously be attributed to either the Red Bed Series or pre-rift rocks. The presence of igneous products in pre-rift rocks is not diagnostic for the formation of the Danakil basin. As

shown in other parts of the Afar Depression, volcanic activity is not always linked with tectonic activity. If the plugs and dykes are linked to a tectonic event, we suggest relating them to the opening of the Red Sea basin occurring around the same time (see sections 4.3.1 and 4.6 above). The deposition of the Red Bed Series, however, is diagnostic for the formation of an incipient Danakil basin (Brinckmann & Kürsten, 1971; Le Gall et al., 2018; Varet, 2018). As discussed above, all datings unambiguously attributed to the Red Bed Series are 8.5 Ma or younger. This younger age is also much more consistent with the kinematic models of Eagles et al. (2002), McClusky et al. (2010) and Reilinger & McClusky (2011) that determined the initiation of the Danakil Block rotation at ~11 Ma (see section 4.6 above) and with the 9 to 3 Ma heating event recorded in fission tracks on Mt Ghedem (Ghebreab *et al.*, 2002). We thus conclude that the rifting of the Danakil did not start before the Mid to Late Miocene, although we cannot exclude prior faulting activity linked with the opening of the southern Red Sea. The late onset of tectonic activity in the Danakil Depression could also explain the unusual N-S fault orientation observed in the Garsat Graben and east of the Mekele Outlier (fig. 4). As shown by Zwaan *et al.* (2020b), the rotation of the Danakil Block changed the stress field from NE-SW to E-W on the western Afar margin. In the southern part of this margin, the NNW-SSE faults were formed by initial stress fields, while near the Danakil Depression, the major faults formed later under the E-W stress field, forming N-S faults.

The occurrence of the Red Bed Series decreases towards the northern part of

the Danakil Depression. On the western margin, no Red Bed sediments are found north of 14.24°N. They are visible further on the eastern margin but not north of 14.50°N. This, along with the change in topography and the larger tectonic uplift in the north, has been used as an argument for the younger age of the rift in the north (chapter III, this volume, see also section 5.2.1), but the absence of dated material prevents any quantification.

II.4.13.3 Buia Graben

The Buia graben is situated between the Danakil Depression and the Gulf of Zula (fig. 1a). We only discuss the opening of the graben itself, representing the northern extension of the Danakil rift. This area experienced denudation since ~15 Ma, which is not linked to the opening of the Danakil Depression but to the opening of the Red Sea discussed in section 4.3.1 (Balestrieri *et al.*, 2005).

The Buia graben is not filled by a large amount of syn-rift sediments or lavas, as testified by the basement outcropping inside the crater of the Alid volcano, on the eastern fault scarp (Duffield *et al.*, 1997) and on the hills separating the Buia graben and the Danakil Depression (see fig. 4). The faulted stratoid lavas found on the shoulders of the graben were dated between 2.1 and 1.1 Ma (Duffield *et al.*, 1997; Buffler *et al.*, 2010). Recent (< 0.5 Ma) lavas were also reported by Buffler *et al.* (2010) in the north of the graben. The central Alid volcano erupted between ~0.212 and 0.015 Ma, and the basaltic flows at the foot of the volcano are believed to be younger (Duffield *et al.*, 1997; Lowenstern *et al.*, 2006).

The SW part of the graben forms the Dandiero basin and features

approximately 500m of Pleistocene sediments studied in detail after the discovery of Hominid fossils (Abbate *et al.*, 2004; Ghinassi *et al.*, 2009, 2014; Papini *et al.*, 2014; Sani *et al.*, 2017). These continental sediments are Pleistocene in age.

The Buia graben started forming at ~1.2 Ma (Sani *et al.*, 2017) with the first phase of N-S faulting followed by NNW-SSE faulting. The graben is nowadays tectonically active as recorded by coral terrace elevations (Buffler *et al.*, 2010; chapter III, this volume), geodetic velocities (Viltres *et al.*, 2020), earthquake swarms and a likely dike intrusion in Bada in 1993 (Ruch *et al.*, 2021).

II.5 TECTONIC STRUCTURE OF THE AFAR REGION

In light of the complex geological history of the Afar region (section 4) and studies on present-day tectonic activity, the tectonic structures and kinematic of that area are discussed in this section.

II.5.1 Plates, microplates, blocks and fragments

Three main tectonic plates play a major role in the region: the *Nubian plate*, the *Arabian plate* and the *Somalian plate* (fig. 2, 8). Several smaller tectonic elements of varying extent and importance are also present. The most important are the *Danakil Block* and the *Ali Sabieh Block*, besides some smaller, minor crustal elements.

II.5.1.1 The Danakil Block

Most authors classify the Danakil Block as a microplate (e.g. Sichler, 1980; Acton *et al.*, 1991; Eagles *et al.*, 2002;

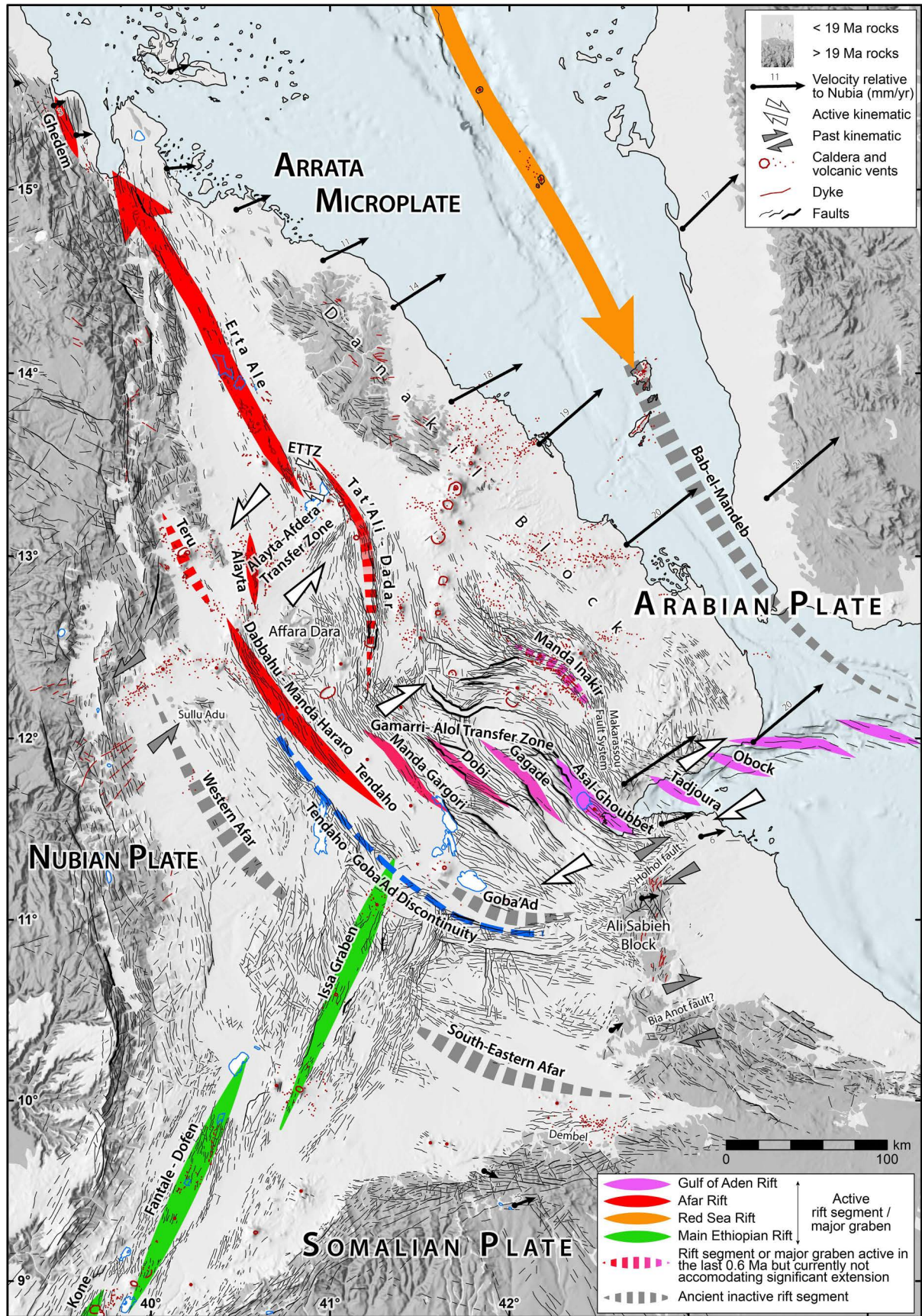


Figure 8: Rift dynamics of the Afar Depression. Rift segments are discussed in section 5.2. After Corti (2009), Pagli et al. (2014), Doubre et al. (2017), Viltres et al. (2020), Moore et al. (2021), Ruch et al. (2021), Sangha et al. (2022) and discussion in section 5.2. ETTZ = Erta Ale – Tat'Ali Transfer Zone.

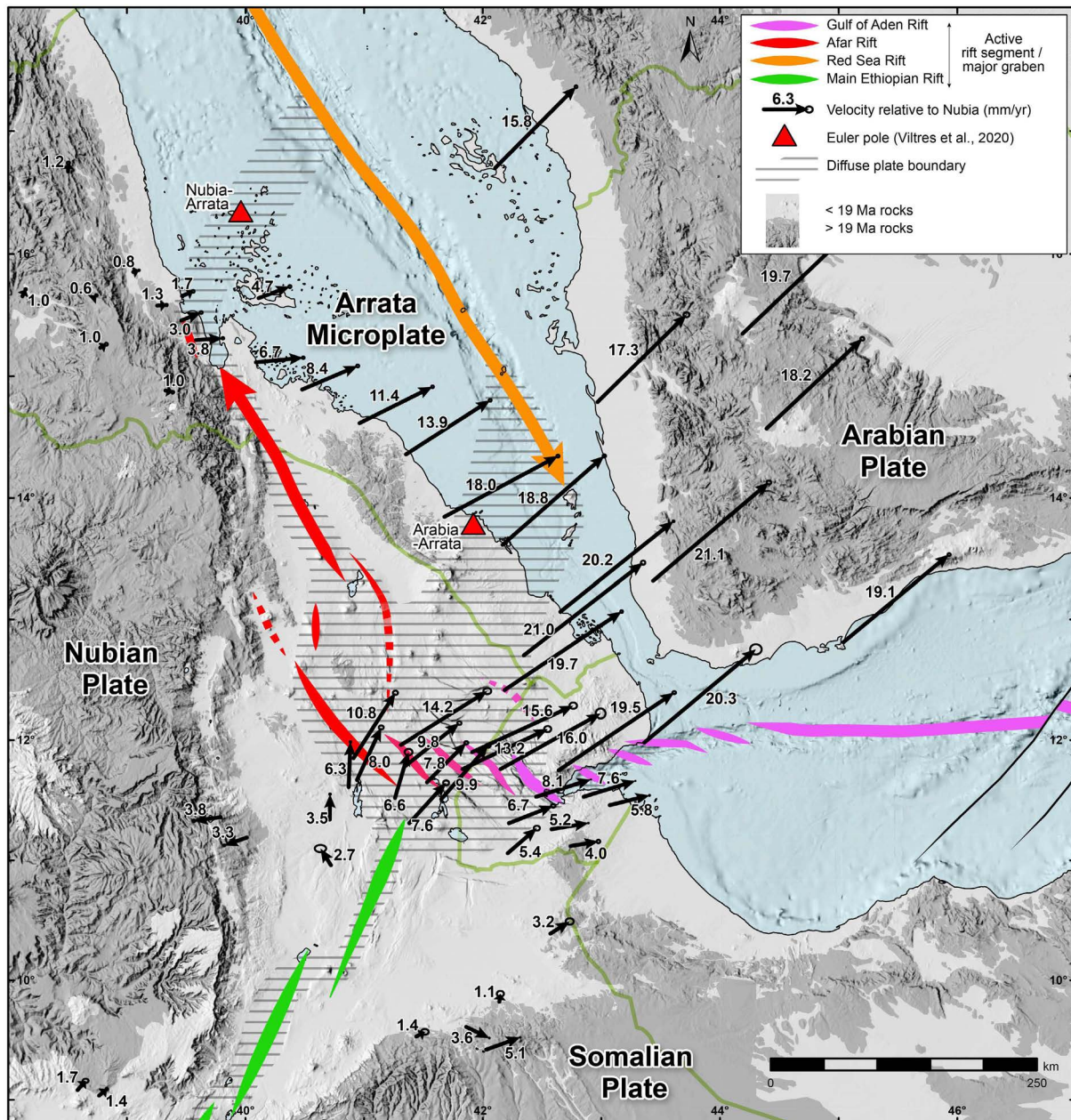


Figure 9: Plate kinematics of the Afar region. GNSS velocities are from McClusky et al. (2010), Doubre et al. (2017) and Viltres et al. (2020). Only datapoints reflecting inter-dyking periods were used in the data of Doubre *et al.* (2017). Rift segments and diffuse plate boundary are discussed in section 5.

McClusky et al., 2010; Keir et al., 2013; Schettino et al., 2016). The counter-clockwise rotation of this Block has been postulated since the early 70s based on geological evidences such as fault orientation, paleo stress analysis, comparison between Ethiopian plateau and Danakil Block geology, the distribution of young magmatic material and early plate tectonic reconstructions

(Mohr, 1970, 1972; Tazieff et al., 1972; Sichler, 1980; Souriot & Brun, 1992; Zwaan et al., 2020b). Paleomagnetic (Schult, 1974; Manighetti et al., 2001) and InSAR studies (Moore *et al.*, 2021), as well as GNSS data (ArRajehi *et al.*, 2010; McClusky *et al.*, 2010; Viltres *et al.*, 2020, see fig. 9) confirmed this hypothesis. However, the assumption of one single rigid Danakil Block is an

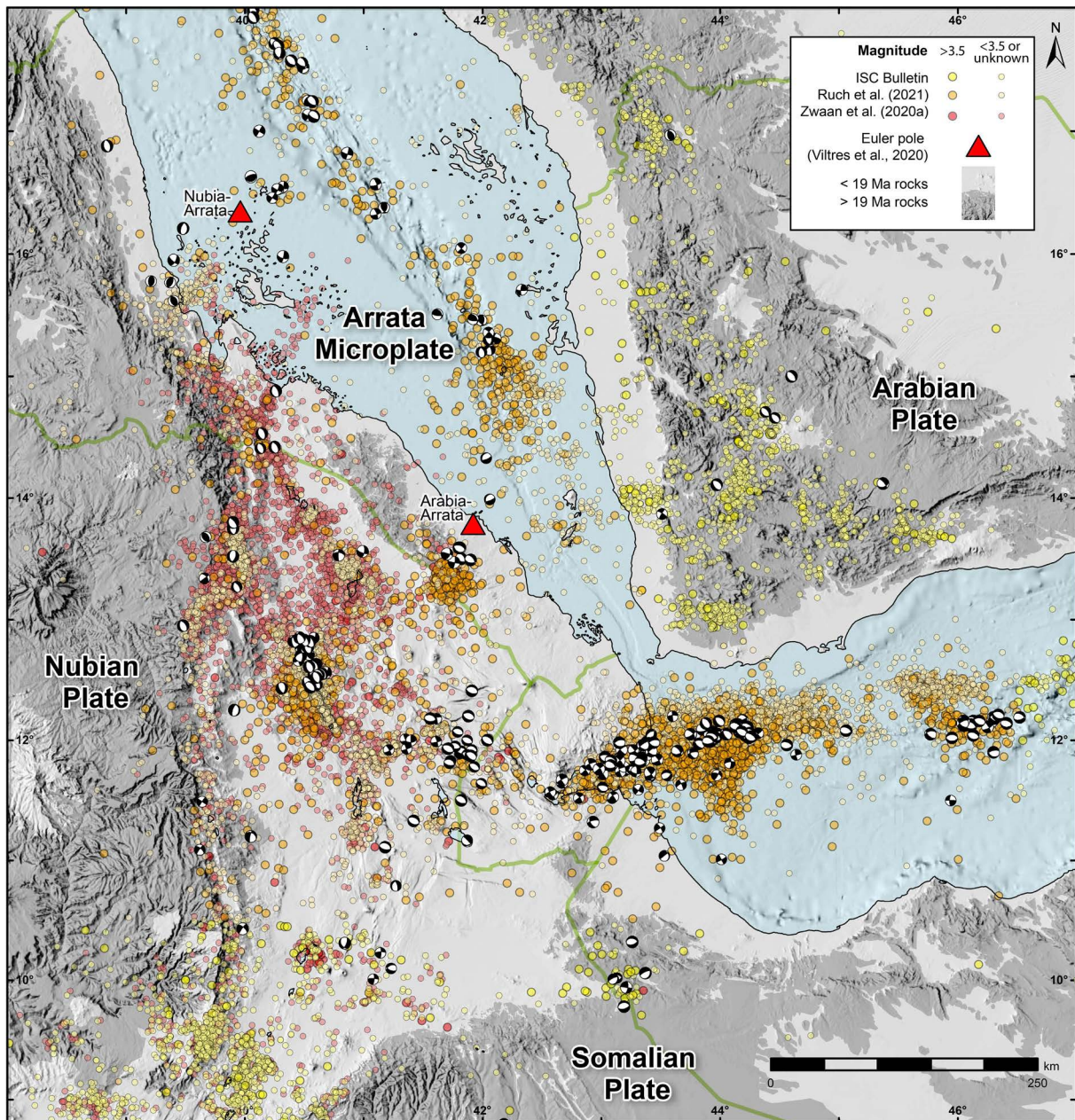


Figure 10: Earthquake map of the Afar region. Earthquakes from Zwaan et al. (2020a), Ruch et al. (2021) and ISC Bulletin (Willemann & Storchak, 2001; Bondár & Storchak, 2011; Storchak et al., 2017, 2020; International Seismological Centre, 2021, retrieved on 22.09.2021). Focal mechanisms are from Hofstetter & Beyth (2003) and Global CMT Catalogue (Dziewonski *et al.*, 1981; Ekström *et al.*, 2012, retrieved on 27.11.2021).

oversimplification. Barberi & Varet (1975, 1977), Barberi *et al.* (1975) and Varet (2018) already mentioned that the southern Danakil Block “has to be considered as an accretion of the Arabian plate” that behaves independently from the northern Danakil Block, which has been named as the *Arrata Microplate*. This configuration also explains the low seismicity (fig. 10) and tectonic

quiescence in the Bab-el-Mandeb Strait (section 4.3.1 and references therein). Recent GNSS data confirm this hypothesis (Viltres *et al.*, 2020). They show the absence of relative movement between the southern Danakil Block and Arabia – both being thus part of the same plate - and confirm that the Arrata Microplate is rotating rigidly and independently from both the Nubian and

Arabian plates (see fig. 9). It should, however, be noted that even if the southern Danakil Block nowadays represents an accretion to the Arabian plate, paleomagnetic (Manighetti *et al.*, 2001) and geological data (see above) indicates that this was not the case in the past when the whole Danakil Block rotated and rifted away from Arabia.

The south-eastern and north-western limits of the Arrata Microplate are poorly defined. As shown by Viltres *et al.* (2020), they are probably situated on the Euler poles of the Arabia-Arrata and Nubia-Arrata rotations, respectively (fig. 9). As such, they should only accommodate little strain to no strain at the Euler pole. The limit between the Arrata Microplate and the southern Danakil Block (part of the Arabian plate) is unclear and is probably situated between the Nabro range, the Hanish island and the Beilul Bay (Barberi & Varet, 1977; Varet, 2018; Viltres *et al.*, 2020, see fig. 9). Earthquake data show seismic activity with sinistral strike-slip focal mechanisms in the Nabro Range (fig. 10). The Hanish Islands show NE-SW fissures and vent alignment (Gass *et al.*, 1973; Viltres *et al.*, 2020, fig. 8). Free air anomaly and vertical gravity gradient data (suppl. mat. 5 and 6) also indicate that the Beilul Bay might represent the southernmost continuation of the Red Sea Rift. However, the onshore area of Beilul Bay is devoid of major fault sets.

The northern limit of the Arrata Microplate is similarly ill-defined and probably runs NE of Massawa (Viltres *et al.*, 2020). Ruch *et al.* (2021) show the occurrence of a WSW-ENE-oriented earthquake swarm in Massawa in 2002. The earthquake cluster can be followed until the Red Sea trough at 17°N where several possibly sinistral strike-slip focal

mechanisms have been recorded (see fig. 10). Compressive focal mechanisms recorded in the vicinity of Massawa (fig. 10) are coherent with a Euler pole situated on the plate boundary. Lineaments in free air anomaly and vertical gravity gradient data can also be observed in the same zone, albeit in a slightly different orientation (SSW-NNE, suppl. mat. 5 and 6). Further lineaments and earthquakes could even indicate that some of the strain is accommodated SE of the Dahlak Islands and the Howlakil Bay.

These elements point towards diffuse strain accommodation patterns for the NW and SE limits of the Arrata Microplate. As such, fig. 9 only presents diffuse plate boundary.

II.5.1.2 The Ali Sabieh Block

The continental Ali Sabieh Block significantly complicates regional plate tectonic reconstructions (Beydoun, 1970; Collet *et al.*, 2000; Purcell, 2018) and was ignored by some authors (Le Pichon & Francheteau, 1978; Redfield *et al.*, 2003).

While some authors consider the Ali Sabieh Block as autochthonous (Christiansen *et al.*, 1975; Kronberg *et al.*, 1975; Purcell, 2018), others inferred that it was part of the southern Danakil Block also rotating counterclockwise (Mohr, 1970; Sichler, 1980; Garfunkel & Beyth, 2006). Other studies proposed a “saloon door mechanism” where both the Danakil and Ali Sabieh Blocks were rotating in opposite directions (Manighetti *et al.*, 2001; Audin *et al.*, 2004). The latter hypothesis was confirmed by paleomagnetic data showing a 29° clockwise rotation for the Ali Sabieh Block between 55 Ma and 8 Ma (Kidane, 2016). While it was since then accreted to the Somalian plate, it should be

considered a distinct block during its geological history.

The southern extent of the Ali Sabieh Block and the mechanisms of movement are poorly defined. Most authors mention the *Bia Anot fault* as its southern limit (Black et al., 1975; Audin et al., 2004). According to the model of Audin *et al.* (2004), a significant amount (several tens of km) of dextral strike-slip movement would have been accommodated by this fault. The absence of major visible faults and the continuity of the basement and Mesozoic structure (fig. 4 and suppl. mat. 7) argues against the presence of both faults indicated by Black et al. (1975) and Audin et al. (2004) (see suppl. mat. 7). We identified a major fault SE of Bia Anot (eastern Bia Anot fault on suppl. mat. 7) that might have accommodated strike-slip movement but that has an orientation (ENE-WSW) that is not strictly parallel to the strike-slip faults of the Gulf of Aden. Another major fault (probably buried beneath younger volcanic rocks) is indicated by the Ayisha geological map (Geological Survey of Ethiopia, 2015), SW of Bia Anot (western Bia Anot fault on suppl. mat. 7). This lineament also marks a major boundary between rocks of different ages and structure orientation (WNW-ESE faulted Pliocene and younger in the West, NNE-SSW faulted and tilted Mid-Miocene to pre-rift in the East, fig. 4). North of Bia Anot, this fault might connect with a WNW-ESE oriented zone at $\sim 10.6^\circ\text{N}$ where no pre-rift rock crops out (northern Bia Anot discontinuity on suppl. mat. 7). This northern Bia Anot discontinuity also coincides with a significant change in the orientation of the structures in the pre-rift rocks from WNW-ESE south of it and NNE-SSW on the Ali Sabieh Block. It might represent the southern limit of the

Ali Sabieh Block. Even though we used the eastern Bia Anot fault in our kinematic model (section 6.1), our work cannot bring a clear conclusion, and further field and paleomagnetic studies would be needed to understand the complex tectonic history of the area.

II.5.1.3 Crustal fragments

Several other continental crustal fragments have been proposed in the Afar Depression, and probably more are present but not reported. The Affara Dara, exposing Trap basalts intruded by a ~ 25 Ma granitic body (Barberi *et al.*, 1972a; C.N.R - C.N.R.S. Afar Team, 1973; Varet, 2018, fig. 4) represents a clear continental fragment. Hammond *et al.* (2011) also shows lower crustal Vp/Vs seismic velocities in the area, confirming the presence of a crustal body. The northern Sullu Adu region also exposes early Oligocene flood basalts and similarly represents a continental fragment. Based on geophysics, Hammond *et al.* (2011) proposed another continental fragment extending further south, near Mile. The presence of this block is however debated (see Wang *et al.*, 2021). Hammond *et al.* (2011) further hypothesize a possible block in central-eastern Afar (roughly centred on Siyyarou).

II.5.2 Rifts, rift segments and rift propagation

We discuss here the *rifts* and *rift segments* of the Afar region. We define *rifts* as strain-accommodating areas separating rigid and extensive blocks, i.e. tectonic plates or microplates.

At a smaller scale, rifts are segmented into *rift segments* (e.g. Nelson *et al.*, 1992). They represent continuous, mainly

linear strain accommodating areas testified by faulting and, in some cases, magmatic activity and/or formation of sedimentary basin. They are the equivalent of the distinct Wonji Fault Belts in the MER.

II.5.2.1 Re-evaluation of rift propagation in the Afar

The literature usually mentions the presence of three rifts in the Afar Depression: the Gulf of Aden Rift, the (Northern) Main Ethiopian Rift ((N)MER), which is part of the East African Rift System, and the Red Sea Rift. They separate the Arabian, Nubian and Somalian plates. These rifts are often described as “propagators” considering the direction of propagation. We call for a re-evaluation of rift propagation in the Afar Depression at a regional scale. For this discussion, we define rift propagation as the along-strike migration of the part of a rift that experienced a certain amount of lateral extension.

The Northern Main Ethiopian Rift (NMER) accommodates extension between the Nubian and Somalian plates. At the scale of the East African Rift System, rifting is propagating southwards (Chorowicz, 2005; Macgregor, 2015), consistent with a Euler pole situated far south (DeMets & Merkouriev, 2016; Stamps et al., 2021) but a northward propagation from the Turkana Depression has also been proposed (Michon *et al.*, 2022). At the scale of the study area, while Tesfaye *et al.* (2003) and Wolfenden *et al.* (2004) propose a northward propagation from the Central MER, most authors suggest a southward propagation from the Afar Depression towards Central MER (Bonini et al., 2005; Rooney et al., 2007; Keranen & Klemperer, 2008; Abebe et al.,

2010). The long distance to the Euler pole implies that, at the scale of our study area, the Nubia-Somalia extension can be considered almost parallel, and rift propagation mechanisms are thus less significant.

It is well-defined that the Gulf of Aden propagated westwards into the Afar Depression. This was shown by the magnetic anomalies in the Gulf of Aden sensu stricto (Fournier *et al.*, 2010; Leroy *et al.*, 2012) and by the presence of a young rift segment in the Gulf of Tadjoura towards Asal-Ghoubbet (Courtillet, 1980; Manighetti et al., 1997, 1998, 2001; Daoud et al., 2011, see section 4.2). This westward propagation is coherent with the anticlockwise rotation of the Arabian plate.

What traditionally is called the “Red Sea rift” is characterized by two distinct rift arms (see fig. 8, 9). One is situated “offshore”, in the Red Sea trough, possibly reaching the Hanish islands as far south as 14°N. The other one comprises the Danakil Depression and central Afar.

The “offshore” Red Sea rift is thought to propagate northward in the central and northern part of the Red Sea (Courtillet, 1982; Courtillet et al., 1987; Izzeldin, 1987; Ghebreab, 1998; Augustin et al., 2016; Schettino et al., 2016; Khalil et al., 2020, see also section 4.3). Conversely, it propagates southward in its southern part (south of approx. 17°N, Courtillet, 1982; Courtillet et al., 1987; Ghebreab, 1998; Schettino et al., 2016; Khalil et al., 2020). The northward propagation from ~17°N is linked to the counter-clockwise rotation of the Arabian plate relative to the Nubian one, and the southward propagation linked to the counter-clockwise rotation of the Danakil

Block and Arrata Microplate with respect to the Arabian plate.

The propagation of the second arm of the Red Sea rift, comprising the Danakil Depression and central Afar, is more controversial. Tazieff et al. (1972) and Barberi & Varet (1975, 1977) proposed decreased spreading rates, crustal separation and opening, as well as a younger ages towards the northern part of the Afar, therefore implying a northward propagation. Since the late 90s', however, rifting in the Afar has mostly been described as a southward propagation of the Red Sea Rift (the "Red Sea propagator"). This concept was first mentioned and published by Manighetti *et al.* (1998). However, the study focused on the Gulf of Aden rift and the authors probably assumed the same concept of inland (centipede) propagation to be true for the other side of the Afar Depression or they extrapolated the southward propagation of the Red Sea Rift (*sensu stricto*) to the Afar. Since its first publication, this hypothesis has been mentioned by many authors (Manighetti et al., 2001; Hofstetter & Beyth, 2003; Kidane et al., 2003; Lahitte et al., 2003b; a; Audin et al., 2004; Wolfenden et al., 2004; Beyene & Abdelsalam, 2005; Rowland et al., 2007; Acocella et al., 2008; Bridges et al., 2012; Muluneh et al., 2013, 2017; Reed et al., 2014; Dumont et al., 2017, 2019; Purcell, 2018; Mologni et al., 2021; Wang et al., 2021; Chauvet et al., 2023). Only few publications consider a northward rift propagation (Gaulier & Huchon, 1991; Collet et al., 2000; Eagles et al., 2002; Ghebreab et al., 2002; Sani et al., 2017; Molnar et al., 2018). The controversy on the northward versus southward rift propagation in the Danakil Depression and central Afar has however never been discussed in detail.

The following facts argue for northward rift propagation in the Danakil Depression and central Afar:

- 1) The counter-clockwise rotation of the Danakil Block postulated for a long time (Mohr, 1970; Tazieff et al., 1972; Sichler, 1980; Souriot & Brun, 1992), is now proven by GNSS velocity measurements, InSAR and paleomagnetic studies, as well as geological evidences (see section 5.1.1 and references therein). Rotation implies rift propagation towards the Euler Pole, i.e. towards the north (see suppl. mat. 16). This geometrical observation is confirmed by numerous analogue and numerical modelling studies of rotational rifting and field analogues (Benes & Scott, 1996; Molnar et al., 2017, 2018, 2020; Mondy et al., 2018; Pourhiet et al., 2018; Zwaan et al., 2019; Khalil et al., 2020; Maestrelli et al., 2020; Zwaan & Schreurs, 2020; Schmid et al., 2022).
- 2) Overall northward younging of the main outcropping magmatic products is observed in the Afar Depression (fig. 4). While the southern Afar is dominated by Pliocene magmatic products, they are mainly Early- to Mid-Pleistocene in central-eastern Afar and mainly Recent (<0.6 Ma) in the Danakil Depression, with the exception of the thin Early Pleistocene basalts east of the Buia Graben.
- 3) The northward younging of the rifting age is noticeable (see section 4, fig. 7). While rifting is reported since 27 Ma in central-western and south-western Afar (Wolfenden et al., 2005; Stab et al., 2016), rifting started probably ~11 Ma ago in the Danakil basin

(section 4.13.2) and at 1.2 Ma ago in the Buia graben (Sani *et al.*, 2017).

- 4) At a smaller scale, this trend of rifting with ages becoming younger towards the N is also observed from Goba'Ad (Pliocene) to Tendaho (~1.1 to 0.6 Ma, Lahitte *et al.*, 2003b) to Manda Hararo (~0.2 Ma, Acocella *et al.*, 2008). Dumont *et al.* (2019) furthermore shows a northward propagation of faulting from Manda Hararo to Dabbahu at ~0.1 Ma.
- 5) The topography corroborates this northward younging trend (suppl. mat. 8). The topography in the south is wider, smoother and less steep, while in the north, it is steeper, indicating a less mature topography. This trend is also noticeable at the scale of the Danakil depression (chapter III, this volume).
- 6) The presence of pre-rift rocks shows a N-S gradient in the Danakil Depression. Basement crops out on the Buri peninsula, the eastern coast of the Gulf of Zula, the Howlakil Bay, and even the rift axis, on the Alid volcano (fig. 4, Duffield *et al.*, 1997). Volcanic material from the axial Maraho volcano in the northern part of the Depression contains xenoliths of the basement and Mesozoic cover (Kürsten, 1975), indicating the deep presence of pre-rift rocks. Further south, the magma of the Erta Ale range does not show any crustal contamination, probably indicating the absence of basement (Barberi & Varet, 1970; Barberi *et al.*, 1970; Barrat *et al.*, 1998; Hagos *et al.*, 2016a; Watts *et al.*, 2020). This evidences an overall crustal separation

gradient, increasing from north to south.

Overall, we consider that rifting in central Afar and Danakil Depression is not propagating southward but northward. The northward propagation implies that the northern Afar has experienced less stretching and lower extension rates than the central Afar (see also sections 6.1 and 7.1). It also implies that the same stage of extension happens later towards the Euler pole, i.e. northwards. Rifting in the Afar Depression would thus represent a large-scale en-echelon relay structure of the Red Sea, with the Danakil Block (nowadays only the Arrata Microplate) rotating between two rifts propagating in opposite directions.

This northward rift propagation calls for reconsidering the “Red Sea Rift” nomenclature referring to rifting in central Afar and the Danakil Depression. Northward propagation implies that the Red Sea Rift did not propagate southward from the Red Sea central trough through the Gulf of Zula into the Danakil Depression, as often described. The southern Red Sea Rift (the offshore arm of the Red Sea Rift) and the described Rift in central Afar/Danakil Depression (interpreted previously as the onshore segment of the Red Sea Rift) mark the SW and NE limit of the rigid (Viltres *et al.*, 2020) Arrata Microplate. Both rifts overlap for > 300 km and are > 100 km apart. Of course, both arms have the same overall orientation and separate the Nubian and Arabian tectonic plates at a global scale. However, at a regional scale, if the Arrata/Danakil Block is considered a microplate, then both rifts (as defined above) should be considered distinct. We suggest calling it the *Afar Rift*. Barberi *et al.* (1972a) already named

this segment the “*Afar Rift*” and “*Danakil Rift*”. Viltres *et al.* (2020) also mention the “*Danakil-Afar rift*” and Le Pichon & Francheteau (1978) the “*Danakil Rift*”. We prefer the “*Afar Rift*” because it encompasses both the Danakil Depression (northern Afar) and central Afar. The authors acknowledge that this distinction is not necessarily useful for global plate tectonic models but valuable at a regional scale to avoid confusion among scientists.

As shown in fig. 8, the Gulf of Aden Rift and the Afar Rift are linked in central-eastern Afar, gradually accommodating extension in a series of en-echelon grabens. The formal distinction between the Gulf of Aden Rift and the Afar Rift at a regional scale is that the former accommodates extension between the Somalian and the Arabian plates while the latter accommodates extension between the Nubian and the Arrata plates. As no precise localization of the MER can be identified north of the Tendaho – Goba’Ad Discontinuity (TGD) and the limit between the Arrata and southern Danakil Block is gradual and ill-defined (see section 5.1.1), the boundary between what belongs to the Afar Rift and the Gulf of Aden Rift is also vague. It highlights the fact that the Gulf of Aden and Afar Rifts are forming a single, near-continuous structure (fig. 8), propagating in the same overall direction. This continuity led Eagles *et al.* (2002) to consider the whole structure (the Afar Depression, including the Danakil Depression) as part of the Gulf of Aden rift.

We conclude that the Afar region is structured by the westward propagation of the Gulf of Aden Rift, the northward propagation of the Afar Rift, the southward propagation of the Red Sea

Rift and close to parallel extension of the Main Ethiopian Rift (fig. 8). This conforms to the counter-clockwise rotation of both the Arabian and Danakil plates. This rift propagation pattern and its associated timing (section 4) confirm the conclusion proposed by Mohr (1970, 1972), Girdler (1991), Manighetti *et al.* (1997), Hébert *et al.* (2001), Wolfenden *et al.* (2004), Purcell (2018), Khalil *et al.* (2020) and Maestrelli *et al.* (2022) that the triple junction models showing three rifts propagating synchronously away from one hotspot does not apply to the Afar region, just as it does not apply to the south Atlantic and the Etendeka and Paraná flood basalts or to the north Atlantic and the Iceland hotspot (Franke, 2013; Foulger *et al.*, 2020; Peace *et al.*, 2020).

II.5.2.2 Active rift segments

The position and configuration of rifts and rift segments in the Afar region varied significantly throughout its history. The most critical change corresponds to the abandonment of the Bab-el-Mandeb rift segment after Mid to Late Miocene times (see section 4.3.1). An extinct rift segment was also active between the Late Miocene and Pliocene in present-day Western and South-Eastern Afar (see sections 4.10, 4.11 and fig. 7). As discussed in section 4.12, the Goba’Ad graben was probably active in the Pliocene and Pleistocene and represents another extinct rift segment.

At a shorter temporal scale, transient rifting processes, where individual rift segments accumulate strain for protracted periods before release during short volcanotectonic events, are known in the Afar Depression, such as in Iceland and in other rifts (Foulger *et al.*, 1992, 2020; Bourgeois *et al.*, 2005; LaFemina *et*

al., 2005; Sigmundsson, 2006; Wright et al., 2006, 2012; Vigny et al., 2007; Grandin et al., 2010; Ebinger et al., 2013; Ruch et al., 2016; Smittarello et al., 2016). For example, Sangha *et al.* (2022) show that the distribution of extension changed before and after the 2005-2010 volcano-tectonic event in Dabbahu-Manda Hararo, and suggest that “surface deformation patterns in Afar changed drastically over periods of less than a million years”. Another example is Manda Inakir which experienced significant Recent volcanic and tectonic activity as shown by the dense fault network in Recent lavas (fig. 4) and by the 1928-1929 tectono-magmatic crisis (Audin *et al.*, 1990). However, GNSS, InSAR and seismicity data show that the Manda Inakir currently hardly accommodates any strain (Pagli et al., 2014, 2019; Doubre et al., 2017; Moore et al., 2021; Ruch et al., 2021; Sangha et al., 2022, see also fig. 9 and 10). This shows that the strain accommodation pattern in the Afar Depression can change over very short time scales.

The following sections discuss the present-day active rift segments. It is mainly based on very short observations of movement by GNSS, InSAR or seismicity methods and are not necessarily representative of the long-term evolution of the area.

II.5.2.2.1 *The Northern Main Ethiopian Rift*

In the studied area, the Northern Main Ethiopian Rift comprises 3 Wonji Fault Belt (WFB) segments: the northernmost part of the Kone segment, the Fantale segment and the Issa Graben. The latter is technically not situated in the NMER but in southern Afar. Nevertheless, as it accommodates

strain between the Nubian and Somalian plates, it is considered part of the MER here. The MER rift ends northward abruptly on the Tendaho – Goba’Ad Discontinuity (TDG). Structures north of this discontinuity are related to the Gulf of Aden and Afar rifts (Hayward & Ebinger, 1996; Tesfaye et al., 2003, fig 4, 10).

II.5.2.2.2 *The Gulf of Aden Rift*

In the Afar Depression, the Gulf of Aden rift is commonly described as comprising the Gulf of Tadjoura, the Asal-Ghoubbet graben and the Manda-Inakir segments, linked by the Makarassou fault system (e.g. Tapponnier & Varet, 1974; Manighetti et al., 1998, 2001; Le Gall et al., 2011, see fig. 8). Two rift segments are identified in the southwestern Gulf of Tadjoura, the Obock and Tadjoura Troughs, linked by the Maskali fault zone (Dauteuil *et al.*, 2001). The Asal-Ghoubbet rift segment witnessed a major rifting episode in 1978 (Abdallah et al., 1979; Kasser et al., 1981; Vigny et al., 2007; Smittarello et al., 2016; Ruch et al., 2021). GNSS and InSAR data indicate that this segment accommodates a significant part of the strain between the Arabian and Somalian/Nubian plates on a narrow zone with approx. 12-14 mm/yr of extension (Pagli et al., 2014; Doubre et al., 2017; Sangha, 2021; Sangha et al., 2022, see also fig. 9), while Moore *et al.* (2021) indicate a more distributed extension.

The Manda Inakir rift segment is traditionally mentioned as an active rift segment linked to the Gulf of Aden rift (Tapponnier & Varet, 1974; Barberi & Varet, 1975, 1977; Manighetti et al., 1998; Lahitte et al., 2003b; Varet, 2018). As discussed above, it does not currently

accommodate measurable extension although it probably did in the last century.

II.5.2.2.3 The link between the Gulf of Aden Rift and the Afar Rift

As discussed in section 5.2.1, the limit between the Gulf of Aden Rift and the Afar Rift is vague. Traditionally, only the Asal-Ghoubbet and Manda Inakir segments are considered part of the Gulf of Aden Rift. However, considering the position of the MER, one could argue that the Gagade and Dobi grabens also belong to the Gulf of Aden Rift. As a formal limit is not useful, the transition area is considered here separately.

Strain in central-eastern Afar is distributed over more than 100 km (Kogan et al., 2012; Pagli et al., 2014; Doubre et al., 2017; Moore et al., 2021; Sangha et al., 2022). Data from Pagli et al. (2014), Doubre et al. (2017), Sangha (2021) and Sangha et al. (2022) show that most of the strain is currently accommodated in a zone going from Assal to Gagade, Dobi, Manda Gargor, Manda Hararo and Dabbahu, forming an en échelon graben pattern, similar to what is observed in the Gulf of Tadjoura (see fig. 8 and 9). It furthermore represents an area with intense faulting (fig. 4); the highest extensional strain (Polun *et al.*, 2018; Sangha, 2021); high fault slip rates (2-3 mm/yr) during the Holocene and Quaternary (Manighetti *et al.*, 2001); the most significant seismic activity (fig. 10); the highest cumulative seismic moment release during the 1950-2017 period (La Rosa, 2020) and earthquake swarms recorded between 1960 and 2017, indicating magmatic processes at depth (Ruch *et al.*, 2021). This broad area linking the Gulf of Aden Rift and the Afar Rift is named the Gamarri-Alol Transfer

Zone. This complex transfer zone has been interpreted as bookshelf faulting, explaining the clockwise rotation of the Stratoid lavas of central-eastern Afar (Courtillet et al., 1984; Tapponnier et al., 1990; Acton et al., 1991, 2000; Manighetti et al., 2001; Kidane et al., 2003). Conversely, Pagli *et al.* (2019) argue for a distributed extension and rift-perpendicular shearing model.

II.5.2.2.4 The Afar Rift

The Dabbahu – Manda Hararo rift segment constitutes the southernmost rift segment that is indisputably part of the Afar Rift. It currently accommodates a significant amount of strain. This area shows low crustal contamination, high partial melting, shallow Moho, linear magnetic anomalies, an extensive and long-lasting magma reservoir and is believed to be heavily intruded, representing an almost oceanic crust, close to continental breakup (Ayele et al., 2009; Hammond et al., 2011; Bridges et al., 2012; Desissa et al., 2013; Didana et al., 2014; Reed et al., 2014; Johnson et al., 2016; Lewi et al., 2016; Ayalew et al., 2019; Ahmed et al., 2022). Towards the southeast, the Dabbahu-Manda Hararo rift segment is prolonged by the Tendaho graben. The Tendaho Graben accommodates approximately 6 mm/yr of extension, mainly along its northern fault (Gamari Fault) and in the Manda Gargori graben (Doubre *et al.*, 2017, see also fig. 9). This area witnesses strong seismic activity (Ruch *et al.*, 2021) and fig. 10), as shown by the deadly 1969 Serdo earthquake (Gouin, 1979).

The Teru Graben and the Ma'Alalta volcano represents the northernmost prolongation of the Dabbahu – Manda Hararo rift segment and is considered to be an active rift segment (Tortelli *et al.*,

2021). The data of Moore *et al.* (2021) show that this segment is probably not accommodating significant extension and that most of the strain is transferred from the Dabbahu – Manda Hararo rift segment to the Tat'Ali – Dadar and Erta Ale rift segment by the Alayta – Afdera Transfer Zone (Barberi *et al.*, 1974; Barberi & Varet, 1977; Schaegis *et al.*, 2021).

The Tat'Ali – Dadar forms a rift segment with active faulting (Sangha *et al.*, 2022), Recent acidic and basaltic volcanism but most of the strain seems to be currently accommodated by the Dabbahu – Manda Hararo rift segment except for the northern part of the segment (Lahitte *et al.*, 2003a; Moore *et al.*, 2021; Sangha *et al.*, 2022). Further north, the Erta Ale – Tat'Ali Transfer Zone links the Tat'Ali and Erta Ale rift segments (Illsley-Kemp *et al.*, 2018; La Rosa *et al.*, 2019, 2021, 2022).

The Erta Ale Rift Segment runs from Lake Afdera in the south to the Gulf of Zula in the north. It shows an alignment of very active Recent (see section 4.13.2) shield volcanoes (Barberi & Varet, 1970, 1972; Barrat *et al.*, 1998; Pagli *et al.*, 2012; Global Volcanism Program, 2013; Hagos *et al.*, 2016a; Magee *et al.*, 2017; Xu *et al.*, 2017, 2020; Watts *et al.*, 2020). Based on the structure of this part of the basin as well as the petrology of the Erta Ale Range, several authors postulated that it represents the point of continental breakup or even proto-oceanic spreading (Barberi & Varet, 1970, 1977; Barberi *et al.*, 1970; Barrat *et al.*, 1998; Watts *et al.*, 2020; Hurman *et al.*, 2023). In the north, the basin is covered by a rapidly subsiding salt plain punctured by the Dallol and Maraho volcanoes (Bastow *et al.*, 2018; Hurman *et al.*, 2023; chapter

III, this volume). In the north, finally, the Erta Ale Rift Segment forms the Recent Buia graben (Sani *et al.*, 2017). The whole rift segment experiences significant seismicity and dyke emplacement (Gouin, 1979; Nobile *et al.*, 2012; Moore *et al.*, 2021; Ruch *et al.*, 2021; La Rosa *et al.*, 2023). While the margins show significant seismicity, extension in the rift axis accommodates most of the long-term extension (Illsley-Kemp *et al.*, 2018; Moore *et al.*, 2021). In the short term, the InSAR data of Moore *et al.* (2021) also show that extension is focused in the centre of the Depression. Further north, Viltres *et al.* (2020) show that the rift axis is situated west of Mt Ghedem (see fig. 8).

II.5.2.2.5 The Red Sea Rift

The Red Sea Rift is probably composed of one single segment in the study area, represented by the central trough of the Red Sea. It likely does not extend further south than the Hanish Islands (see sections 4.3.1 and 5.1.1). It shows significant seismic and magmatic activity, as testified by the formation of new volcanic islands (Xu *et al.*, 2015; Ruch *et al.*, 2021).

II.5.3 Structural inheritance

Structural inheritance is known to control rift localisation, geometry and evolution (e.g. Buiter & Torsvik, 2014; Molnar *et al.*, 2019; Camanni & Ye, 2022; Zwaan *et al.*, 2022; Brune *et al.*, 2023). In the Afro-Arabian Rift System, there is a general agreement that older structures influence rift development (e.g. Black *et al.*, 1974; Dixon *et al.*, 1987; Beyth, 1991; Ghebreab, 1998; Korme *et al.*, 2004; Corti *et al.*, 2018a, 2022; Varet, 2021). These structures were therefore mapped (fig. 11) to understand their potential

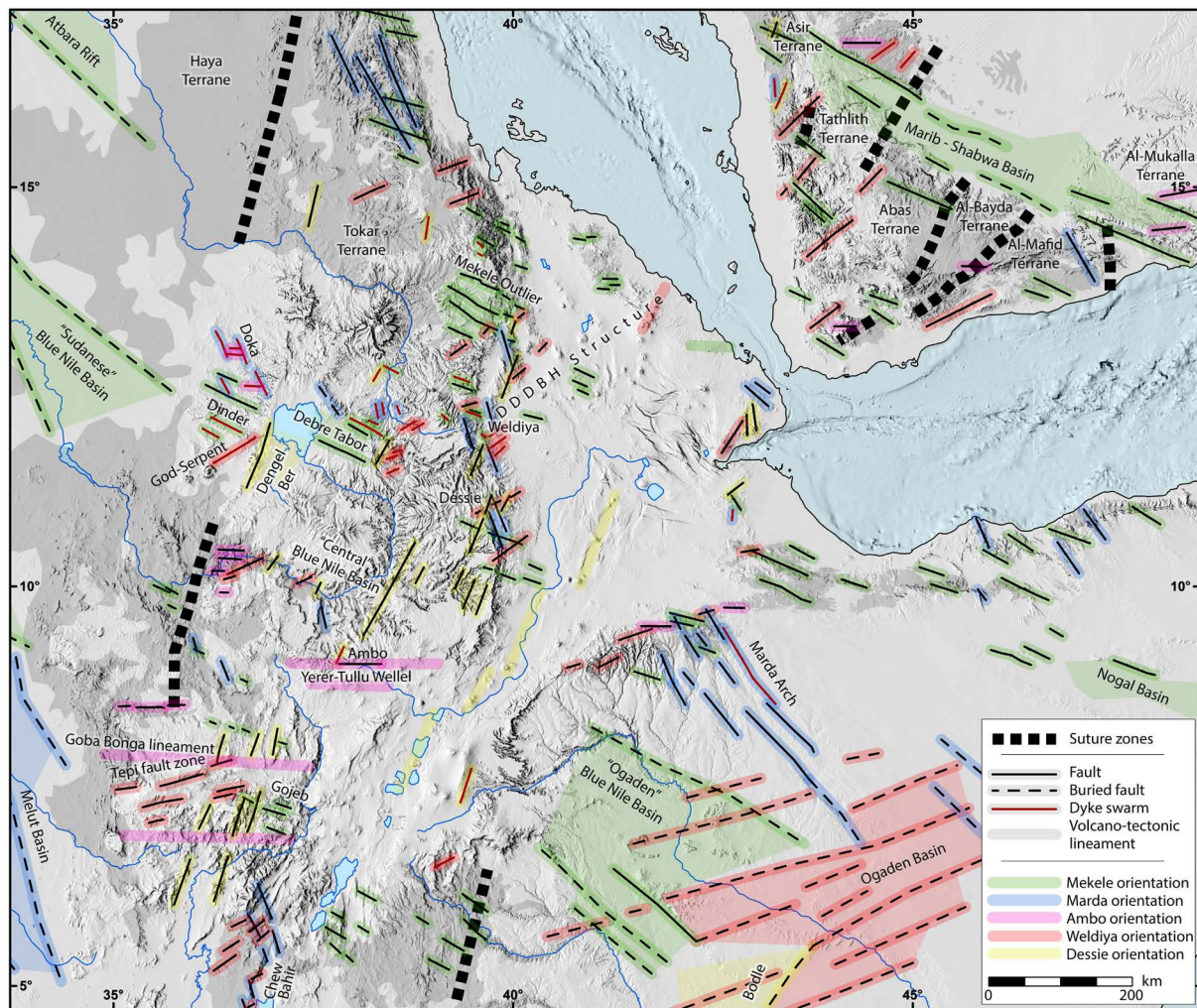


Figure 11: Main potential inherited structures. Suture zones and linear features are represented. Greyed out areas represent basement outcrop. The restoration of this map to pre-rift configuration is presented in suppl. mat. 9. Faults after Moore & Davidson (1978), Salama (1985), Mohr & Zanettin (1988), McHargue et al. (1992), Redfern & Jones (1995), Geological Survey of Ethiopia (1996), Chorowicz et al. (1998), Hunegnaw et al. (1998), Mège & Korme (2004), Dou et al. (2007), Gani et al. (2009), Philippon et al. (2014), Ali & Watts (2016), Mohamed et al. (2016), Corti et al. (2018b, 2022), Rooney et al. (2018), Ali & Lee (2019) and Rime et al. (2022a; b), SRTMGL1 and WGM2012 Free Air gravity data (Bonvalot *et al.*, 2012). Suture zones after sources presented in table 3.

influence on the evolution of the Afar region.

Two types of structures form crustal weaknesses in the study area and were potentially reactivated during the Cenozoic rifting phase: The first type are lineaments and suture zones found in the basement. These structures are often curved and mainly represent ductile deformations developed during the Precambrian East African orogeny cycle (see section 4.1). The prominent sutures relevant to the Afar region are the ones

separating the Asir, Tathlith, Abas, Al-Bayda, Al-Mafid and Al-Mukalla terranes (Quick, 1991; Windley et al., 1996; Johnson et al., 2011; Flowerdew et al., 2013; Fritz et al., 2013; Johnson, 2021; Osman & Fowler, 2021). These structures are represented in fig. 11 and suppl. mat. 3 (see Table 3 for references). The second type are straight linear features that developed after this orogeny. Most of them are faults but also dykes that are interpreted to reactivate pre-rift weaknesses. Five major recurrent linear structure orientations have been

recognized in the literature and/or mapped in the study area (fig. 11). For convenience, we name those orientations after localities or names used in the literature.

- **Mekele orientation, NW-SE:**

This orientation is ubiquitous throughout the study area (fig. 11) and was formed during the Upper Jurassic – Lower Cretaceous rifting phase (see section 4.1). Rifting resulted in many grabens and basins, including the Mekele Basin (Beyth, 1972; Black *et al.*, 1974; Sembroni *et al.*, 2017), the Blue Nile Basin, spanning from Sudan to SE Ethiopian (Salama, 1985; McHargue *et al.*, 1992; Hunegnaw *et al.*, 1998; Gani *et al.*, 2009; Chernnet *et al.*, 2020), we arbitrarily name them “Sudanese”, “Central” and “Ogaden” Blue Nile on fig. 11), numerous rifts in Sudan and South Sudan (e.g. Atbara rift, Melut Basin: Salama, 1985; McHargue *et al.*, 1992; Guiraud *et al.*, 2005; Dou *et al.*, 2007; Mohamed *et al.*, 2016), the Somaliland basins (e.g. Nogal basins: Ali & Watts, 2016; Ali & Lee, 2019) and the southern Yemenites basins (e.g. Marib-Shabwa basin: Redfern & Jones, 1995; As-Saruri *et al.*, 2010). Due to their orientation roughly perpendicular to the Arabian - Nubian/Somalian extension, many of these structures were reactivated during the Cenozoic, mainly in the Gulf of Aden (Beydoun, 1970; Leroy *et al.*, 2012; Ali, 2015), but also in the Mekele outlier (Hagos *et al.*, 2020). These structures are also observed in areas covered by syn-rift rocks, e.g. in the Debre Tabor graben (Chorowicz *et al.*, 1998; Hautot *et*

al., 2006), Dinder Dyke Swarm (Mège & Korme, 2004; Rooney *et al.*, 2018) and the Gojeb graben (Corti *et al.*, 2018b, 2022).

- **Marda orientation, NNW-SSE:**

This orientation has been recognized for a long time (Black *et al.*, 1974; Purcell, 1976; Ghebreab, 1998; Korme *et al.*, 2004; Mège *et al.*, 2016), forming a prominent fault zone on the Somalian plateau, prolonging in the Ogaden basin (Hunegnaw *et al.*, 1998). It is believed to represent a Precambrian fault zone being reactivated in the late Oligocene resulting in an major dyke swarm (also called Ogaden Dyke Swarm). The orientation is similar to the main Red Sea orientation, structuring most of the Afar western margin (Chorowicz *et al.*, 1999; Zwaan *et al.*, 2020b and fig. 11). This orientation also forms the Doke Dyke swarm (Rooney *et al.*, 2018), contributes to the structure of the Chew Bahir basin (e.g. Philippon *et al.*, 2014; Corti *et al.*, 2022), the Melut Basin (Dou *et al.*, 2007; Mohamed *et al.*, 2016) and some faults of the Gulf of Aden (fig. 11).

- **Ambo orientation, E-W:** This orientation is not pervasive over the whole study area but represents a major discontinuity in the western MER margin, forming the *Yerer-Tullu Wellel volcano-tectonic lineament*, of which the Ambo fault is part. This lineament, and the similar *Goba Bonga lineament*, represent major crustal discontinuities, possibly delineating Precambrian crustal weakness (Mohr, 1967; Berhe, 1986; Abebe *et al.*, 1998; Gashawbeza *et al.*, 2004; Korme *et al.*, 2004; Keranen & Klemperer, 2008; Corti, 2009; Corti *et al.*,

2018a; b, 2022; Bowden et al., 2022). They have an significant influence on the structure of the Main Ethiopian Rift. E-W oriented structures are also found on the Afar south-eastern margin, NW of Lake Tana (Rooney *et al.*, 2018) and in south-eastern Yemen, but a clear link is difficult to establish.

- **Weldiya orientation, NE-SW:**

This orientation is represented by many faults and dykes throughout the study area (fig. 11). Faults of this orientation structure the Yemenite eastern margin (fig. 11, Collet *et al.*, 2000). In central Ethiopia, this orientation represents many dykes, including the God-Serpent dyke swarm (Chorowicz *et al.*, 1998; Mège & Korme, 2004; Rooney *et al.*, 2018), dykes east of Debre Tabor and the ones present near Weldiya, on the central-western Afar margin (fig. 4 and 11). Multiple faults cutting through the western Afar margin have the same orientation (fig. 11, Kronberg *et al.*, 1974b, 1975; Collet *et al.*, 2000). It furthermore corresponds to the orientation of the Tepi fault zone (Moore & Davidson, 1978; Corti *et al.*, 2018b, 2022), one of the fault sets structuring the Chew Bahir basin (Philippon *et al.*, 2014; Corti *et al.*, 2022), the south eastern margin of the Afar (fig. 4 and 11) and the Ogaden basin (Worku & Astin, 1992; Hunegnaw *et al.*, 1998). The widespread nature of this orientation hints at a pre-rift age (Kronberg *et al.*, 1975) and the similarity with the structures of the Ogaden basin could indicate that this structural trend formed during the Karoo rifting phase in eastern Africa, i.e. being Permian to Triassic in age (see section 4.1).

- **Dessie orientation, NNE-SSW:**

This orientation is mainly present in the Nubian part of the study area. It is predominantly present in the Central Blue Nile Basin (Kronberg *et al.*, 1974b, 1975; Gani *et al.*, 2009), reaching the western Afar margin, notably in the vicinity of Dessie (fig. 4 and 11), but it is also visible in southern Ethiopia and forms the Dengel Ber graben (Chorowicz *et al.*, 1998). It is unclear when these structures formed and whether they represent a structural inheritance. As this orientation is roughly perpendicular to the Nubia – Somalia extension and parallel to the Wonji fault belt, it is plausible that it was formed since the Miocene, when the strain was not yet focused in the MER. The possibility of pre-rift influence is, however, plausible as such structures are also found in northern Ethiopia. Chorowicz *et al.* (1998) furthermore noted that the Dengel Ber graben predates the termination of flood volcanism indicating at least an Oligocene age. (Hunegnaw *et al.*, 1998; Shawel *et al.*, 2022) show that the *Mandera Bodle Rift* in southern Ogaden has the same orientation and is floored by Carboniferous to Permian sediments. Similarly, Hercynian (Devonian to Carboniferous) structures in eastern Arabia show a similar N-S to NNE-SSW orientation (e.g. Qatar Arch: Geert *et al.*, 2001; Faqira *et al.*, 2009; Stewart, 2016).

Both basement lineaments/suture zones and linear discontinuities certainly play a role in the localisation and segmentation of the rift. When brought back to pre-rift position (section 6.1 and

suppl. mat. 9), the major sutures between Abas, Al-Bayda and Al-Mafid terranes align with the NMER and could potentially extend under the Yerer-Tullu Wellel volcano-tectonic lineament that represents a crustal-scale discontinuity (see above). These terrane boundaries also align with the pre-rift boundary between the Danakil Block and the Ali Sabieh Block (suppl. mat. 9). The presence of these suture zones combined with different linear discontinuities (see suppl. mat. 9) formed an significant weakness zone in present-day Afar Depression which probably contributed to the localization of the Afar Rift.

The contribution of the Dessie, Weldiya and Mekele discontinuities to the segmentation of the Afar Depression and NMER is noticeable. The step from the Fantale-Dofen and Issa rift segments corresponds to the presence of Mekele discontinuities potentially extending up to Dire Dawa (fig. 4 and 13). The difference between the south-western and central-western Afar experiencing slightly different geological histories (fig. 7) could also be explained by the prominent Weldiya-oriented structures. SE of Dessie, Wolfenden *et al.* (2005) measured an ENE-striking fault zone (the “Dese-Bati accommodation zone”) featuring at least 800 m of sinistral offset. The major eastward step of the western Afar margin between the central-western Afar and the Danakil Depression also coincides with Weldiya and Dessie-oriented faults and dykes (fig. 4 and 11). This Weldiya-oriented structure can be traced further away from the western margin to the active volcanic range of Dabbayra, the Alayta-Afdera Transfer Zone and the Nabro volcanic range. It is also along this lineament that Tertiary intrusive rocks are outcropping (Limmo,

Asa Ale, Affara Dara and Dioita granites). This alignment is believed to represent a transfer zone along a pre-rift crustal weakness (Black *et al.*, 1972; Tazieff *et al.*, 1972; Barberi & Varet, 1977; Varet, 2018). Varet (2018) calls it the *DDDBH Fracture Zone* (Dabbayra-Dababhu-Dubi-Bidu-Hanish) but we instead call it the *DDDBH Structure* to avoid any interpretation. The kinematic reconstruction (section 6.1) postulates that this structure extends to a fault zone of similar orientation identified in Yemen (suppl. mat. 9). The DDDBH structure also roughly coincides with the limit between the Arrata Microplate and the southern Danakil Block, i.e., the present-day boundary of the Arabian plate (see section 5.1.1).

It can therefore be postulated that the structural inheritance of both Precambrian lineaments/suture zones and younger faults sets played a role in the localization of the Afar Rift and the segmentation of the rift.

II.6 EVOLUTION OF THE AFAR REGION

II.6.1 Kinematic modelling

A wide range of different models for the evolution of the Afar region has been presented in the literature (Mohr, 1970, 1972; Tazieff *et al.*, 1972; Barberi & Varet, 1977; Le Pichon & Francheteau, 1978; Sighler, 1980; Acton *et al.*, 1991, 2000; Gaulier & Huchon, 1991; Crossley *et al.*, 1992; Collet *et al.*, 2000; Manighetti *et al.*, 2001; Eagles *et al.*, 2002; Wolfenden *et al.*, 2004; Beyene & Abdelsalam, 2005; Bosworth *et al.*, 2005; Garfunkel & Beyth, 2006; McClusky *et al.*, 2010; Hammond *et al.*, 2011; Schettino *et al.*, 2016; Stab *et al.*, 2016;

Sani *et al.*, 2017; Purcell, 2018; Zwaan *et al.*, 2020b). However, there is currently no holistic, detailed model for the evolution of the Afar region spanning the complete rift history, considering the complex geological structure of the Depression and including the Red Sea – Gulf of Aden rift system.

The new geological maps of the Afar Depression (fig. 4), of the southern Red Sea & western Gulf of Aden region (fig. 6), and the map of basement structures (suppl. mat. 3), along with the review of the geological history and tectonic structure of the region (sections 4 and 5) allow proposing a new model for the kinematic evolution of the Afar region. In order to present a kinematically and geometrically coherent and plausible model, modelling was performed using GPlates 2.3.0 (Müller *et al.*, 2018). We did not model complex intraplate deformation but only constructed a model including rigid plates, microplates, blocks or *tectonic elements* (a more generic term than plate or block, see e.g. V  rard, 2021). A movie of the kinematic model is presented in suppl. mat 18, and data are available in suppl. mat. 19.

The model is constrained by the position of the three main plates (Nubian, Somalian and Arabian) through time. We used the Arabian plate as the reference plate. The Somalian plate was then moved relative to the Arabian plate according to the relatively well-constrained magnetic anomalies of the Gulf of Aden presented by Fournier *et al.* (2010) with ages updated according to Ogg (2020) (suppl. mat. 10). Movement before the last recognized magnetic anomaly is modelled to match structures on both sides of the Gulf of Aden (see suppl. mat. 11).

Because of the lack of magnetic anomalies in the Red Sea older than 6.3 Ma (Izzeldin, 1987) or 4.3 Ma (Schettino *et al.*, 2016), the movement of the Nubian plate was tied to the Somalian plate. A study of magnetic anomalies in the southern Indian Ocean defines the relative movement of the two plates, but this movement does not correspond to geological evidence, probably due to structures outside of the MER accommodating a significant amount of strain (DeMets *et al.*, 2021). We thus modelled the movement of the Nubian plate relative to the Somalian one with the following assumptions. The Precambrian basement suture zones and other distinctive features (fig. 6 and suppl. mat. 3) must align between the Nubian and Arabian plates before rifting (see suppl. mat. 11, 12 and 13). The movement also needs to be coherent with the opening of the Gulf of Suez and the movement along the Dead Sea fault (section 4.3.1). This defines the position of the Nubian plate relative to the Somalian plate in pre-rift times. They remain fixed relative to each other until rifting begins in the NMER around ~ 13 Ma (see section 4.9). The present-day position of the plate defines its movement between 13 Ma and the Present. The magnetic anomalies of the Red Sea (Izzeldin, 1987; Schettino *et al.*, 2016, suppl. mat. 10) were therefore not used to construct this model but the result led to an excellent match of these anomalies (see suppl. mat. 18), highlighting the plausibility of the model.

The movement of the other tectonic elements of the Afar region was then modelled according to the geological evidences described in sections 4 and 5 and considering the correlation of transverse and potentially pre-rift structures (section 5.3, suppl. mat. 11

and 12). We used 14 tectonic elements to model the deformation of the Afar region (suppl. mat. 14). While it is known that the central-eastern Afar does not behave as rigid blocks (Manighetti *et al.*, 2001), these blocks were necessary for the modelling. This is to avoid kinematic incoherence and to serve as a guide to produce a coherent graphical output for the paleogeological maps (fig. 12).

Unlike all models except the one by Gaulier & Huchon (1991), we did not model the Danakil Block as a single rigid block. Considering the pre-rift configuration of the three main tectonic plates described above and the critical step on the western Afar margin between 12 and 13°N linked to the DDBH structure, the consideration of a rigid Danakil Block would have implied significant shortening (compression) in the northern part of the Danakil Depression. Conversely, it is known that the Danakil nowadays represent a plate boundary (see section 5.1.1 and fig. 9) and that the NW-SE-orientated, probably pre-rift DDBH structure strongly influences the structure of the region (see section 5.3). Near the town of Weldiya, the dykes with the same orientation were dated between 21 and 17 Ma by Ayalew *et al.* (2019), testifying to the activity of the structure during early rift development. We thus segmented the Danakil Block according to this orientation in the Nabro range and across Beilul Bay. Like other studies (Chu & Gordon, 1998; Eagles *et al.*, 2002;

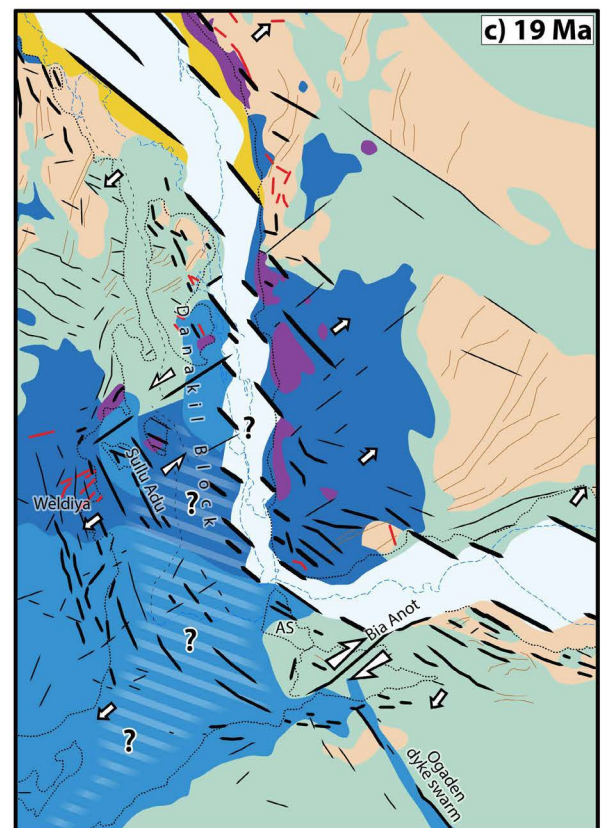
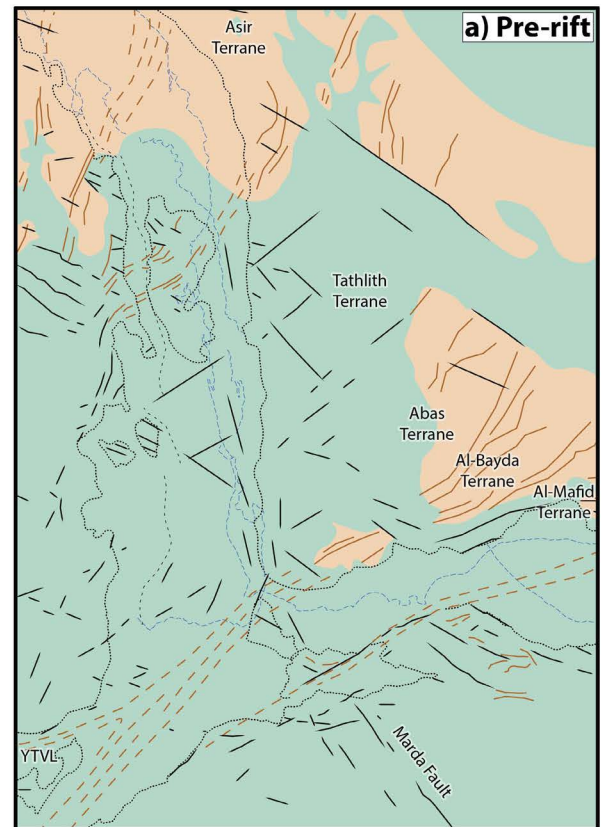
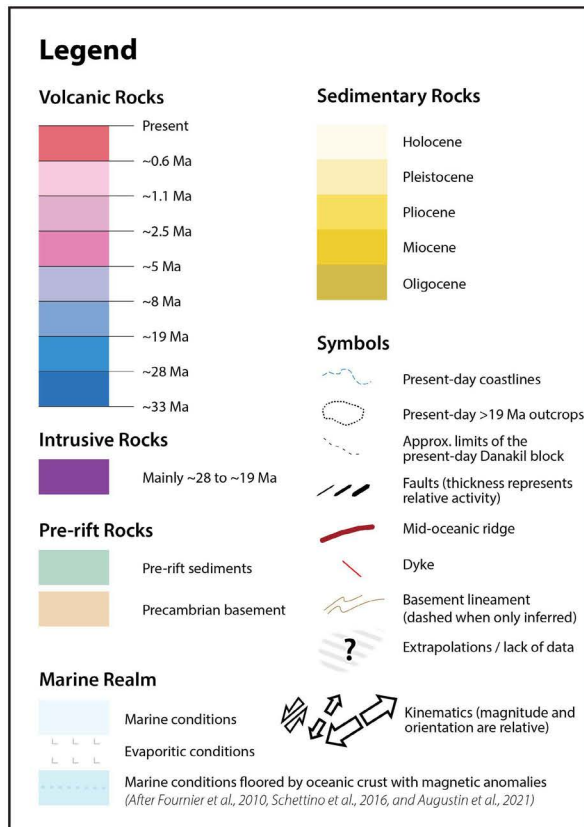
Schettino *et al.*, 2016), we also separated the Buri peninsula from the Danakil Block in the Howlakil Bay that is marked by a positive Bouguer gravity anomaly (suppl. mat. 4) and by seismic activity (fig. 10). This avoids any compression in the north during the rotation of the Danakil Block.

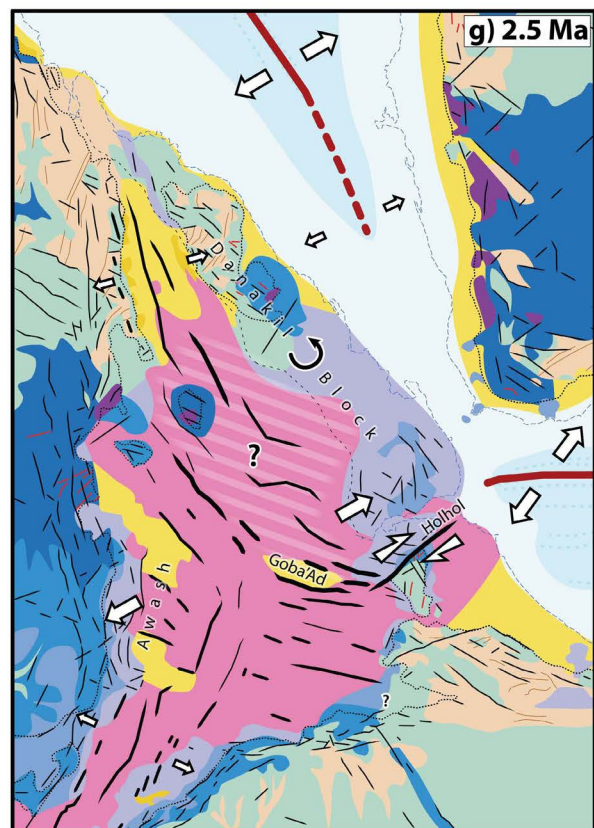
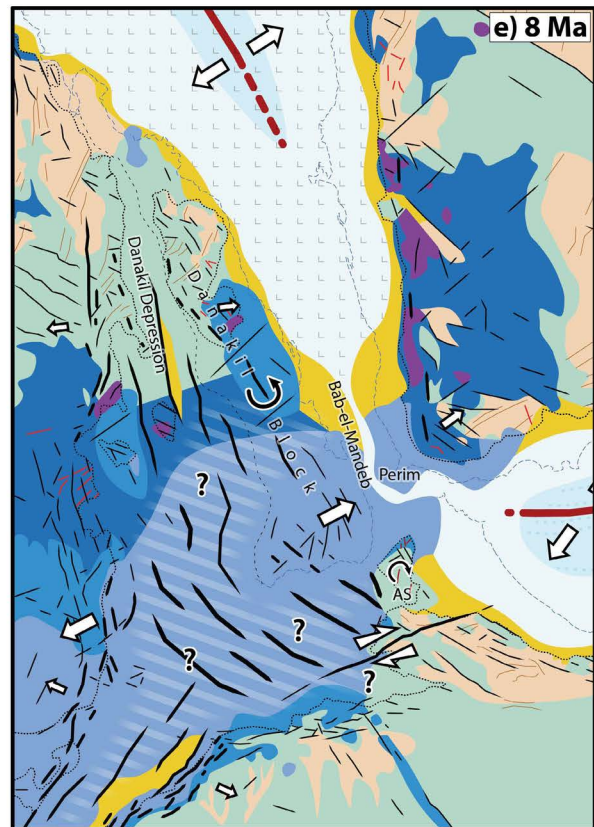
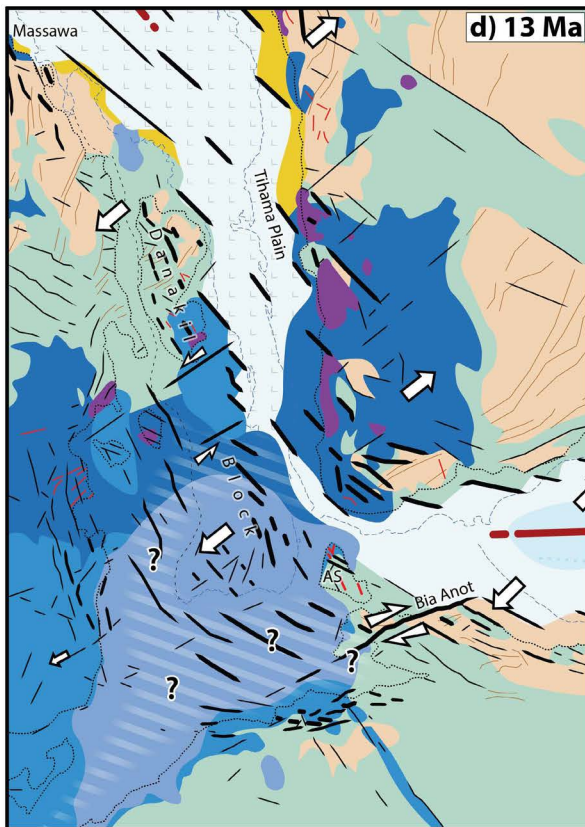
As discussed in section 5.1.2, the kinematics and timing of movement of the Ali Sabieh Block are poorly constrained. The pre-rift configuration suggests that this Block was situated in present-day south-eastern Afar. We modelled movement along the eastern Bia Anot fault since early rifting followed by rotation accommodated in the northern Bia Anot discontinuity. To avoid overlap of the Danakil and Ali Sabieh Block, the latter must rotate ahead of the Danakil Block, consistently with the observation of no rotation since at least 8 Ma (Audin *et al.*, 2004).

II.6.2 Tectonostratigraphic evolution of the Afar region

The tectonostratigraphic evolution of the Afar region is discussed based on the geological history of the different parts of the Afar (section 4, fig. 7), the tectonic structure of the Afar (section 5), and the new kinematic model (section 6.1). Note that despite the consideration of many sources and references, the kinematic model has to be considered conceptual, schematic and hypothetical.

Figure 12 ► : Tectonostratigraphic evolution of the Afar region. Paleogeological map of the evolution of the Afar Depression and surrounding regions for pre-rift times (a), ~30 Ma (b), ~19 Ma (c), ~13 Ma (d), ~8 Ma (e), ~5 Ma (f), ~2.5 Ma (g), ~1 Ma (h), and present day (i). This model is based on the GPlates kinematic model, our review of the geological history of the region (section 4 and fig. 7) and on the geological maps presented in fig. 4 and 6. Note that the model has to be considered as conceptual, schematic and hypothetical. AG= Asal-Ghoubbet, AS = Ali Sabieh Block, F-D= Fentale-Dofen, GoT = Gulf of Tadjoura, YRVL = Yerer-Tullu Wellel volcanotectonic lineament.





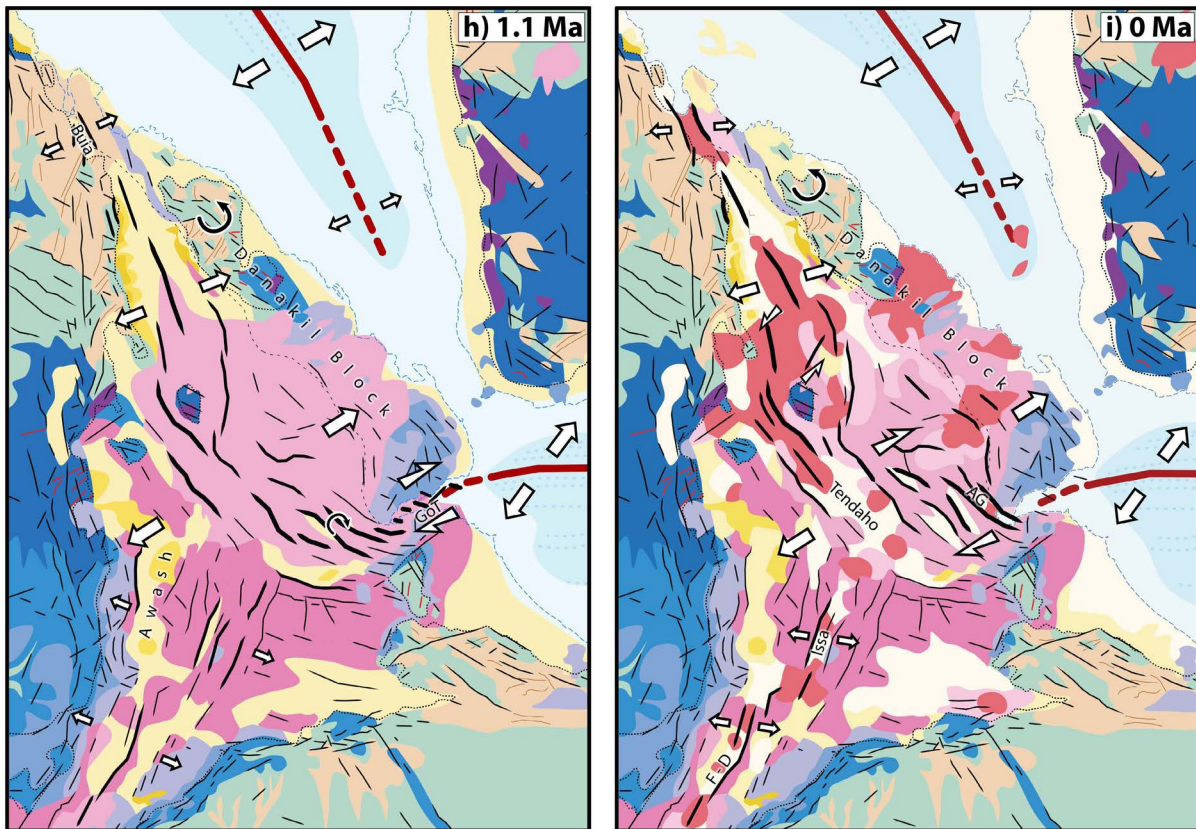


Figure 12 (continued)

Prior to rifting and the eruption of the Traps (fig. 12a), the region was structured by Precambrian and Phanerozoic inherited structures (section 5.3). The suture zones between the Abas, Al-Bayda, Al-Mafid and Al-Mukalla terranes probably extended into the future southern Afar, NMER and possibly continued under the Yerer-Tullu Wellel volcano-tectonic lineament (YTVL). The Marda fault potentially extended under the future Red Sea Rift.

Between 33 and 28 Ma (fig. 12b), a short pulse of hotspot-related volcanism led to the deposition of a significant thickness of volcanic material (the "Traps") in the northern part of Ethiopia and in Yemen (sections 4.4 and 4.5 and references therein). Before 30 Ma, tectonic movements are only recorded in the eastern and central Gulf of Aden (fig. 7).

During the late Oligocene and Early Miocene (~28 to ~19 Ma, fig. 12c), Trap basalts erupted further towards the south (sections 4.4, 4.8 and references therein). The Ogaden dyke swarm, reactivating the Marda fault zone, was also emplaced, along with several plutonic bodies and dykes in the Afar region and along the Yemenite and Saudi coast (sections 4.8, 4.3.1, 4.11 and references therein). Rifting was mainly concentrated in the Gulf of Aden and Red Sea margins, affecting the Eritrean, Somalian and Yemenite coasts, as well as the Danakil Block (sections 4.2, 4.3.1, 4.6 and references therein). In the Afar Depression, the first tectonic movements are reported at the south-western Afar margin and in Sullu Adu (section 4.10 and references therein). As no movement is recorded in the Danakil Depression, a transfer zone probably developed between Sullu Adu and the Red Sea,

potentially reactivating Weldiya (SW-NE) oriented inherited structures, as shown by the ~20 Ma dykes of the same orientation observed near the eponym city. It is plausible that the Ali Sabieh Block (AS on fig. 12) began separating from the Somalian plateaux around the same time, potentially activating the Bia Anot Fault. The different rift segments probably reactivated some of the Mekele (NW-SE) and Marda (NNW-SSE) oriented faults. At that time, an early sea probably made its way from the Gulf of Aden to the Red Sea but the configuration of this sea is unknown (section 4.3.2 and references therein). Oceanic spreading had just begun by ~19.5 Ma east of the Sheba ridge (section 4.2 and references therein).

Between ~19 and ~13 Ma (fig. 12d), the eruption of volcanic material continued on the present-day Ali Sabieh and Southern Danakil Blocks, as well as in the eastern part of the future NMER and the south-western Afar margin (sections 4.7, 4.6, 4.9, 4.10 and references therein). Tectonic movement accelerates with the westward propagation of the Gulf of Aden Rift. At ~16 Ma, oceanic spreading began in the western Gulf of Aden (Fournier *et al.*, 2010). The majority of the tectonic movements in the Afar region are concentrated on the Red Sea and Gulf of Aden margins. The central-western Afar margin experiences a period of volcanic quiescence, but some strain was probably accommodated in present-day southern/central Afar (sections 4.11 and references therein). According to our kinematic model, the Ali Sabieh Block moved away from the Somalian Plateau during this period, possibly dragged by the Arabian plate. As discussed in section 5.1.2, it is unclear how this movement is accommodated, but we model a strike-slip

motion on the eastern Bia Anot fault. This is consistent with Gaulier & Huchon (1991) reporting sinistral movements during this period. The marine realm expanded significantly. During the Mid-Miocene, the Sea extended further than today under the Tihama plain (Hakimi *et al.*, 2021) and near Massawa (Sagri *et al.*, 1998), but during the Langhian-Serravallian (~16-12 Ma), conditions became increasingly restricted (section 4.3.2 and references therein).

The period between 13 and 8 Ma (fig. 12e) corresponds to the most critical changes in the regional geodynamics since rifting initiation. It coincides with a major change in the dynamic of the East African Rift System (Michon *et al.*, 2022). Oceanic spreading in the central Red Sea is believed to begin around 13 Ma (section 4.3.1 and references therein). This period also corresponds to a major reorganisation in the northern part of the Red Sea with the abandonment of the Gulf of Suez and the full strain transfer to the Dead Sea Fault (section 4.3.1). The deposition of sediments of the Chorora Fm in an initial depression marks the initial development of the NMER (section 4.9 and references therein). This changes the Afar region from a two plates system to a triple junction with three distinct major tectonic plates. Another major change is the individualization of the Danakil Block and the beginning of its rotation, caused by the significant reduction in the tectonic activity of the southern Red Sea rift and the related increase in activity of the Afar Rift (sections 4.3.1, 4.13.2 and references therein). This constitutes a major rift jump. The Afar Rift begins its northward propagation, and rifting initiates in the Danakil Depression (section 4.13.2). The Bab-el-Mandeb area is probably devoid of

tectonic activity since these times (section 4.3.1). According to our model, the Ali Sabieh Block rotated to reach its present-day position during this period. Eruptions are reported on the southern Ethiopian plateaux, in the NMER, on the southern Danakil Block and near the Bab-el-Mandeb strait (sections 4.4, 4.9, 4.6, 4.3.2 and references therein). The connection between the Gulf of Aden and the Red Sea was strongly reduced, leading to evaporitic conditions throughout the Red Sea (section 4.3.2 and references therein). The temporal coincidence between the end of active rifting of the southernmost Red Sea and the restricted conditions in the Red Sea evidence a potential link. The absence of active rifting coupled with dynamic topography (Moucha & Forte, 2011) caused by the plume probably triggered uplift or at least reduced subsidence, which, coupled with global relative sea level fall during the Mid-Miocene Climatic Transition (MMCT, Miller *et al.*, 2020), led to the isolation of the basin. Sedimentation at the foot of the Yemenite escarpment and the only slightly younger eruption of the Perim volcano might also have played a role in this isolation.

The period between 8 and 5 Ma (fig. 12f) is marked by the eruption of the Dahla basalts across the Afar Depression and possibly along the Somalian coast (sections 4.6, 4.7, 4.8, 4.9, 4.10, 4.11, 4.13.2, references therein and Ali, 2009, 2015; Ali *et al.*, 2018). Significant tectonic movements are also reported in the western Afar margin, in Sullu Adu and probably in southern Afar (sections 4.8, 4.10, 4.11 and references therein). The development of the Danakil Depression continues, accommodating sediments and a few lava flows (sections 4.13.2 and references therein). The rotation reported

at the NW margin of the Ali Sabieh Block probably indicates that strain between the Danakil and Ali Sabieh Blocks was accommodated in that area (Audin *et al.*, 2004). In the NMER, the border faults are gradually less active as strain is concentrated in the central part rift (section 4.9 and references therein). The Red Sea margins were probably nearly devoid of tectonic activity (section 4.3.1 and references therein), the strain being accommodated by the oceanic spreading in the north and by the Afar Rift in the south, forming a relay structure. After the deposition of significant evaporites, the Red Sea again experiences marine conditions (section 4.3.2 and references therein).

Between 5 and 2.5 Ma (fig. 12g), the Afar Depression is marked by the eruption of the lower Stratoid Fm nowadays ubiquitously outcropping south of the Tendaho – Goba'Ad discontinuity (section 4.7, 4.8, 4.9, 4.10, 4.11 and references therein). The rotation of the Danakil Block is accommodated along the Holhol fault (Audin *et al.*, 2004) and in southern and central Afar, where sedimentary basins are forming in Goba'Ad (section 4.12), as well as in the Middle and Lower Awash valley. Last being important for the first Hominine ancestors thriving in a lacustrine and fluvial environment (section 4.10 and references therein). The Danakil Depression accumulated sediments and magmatic material (section 4.13.2 and references therein). Extension is primarily accommodated by seafloor spreading in the Red Sea and the Gulf of Aden, (sections 4.2, 4.3 and references therein).

Between 2.5 and 1.1 Ma (fig. 12h), the central-eastern Afar experiences voluminous volcanic eruptions forming

the upper Stratoid Series (section 4.12 and references therein) and corresponding to a northward shift of tectonic activity from the southern Afar to the central-eastern Afar (Tortelli *et al.*, 2022). It also constitutes a triple junction migration (Kalb, 1995; Tesfaye *et al.*, 2003; Wolfenden *et al.*, 2004; Hammond *et al.*, 2011), probably caused by the significantly higher velocities of the Arabian plate compared to the Somalian one (Maestrelli *et al.*, 2022). The Afar Rift continues its northward propagation with sedimentation in the Danakil basin and the beginning of the opening of the Buia graben (sections 4.13.2, 4.13.3 and references therein). The rotation of the Danakil Block is accommodated by block rotation in central-eastern Afar, and strain is transferred from the Holhol fault to the present-day Gulf of Tadjoura (GoT on fig. 12, see sections 4.7, 4.12 and references therein). Significant sedimentation is also ongoing in the Awash valley and the Danakil Depression (sections 4.10, 4.13 and references therein).

Between 1.1 Ma and the present day (fig. 12i), the Afar Depression witnesses the propagation of the Gulf of Aden Rift, forming the Asal-Ghoubbet graben (AG on fig. 12, see section 4.3 and references therein). The linkage between the Gulf of Aden Rift and the Afar Rift is achieved through a series of en-echelon faults block rotation and rift focussing in central-eastern Afar (sections 4.12 and references therein). Volcanic eruptions focus in active rift segments (Asal, Manda Inakir, Dabbahu-Manda Hararo, Erta Ale, Issa Graben, Fentale, Hanish, Zubair, Jebel at Tair) and volcanoes at the margins (Nabro, Ma'Alalta, Moussa Ali, Dembel, Dabbayra) (sections 4.2, 4.3.1, 4.6, 4.8, 4.9, 4.10, 4.11, 4.12, 4.13

and references therein). Continental sedimentation fills the Tendaho graben and other grabens, as well as the Danakil Depression. The sea starts invading the latter, and the Asal-Ghoubbet graben during sea-level highstands, forming substantial evaporite deposits (sections 4.2, 4.12, 4.13.2 and references therein). In the NMER and southern Afar, the tectonic and volcanic activity is mainly restricted to the Wonji segments (Fantale-Dofen (F-D on fig. 12) and Issa).

II.7 DIFFERENT RIFTING STYLES IN THE AFAR DEPRESSION

II.7.1 Central Afar domain versus Danakil domain

Similar to Keir *et al.* (2013), Stab *et al.* (2016), Le Gall *et al.* (2018) and Ahmed *et al.* (2022), we identify two different domains featuring different rifting styles, magmatic and sedimentary histories: the “Danakil” domain and the “Central Afar” domain (see Table 4, fig. 13). Note that the “Central Afar” is meant here in its broad sense, including the central-eastern, central-western, south-western Afar, and possibly the south-eastern Afar. The boundary between these two domains broadly corresponds to the DDBH Structure (see section 5.3 and fig. 11).

Central Afar is shaped by a wide depression cut by numerous grabens. Elevations in the Central Afar Depression are usually between 350 and 1000 m, while a single, well-defined valley shapes the Danakil Depression with elevations reaching 125 m below sea level at lake Karum (fig. 1). According to our kinematic model (section 6.1) and coherently with the northward

	Central Afar Domain	Danakil Domain
Topography	Wide depression with grabens. Elevation between 350 and 1000 m	Valley shape with elevation below sea level (down to -125 m)
Extension	Important extension (~200-250 km)	Lower extension (~10 - 100 km)
Stretching factor	Higher	Smaller
Relative age of rifting	Older	Younger
Crustal thickness	Relatively thick crust with Moho depth around 24 km	Relatively thin crust with Moho depth around 22 km, possibly as low as 14 km.
Oligocene to Mid-Miocene magmatism	Voluminous (between 1 and 2 km thickness) flood basalt deposits covering the margin	Only one large magmatic deposit in Adigrat (0.6 km thick). Dykes and plugs on the margins of the Depression.
Basin fill	Vastly dominated by lavas	Dominated by sediments intercalated by lava flows
Syn-rift magmatism	at least 1.5 km of Pleistocene lavas and possibly ~1 km of Mio-Pliocene lavas	Metric- to decametric-scale flows and sills in the Red Bed Series
Recent (<0.6 Ma) magmatic history	Voluminous lava deposits, mainly in the Dabbahu-Manda Hararo range	Voluminous lava deposits in the southern part of the rift axis (Erta Ale range)
Underplating	Evidences for magmatic underplating	No evidence for magmatic underplating
Strain accommodation	Important magma-accommodated extension	Mechanical extension in the N, magma-accommodated extension in the Erta Ale Range.
Thinning	Magma-compensated thinning	Non-compensated thinning
Rift localization	Distributed extension, incorporation of continental fragments	Clear rift localization in the centre of the rift valley
Hotspot influence	Strong	Less important
Passive margin equivalent	Oceanic/marginal plateau	Magma-rich margin

Table 4: Comparison between the Central Afar and Danakil domains.

propagation of the Afar Rift, Central Afar experienced significant extension with high stretching factors (~200-250 km and ~2.5 – 3.3 respectively, see suppl. mat. 15) while both the extension and the stretching factors are lower for the Danakil Depression (~10 – 100 km and ~1.2 – 1.8 respectively). The age of rifting in Central Afar is also older than in the Danakil Depression (see sections 4 and 5.2.1).

While measures of crustal thickness vary considerably between authors (see suppl. mat. 16, appendix A), the average thickness (Dugda et al., 2005; Stuart et al., 2006; Hammond et al., 2011; Reed et al., 2014; Lavayssière et al., 2018; Kibret et al., 2019; Wang et al., 2021; Ahmed et al., 2022) for the central-western and

central-eastern Afar or these zones plus the south-western and south-eastern Afar is 24 ± 3 km (N=90 resp. 103, see suppl. mat. 16 for zones considered). In the Danakil Depression, some authors indicate a very shallow Moho depth (14 km by Makris and Ginzburg, 1987; 16 km by Berckhemer et al., 1975 and Hammond et al., 2011 and the average crustal thickness is 22 ± 3 km (N=19 see suppl. mat. 16, Hammond et al., 2011; Lavayssière et al., 2018; Wang et al., 2021; Ahmed et al., 2022), being slightly thinner than for Central Afar. Studies that encompass both the Danakil and Central Afar evidence the presence of a shallower Moho under the Danakil Depression (Berckhemer et al., 1975; Makris & Ginzburg, 1987; Mammo, 2004; Tiberi et al., 2005; Hammond et al., 2011;

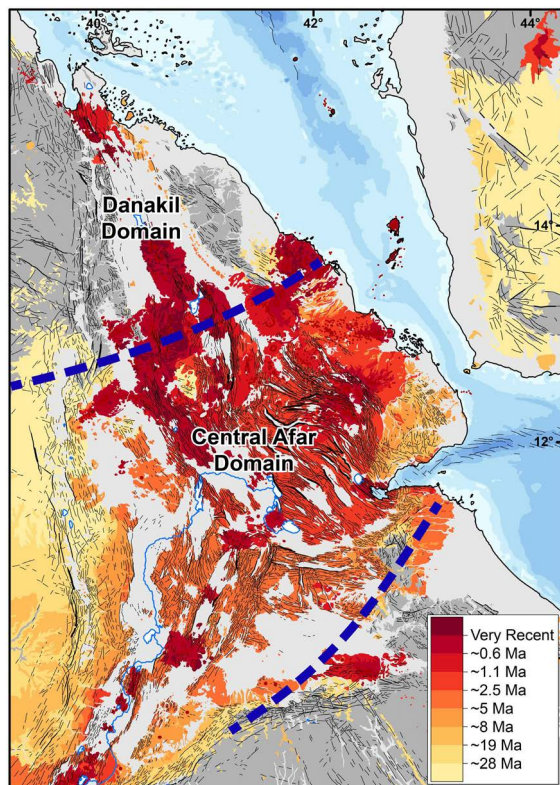


Figure 13: Illustration of the two main domains of rifting style and magmatic history. Volcanic and intrusive rocks are shown in shades of red indicating their age. Dark grey indicates pre-rift rocks. Dashed blue lines indicate the approximate boundaries between the Danakil and Central Afar Domains. The Central Afar Domain features a much more extensive pre- and syn-rift magmatic products. Bathymetry similar to fig. 4.

Lavayssière *et al.*, 2018; Chambers *et al.*, 2019) with the exception of Wang *et al.* (2021) (see also appendix A). The lower Bouguer gravity anomaly under the Danakil Depression (suppl. mat. 4) is coherent with the presence of a relatively shallow Moho.

The pre- to early syn-rift (Oligocene to Early Miocene) volcanic activity also differs between both domains (fig. 13). The margins and plateaus west of Central Afar experienced very significant volcanic activity. Beccaluva *et al.* (2009) mention 1.8 km of lavas in Lalibela, and Mohr & Zanettin (1988) report thicknesses between >1290 and >1870 m on the central western Plateau and

between >1030 and <1980 m in the southern western Plateau. Geophysical data shows an altered crust, possibly caused by these magmatic intrusions (Ahmed *et al.*, 2022). Similarly, on the Yemenite plateau, the thickness of magmatic material locally exceeds 1.5 km (Baker *et al.*, 1996; Ukstins Peate *et al.*, 2005). On the margins of the Danakil Depression, however, significant Oligocene to Mid-Miocene lava is only outcropping in the region of Adigrat (approx. 500 m to 650 m thick according to Mohr & Zanettin, 1988; Beccaluva *et al.*, 2009). Few dykes and plugs are mentioned on the margins of the Depression (Brinckmann & Kürsten, 1971; Geological Survey of Ethiopia, 1978a). Ahmed *et al.* (2022) evidence unmodified crust in this area. Overall, the Danakil domain experienced less magmatic influence during pre- and early syn-rift times compared to Central Afar.

The syn-rift architecture of both basins is also distinctive. Central Afar is dominated by magmatic products with at least 1.5 km of Pleistocene (Lahitte *et al.*, 2003b) and possibly ~1 km of Mio-Pliocene lavas (Stab *et al.*, 2016). The syn-rift deposits of the Danakil Depression (i.e. the Red Bed Series) are conversely dominated by sedimentary products (Brinckmann & Kürsten, 1971; C.N.R - C.N.R.S. Afar Team, 1973; Le Gall *et al.*, 2018). Le Gall *et al.* (2018) mention two lava flows (average thickness of 10 m) and seven sills (3-10 m thick) in a ~250 m sedimentary succession. Brinckmann & Kürsten (1971) similarly mention one 20-30 m flow and three ≤10 m flows in a ~270 m sedimentary log. Our observations match these descriptions.

Both regions experienced significant volcanic activity during Recent times (<

0.6 Ma, fig. 4 and 13). In Central Afar, eruptions are mainly concentrated in the Dabbahu-Manda Hararo Range and Asal Graben. In the Danakil Depression, voluminous magmatic products are present in the Erta Ale range, interpreted to represent incipient break-up (Barberi et al., 1972a; Bastow & Keir, 2011; Keir et al., 2013; Le Gall et al., 2018).

Central Afar is believed to feature significant magmatic underplating (Stab et al., 2016; Wang et al., 2021), while none have been detected under the Danakil Depression (Wang et al., 2021). This is consistent with the reported extensive magmatic activity of Central Afar throughout its rift history. The volume of underplated material is probably at least as large as the volume of erupted material (Cox, 1980). Underplating may explain the higher elevations by increasing crustal thickness and providing isostatic buoyancy.

A significant part of the strain is believed to be accommodated by magmatic intrusions in Central Afar (Hofstetter & Beyth, 2003; Bastow & Keir, 2011; Keir et al., 2011, 2013; Ebinger et al., 2013; Ferguson et al., 2013a) even though faulting still play an important role outside the magmatic segments (Polun et al., 2018). In the Danakil Depression, extension is likely accommodated by mechanical extension (faulting and ductile stretching) in the north and magmatic intrusion in the Erta Ale range since Recent times (Bastow & Keir, 2011; Keir et al., 2013; Bastow et al., 2018; Illsley-Kemp et al., 2018; Le Gall et al., 2018; Ahmed et al., 2022; Hurman et al., 2023; La Rosa et al., 2023). The crust in Central Afar is significantly thicker (between 17% and 55%) than expected based on mechanical extension solely (McKenzie, 1978)

considering the extension predicted by our tectonic plate reconstruction (see Appendix A) but also the reconstruction of Eagles *et al.* (2002) (discussion in Bastow & Keir, 2011; Keir et al., 2013; Stab et al., 2016; Ahmed et al., 2022). This indicates that crustal thinning is compensated by underplating, magmatic intrusions and erupted material, a process known as *magma compensated thinning* (Thybo & Nielsen, 2009; Bastow & Keir, 2011; Ferguson et al., 2013b; Stab et al., 2016). It corroborates the hypothesis of magma-accommodated extension. Conversely, the Danakil Depression shows a crustal thickness compatible with mechanical extension (see Appendix A). Thinning in the Danakil Depression is thus probably not compensated, with only localized magmatic intrusions, supporting the hypothesis of extension dominated by mechanical processes.

In Central Afar, the extension is distributed over a large area (>100 km) and vary significantly over times of less than 1 Ma (Kogan et al., 2012; Pagli et al., 2014; Doubre et al., 2017; Moore et al., 2021; Sangha et al., 2022, see also section 5.2.2). In the Danakil Depression, rifting is localized in the centre of the Depression, accommodating most of the long-term extension, even if the margins still experience seismic activity (Illsley-Kemp et al., 2018; Moore et al., 2021).

The contrasting rift styles between Central Afar and the Danakil domain can be explained by distinct volumes of magmatic accretion until Recent times. We tentatively explain this by a stronger influence of the Afar plume in Central Afar compared to the Danakil Depression. This difference in hotspot influence can explain:

- 1) The abundance of Oligocene to Mid-Miocene volcanic rocks at the margins of Central Afar.
- 2) The difference in basin fill, including the abundance of Late Miocene to Middle Pleistocene volcanic rocks in Central Afar and the sediment-filled Danakil Depression.
- 3) The lower crustal thickness of the Danakil Depression and its lower elevation due to the absence or limited magma-compensated thinning.
- 4) The importance of mechanical extension in the N-Danakil Depression compared to Central Afar, where magmatic intrusion accounts for a significant part of strain accommodation.
- 5) The apparent contradiction between the near-oceanic state of the Danakil Depression and its younger age, smaller stretching factor and smaller total extension compared to Central Afar. The significant magmatic production of the former compensates for thinning and delays breakup.
- 6) The absence of penrose-like oceanic crust in the Central Afar – Bab-el-Mandeb region due to magma-compensated thinning. The rest of the Gulf of Aden – central/southern Red Sea rift system experienced comparable extension but feature oceanic crust.

II.7.2 Stages of rift evolution

One hypothesis presented in the literature is that Central Afar and the Danakil Depression represent two subsequent stages of rift evolution. The Danakil Depression would represent the evolution of what Central Afar is nowadays (Ebinger & Hayward, 1996; Hayward & Ebinger, 1996; Bastow &

Keir, 2011; Keir et al., 2013; Moore et al., 2021). Based on this hypothesis, Bastow & Keir (2011) and Keir et al. (2013) postulate that the Danakil Depression experienced, like Central Afar, a long and significant history of magma intrusions. The same authors suggest that the current ductile plate stretching in the Danakil Depression is caused by its history of magmatic intrusions.

The idea of the Danakil Depression representing a more developed stage of the current Central Afar is inconsistent with the northward propagation of the rift (discussed in section 5.2.1), with the younger age of the Danakil Depression (section 4.13.2) and with the lower stretching factor and extension rate it experienced (suppl. mat. 15). It is also inconsistent with their contrasting magmatic history and the much more voluminous erupted, intruded and underplated material reported in Central Afar (section 7.1 and references therein). These observations invalidate the idea that the Danakil Depression experienced the same geological history as Central Afar in the past. We therefore consider that the Danakil Depression is not representing the direct evolution of what Central Afar is nowadays. The Danakil Depression is more evolved or more mature in the sense that it is closer to oceanic spreading, but not in the sense that the Danakil Depression was structured like, or was looking like Central Afar a few million years ago. The Danakil Depression and Central Afar represent two different types of rift and not two different stages of evolution of the same rift type. The observations on volumes of magmatic products also preclude that strain was accommodated by significant magma intrusion for a protracted period during the early phase

of rift development in the Danakil Depression. This, in turn, calls for a reassessment of the idea that the current ductile plate stretching in Danakil follows - or even is a consequence - of this magmatic intrusion history. We thus argue that the Danakil Depression first experienced mechanical extension since Late Miocene times and that extension in the southern part of the Depression only recently began to be accommodated by magmatic intrusions.

II.7.3 The Afar Depression as an analogue for understanding the development of magma-rich margins and oceanic plateaus?

As Central Afar and the Danakil Depression represent two different domains, it is interesting to compare these with older, well-studied passive margins where most concepts of rifted margins have been developed.

The Danakil Depression feature several strong similarities with magma-rich margins. The southern part of the Depression features a thin crust, an important central volcanic range interpreted (Erta Ale Range) and the Red Bed Series on the margins show a seaward-dipping reflectors (SDR) structure (Keir *et al.*, 2013; Le Gall *et al.*, 2018; Watts *et al.*, 2020). In the northern part of the Danakil Depression, magmatism plays a more minor role. The faults are dipping towards the rift axis, and the basin presents a graben structure. However, unlike Le Gall *et al.* (2018), we would not call it a magma-poor margin as volcanic activity and intrusions are present in Alid, Dallol, Maraho and Bada (Lowenstern *et al.*,

2006; Nobile *et al.*, 2012; Ruch *et al.*, 2021). Furthermore, the northward propagation of the Afar Rift will probably lead to more important volcanism in this part of the Depression in the future. This style of rifting is perhaps comparable to the southern Red Sea or the western Gulf of Aden.

With an even more pronounced magmatic production, the Central Afar domain could definitely fit in the magma-rich category but several elements point towards a potential different future evolution than the classical magma-rich margins, possibly forming the equivalent of an oceanic plateau.

Oceanic plateaus are areas of positive topography in an oceanic basin featuring a crust thicker than typical oceanic crust. They are commonly linked to hotspot activity and flood basalts, e.g. the Greenland-Iceland-Faroe Ridge linked with the Iceland hotspot and the Greenland flood basalts, the Rio Grande Rise linked with the Tristan hotspot and the Parana flood basalts, the Mascarene plateau linked with Réunion hotspot and the Deccan flood basalts, the Kerguelen linked with the Kerguelen hotspot and the Rajmahal Flood basalts (Coffin & Eldholm, 1992; Courtillot *et al.*, 2003). Because oceanic plateaus are sometimes situated far offshore in the oceanic domain, they were thought to be purely magmatic, but it was recently shown that several (e.g. Rio Grande rise, Mauritius and the Mascarene Plateau, Elan Bank and William's Ridge in the vicinity of the Kerguelen Plateau, possibly Iceland and the Beira High offshore Mozambique) contain continental material (Bott *et al.*, 1974; Frey *et al.*, 2000; Weis *et al.*, 2001; Exon *et al.*, 2007; Torsvik *et al.*, 2013; Ventura Santos *et al.*, 2019; Foulger *et al.*, 2020, 2022; Yuan *et al.*, 2020; Coffin

et al., 2021; Tomasi et al., 2021). They are also linked with the formation of microcontinents (e.g. Jan Mayen, Sao Paolo High, Laxami-Laccadive Ridge, Seychelles, possibly Elan Bank and *Icelandia*). The formation of these microcontinents and the incorporation of continental fragments are explained by overlapping rift propagation and rift jumps (Frey et al., 2000; Müller et al., 2001; Borissova et al., 2003; Gaina et al., 2003; Collier et al., 2008, 2009; Minshull et al., 2008; Bénard et al., 2010; Mohriak et al., 2010; Gernigon et al., 2012; Peron-Pinvidic et al., 2012; Hammond et al., 2013a; Heine et al., 2013; Nair et al., 2013; Torsvik et al., 2013, 2015; Karson, 2016; Hjartarson et al., 2017; Kukla et al., 2018; Schiffer et al., 2018; Foulger et al., 2020, 2022; Geoffroy et al., 2020; Mishra et al., 2020; Peace et al., 2020; Neuharth et al., 2021). These rift jumps are believed to be related to the presence of mantle plumes (Hill, 1991; Müller et al., 2001; Gaina et al., 2003, 2009; Mittelstaedt et al., 2008; Martin et al., 2011; Hammond et al., 2013a).

Similar to oceanic plateau, Central Afar is wide, relatively flat, situated at higher elevation than the Danakil Depression or Red Sea and Gulf of Aden basin and features a relatively thick crust. This crust is heavily structured by magmatic material, including up to several kilometres of lava flows in the upper crust and at least as much underplated and intruded material (see section 7.1). It therefore has similarities with so-called *Icelandic-type crusts* dominated by magmatic material in the upper crust (3-10 km) and by magma-inflated lower crust with remnant blocks of continental material (10-30 km) (Foulger et al., 2003, 2020). While Iceland represents a more developed stage than

Central Afar, it is believed that the structure of the entire Greenland-Iceland-Faroe Ridge is comparable (Hjartarson et al., 2017; Foulger et al., 2020). The shear velocities of the lower crust in Central Afar (7.0 - 7.2 km/s reported by Hammond *et al.*, 2011; 7.0 - 7.3 km/s reported by Bastow & Keir, 2011; 7.3 km/s reported by Makris & Ginzburg, 1987) are also similar to the ones reported for Icelandic-type lower crust (7.0-7.3 km/s, Foulger *et al.*, 2020). The Jan Mayen, Seychelles, Sao Paolo High and Elan Bank microcontinents have sizes comparable to the Danakil Block, while the continental fragment of the Rio Grande Rise and Mauritius might be compared to the Affara Dara fragment. In a similar way as other microcontinents, the Danakil Block was isolated due to the propagation of two overlapping rifts (see section 5.2) and the isolation of fragments of continental material in Central Afar was similarly caused by rift segments jump (see sections 5.2.2, fig. 8).

The Central Afar domain may therefore represent the early development of an oceanic plateau that might not transition to normal oceanic spreading in the near or medium term. Using the Central Afar domain as an analogue to understand the formation of classic passive margins should therefore be considered with caution.

II.8 CONCLUSION

This study presents new and revised geological maps of the Afar Depression and the Afro-Arabian Rift System. The maps materialize an up-to-date and integrative representation of the geology of the region. The tectonostratigraphic evolution of the Afar Depression and

surrounding regions has been reviewed, allowing for a detailed understanding of the chronology and localization of major tectonic, magmatic and sedimentary events. The review highlights a complex geological history and shows that most geological events are diachronous throughout the different parts of the region.

The Afar region is tectonically structured by four rifts, the Gulf of Aden Rift, the Main Ethiopian Rift, the Afar Rift and the Red Sea Rift separating the Nubian, Arabian and Somalian plates, as well as the Arrata Microplate. Unlike commonly presented, the integrative dataset shows that the Afar Rift propagates northwards and not southwards. It forms a relay structure with the Red Sea Rift and is linked with the counter-clockwise rotation of the Danakil Block. The segmentation and localisation of the different rifts and rift segments is partially inherited by Precambrian and Phanerozoic structures.

The combination of maps, the review of the geological history and the current structure of the Afar region allows us to propose a new kinematic model and paleogeological maps of the evolution of the Afar region. The eruption of the flood basalts began ~33 Ma ago in northern Ethiopia and in Yemen. Rifting began approx. 27 Ma ago in the Afar region. During the first phase, rifting was mainly concentrated in the Gulf of Aden and the Red Sea with more limited extension in Central Afar. This situation changed between 13 and 8 Ma with a major tectonic reorganization of the region. Tectonic activity significantly decreased in the southernmost Red Sea as the Afar Rift began to accommodate most of the extension while oceanic spreading

initiated in the central Red Sea. At the same time, the Northern Main Ethiopian Rift formed and the Afar region developed into a triple junction. Significant magmatism accompanied distributed extension in the Depression since the Pliocene, forming km-thick lava deposits.

The Afar Depression can be segmented into two distinct domains: the Danakil domain and the Central Afar domain. Central Afar experienced significant extension as well as protracted and extensive magmatism during pre- and syn-rift times. It is believed to be strongly influenced by the Afar hotspot. Magma-compensated thinning prevented full breakup. As such, Central Afar cannot be considered as a conventional rift system but probably represents the early development of an oceanic plateau. Conversely, the Danakil Depression is younger and experienced less magmatic activity until Recent times. Although the extension was lower than in Central Afar, the absence of magma-compensated thinning allowed it to reach an evolved stage of continental breakup. The Danakil Depression probably represents the equivalent of a magma-rich margin. As the Danakil Depression experienced a different geological history compared to Central Afar, it cannot be considered as the simple further evolution of the latter.

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AUTHOR CONTRIBUTIONS

VR was responsible for the conceptualization, investigation, literature review, fieldwork, compilation of the database, compilation of the maps, figures, discussion and writing. AF was responsible for the conceptualization, investigation, literature review, fieldwork, discussion, writing, project administration, funding acquisition and supervision. JR was responsible for

discussion and writing. TK was responsible for the conceptualization, fieldwork, discussion, writing and supervision.

DECLARATION OF COMPETING INTERESTS

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

DATA AVAILABILITY

Datasets related to this article can be found at <https://doi.org/10.5281/zenodo.7351643> and <https://doi.org/10.5281/zenodo.7351765>, hosted at Zenodo (Rime et al., 2022a; b).

APPENDIX A: CRUSTAL BALANCING

Introduction

To quantify the mechanisms of accommodation of extension that are important for the Afar Depression, we propose a simple approach of crustal balancing. We distinguish between magma-accommodated rifting and mechanically-accommodated rifting.

In the case of 100% magma-accommodated rifting (fig. A1a), the theoretical space created by the extension (A_e , defined as the amount of extension (e) multiplied by the initial crustal thickness (T_0)) is entirely compensated by magmatic addition. The crustal thickness remains constant, and the present-day (syn-rift) crustal surface over the cross-section (A_1) is increased by A_e compared to the initial (pre-rift) surface (A_0).

In the case of 100% mechanical accommodation (fig. A1b), the extension is entirely compensated by crustal thinning. The crustal surface over the cross-section remains constant between pre- and syn-rift ($A_0 = A_1$).

Methods

The ratio of magma-accommodated rifting (m) is calculated as:

$$m = \frac{A_1 - A_0}{A_e}$$

$$m = \frac{A_1 - [(L_1 - e)T_0]}{e T_0}$$

With L_1 the cross-section length and the other terms defined above.

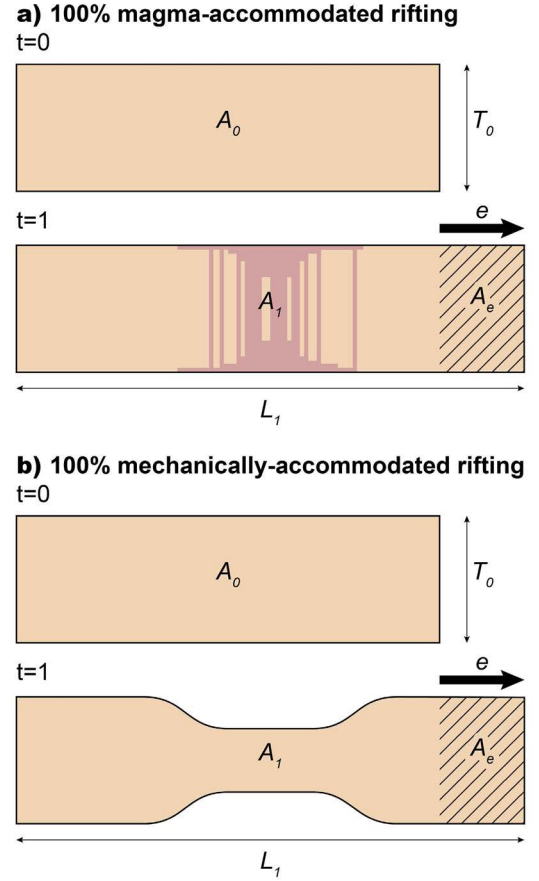


Figure A1: Schematic representation of **a)** magma-accommodated rifting and **b)** mechanically-accommodated rifting.

A_1 is estimated based on the published Moho depths for the Afar Depression (Berckhemer et al., 1975; Makris & Ginzburg, 1987; Tiberi et al., 2005; Dugda & Nyblade, 2006; Hammond et al., 2011; Reed et al., 2014; Lavayssière et al., 2018; Kibret et al., 2019; Wang et al., 2021; Ahmed et al., 2022, see also suppl. mat. 14). The data points situated at less than 75 km distance of two cross-sections (one in the Danakil Depression, one in Central Afar) are projected onto them (fig. A2a, b, c). As the data are very scattered, three estimates of the crustal geometry are suggested: a minimal, best and maximum estimate. The extension (e) is calculated from the GPlates kinematic model (see section 6.1). Initial crustal thicknesses (T_0) are assumed to be 38 km and 42 km for the Danakil and

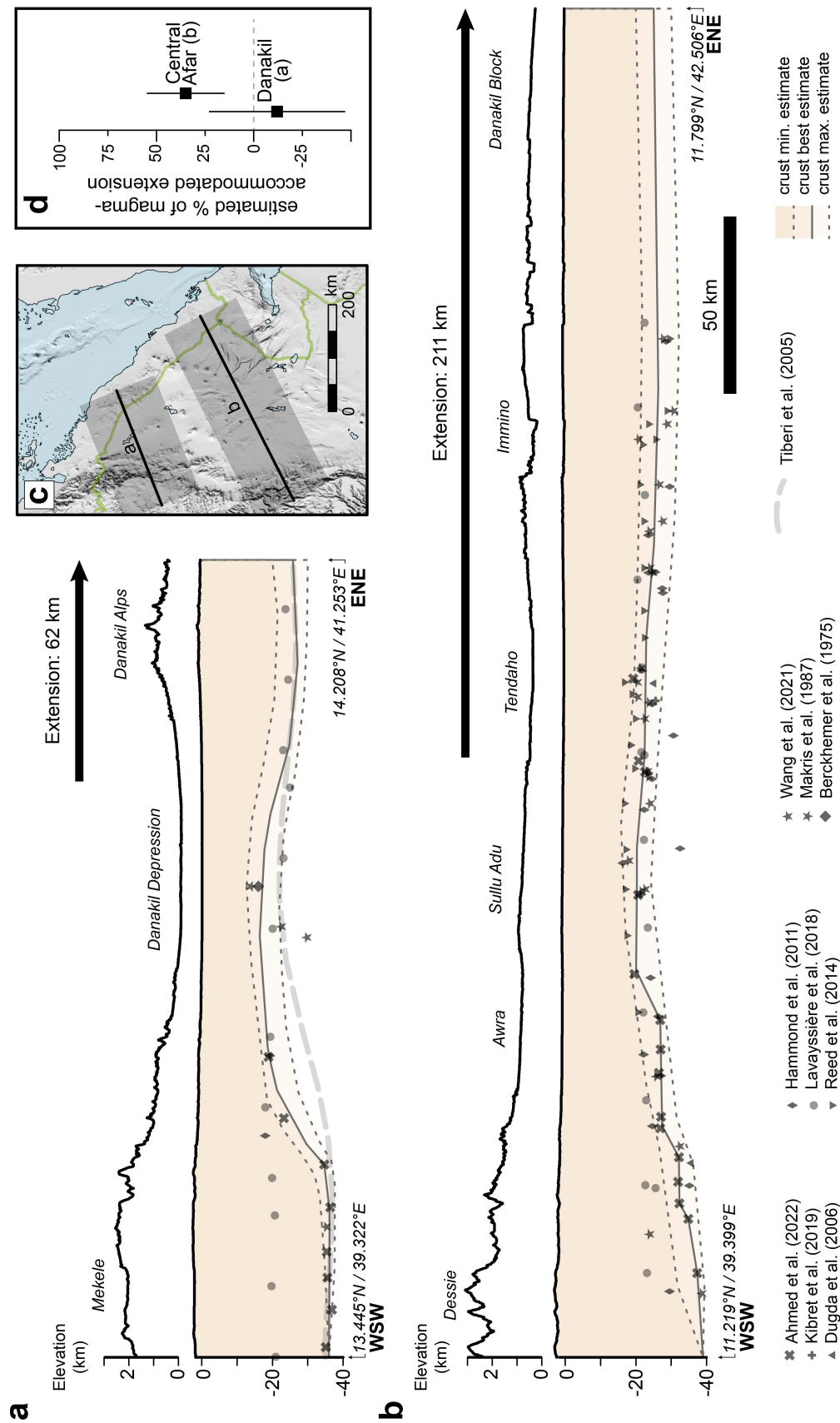


Figure B2: Crustal balancing in the Danakil Depression (a) and Central Afar (b). Symbols represent projected Moho depth by different authors. Coloured polygons indicate minimum, best and maximum estimates of crustal geometry. Surface topography is also indicated. c) Position of the cross-sections. Only the data situated in the shaded areas were projected on the cross-sections. d) Results of the proportion of magma-accommodated extension. Black square represents the best estimates and bars indicate min. and max. estimates.

Central Afar, respectively (Ahmed *et al.*, 2022).

Results

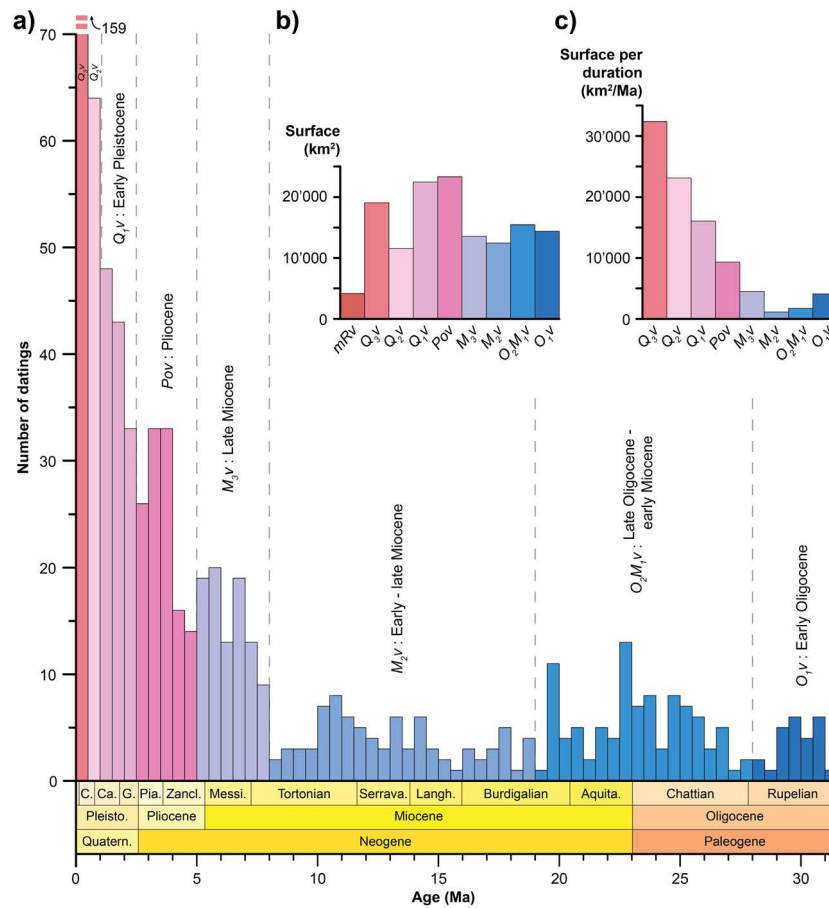
For the Danakil cross-section (fig. A2a), the results (fig. A2d) show that the percentage of magma-accommodated extension ranges from -47% (minimum estimate of crust surface), -12% (best estimate) and +23% (maximum estimate). For Central Afar (fig. A2b), the percentages are 17% (minimum estimate), 36% (best estimate) and 55% (maximum estimate).

Discussion

In accordance with the scattered Moho depth data, the results show a large spread and should thus be interpreted with caution. The best estimate for the Danakil Depression tend to indicate slightly negative values. This would mean that material was removed from the crust, which we consider not realistic. These low values could indicate that 1) the Moho depth data are underestimated, 2) the extension value is underestimated, 3) the initial crustal thickness is overestimated. However, considering that positive values are within the range of calculated uncertainty, we interpret these results as valid. They show low magma addition to the crust, consistently with outcrop evidences (see section 7.1). Conversely, calculations for Central Afar shows that a significant proportion of extension is accommodated by magmatic addition to the crust, consistently with other evidences discussed in section 7.1.

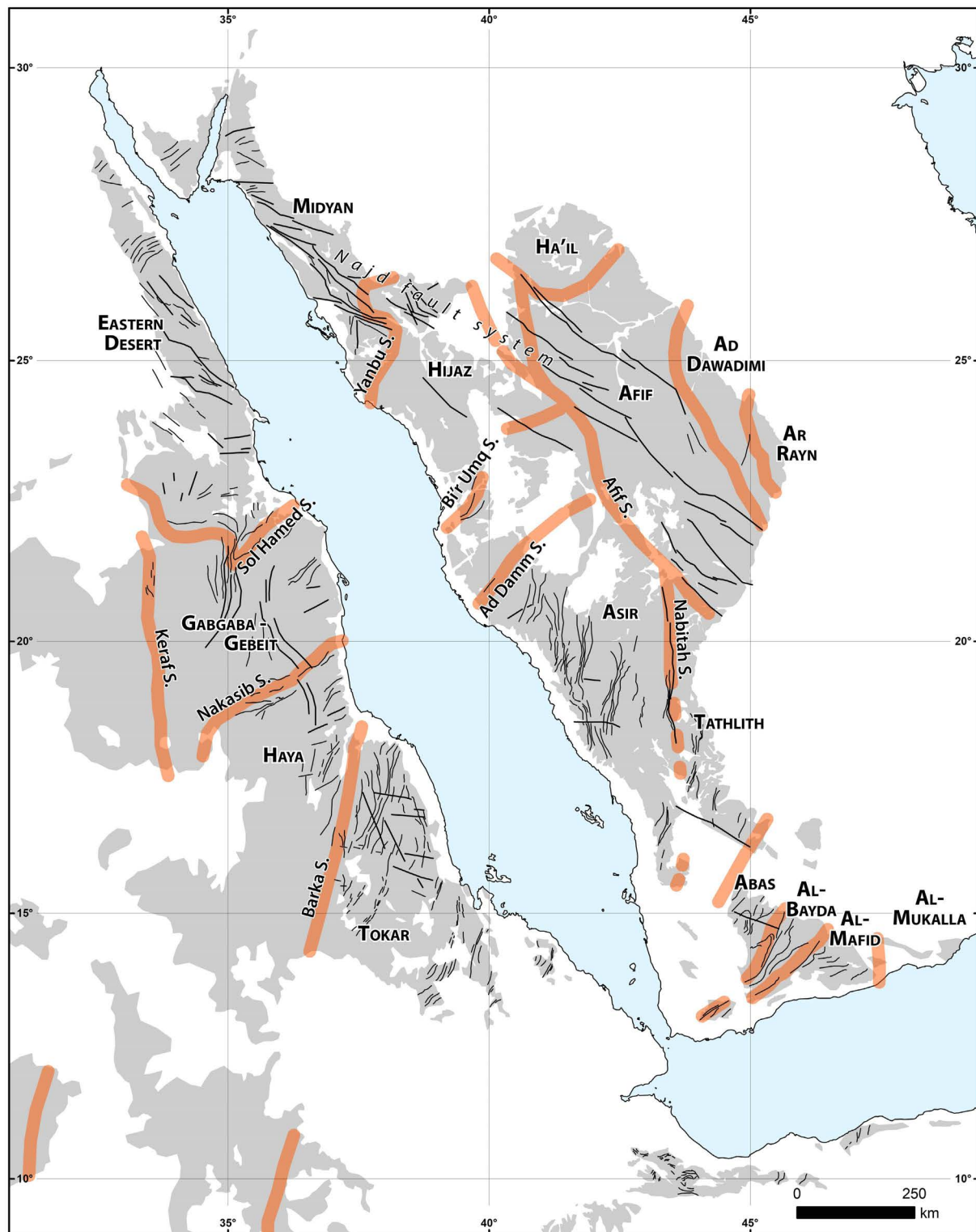
Even considering the large uncertainties, these results show a remarkable difference between the Danakil and Central Afar domains.

APPENDIX B: SUPPLEMENTARY DATA

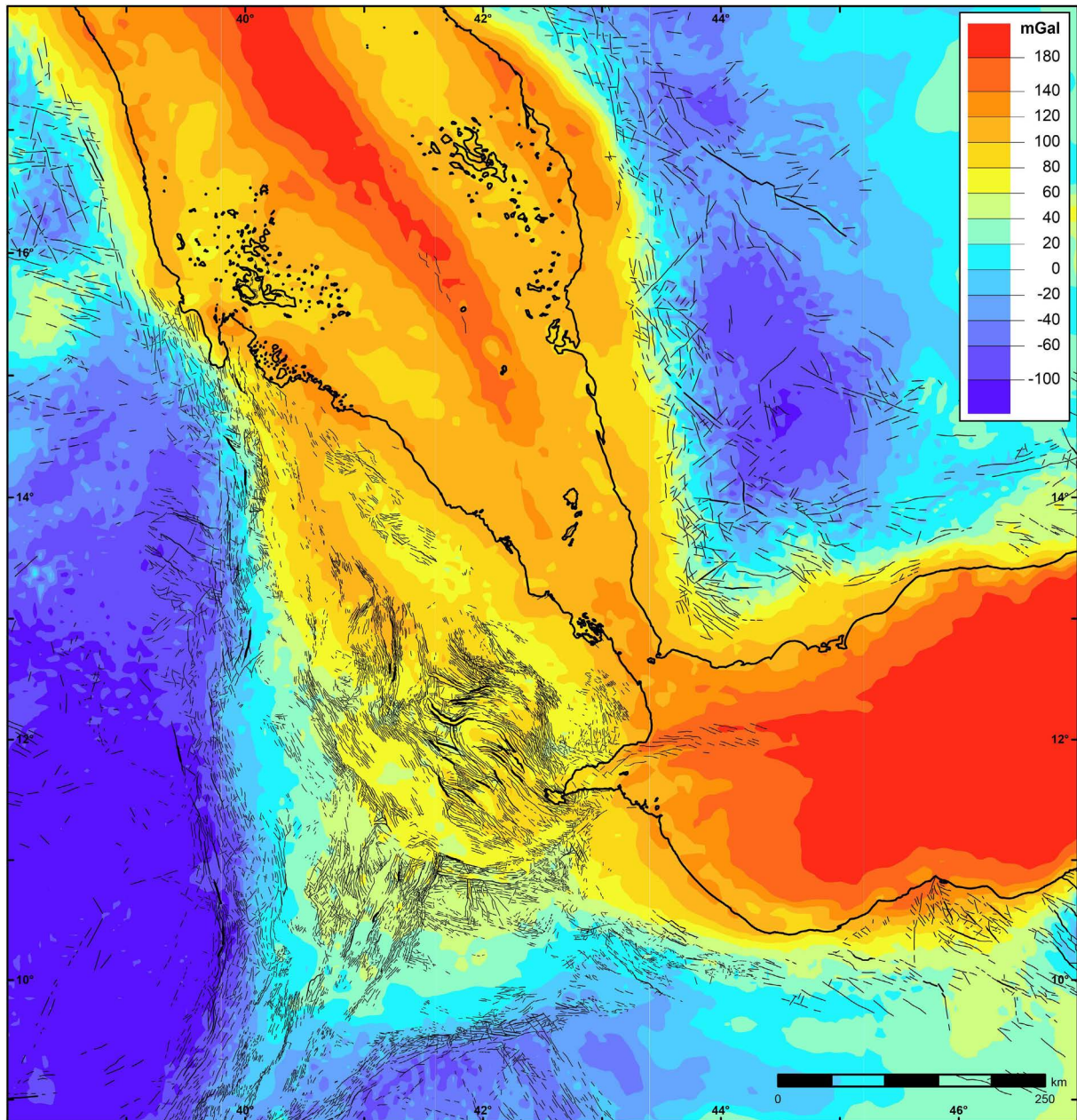


Supplementary material 1: Analysis of age distribution: **a)** Histogram of the Cenozoic dating database. N=759. Two data points at 32.9 and 43 Ma are not shown for clarity. Note that the first bin (0-0.5 Ma) is truncated but reaches 159. **b)** Surface covered by each magmatic stratigraphic unit on the geological map of the Afar Depression (fig. 4). **c)** Mapped surface per duration for each unit (km²/Ma).

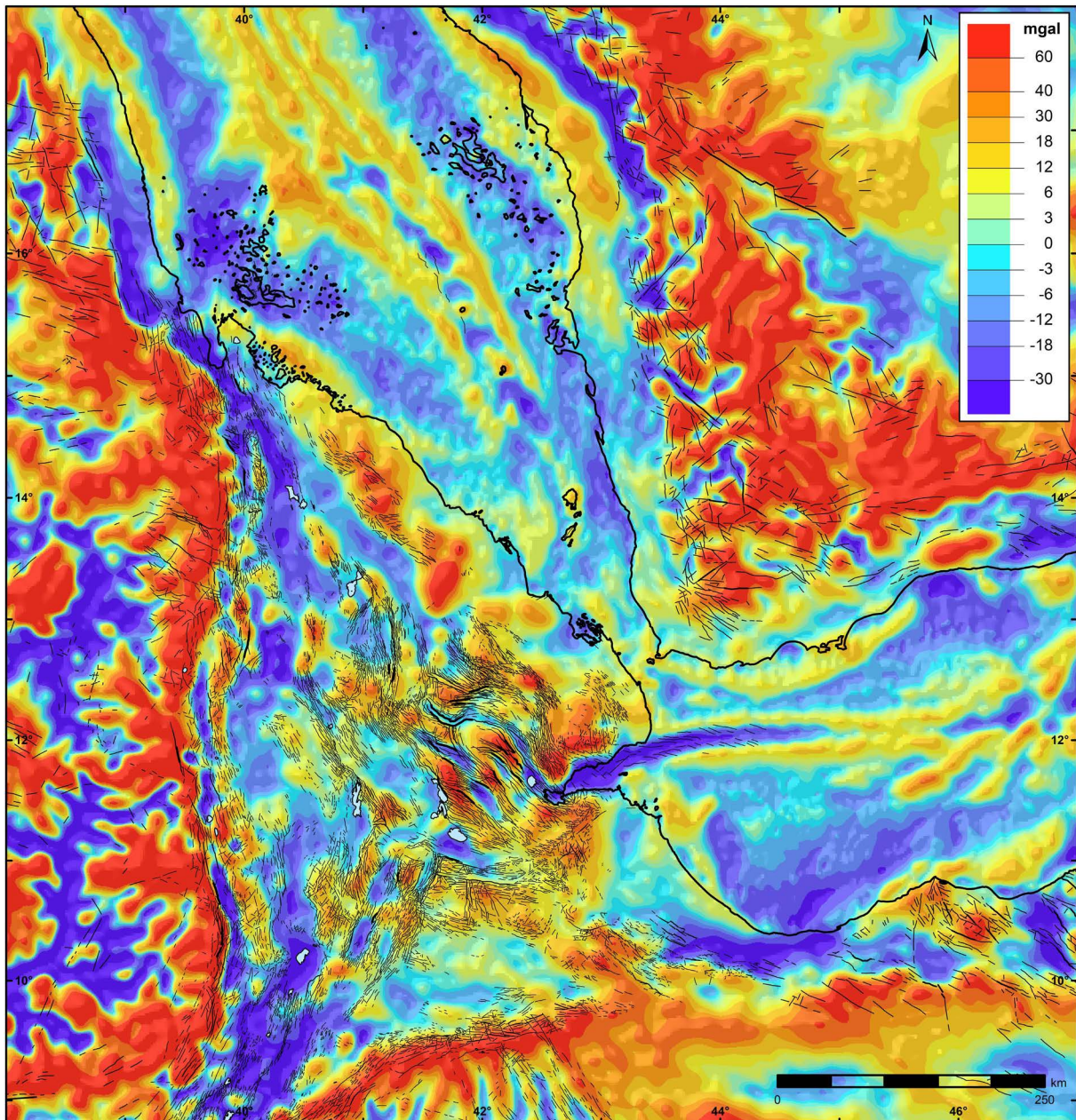
Supplementary material 2 ► : Most important syn-rift lithological units of the Afar Depression and margins. The green rectangles represent lava-dominated units, the yellow rectangles sediment-dominated units and the blue rectangles volcanic phases (which are not stratigraphical units *sensu stricto* but linked to different types of deposits and thus relevant for the reader). Note that the chronostratigraphy is often poorly constrained. Abbreviations: Ax. = Axial; B. = Basalts; E. = Early; Fm. = Formation; Gr. = Group; L.= lower La.=Late; M.= Middle; R. = Ranges; Re. = Recent; Rhy. = Rhyolites; S.= Series; Str. or Strat. = Stratoid; Tra. = Trachytes; U.= Upper. Sources: Geological Survey of Ethiopia (1973) (only units largely present in study area are shown on the figure); Varet *et al.* (1975) (R/Q. = Recent Quaternary volcanic Fm. U.S = Upper part of the Stratoid series.); Geological Survey of Ethiopia (1996) (only units largely present in study area are shown on the figure); Kidane *et al.* (2003); Rooney (2017, 2020b; a); Brinckmann & Kürsten (1971) (late quaternary sedimentary Fm.: Zariga Fm., Young terraces gravel and Afrera Fm.); C.N.R. - C.N.R.S. Afar Team (1973); Blanford (1869); Mohr (1965); Zanettin & Justin-Visentin (1975); Wolfenden *et al.* (2005) (different transects); Quade *et al.* (2008); Regional geological maps: (Geological Survey of Ethiopia, 2009a, 2010b, 2011); Stab *et al.* (2016); Morbidelli *et al.* (1975); Juch (1975) (with age limits of Mohr & Zanettin (1988); Regional geological maps: (Geological Survey of Ethiopia, 1978b, 1985); Berhe (1986); Chernet *et al.* (1998); Wolfenden *et al.* (2004); Chessex *et al.* (1975); Zumbo *et al.* (1995b); Le Gall *et al.* (2015) (most important units only).



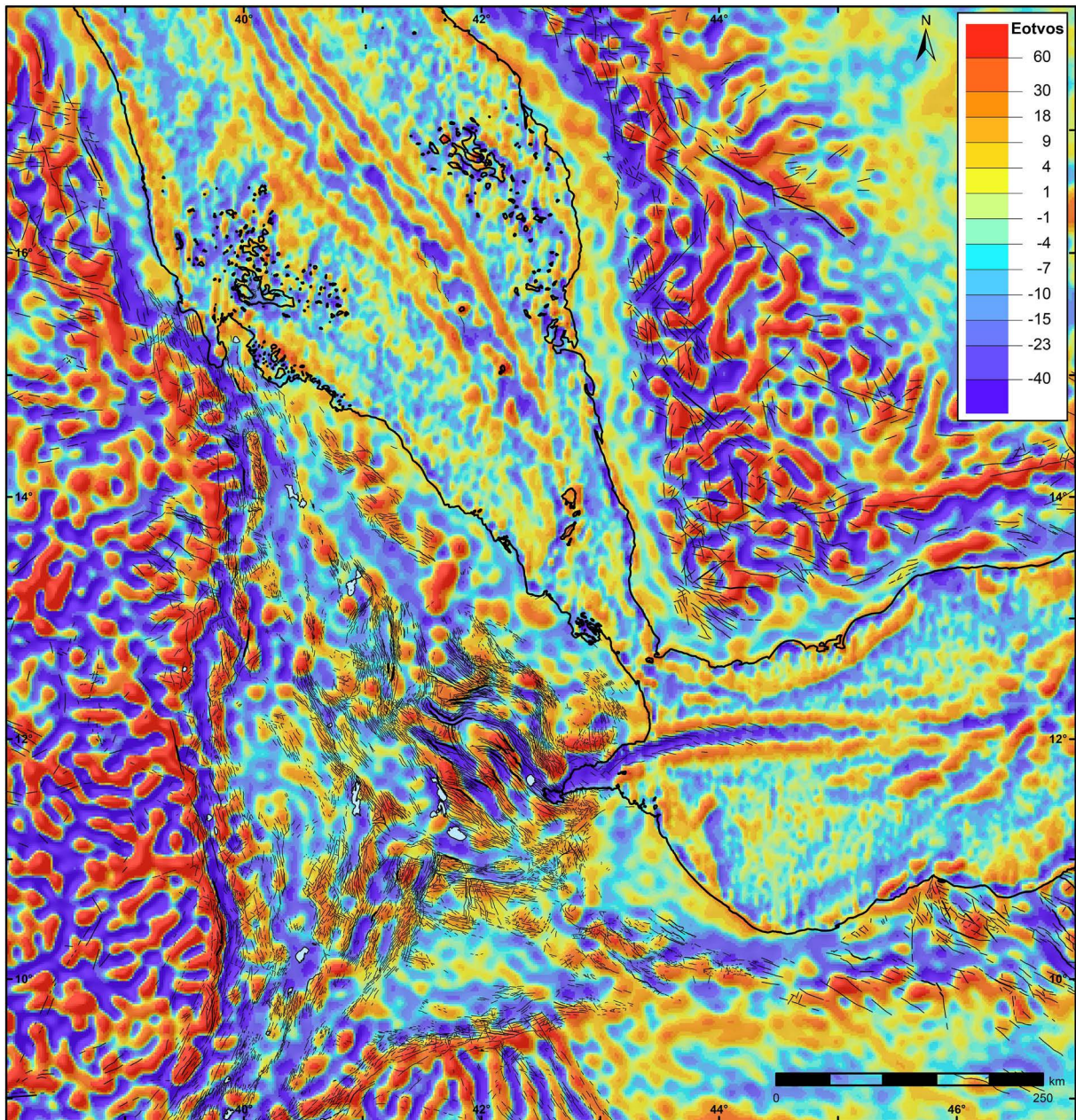
Supplementary material 3: Simplified basement map of the Red Sea and surrounding regions. Orange lines indicate Precambrian suture zones. Black lines indicate lineaments or faults. Full list of references is available in Table 3.



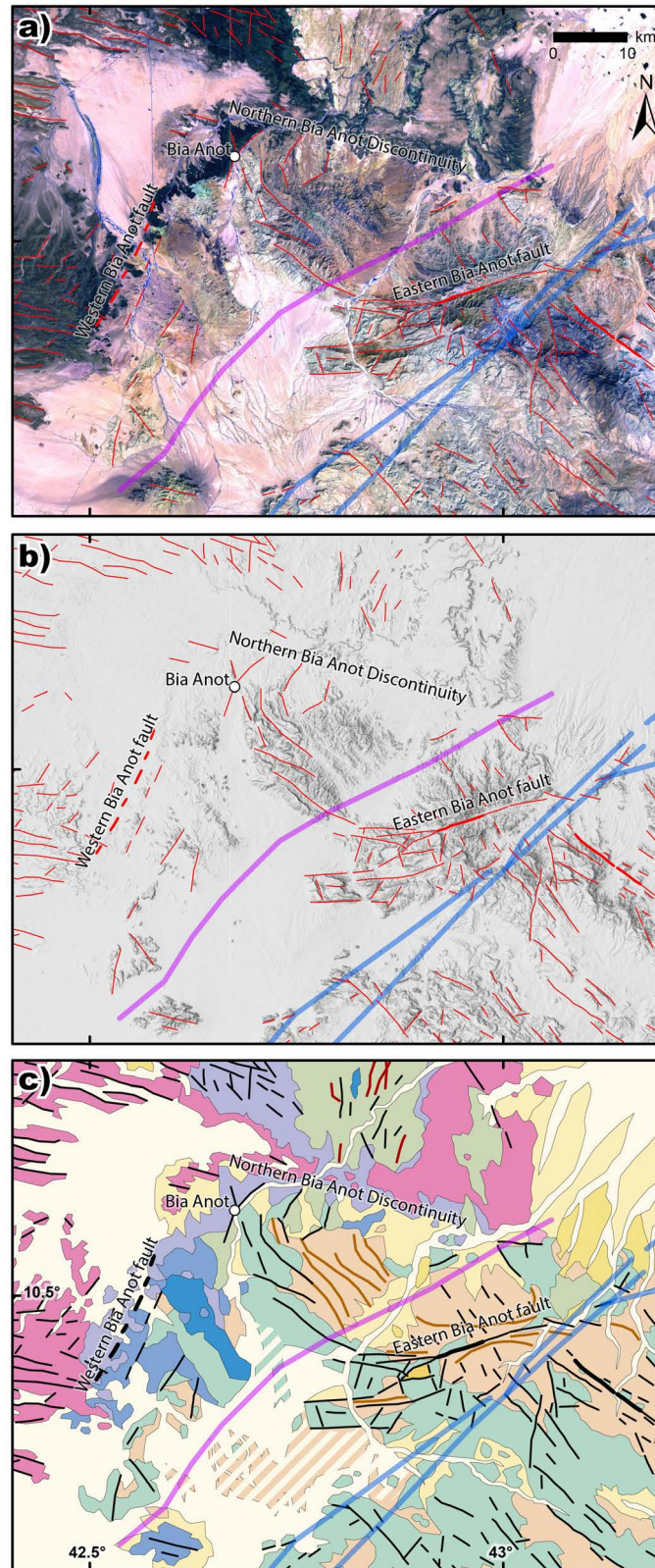
Supplementary material 4: Bouguer gravity anomaly map of the Afar. Geophysical data from WGM2012 (Bonvalot *et al.*, 2012). Faults from fig. 6.



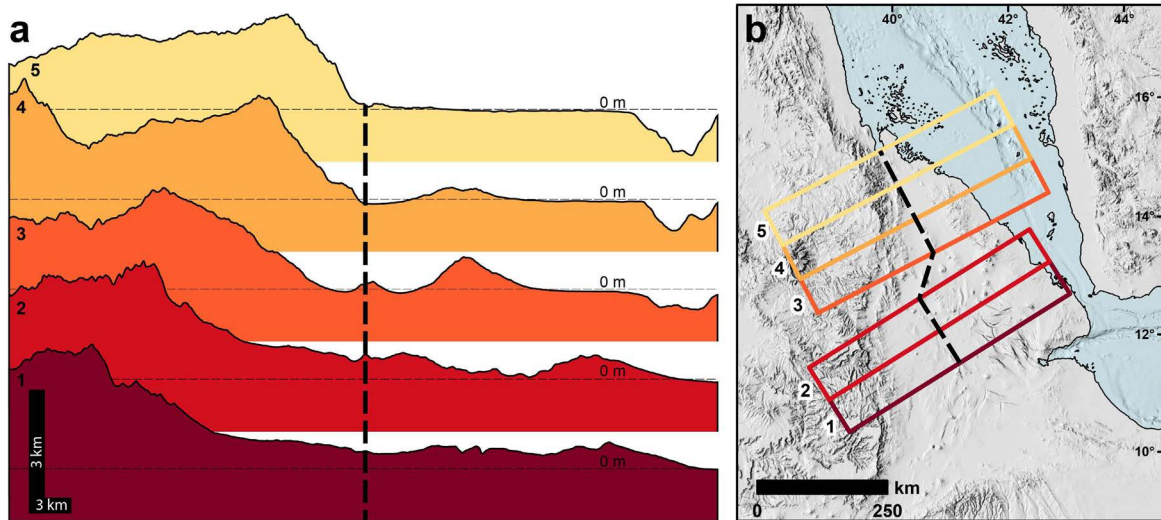
Supplementary material 5: Free air anomaly map of the Afar. Geophysical data from Sandwell *et al.* (2014). Faults from fig. 6.



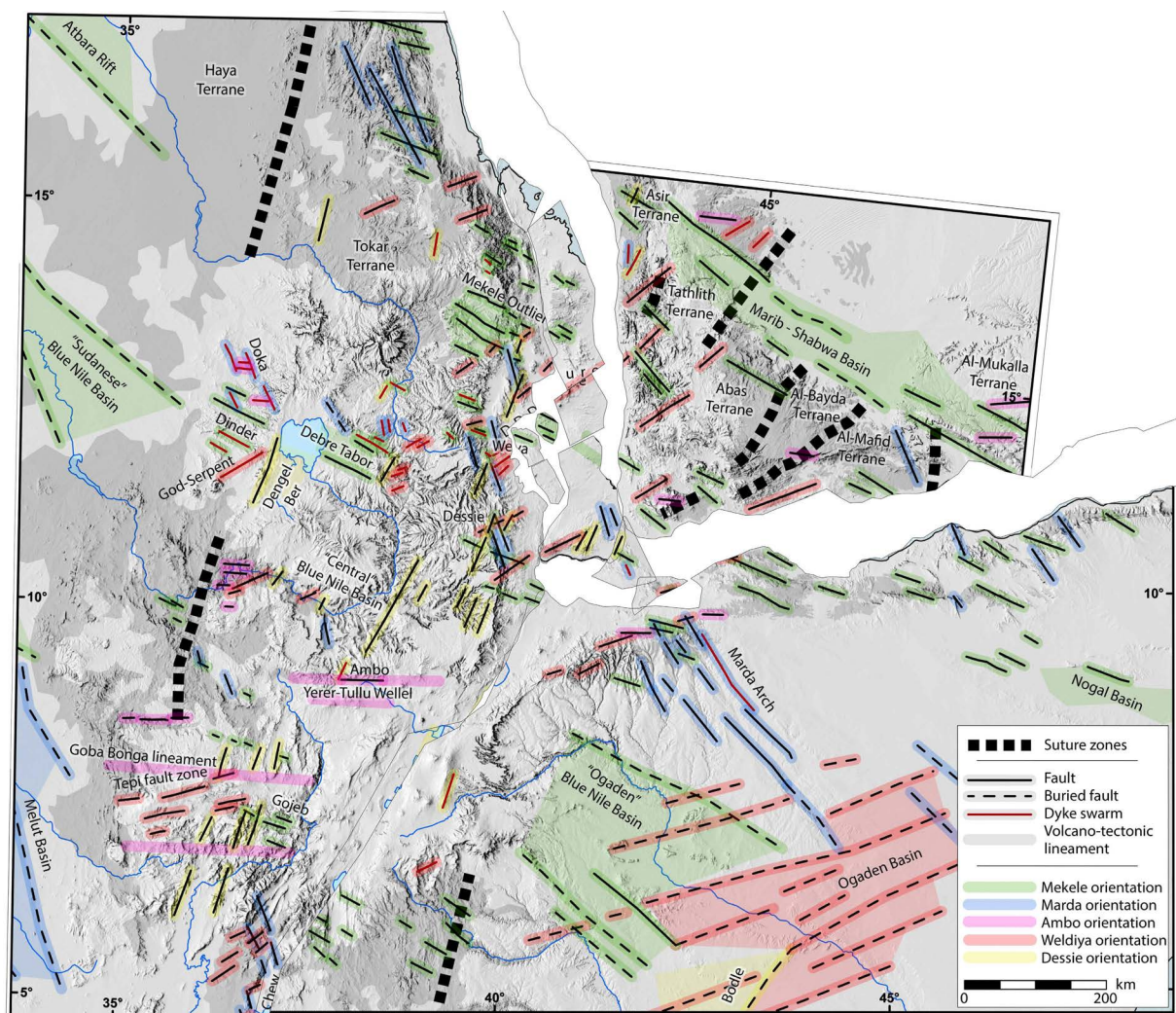
Supplementary material 6: Vertical gravity gradient map of the Afar. Geophysical data from Sandwell *et al.* (2014). Faults from fig. 6.



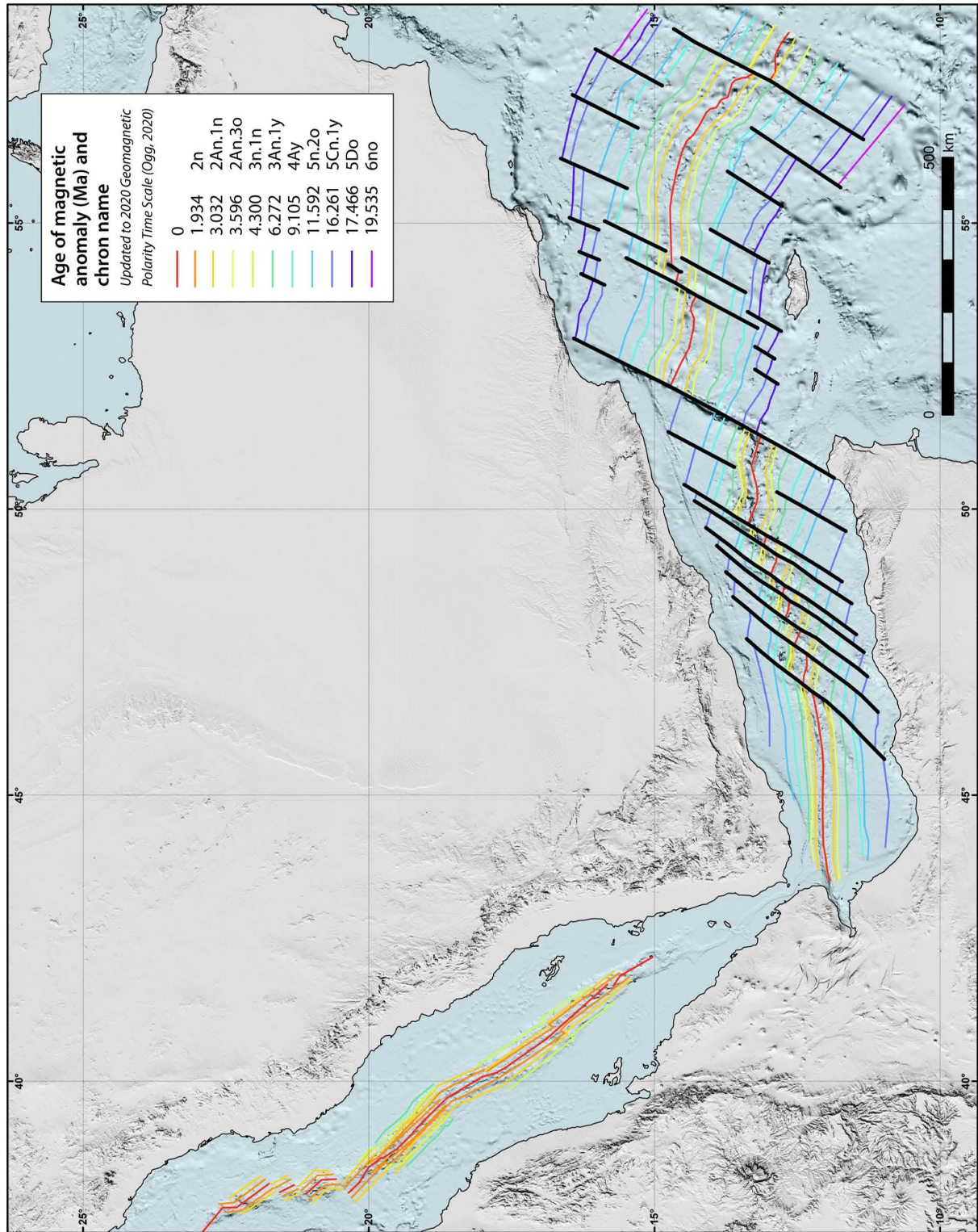
Supplementary material 7: Analysis of the Bia Anot region. a) Landsat image band 7-6-5. b) Hillshade of SRTM digital elevation model. c) Geological map (fig. 4). The large magenta and blue lines indicate the Bia Anot fault as proposed by Black *et al.* (1975) and Audin *et al.* (2004) respectively.



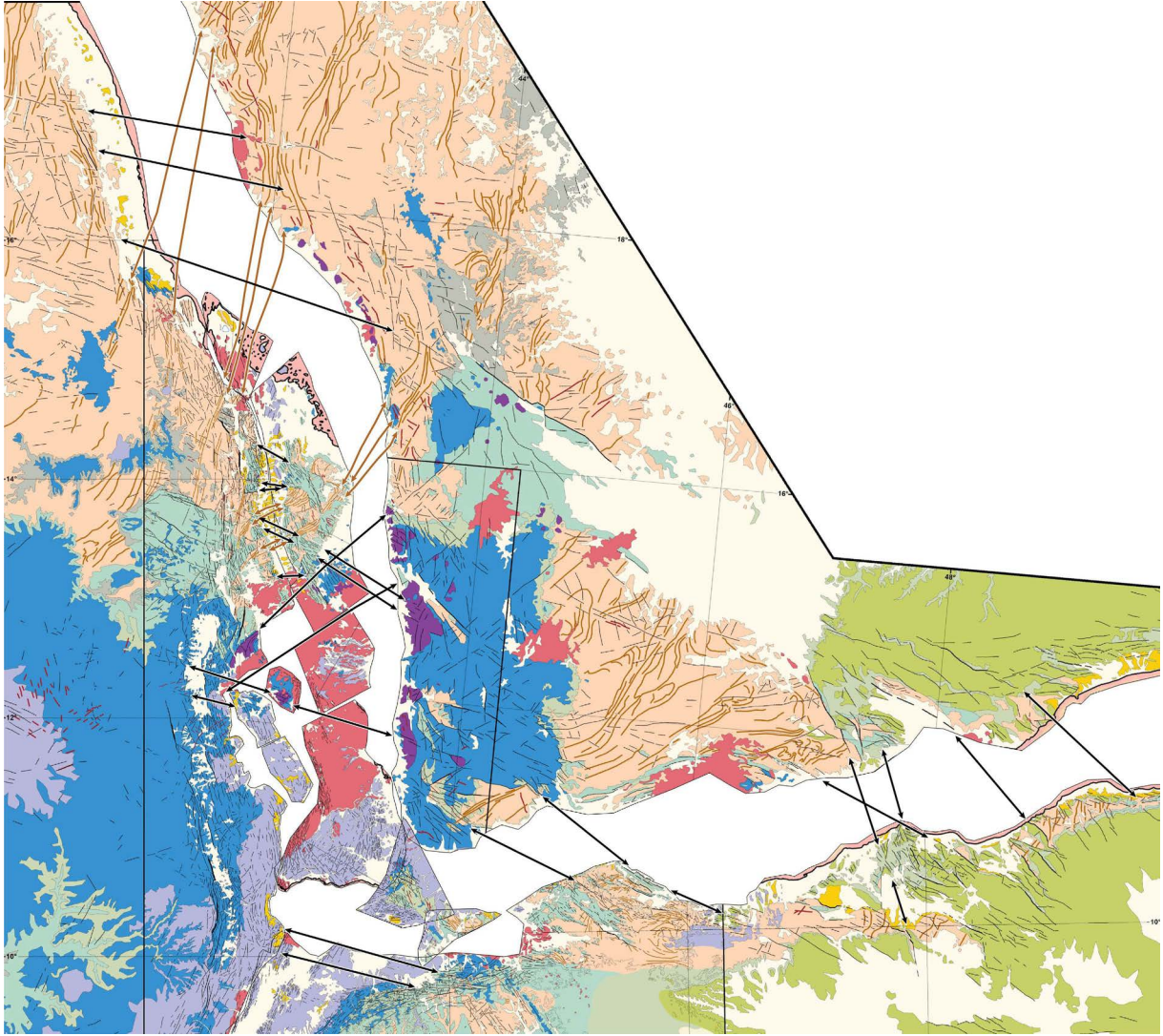
Supplementary material 8: Swath profiles across the Afar Depression. They show a steeper and less mature topography towards the north. Produced with Topographic Analysis Kit (TAK, Forte & Whipple, 2019).



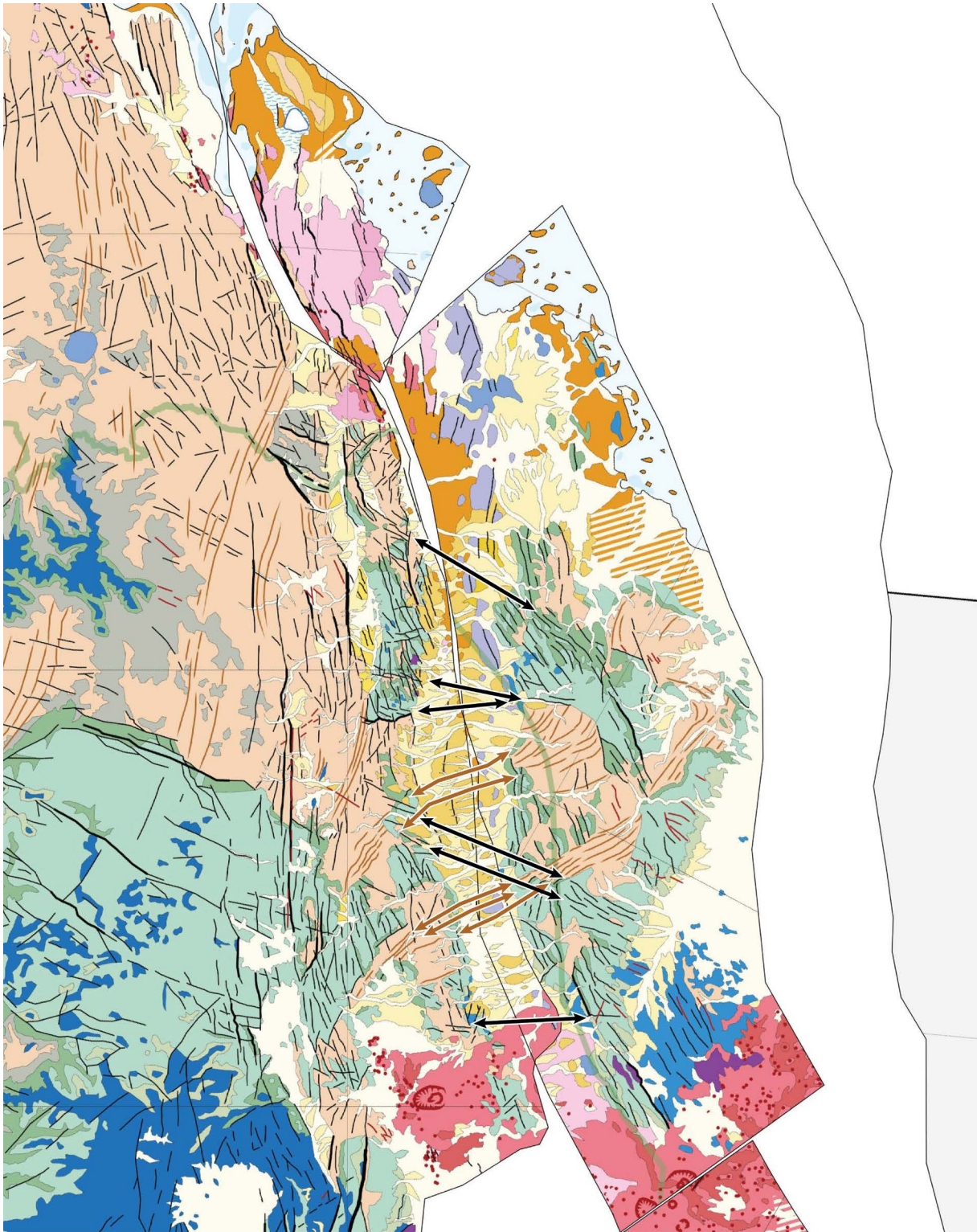
Supplementary material 9: Structural inheritance map restored to pre-rift configuration.



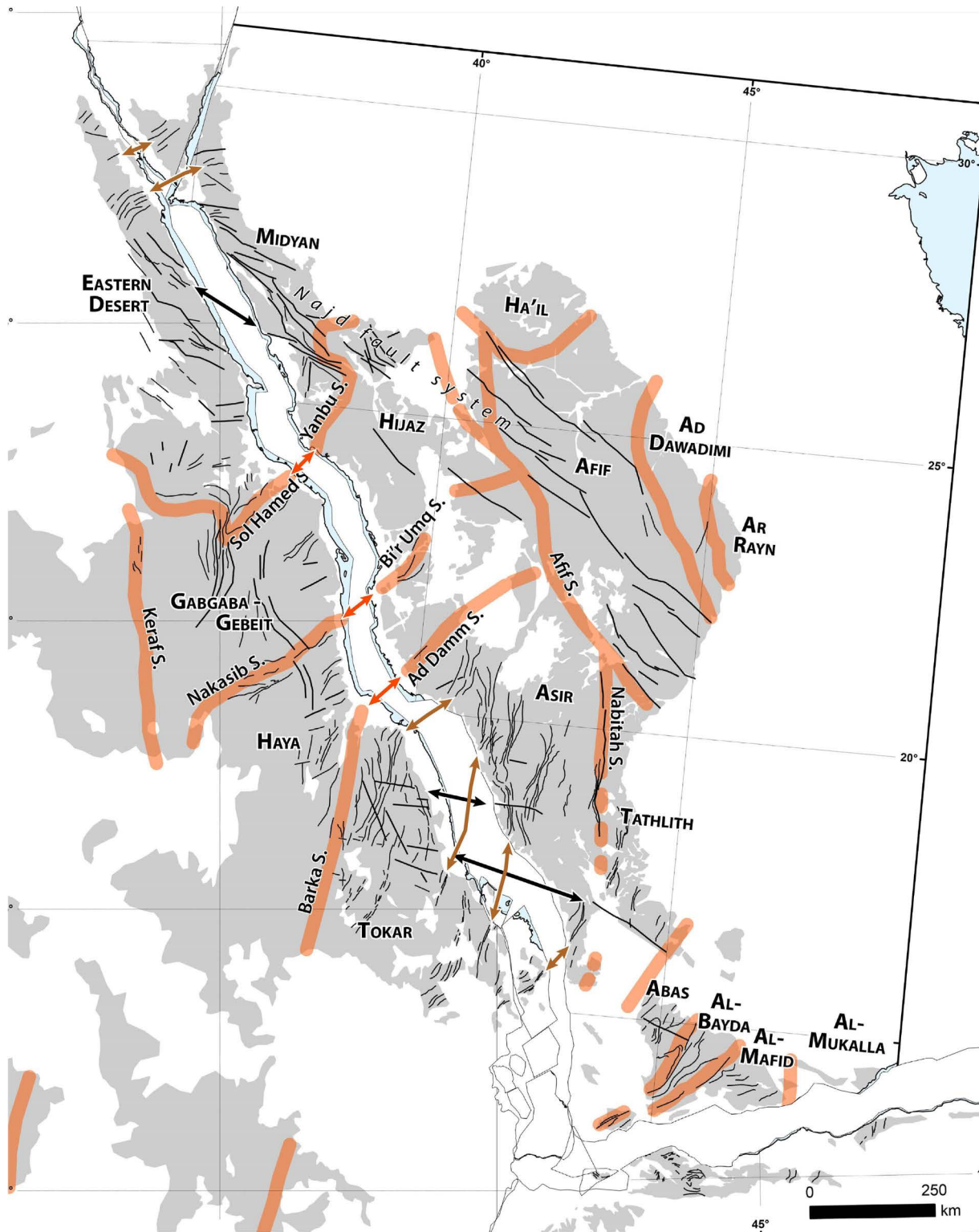
Supplementary material 10: Oceanic magnetic anomalies used in the kinematic model. Anomalies from Izzeldin (1987), Fournier et al. (2010) and Schettino et al. (2016).



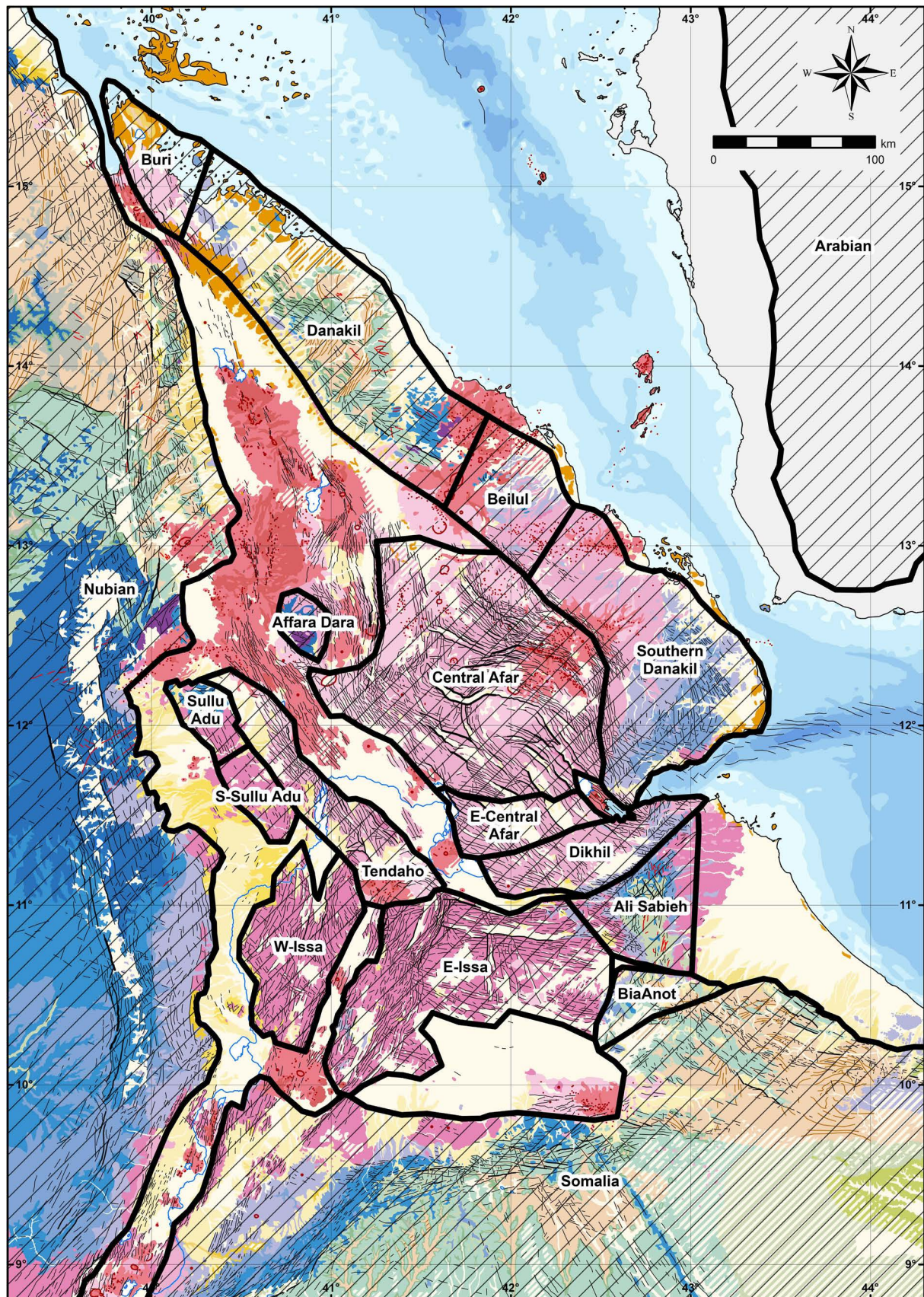
Supplementary material 11: Pre-rift correlation of structures in the Afar Region. Black arrows indicate potential matching fault patterns. Brown arrows indicate potential matching basement lineaments patterns. See fig. 6 for the legend of the map.



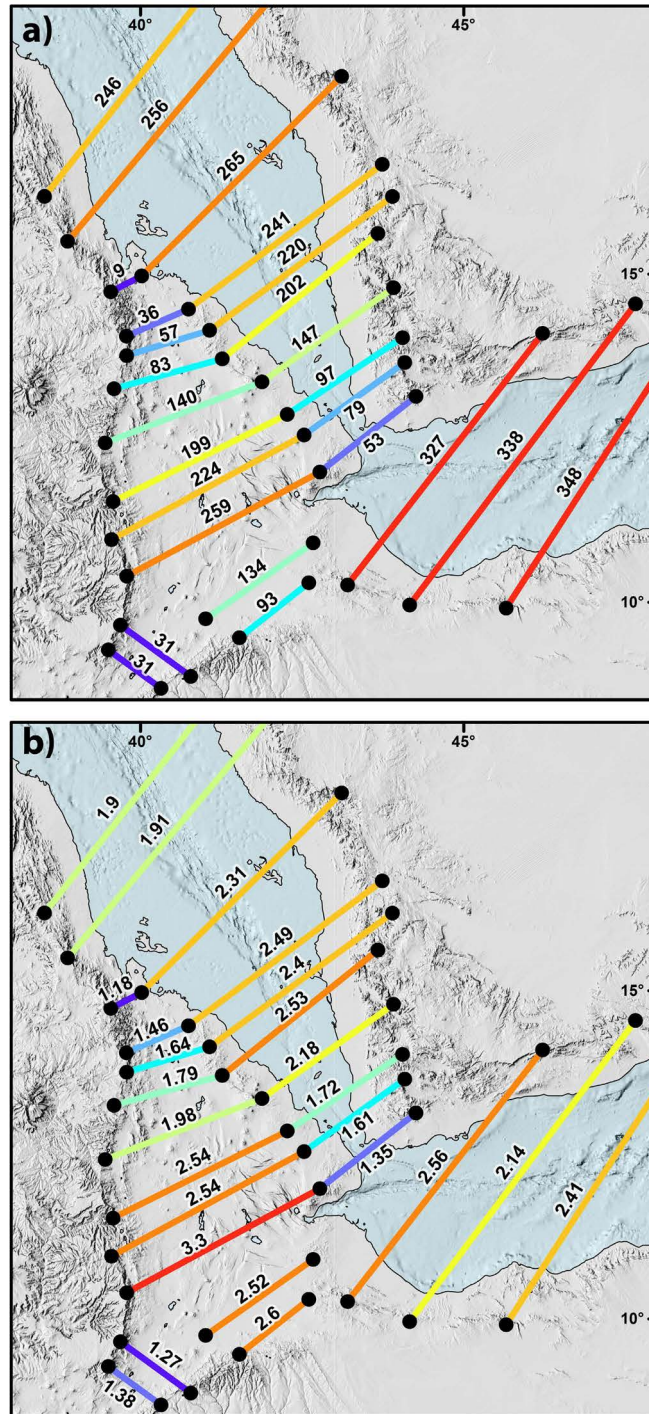
Supplementary material 12: Pre-rift correlation of structures in the Danakil Depression. Black arrows indicate potential matching fault patterns. Brown arrows indicate potential matching basement lineaments patterns. See fig. 4 for the legend of the map.



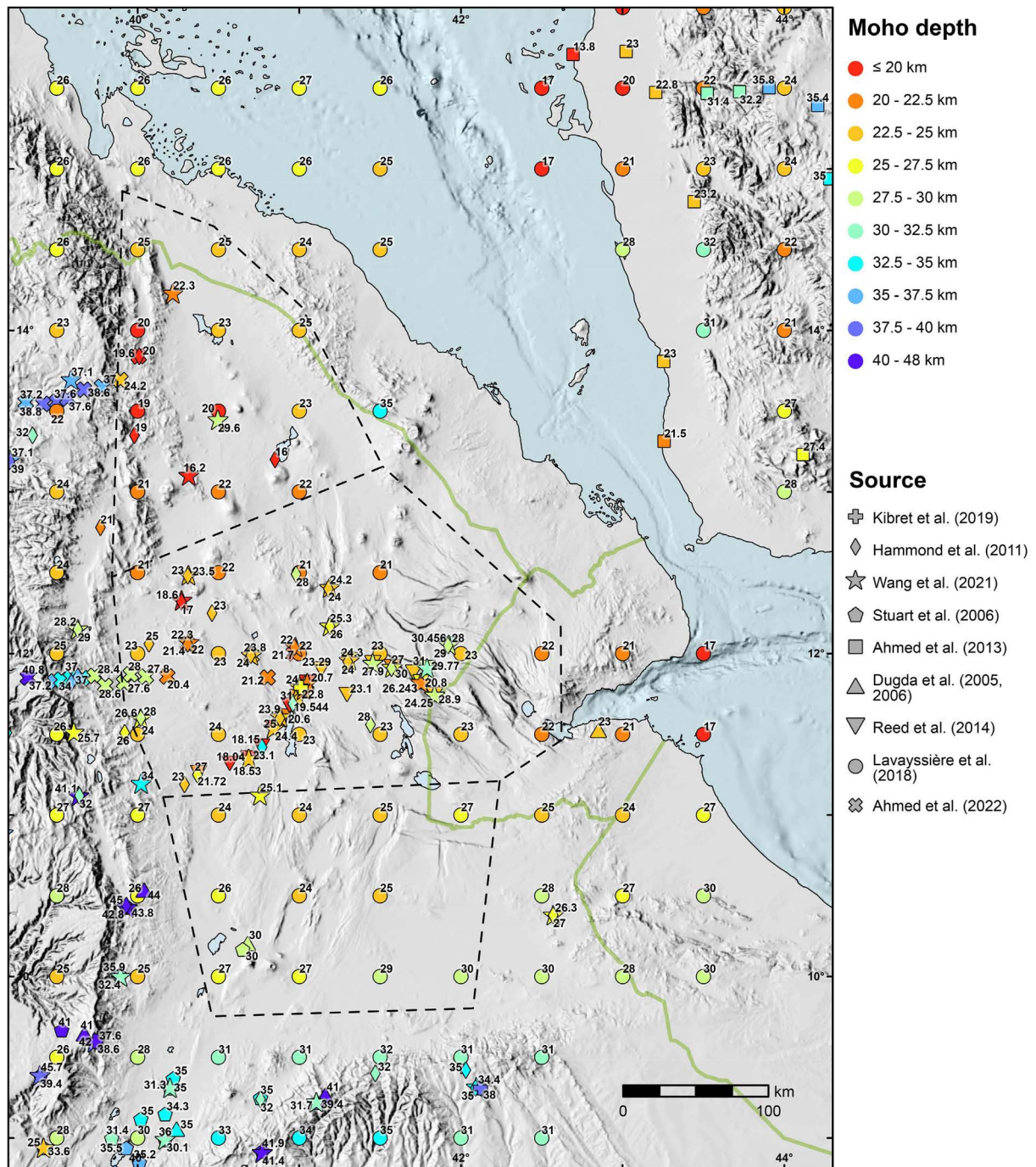
Supplementary material 13: Pre-rift correlation of basement structures. Black arrows indicate potential matching fault patterns. Brown arrows indicate potential matching basement lineaments patterns. Orange arrows indicate potential matching terrane boundary.



Supplementary material 14: Tectonic elements used for the GPlates reconstruction in Afar.
Shapefile is available in suppl. mat. 19.



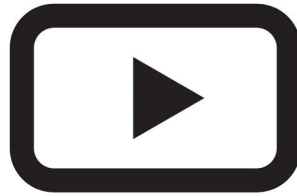
Supplementary material 15: Motion predicted by the kinematic model. a) Total extension (blue = low, red = high), values indicated in km. **b)** Stretching factor (blue = low, red = high). Note that this value is very sensitive to the position of the anchor points, and therefore the zone considered to be affected by the extension.



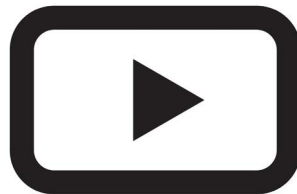
Supplementary material 16: Moho depth map. Compilation of Moho depths published in the literature (Dugda et al., 2005; Dugda & Nyblade, 2006; Stuart et al., 2006; Hammond et al., 2011; Ahmed et al., 2013, 2022; Reed et al., 2014; Lavyssière et al., 2018; Kibret et al., 2019; Wang et al., 2021). Numbers indicate Moho depth in km. Dashed lines indicate the areas considered in section 7.1 and Table 4 for calculating of the average Moho depth.

The following supplementary material are available in the numerical version of this thesis or on the website of Earth-Science Reviews :

<https://doi.org/10.1016/j.earscirev.2023.104519>



Supplementary material 17: Simplified schematic and didactic wood model showing the main concept of rift propagation in the Afar.



Supplementary material 18: Kinematic model of the evolution of the Afar Depression and surrounding regions.



Supplementary material 19:] Kinematic model files. GPlates rotation file and shapefile of the tectonic elements.

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III THE FLOODING OF A YOUNG RIFT BASIN: PAST AND FUTURE OF THE DANAKIL DEPRESSION, AFAR

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ABSTRACT

The Danakil depression is a rare example of an active rift at the transition from continental to marine conditions, a crucial stage in rift and passive margin development, as well as for the development of salt giants. Recent field expeditions shed new light on the basin dynamics. We show strong uplift of the rift shoulders (> 1 mm/yr) in an ultra-slow rift setting. The flooding cycles of the Depression are controlled by the interaction of eustatic sea-level variations, tectonic processes, sedimentation and volcanic events, resulting in the deposition of thick sedimentary units during very short time periods. The two last interglacials are marked by Red Sea incursions. Uplift of the margins prevented flooding during the Holocene but anthropogenically-induced sea-level rise might cause flooding within six centuries. On longer timescales, continuous rifting will not only affect further flooding and desiccation cycles but also relocate of the gateway to the Gulf of Zula.

III.1 INTRODUCTION

Continental breakup is one of the most important features of global plate tectonics. The transition from continental rifting towards the development of ocean basins is a complex process that is not yet fully understood. The understanding of the development and nature of rift and passive margins is very important as they host natural resources, have strong potential for geothermal energy production and are unique paleoenvironmental and paleoclimatic archives. A critical stage in this evolution is the transition between continental and marine conditions. This phase of rifting is often complex but essential to understand the stratigraphic architecture of future passive margins. They often produce organic-rich sediments and voluminous evaporites. Many examples exist in the Atlantic, in the Gulf of Mexico and in the Red Sea but the complexity of the processes and the lack of well-studied Recent analogues render their interpretation difficult. In particular, the quantification of the rates of tectonic and sedimentological processes in older passive margins is very difficult. One of the few modern examples of a rift undergoing transition from continental to marine conditions is the Danakil Depression in the Afar which was flooded several times by the Red Sea during the Pleistocene (Jaramillo-Vogel et al., 2019; Foubert et al., in prep.). The study of this recent and active site is therefore precious to understand the processes that shape rift basins and passive margins and to quantify the rates of geological processes in such settings.

The Afar depression is a complex triple junction formed by the interaction of the Red Sea, Gulf of Aden and East African

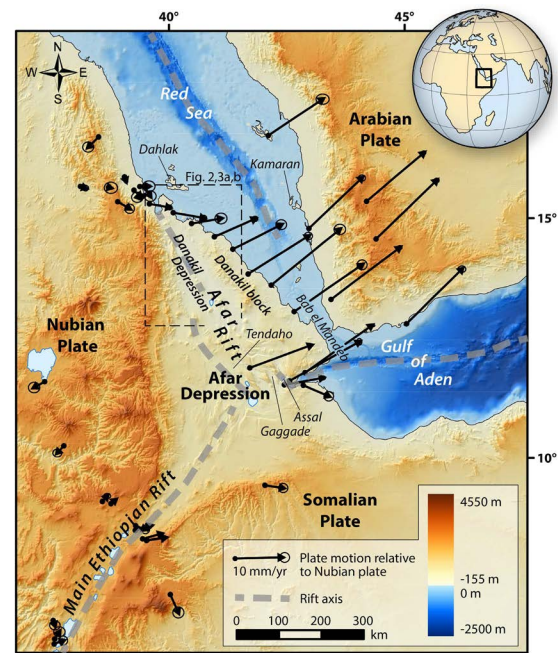


Figure 1: Topographic map of the Afar triple junction. The GNSS velocities (McClusky et al., 2010) show the anticlockwise rotation of the Danakil block.

rift systems (Mohr, 1970; Tazieff et al., 1972; Barberi & Varet, 1977; McClusky et al., 2010; Varet, 2018; Viltres et al., 2020; Rime et al., 2023; chapter II, this volume; Fig. 1). The northward propagating Afar Rift results from the counter clockwise rotation of the Danakil block with respect to the Nubian plate since approximately 10 Ma (Souriot & Brun, 1992; Eagles et al., 2002; McClusky et al., 2010; Reilinger et al., 2015; Rime et al., 2023; chapter II, this volume). The Danakil Depression forms its northern part and is considered as an analogue for magma-rich margins. It currently experiences between 19 (in the south) and 6 (in the north) mm/yr full rate extension (McClusky et al., 2010; Viltres et al., 2020; fig. 1), considered as slow to ultra-slow. Recent thinning of the crust by mechanical extension triggered decompression melting and voluminous basalt flows in the Danakil Depression, the latter is being considered as an analogue for the formation of seaward

dipping reflectors on passive margins (Bastow & Keir, 2011; Keir *et al.*, 2013; Le Gall *et al.*, 2018; Watts *et al.*, 2020; Rime *et al.*, 2023).

The elevation of the Danakil depression reaches 125 m below sea level. Flooding of the depression is testified by the presence of thick (>975 m) evaporite deposits in the centre of the basin (Holwerda & Hutchinson, 1968; Bastow *et al.*, 2018) and Pleistocene fringing reef terraces and gypsum deposits around the margins of the depression (Faure & Roubet, 1968; Holwerda & Hutchinson, 1968; Lalou *et al.*, 1970; Bonatti *et al.*, 1971, 1972; Brinckmann & Kürsten, 1971; Beyth, 1978; Atnafu *et al.*, 2015; Jaramillo-Vogel *et al.*, 2019; Foubert *et al.*, in prep.; Fig. 2).

The scarcity of field studies since the 70s' prevented the detailed understanding of the stratigraphic history of the Danakil basin. The timing of flooding, the localisation of the gateway and the mechanisms leading to flooding and desiccation remained unsolved and controversial (Holwerda & Hutchinson, 1968; Frazier, 1970; Hutchinson & Engels, 1970; Lalou *et al.*, 1970; Bonatti *et al.*, 1971; Brinckmann & Kürsten, 1971; Duffield *et al.*, 1997). During the last decade, scientific studies in the area accelerated again due to improved field access and facilities. Extensive fieldwork resulted in the detailed mapping, description and sampling of Quaternary sediments in the depression to determine the flooding history of the basin (6 field expeditions from 2013 – 2020). This allowed understanding the occurrence of corallgal reefs, oolitic shoal and beach complexes, shallow marine deposits and gypsum at the basin margins and forming the Zariga Fm. (Jaramillo-Vogel *et al.*, 2019;

Endeshaw *et al.*, in prep.; Foubert *et al.*, in prep.; Negga *et al.*, in prep.). Up to four distinct corallgal reef terraces were observed (Supplementary Fig. 1). The two youngest terraces were dated to Marine Isotopic Stage (MIS) 5e (~130 ka) and MIS7 (~200 ka) by Jaramillo-Vogel *et al.* (2019) and Foubert *et al.* (in prep.), while the alteration of the two oldest terraces prevented any dating until now. Most of the outcropping corallgal reef terraces were uplifted or subsided by tectonic movements. This study focusses on the elevations of the corallgal reef terraces based on GNSS measurements combined with geological mapping, remote sensing and topographic analysis to quantify Pleistocene to Holocene tectonic movements along the Depression and its impact on the flooding history of the basin. Based on those results, the evolution of the Depression is discussed, as well as its risk for future flooding due to anthropogenic sea-level rise.

III.2 METHODS

III.2.1 Field investigations

Field exploration was conducted in several regions of the Danakil depression, from the east of the Tat'Ali range to the northern tip of western margin of the Danakil depression. The age of the reef terraces was determined based on U/Th dating on pristine coral skeletons by Jaramillo-Vogel *et al.* (2019) and Foubert *et al.* (in prep.) combined with mapping and lithological logging of sedimentary units allowing the distinction of well-defined lithofacies. Distinct corallgal reef units were dated (absolute and relative) as MIS5e and MIS7 reefs. Other terraces were considered as of unknown age. The position of the paleo shoreline was determined based on sedimentological

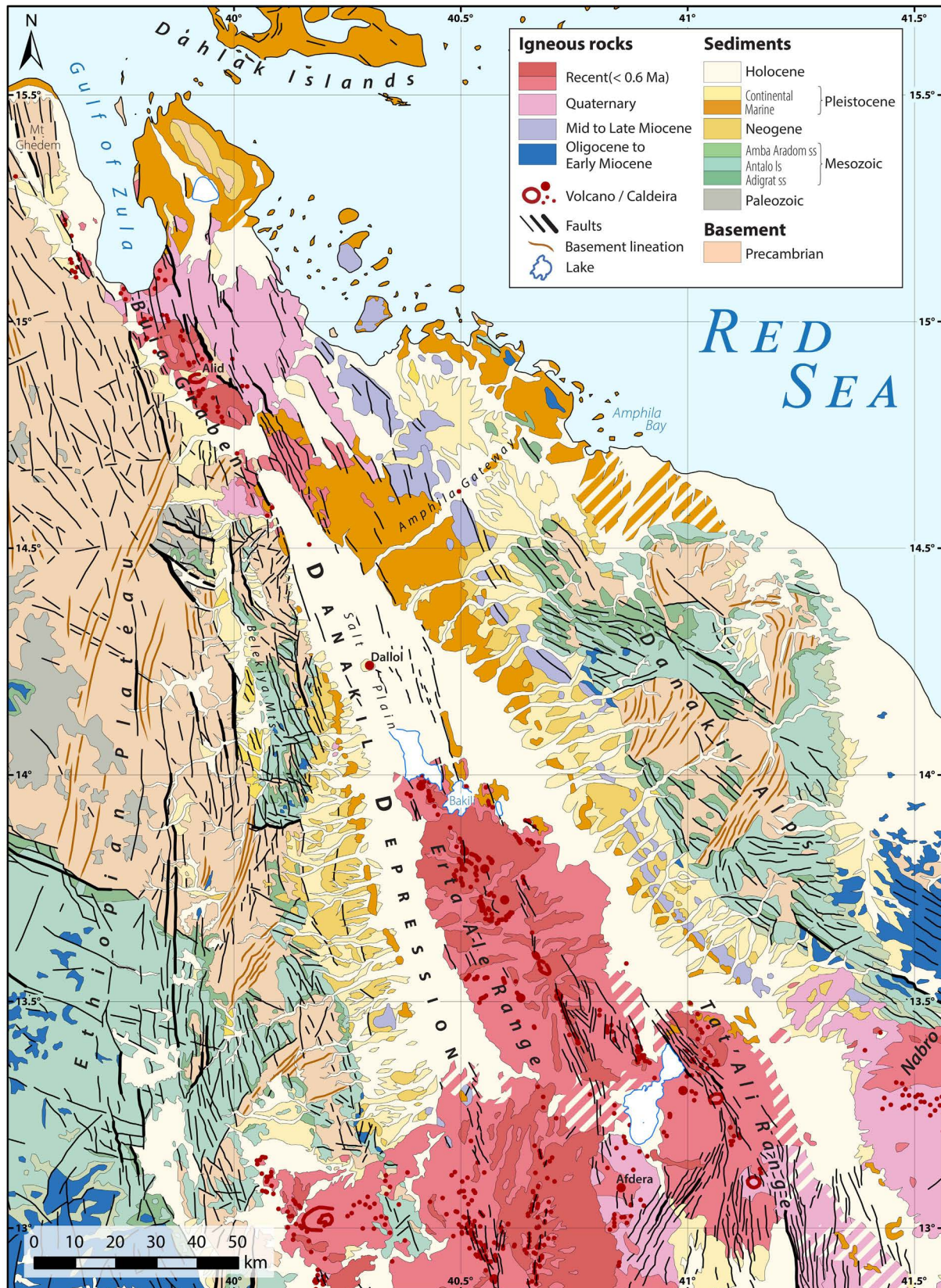


Figure 2: Geological map of the Danakil depression. This map is part of a larger map of the Afar Depression (Rime *et al.*, 2022). A list of sources and the detailed description of the methodology is available in Supplementary Methods 2.

criteria, such as the presence of beach facies, beach-related facies or reef flat deposits. For each outcrop, the sedimentological uncertainty was determined, corresponding to the uncertainty of the position of the paleo shoreline relative to the observed sediments. The height of distinct sedimentary units/marine sediments were measured with a Garmin GPSMAP 66i or with a Leica Viva GS10, used either as a differential GPS or as conventional GPS. The elevations of one outcrop were on a 3D model acquired through a Wintra drone survey using a Trimble R2 GNSS receiver as base station. More details are available in the Supplementary Methods 1. All elevation data are presented as height above Geoid (EGM96), thus representing elevations above mean sea level.

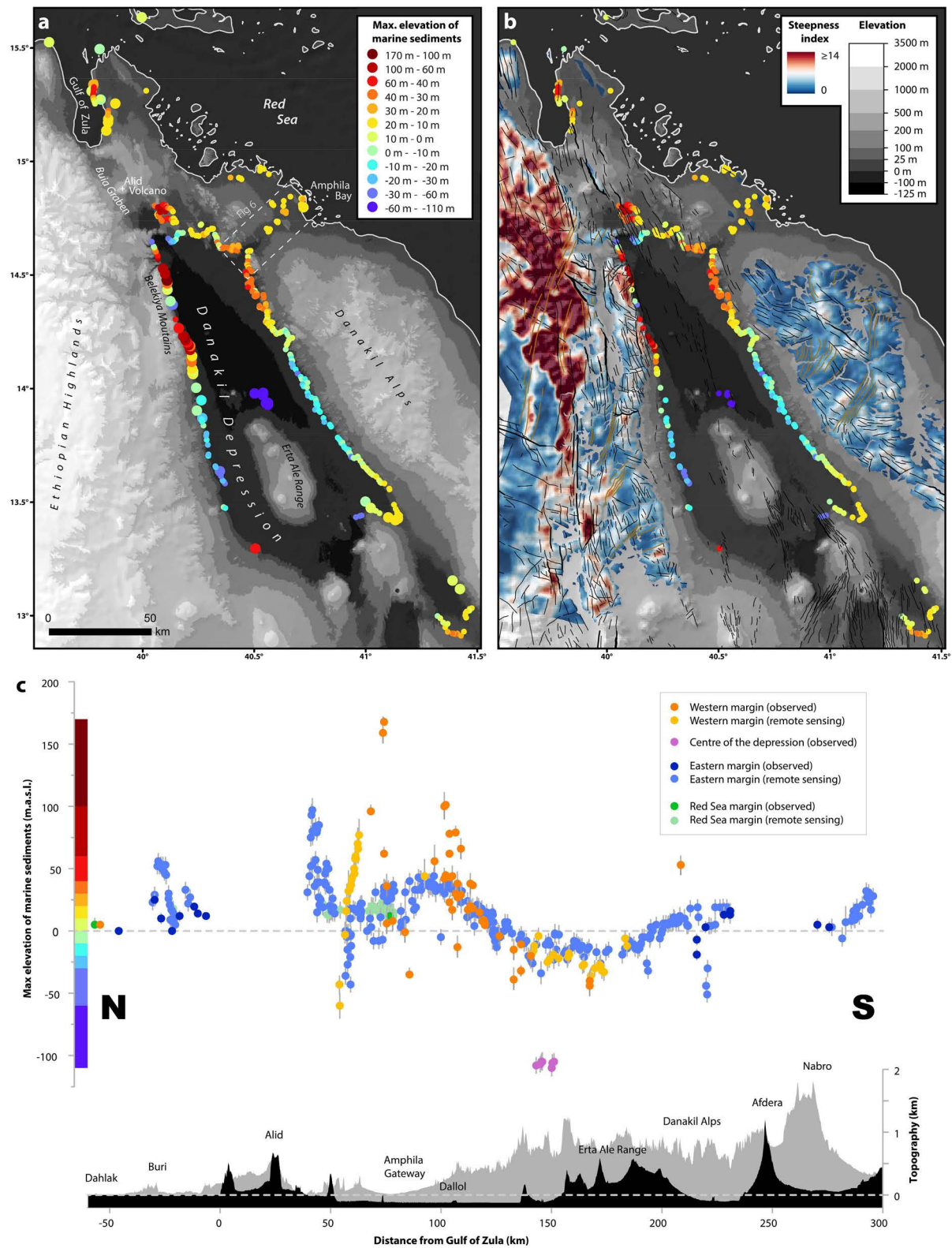
We also reviewed the elevation of marine sediments described in the literature. We only considered information that presents clear paleo shoreline indicators and precise location. These described outcrops are situated around the Gulf of Zula (Bruggemann *et al.*, 2004; Buffler *et al.*, 2010; completed by pers. comm. with J. H. Bruggemann), on the Dahlak Islands (Conforto *et al.*, 1976), in the Amphila Bay (Holwerda & Hutchinson, 1968), in the vicinity of lake Bakili in the centre of the depression (Barberi & Varet, 1970) and in the southernmost outcrops of the western margin (Brinckmann & Kürsten, 1971).

III.2.2 Remote sensing

We used remote sensing to extrapolate field observations to other parts of the depression where no fieldwork was possible because of logistical issues. Multispectral Landsat 8 imagery and other publicly-available aerial imagery

from different providers (Google Earth, Bing Maps) were used for geological mapping. Ground-truthing on the sediments observed along the western margin allowed to distinguish the different spectral signatures of gypsum, carbonate and siliciclastic sediments. The highest position of either gypsum or carbonate sediments around the depression was mapped (Fig. 3) and their elevation was determined with SRTMGL1 (1 arc-second) Digital Elevation Model (details in Supplementary Methods 3). We also use this technique for a few outcrops we visited but where we could not measure the elevation or when no elevation value but precise descriptions of an outcrop were given in the literature.

We performed a topographic analysis of the basin with TAK (*Topographic Analysis Kit*, Forte & Whipple, 2019) using the SRTMGL1 elevation dataset. The normalized steepness index for streams flowing in the depression and in the Red Sea was calculated. Due to the high number of rivers analysed, the results are smoothed on the map for visualization (fig. 3b). The analysis of 9 important river profiles in the region showed that a reference concavity of 0.3 is a suitable fit for the river profiles. This is lower than the typical value of 0.45 but such low concavities have been also reported by Sembroni *et al.* (2017, 2021) in the nearby Mekele Outlier and Tekeze basin. For comparison, the same analysis was performed with a concavity value of 0.45, showing exactly the same trends except for a positive anomaly along the longest rivers. As mentioned in the results section, Fig. 3b only represents data for river systems on pre-rift bedrock. A complete representation of the computation is presented in



◀ **Figure 3: Distribution of marine sediments in the Danakil depression.** Each point represents the maximum elevation of marine sediments. DEM is the SRTMGL1. **a)** Small points represent remotely interpreted outcrops. Larger points represent outcrops visited by the authors or described in the literature. **b)** Results of the topographic analysis with steepness index (ksn) represented only for pre-rift rocks. Note that the values are lower than usual as a concavity of 0.3 has been used. Full results in Supplementary Fig. 2. Black lines represent faults and brown lines basement lineations as represented on Fig. 2. **c)** Vertical profile through the depression with maximum elevation of marine sediments projected on the rift axis. The datapoints are distinguished between remotely interpreted outcrops and outcrops visited by the authors or described in the literature. When several generations of sediments were observed in one outcrop, each generation is represented. Grey bars indicate standard uncertainty. The black profile represents the topography along the rift axis and the grey profile the topography on the Danakil block projected on the rift axis.

Supplementary Fig. 2. Swath profiles along the western margin of the depression were also calculated with TAK.

III.2.3 Uplift and uplift rates

Based on the measured paleo shoreline elevations and the age of the sediments, we calculated uplift and uplift rates (Fig. 4). The relative sea-level is taken from Grant *et al.* (2014). As their values are relative to the nearby Hanish Sill, no additional glacio-hydro-isostatic corrections were necessary. Note that the results presented here are therefore only local variations relative to the Red Sea area. These calculations are based on the 128 ± 2 ka peak for MIS 5e, with a relative Red Sea-level of 5 ± 4 m. As it is unknown to which exact peak the MIS 7 sediments correspond, we used an average value of 219 ± 23 ka, comprising the MIS 7a, c and e peaks. MIS 7a and 7e relative sea-levels were at -12 ± 4 m and MIS 7c was at -27 ± 4 m. We therefore used an average value of -19 ± 11 m. Detailed calculation methodology, including uncertainty, is presented in Supplementary Methods 3.

We calculated a possible Amphila gateway elevation curve by analysing the elevation of marine sediments around the gateway. Gypsum and carbonates were differentiated by their different spectral signals on Landsat images. The vertical

movements of the gateway were then calculated based on assumptions about the age of the deposits explained in the discussion. The future gateway elevation was calculated assuming a constant uplift rate since MIS5e. The details of the calculations are available in Supplementary Methods 4.

III.3 RESULTS

The analysis of the elevation of marine sediments shows a maximum amplitude of 280 m (Fig. 3, Supplementary Tables 1, 2 and 3). The lowest elevations (-110 m) are found around lake Bakili, NE of the Erta Ale range and the highest elevations (170 m) are located on the northern part of the western margin. Uplift and uplift rates are calculated (Fig. 4, Supplementary Table 1). In general, subsidence is observed for outcrops situated towards the centre of the basin while varying uplift values and rates are observed at the margins. Variations can also be observed along the axis of the rift. While uplift is moderate at the margins in the south, subsidence is observed at the longitude of the Erta Ale range. Uplift values gradually increase towards the northern tip of the Danakil depression and are moderate around the Gulf of Zula. Elevations are constant and modest along the Red Sea coast. The highest uplift values are observed on the northern western margin. Results for the

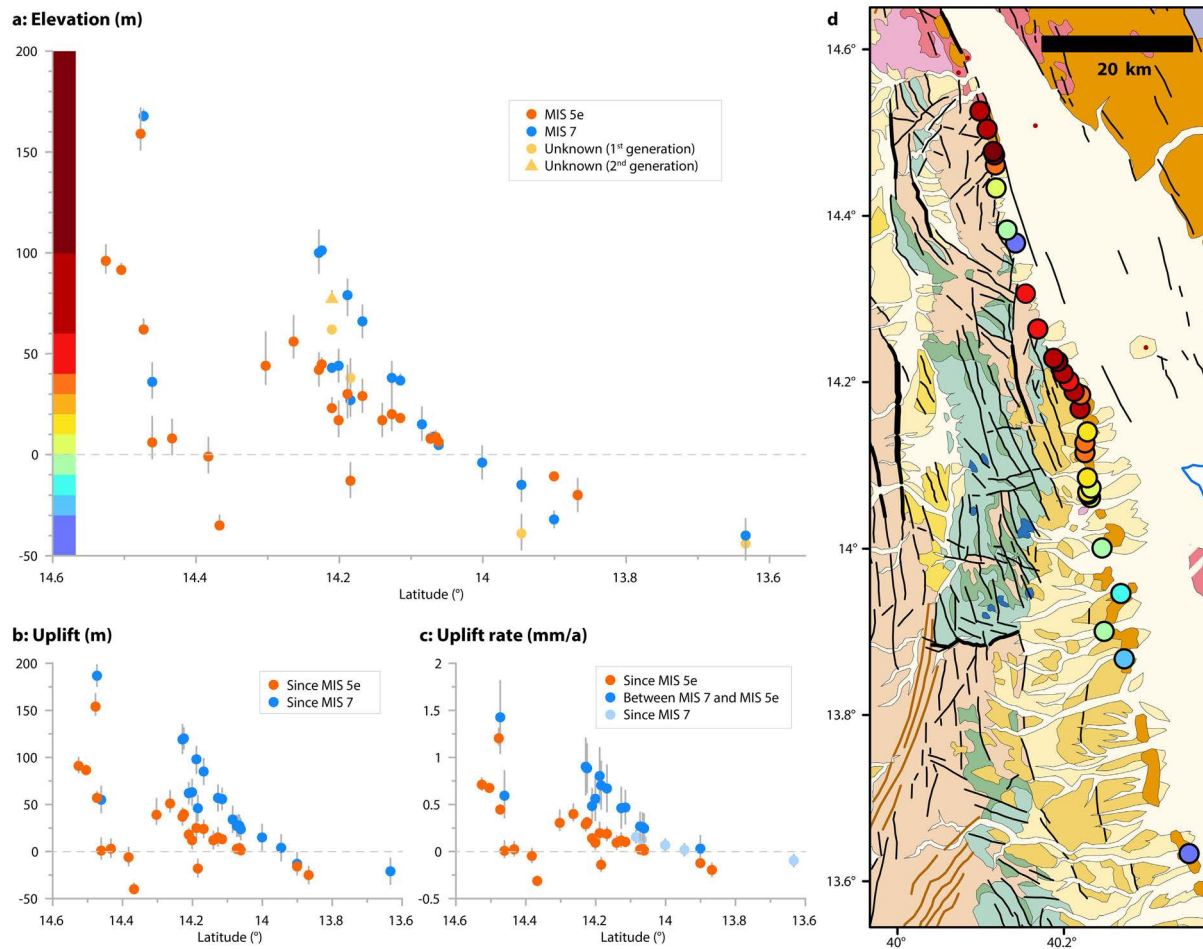


Figure 4: Analysis of paleo shorelines of the western margin. Data shown here represent outcrops studied in the field and discriminated by age. Grey lines represent standard uncertainty **a)** Elevation of the measured paleo shorelines of different generations. **b)** Uplift since deposition, assuming relative sea level of $+5 \pm 4$ m for MIS 5e and -19 ± 11 m for MIS 7. **c)** Uplift rates since MIS 5e, between MIS 7 and MIS 5e sediments and since MIS 7 (only for outcrops without MIS 5e sediments). **d)** Geological map of the western margin. See fig. 2 for the legend. The points represent the highest elevation of the measured paleo shorelines in each outcrop. Color scale identical to Fig. 3 and represented on panel a). Methodological details are available in Supplementary Methods 3.

western margin also show that the uplift pattern is irregular, particularly on the heavily faulted northern part of the margin (Fig. 4).

Uplift rates are varying significantly through time, being smaller between MIS 5e and present than between MIS 7 and MIS 5e. The maximum uplift rate reaches 1.43 mm/yr, with the bulk of the values comprised between 0 and 1 mm/yr.

The steepness index of streams and rivers are used as an independent proxy for relative rock uplift rate (Kirby &

Whipple, 2001; Whipple, 2002; Whittaker, 2012) and was used to complement elevation measurements localized further away from the paleo shorelines. Steepness index calculations are only valid for detachment-limited streams, i.e. bedrock rivers (Whipple, 2002; Whittaker, 2012), which excludes most streams flowing on loose syn-rift sediments. Streams in recent lava might also not have yet reached a steady state and therefore not be valid for interpretation. However, rivers flowing on Mesozoic and Paleozoic sediments,

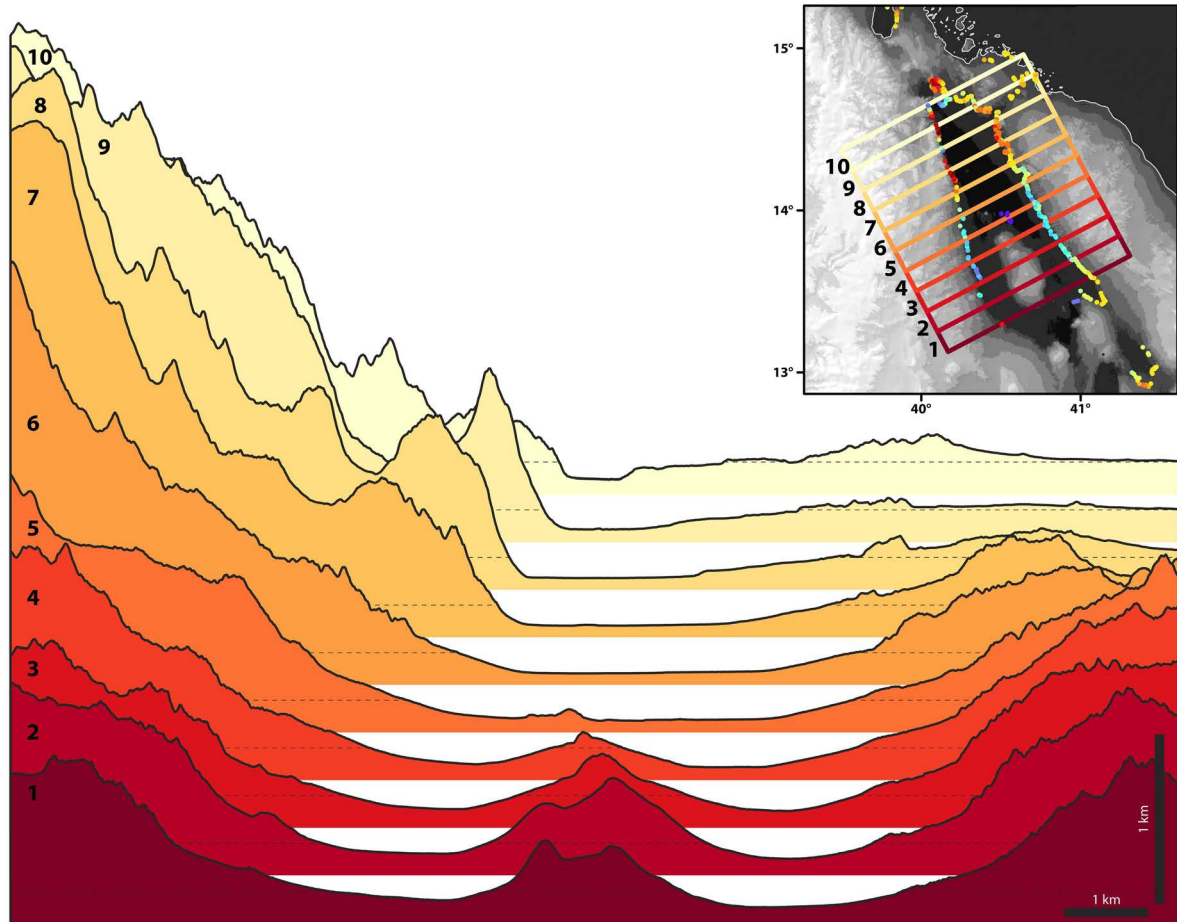


Figure 5: Topography of the Danakil depression. Swath profiles (representing mean elevation) along the Danakil depression. The western margin shows steeper and less mature topography towards the N (up). The topographic asymmetry can be observed on the profiles 7 to 10. Insert map show the position of the swath profiles and the elevation of marine sediments (same colour as fig. 3).

trap basalts, as well as basement rocks are often bedrock rivers and are assumed to be in steady state and therefore trustworthy. Fig. 3b thus only represent steepness indexes for those bedrock lithologies but the full processed dataset is available in Supplementary Fig. 2. Steepness index calculations for the northern part of the western margin mimic the trends observed in reef elevation and uplift rates (Fig. 3b). Overall, the highest index values are present on the northern part of the escarpment between 14 and 15°N.

III.4 DISCUSSION

III.4.1 Uplift and uplift rates

The paleo shoreline uplift pattern, steepness index map and uplift rate values (Fig. 3 and 4) are interpreted in terms of basin dynamics. The significant uplift of the shoulders and subsidence in the basin indicates active rifting. Uplift rates (Fig. 4c) are comparable to the ones found for the Corinth rift over similar time periods (Armijo *et al.*, 1996; McNeill & Collier, 2004; de Gelder *et al.*, 2019; Fernández-Blanco *et al.*, 2019) but one order of magnitude higher than what is typically found along the Red Sea, with

most of the values less than 0.1 mm/yr (Lambeck *et al.*, 2011; Bosworth *et al.*, 2019) and the highest being 0.23 mm/yr on Tiran Island (northernmost Red Sea; Ribot *et al.*, 2021). In particular, they are much higher than the values measured in the southern Red Sea, e.g. on the Dahlak Islands (Conforto *et al.*, 1976), in the Bab-el-Mandeb strait (Faure *et al.*, 1980; Al-Mikhlaifi *et al.*, 2018) or on the Kamaran Island (Al-Mikhlaifi *et al.*, 2021) (Fig. 1) where MIS 5e terraces are all localized between 1 and 9 m.a.s.l.. Uplift rates are however comparable with estimated subsidence rates in the centre of the basin of ~400 m in 100-120 ky (Bastow *et al.*, 2018), i.e. ~3-4 mm/yr. The combination of strong uplift of the margins and subsidence in the basin creates significant accommodation space. Higher rates have been reported in central Afar with vertical fault slip rates of ≥ 3 mm/yr in the Gaggade graben (Manighetti *et al.*, 2001) or subsidence rates between 4.7 and 9.1 mm/yr in the Tendaho graben (Mologni *et al.*, 2021) and 8-9 mm/yr in the Asal rift (Stein *et al.*, 1991), which is expected due to higher extension in the south (McClusky *et al.*, 2010; Viltres *et al.*, 2020).

The high uplift rates in the northern tip of the depression suggest that this area experiences the most important tectonic activity. In the Belekiya Mountains, these high uplift rates correlate with high steepness index (fig. 3b), and steeper topography (fig. 5), corroborating this hypothesis. In the northern part of the Depression, most reefs developed on the footwall of major active faults with steep scarps (Fig. 3, 6a) resulting in significant uplift. However, some reefs (e.g. between 14.35 and 14.47°N, fig. 4) formed in the hanging wall of these major basin-bounding

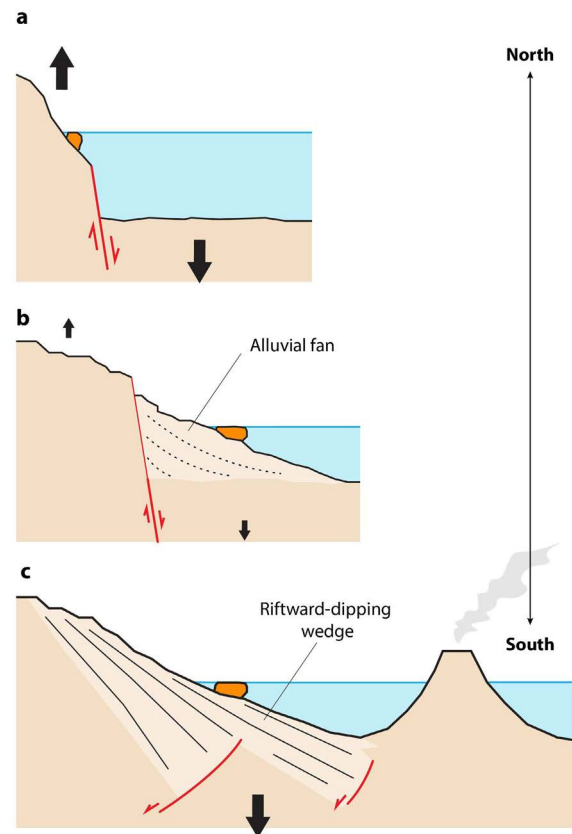


Figure 6: Schematic representation of the different uplift and subsidence patterns across the strike of the basin. In the northern part of the Depression (a), most reefs developed on the footwall of major active faults with steep scarps resulting in significant uplift. In the central part of the western margin (b), the local topography more mature with the development of large alluvial fans, possibly indicating the presence of older and/or less active faulting. The reefs colonizing these fans are probably situated in the hanging-wall of major basin-bounding faults. Both factors can explain the smaller uplift or subsidence of the reefs. In the southern part of the Depression (c), subsidence is interpreted to be caused by magmatic loading of the Erta Ale range and continentward-dipping faults forming the equivalent of seaward-dipping reflectors.

normal faults and experienced smaller uplift or even subsidence. In the central part of the western margin (~ between 13.9 and 14°N), the steepness index is lower and the local topography more mature with the development of large alluvial fans, possibly indicating the presence of older and/or less active faulting. The reefs colonizing these fans

are probably situated in the hanging-wall of major basin-bounding faults (Fig. 6b). Both factors can explain the smaller uplift or subsidence of the reefs. The subsidence trend observed between 13.4°N and 14°N (fig. 3) is related to the presence of the Erta Ale range and with continentward-dipping faults (e.g. 13.78N, 40.22E, Fig. 2). As shown by numerical models (Corti *et al.*, 2015), several kilometres of subsidence can be accommodated by magmatic loading. We interpret this as an important process impacting the local subsidence trend, along with continentward-dipping faults forming a riftward-dipping wedge, similar to seaward-dipping reflectors observed in passive margins (Fig. 6c). This interpretation has been already postulated for the Erta Ale range by several authors (Keir *et al.*, 2013; Le Gall *et al.*, 2018; Watts *et al.*, 2020). Overall, these N-S changes of uplift rates, topographic maturity and basin structure may be explained by the northward rift propagation in the Danakil Depression, the southern part thus representing a more evolved rifting stage compared to the northern part (Rime *et al.*, 2023, chapter II, this volume).

While the elevation of marine sediments in the south of the basin is similar on the eastern and western margin, they differ significantly in the northern part of the basin (Fig 3c). This is probably related to asymmetric rifting in this part of the basin which is also expressed by a much steeper topography of the western margin compared to the eastern one (Fig. 5, profiles 7 to 10). Interpretation of seismic reflection data in the same area also indicate more active western faults (Bastow *et al.*, 2018). This can also explain the irregularity of the uplift pattern on the western margin

where more intense faulting generates higher but varied uplift compared to the eastern margin and the Red Sea coast. In the Buia graben, this asymmetry seems less pronounced or even of opposite polarity as expressed by the large fault structuring the eastern margin of the graben (Duffield *et al.*, 1997; Sani *et al.*, 2017). This is coherent with the highest elevation of coral terraces at the eastern margin (Fig. 3c). In the southern part of the depression, the rift is wider and the topography more symmetric.

Abrupt changes in uplift amplitude along the north-western margin of the depression can be correlated with faults oblique to the rift (Fig. 4d) with orientations ranging from WSW-ENE to N-S. These faults often vertically offset Mesozoic sediments and basement by up to several hundred of meters (Fig. 2) and divide zones with different steepness index and paleo shoreline uplift values (Fig. 3b). This indicates that oblique faults segment the margin in blocks experiencing different tectonic movements.

III.4.2 The gateway between the Red Sea and the Danakil Depression

The position of the gateway through which the Red Sea flooded the depression is debated. Many authors (Frazier, 1970; Lalou *et al.*, 1970; Bonatti *et al.*, 1971; Beyth, 1978) argued for the presence of a gateway through the Buia graben connecting the depression to the Gulf of Zula. This gateway would have been closed by the volcanic activity of the Alid volcano. The description of “marine sediments” observed on the flanks of the Alid volcano and dated between 300 ka and 212 ka by Duffield *et al.* (1997)

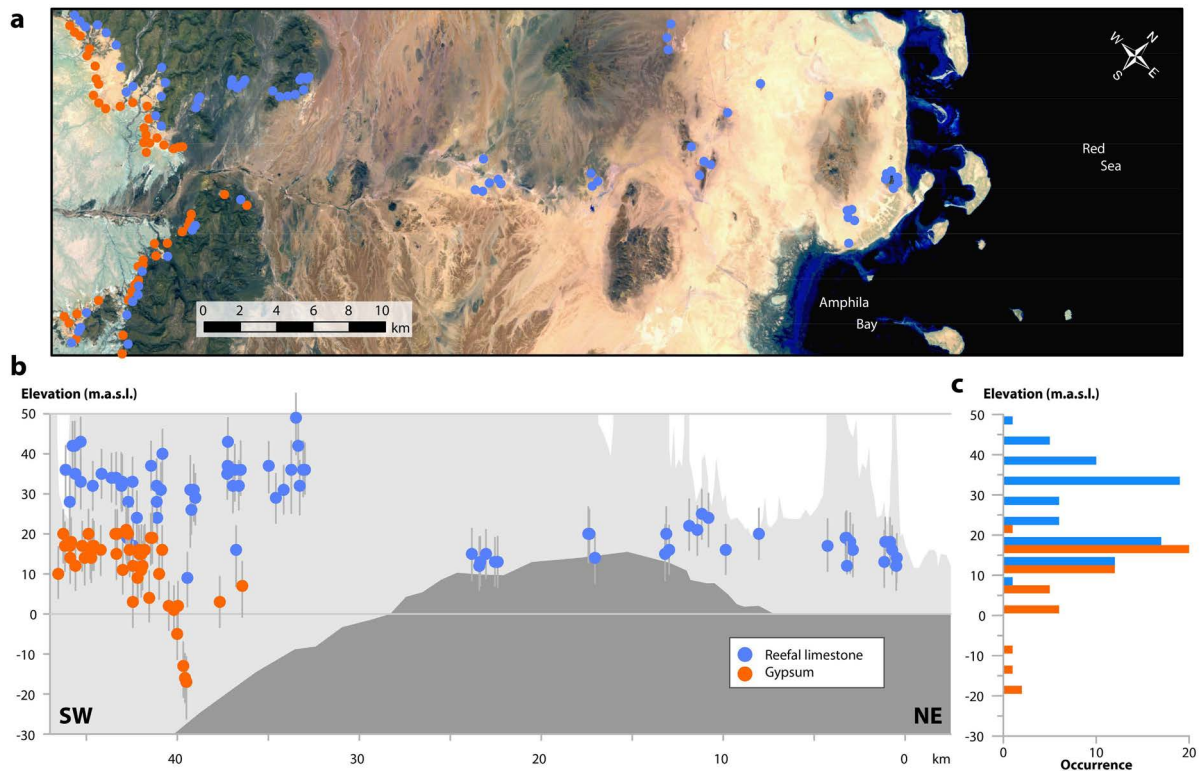


Figure 7: Analysis of the Amphila gateway. Analysis of outcrops mentioned by Holwerda & Hutchinson (1968) and identified by remote sensing. We discriminate between gypsum and reefal limestones. Gypsum outcrops are interpreted to be MIS5e. The limestones age is unknown. **a)** Map of the Amphila gateway (Landsat 8 bands 7-6-5) Localisation of the map is presented on Fig. 3a. **b)** Profile through the gateway. Light and dark grey shading represent the maximum and minimum topographic elevation of a swath profile covering the area of a). Grey vertical lines represent the standard uncertainty. **c)** Frequency distribution of the elevations of both lithologies.

strengthened this hypothesis. Only few authors argued for a flooding from the Amphila Bay (Holwerda & Hutchinson, 1968; Hutchinson & Engels, 1970). We think the latter hypothesis is more plausible for the following reasons (see also Foubert et al., in prep. for further discussion).

- 1) The elevation of the Amphila gateway (16 m) is much lower than those of the Buia gateway (215 m).
- 2) The Amphila gateway is currently experiencing uplift (Fig. 3, Fig. 7 and following sections) and was thus situated at lower elevations in the past, while the Buia graben is believed to experience subsidence (Sani *et al.*, 2017) and was thus

situated at higher elevations in the past.

- 3) Reefal limestones can be spotted along the Amphila gateway by remote sensing (Fig. 7) and have also been unambiguously described by Holwerda & Hutchinson (Holwerda & Hutchinson, 1968).
- 4) The doming of the Alid volcano has now been constrained between 36 ± 9 and 15.2 ± 2.9 ka (Duffield *et al.*, 1997; Lowenstern *et al.*, 2006), i.e. long after the last marine incursion and can therefore not be the cause of its desiccation.

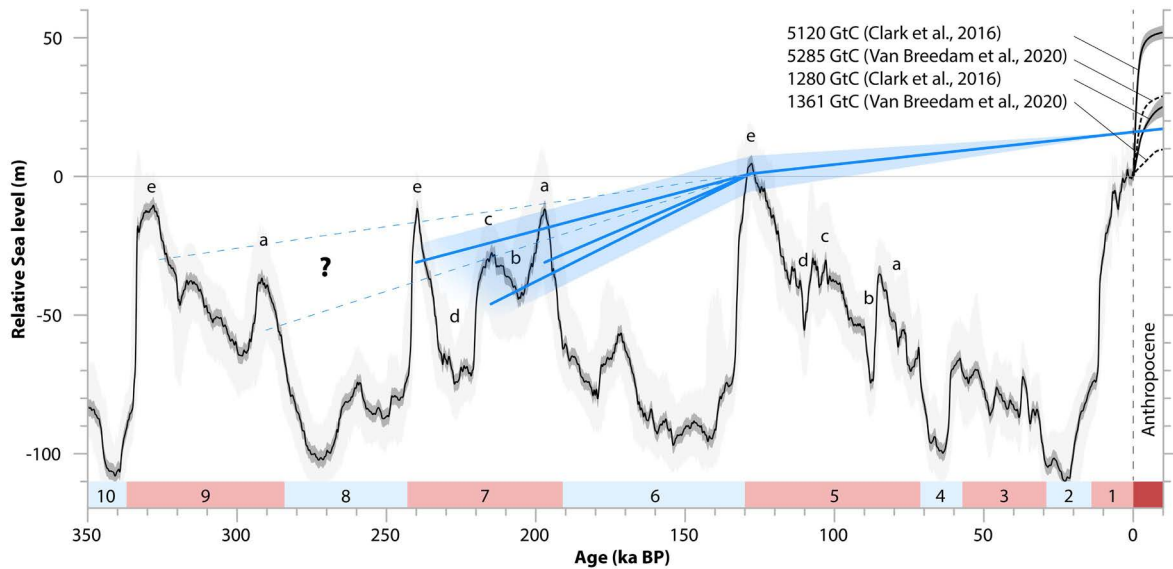


Figure 8: Interpretation of the Amphila gateway evolution. The black line represents the relative sea-level. Data from Grant *et al.* (2014) for past relative sea-level and from Clark *et al.* (2016) (solid lines) and Van Breedam *et al.* (2020) (dashed lines) for future relative sea-level in high and low emission scenarios. Light grey indicates 95% probability interval of RSL dataset. Dark grey indicates 95% interval for the probability maximum for paleo sea-level and 1 σ uncertainty for future sea-level of Clark *et al.* (2016). The thick blue line shows the gateway elevation evolutions scenario discussed in the text. The light blue represents its standard uncertainty. The dashed blue lines represent rejected hypothesis discussed in the text. Marine Isotopic stages and substages are also shown (Lisiecki & Raymo, 2005). Details of the calculations can be found in Supplementary Methods 4.

5) We strongly doubt the marine origin of the sediments observed on the flanks of the Alid volcano which lack diagnostic marine organisms and facies. Moreover, detailed sedimentological studies in the Dandiero Basin, the SW part the of Buia graben, indicate the presence of a continental fluvio-lacustrine environment throughout the Pleistocene, without any marine sediments (Abbate *et al.*, 2004; Sani *et al.*, 2017).

We further analysed the possibility of flooding through the Amphila Bay gateway by looking at the elevation of distinct sedimentary deposits. Gypsum and limestone deposits were distinguished (Fig. 7a,b). The vast gypsum cover capping the eastern margin is situated at an average elevation of ~20m (Fig. 7c) i.e. above the gateway, indicating that the water body in which they formed was connected to the Red

Sea. Considering the age of the last marine incursion, those deposits cannot be younger than MIS 5e (Foubert *et al.*, in prep.). We also reject any older age as these deposits would have been covered by sediments produced by the MIS5e flooding. We therefore interpret the gypsum deposits to be MIS5e. This is corroborated with observations on the western margin where extensive gypsum deposits are clearly correlated with MIS 5e sediments. The age of the limestone deposits, however, cannot be determined by remote sensing. They are separated in two elevation clusters (Fig. 7c), one around 20 m and one around 35 m. We hypothesize that the 20 m cluster is correlated with the MIS 5e gypsum situated at the same elevation, a configuration often observed on the western margin. We speculate that the higher 35 m cluster is older and probably of MIS 7 age. These carbonate deposits

form extensive fringing reef deposits, a geomorphic feature commonly observed at the western margin. We strongly doubt the possibility of MIS 8 or 9 reefs being so well preserved. Assuming that the topography of the Amphila gateway did not change substantially and considering the uplift for the eastern margin, the gateway was significantly shallower during MIS5e than during MIS7. These results are consistent with the observation of much more developed MIS 7 reefs on the western margin (Jaramillo-Vogel *et al.*, 2019; Negga *et al.*, in prep.).

We assume that uplift was approximately constant across the gateway, as indicated by the cluster of marine sediments at 20 m and 35 m throughout the gateway and by absence of significant variation of the steepness index (Fig. 3b) around the gateway. With this assumption and the inferred ages of these terraces, it is possible to reconstruct curves for the elevation of the present-day gateway (Fig. 8). It evidences a 15 ± 6 m uplift (0.12 ± 0.05 mm/yr uplift rate) since MIS 5e, and 37 ± 9 m of uplift between MIS 7 and MIS 5e (0.33 ± 0.08 to 0.6 ± 0.1 mm/yr uplift rate depending on the different scenarios). It shows that uplift is the cause of the non-flooding during the current interglacial. In the absence of uplift since MIS 5e, the gateway would now be situated at sea-level, probably triggering a new flooding since the late Holocene. It further indicates a decrease in uplift rate after MIS 5e, in accordance with observation on the western margin (Fig. 4). Flooding during MIS 2, 3, 4 or 5a appears virtually impossible. Some datings of marine sediments presented in the literature (25 ka by Brinckmann & Kürsten, 1971; 32 ka by Bonatti *et al.*, 1971; 34 ka by Brinckmann & Kürsten, 1971; 80 ka by

Lalou *et al.*, 1970 and 76-88 ka by Hutchinson & Engels, 1970) are therefore considered as incorrect (see also Foubert *et al.* in prep.).

III.4.3 Future flooding of the Depression

With the understanding of the tectonic movement acting on the gateway, we can try to predict the future flooding of the depression. On (geological) short term, anthropogenic global warming already triggered a global sea-level rise that is expected to continue for several centuries to millennia (IPCC, 2014; Clark *et al.*, 2016; DeConto & Pollard, 2016; Van Breedam *et al.*, 2020). Only two recent studies (Clark *et al.*, 2016; Van Breedam *et al.*, 2020) modelled sea level rise over the next 10+ kyr and the results have low confidence levels. The high-emission scenarios of both Clark *et al.* (2016) and Van Breedam *et al.* (2020) imply a flooding of the depression, forming a 9000 km² wide sea. The 5120 GtC scenario of Clark *et al.* (2016) indicates a flooding in 2600 ± 50 AD (see Fig. 8). At that time, assuming a continuous uplift rate, the gateway would only have risen by less than 11 cm. The comparable MMCP8.5 model (5285GtC) of Van Breedam *et al.* (2020) indicate a flooding around 3400 AD, with less than 25 cm gateway uplift. Conversely, the low-emission scenarios differ. The 1280 GtC emission scenario of Clark *et al.* (2016) corresponds to a world with important greenhouse gases emission reduction meeting the Paris Agreement goal of not exceeding 2° global warming. Under this scenario, the depression would be flooded around 5200 ± 400 AD, when the gateway would be 22 to 54 cm higher than today. The modelling of Van Breedam *et al.* (2020) (1361 GtC) indicate no flooding with a

maximum global sea-level rise of approx. 10 m in 12ka. The flooding of the depression and its timing is therefore heavily dependent on greenhouse gases emissions but likely to happen during the Anthropocene.

The longer-term evolution of the rift might lead to a shift in the localisation of the flooding gateway. Continued uplift on the Danakil block and subsidence in the Buia graben associated with the northward rift propagation (Rime *et al.*, 2023, chapter II, this volume) could lead to a future opening through the Gulf of Zula. We roughly estimate subsidence in the Buia graben by extrapolating the uplift on the shoulder of this basin to subsidence in the centre of the basin. Uplift to subsidence ratio determined in the Corinth rift range from 1:1.1 to 1:3.5 (Armijo *et al.*, 1996; McNeill & Collier, 2004; de Gelder *et al.*, 2019). Assuming that the marine carbonate deposits interpreted at 100 m.a.s.l. the SE shoulder of the Buia Graben (approx. 14.78°N, 40.08°S, see fig. 3) are MIS 5e or MIS 7 in age and that rates are constant, we can estimate the time until this gateway could reach present-day sea level between 80 ka and 420 ka (see Supplementary Methods 5). Volcanism might play an important role in this gateway with the Alid volcano potentially damming the basin and delaying a flooding, comparable to the Walvis Ridge damming the Proto-Atlantic in the Cretaceous (Jackson *et al.*, 2000). Nevertheless, if the drift of the Arabian plate remains constant and if sea-level variations remain in the ranges observed during the Quaternary, we can assume that floodings through the Gulf of Zula will happen after a few hundred of thousand of years. This will probably result in the deposition of several

hundred of meters of marine sediments and evaporites in the basin before possibly reaching a continuous connection with the Red Sea within the next one million years.

III.4.4 Basin dynamics

The evolution of the Danakil depression illustrates that the transition from continental to marine conditions in a rift setting is not gradual but punctuated by several stages of opening and closure of the basin. The depression already experienced at least 4 cycles of marine incursion and desiccation and will probably experience several more before a continuous connection with the Red Sea is established. The total duration of this period of alternation will possibly last less than 1.5 Myr but will result in the accommodation and deposition of a significant thickness of diverse sediments, some of them being of economic significance. These alternations are controlled by the interaction of several processes. Tectonic movements including uplift of the rift margins and subsidence of the rift axis over long timescales (several 100s of kyr) play a first order role. Eustatic sea-level variations have a second order role but over shorter timescales (100 kyr), such as considered in this work, it constitutes the main control. It is plausible to think that bioconstructions during the open-marine stage, as well as the deposition or erosion of sediments by fluvial processes, might have played a third order control on the elevation of the gateway. Finally, volcanism did not yet play an important role but might become important in the future evolution of the depression.

III.5 CONCLUSIONS

The Danakil depression is one of the few modern examples of a rift currently undergoing the transition from continental to marine settings. Field mapping and remote sensing allowed constraining its past and future flooding history. The Danakil rift basin is sustaining active tectonism with uplift on the margins and subsidence in the basin, creating important accommodation space. The northern part of the western margin of the depression experiences the highest uplift, reaching more than 1 mm/yr. This important tectonic activity is also expressed by a steeper and younger topography and is probably related to the northward propagation of the rift and rift asymmetry. The Pleistocene flooding of the depression was mainly controlled by eustatic variations and tectonic movements of the eastern margin. Uplift prevented renewed flooding during the Holocene. On geological short time-scales, future flooding of the depression will depend on anthropogenic greenhouse gas emission levels. In the worst-case scenario, a flooding could happen within six centuries. Continued flank uplift and northward rift propagation will probably relocate the flooding gateway to the Buia graben and Gulf of Zula before continuous marine conditions are reached, possibly within the next one million years.

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AUTHOR CONTRIBUTIONS

- V. Rime: Design, field exploration, outcrop interpretation, data collection, mapping, data processing and analysis, interpretation, figures, writing.
- Foubert: Supervision, concept, design, field exploration, data collection, outcrop interpretation, interpretation, critical review.
- L. Perrochet: Field exploration, outcrop interpretation, data collection, data processing and analysis, interpretation, critical review.
- D. Jaramillo-Vogel: Field exploration, outcrop interpretation, critical review.

- H. Negga: Field exploration, outcrop interpretation, critical review.
- Atnafu: Field exploration, critical review.
- T. Kidane: Supervision, Concept, field exploration, interpretation, critical review.

DECLARATION OF COMPETING INTERESTS

The authors declare that the research was conducted in the absence of any financial or non-financial relationships that could be construed as a potential conflict of interest.

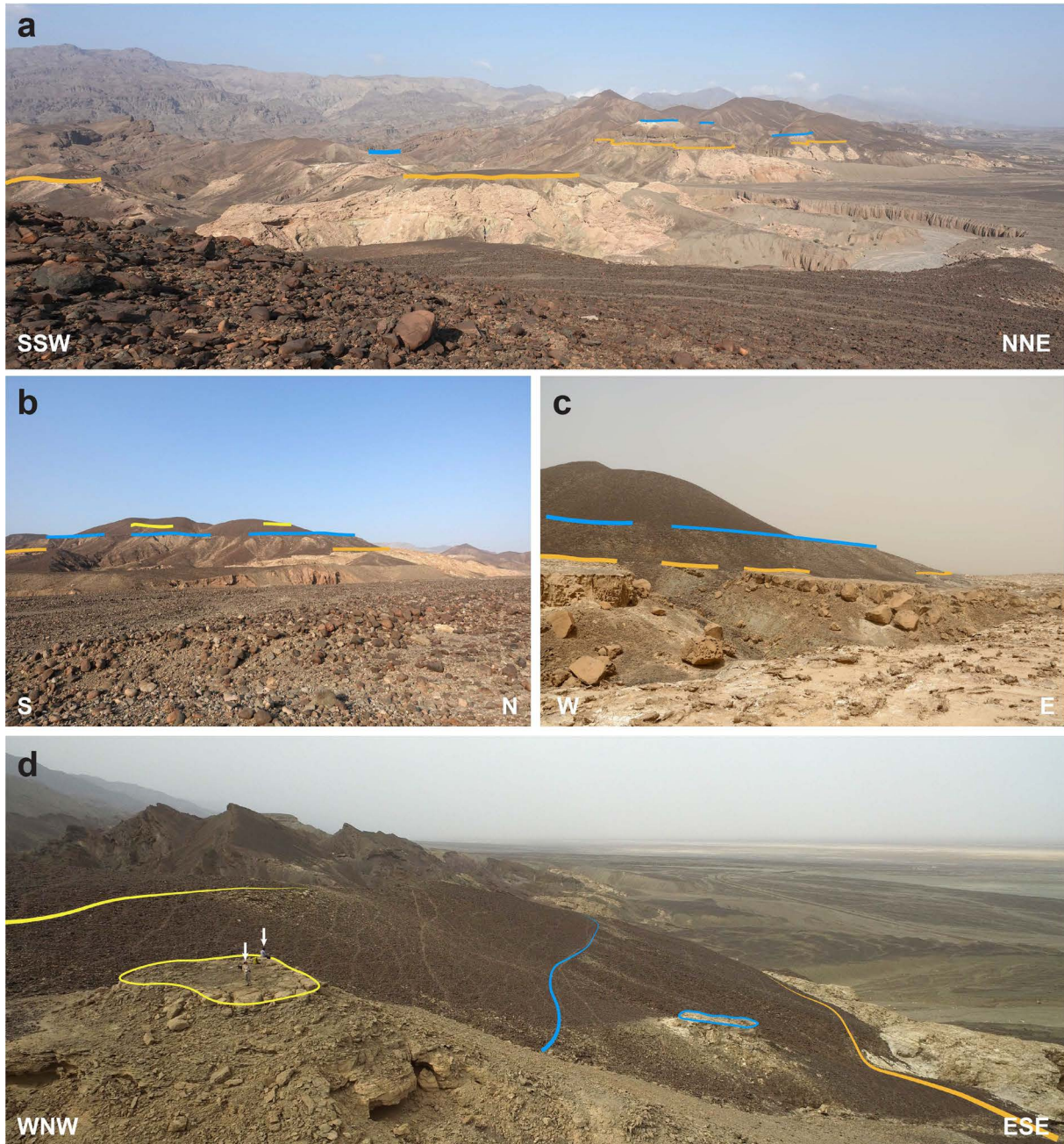
DATA AVAILABILITY

The SRTMGL1 (NASA Shuttle Radar Topography Mission Global 1 arc second) data that were used in this study are available on <https://earthexplorer.usgs.gov/> with the identifier <https://doi.org/10.5067/MEaSUREs/SRTM/SRTMGL1.003>.

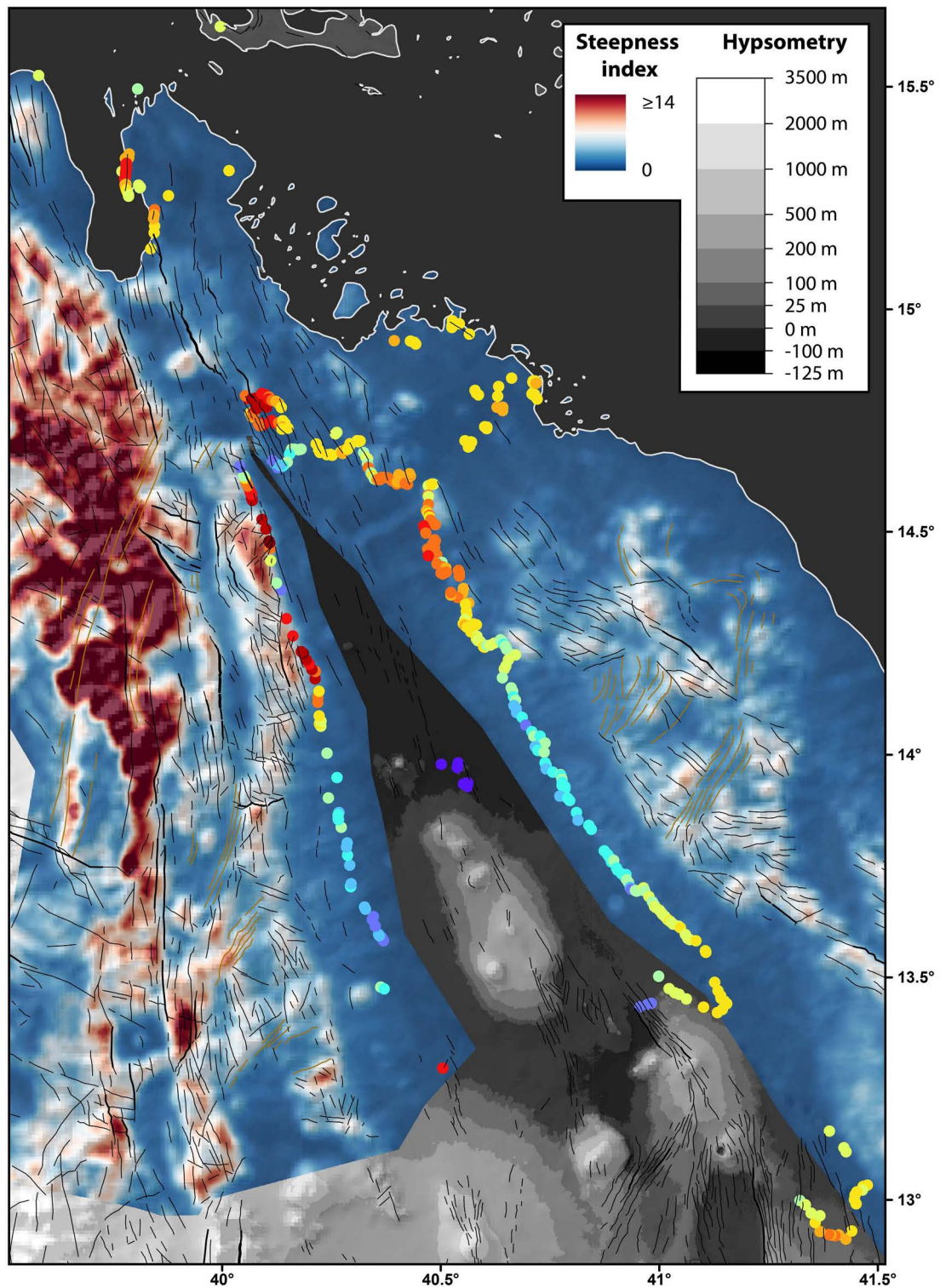
The elevation measurements resulting from this study and elevation data from the literature are available in the supplementary materials.

Other data are available from the corresponding author upon request.

SUPPLEMENTARY FIGURES



Supplementary figure 1: Overview of the different generations of marine terraces. The terraces are found at different elevations due to tectonic uplift. The orange lines represent the MIS 5e paleo shoreline, blue lines MIS 7 shorelines and yellow lines represent the first undated terrace. **a)** Picture taken from 14.184N, 40.219E and showing outcrops CO24 and CO10 from CO26. **b)** Picture taken from 14.187N, 40.224E showing CO26. **c)** Picture taken from 14.115N, 40.226E showing CO4. **d)** Picture taken by drone from 14.209N, 40.198E showing CO27. The arrows point towards two geologists. While only three terraces are shown, a fourth one is present near the summit of the hill (not visible on the picture).



Supplementary figure 2: Full results of the topographic analysis. This map is similar to fig. 3b but shows all results, including for rivers that likely do not meet the detachment-limited criteria or that are likely not in equilibrium (see Results).

SUPPLEMENTARY TABLES

Outcrop	latitude	longitude	Interpreted age	Elevation (masl)	Sedimentological uncertainty (m)	Measurement uncertainty (m)	Total elevation uncertainty (m)	Method	Uplift (m)	Uplift uncertainty (m)	Uplift rate since MIS 5e (mm/yr)	Uncertainty uplift rate since MIS 5e (mm/yr)	Uplift rate between MIS 7 and 5e (mm/yr)	Uncertainty uplift rate between MIS 7 and 5e (mm/yr)	Uplift rate since MIS 7 (mm/yr)	Uncertainty uplift rate since MIS 7 (mm/yr)
C034	13.6384	40.3506	MIS 7	-40	+2	±8	+8/-8	Garmin GPS	-21	+14/-14	-	-	-	-	-0.10	+0.06/-0.06
			Unknown	-44	+2	±8	+8/-8	Visual estimation	-	-	-	-	-	-	-	-
C015	13.8676	40.2726	MIS 5e	-20	+1	±8	+8/-8	Garmin GPS	-25	+9/-9	-0.20	+0.07/-0.07	-	-	-	-
C0311	13.9006	40.2485	MIS 5e	-11	+1	±1	+1/-1	Differential GPS	-16	+1/-4	-0.12	+0.03/-0.03	0.03	+0.14/-0.14	-0.06	+0.05/-0.05
			MIS 7	-32	+1	±4	+4/-4	Leica GPS	-13	+12/-12	-	-	-	-	0.02	+0.06/-0.06
C033	13.9461	40.2686	MIS 7	-15	+2	±8	+8/-8	Garmin GPS	4	+14/-14	-	-	-	-	-	-
			Unknown	-39	+5	±8	+9/-8	Garmin GPS	-	-	-	-	-	-	-	-
C032	14.0069	40.2459	MIS 7	-4	+2	±8	+8/-8	Garmin GPS	15	+14/-14	-	-	-	-	0.07	+0.06/-0.06
			MIS 5e	6	+3	±1	+3/-1	Differential GPS	1	+5/-4	0.01	+0.04/-0.03	0.25	+0.15/-0.14	0.11	+0.05/-0.05
C01	14.0616	40.2321	MIS 7	5	+1	±1	+1/-1	Differential GPS	24	+11/-11	-	-	-	-	-	-
			MIS 5e	9	+3	±1	+3/-1	Differential GPS	4	+5/-4	0.03	+0.04/-0.03	0.26	+0.15/-0.15	0.12	+0.05/-0.05
C012	14.0654	40.2596	MIS 7	8	+1	±1	+1/-1	Differential GPS	27	+11/-11	-	-	-	-	-	-
			MIS 7	9	+1	±1	+1/-1	Differential GPS	28	+11/-11	-	-	-	-	0.13	+0.05/-0.05
C07	14.0729	40.2327	MIS 5e	8	+3	±1	+3/-1	Differential GPS	3	+5/-4	0.02	+0.04/-0.03	0.27	+0.15/-0.15	0.12	+0.05/-0.05
			MIS 7	8	+1	±1	+1/-1	Differential GPS	27	+11/-11	-	-	-	-	0.16	+0.07/-0.06
C014	14.0849	40.2283	MIS 7	15	+3	±8	+9/-8	Garmin GPS	34	+14/-14	-	-	-	-	-	-
			MIS 5e	18	+1	±1	+1/-1	Differential GPS	13	+1/-4	0.10	+0.03/-0.03	0.47	+0.18/-0.18	-	-
C04	14.1151	40.2554	MIS 7	37	+3	±1	+3/-1	Differential GPS	56	+11/-11	-	-	-	-	0.25	+0.06/-0.06
			MIS 5e	20	+5	±8	+9/-8	Garmin GPS	15	+10/-9	0.12	+0.08/-0.07	0.46	+0.22/-0.21	0.28	+0.07/-0.07
C031	14.1269	40.2254	MIS 7	38	+1	±8	+8/-8	Garmin GPS	57	+14/-14	-	-	-	-	-	-
			MIS 5e	17	+2	±8	+8/-8	Garmin GPS	12	+9/-9	0.09	+0.07/-0.07	-	-	-	-
C030	14.1402	40.2283	MIS 7	29	+2	±8	+8/-8	Garmin GPS	24	+9/-9	0.19	+0.07/-0.07	0.67	+0.23/-0.23	0.39	+0.07/-0.07
			MIS 7	66	+1	±8	+8/-8	Garmin GPS	85	+14/-14	-	-	-	-	-	-
			MIS 5e	-13	+4	±8	+9/-8	Garmin GPS	-18	+10/-9	-0.14	+0.08/-0.07	0.70	+0.26/-0.25	0.21	+0.07/-0.07
C026	14.1846	40.2302	MIS 7	27	+1	±8	+8/-8	Garmin GPS	46	+14/-14	-	-	-	-	-	-
			Unknown	38	+5	±8	+9/-8	Garmin GPS	-	-	-	-	-	-	-	-
			Unknown (2)	-	-	-	-	-	-	-	-	-	-	-	-	-
C024	14.1887	40.2124	MIS 5e	30	+10	+14/-12	+14/-12	DEM	25	+18/-12	0.20	+0.11/-0.1	0.80	+0.30/-0.29	-	-
			MIS 7	79	+5	+8/-10	+8/-10	DEM	98	+14/-15	-	-	-	-	0.45	+0.08/-0.08
C010	14.2011	40.2061	MIS 5e	17	+5	±8	+9/-8	Garmin GPS	12	+10/-9	0.09	+0.08/-0.07	0.56	+0.24/-0.23	-	-
			MIS 7	44	+1	±8	+8/-8	Garmin GPS	63	+14/-14	-	-	-	-	0.29	+0.07/-0.07
			MIS 5e	23	+5	±1	+5/-1	Drone survey	18	+6/-4	0.14	+0.05/-0.03	-	-	-	-
			MIS 7	43	+1	±1	+1/-1	Drone survey	62	+11/-11	-	-	0.48	+0.19/-0.18	0.28	+0.06/-0.06
C027	14.2105	40.1892	Unknown (1)	62	+1	±1	+1/-1	Drone survey	-	-	-	-	-	-	-	-
			Unknown (2)	78	+3	±1	+3/-1	Drone survey	-	-	-	-	-	-	-	-
			MIS 5e	45	+3	±1	+3/-1	Differential GPS	40	+5/-4	0.31	+0.04/-0.03	0.88	+0.26/-0.26	-	-
C06	14.2246	40.1928	MIS 7	101	+1	±1	+1/-1	Differential GPS	120	+11/-11	-	-	-	-	0.55	+0.08/-0.08
			MIS 5e	42	+2	±8	+8/-8	Garmin GPS	37	+9/-9	0.29	+0.07/-0.07	0.90	+0.30/-0.30	0.54	+0.09/-0.09
C025	14.2286	40.1578	MIS 7	100	+5	±10	+11/-10	Visual estimation	119	+16/-15	-	-	-	-	-	-
C013	14.2639	40.1687	MIS 5e	56	+10	±8	+13/-8	Garmin GPS	51	+13/-9	0.40	+0.11/-0.07	-	-	-	-
C0X	14.3029	40.1540	MIS 5e	44	+10	+17/-9	+17/-9	DEM	39	+17/-10	0.30	+0.13/-0.08	-	-	-	-
C018	14.3672	40.1419	MIS 5e	-35	+5	±1	+5/-1	Differential GPS	-40	+6/-4	-0.31	+0.05/-0.03	-	-	-	-
C028	14.3828	40.1317	MIS 5e	-1	+5	±8	+9/-8	Garmin GPS	-6	+10/-9	-0.05	+0.08/-0.07	-	-	-	-
C035	14.4334	40.1185	MIS 5e	8	+5	±8	+8/-8	Garmin GPS	3	+10/-9	0.02	+0.08/-0.07	-	-	-	-
			MIS 5e	6	+10	±8	+13/-8	Garmin GPS	1	+13/-9	0.01	+0.1/-0.07	0.59	+0.26/-0.23	0.25	+0.07/-0.07
C029	14.4611	40.1177	MIS 7	56	+5	±8	+9/-8	Garmin GPS	55	+14/-14	-	-	-	-	-	-
			MIS 5e	62	+5	±1	+5/-1	Differential GPS	57	+6/-4	0.45	+0.05/-0.03	1.43	+0.39/-0.38	-	-
C016	14.4731	40.1167	MIS 7	168	+3	±1	+3/-1	Differential GPS	187	+11/-11	-	-	-	-	0.85	+0.10/-0.10
C019	14.4773	40.1150	MIS 5e	159	+10	±8	+13/-8	Garmin GPS	154	+13/-9	1.20	+0.11/-0.07	-	-	-	-
C017	14.5043	40.1077	MIS 5e	92	+3	±1	+3/-1	Differential GPS	87	+5/-4	0.68	+0.04/-0.03	-	-	-	-
C021	14.5258	40.0990	MIS 5e	96	+5	±8/-6	+8/-6	DEM	91	+9/-7	0.71	+0.07/-0.06	-	-	-	-

Supplementary Table 1: List of studied outcrops on the western margin

Outcrop	latitude	longitude	Interpreted age	Elevation (masl)	Sedimentological uncertainty (m)	Measurement uncertainty (m)	Total elevation uncertainty (m)	Method used
COE16	13.1441	41.3921	Unknown	5	+ 2	± 4	+ 4/ - 4	Leica GPS
COET1	13.1065	41.4248	Unknown	3	+ 1	± 4	+ 4/ - 4	Leica GPS
COET5	13.4671	41.0276	Unknown	3	+ 2	± 4	+ 4/ - 4	Leica GPS
COET5	13.4963	41.0051	Unknown (1)	-19	+ 1	± 4	+ 4/ - 4	Leica GPS
			Unknown (2)	-7	+ 1	± 4	+ 4/ - 4	Leica GPS

Supplementary Table 2: List of studied outcrops on the eastern margin.

Outcrop	Latitude	Longitude	Elevation (m.a.s.l.)	Uncertainty	Reference	Elevation data source
AN1-4	15.171	39.853	14	+ 2 / - 2	Buffler et al. (2010)	Original paper
AN13-14-7-8-10	15.186	39.854	20	+ 2 / - 2	Buffler et al. (2010)	Original paper
Abdur South	15.135	39.845	12	+ 2 / - 2	Buffler et al. (2010)	Original paper
Melita Bay	15.272	39.820	0	+ 2 / - 1	Buffler et al. (2010)	Original paper
Northern End of Aleite	15.338	39.787	25	+ 5 / - 5	Buffler et al. (2010)	Original paper
Buri	15.254	39.886	12	+ 4 / - 4	Buffler et al. (2010)	Original paper
Hoda Island	15.494	39.816	0	+ 2 / - 1	J.H. Bruggemann (pers. comm.)	Pers.comm
Foot of Aleite	15.309	39.779	10	+ 4 / - 4	J.H. Bruggemann (pers. comm.)	Pers.comm
Foot of Mt. Ghedem	15.525	39.587	5	+ 1 / - 2	J.H. Bruggemann (pers. comm.)	Pers.comm
Bakili	13.975	40.509	-108	+ 6 / - 6	Barberi & Varet (1970, p. 869)	Measure on SRTMGL1 based on outcrop description
Bakili	13.977	40.547	-107	+ 6 / - 6	Barberi & Varet (1970, p. 869)	Measure on SRTMGL1 based on outcrop description
Bakili	13.972	40.546	-105	+ 7 / - 6	Barberi & Varet (1970, p. 869)	Measure on SRTMGL1 based on outcrop description
Bakili	13.967	40.545	-105	+ 7 / - 6	Barberi & Varet (1970, p. 869)	Measure on SRTMGL1 based on outcrop description
Bakili	13.930	40.559	-110	+ 7 / - 7	Barberi & Varet (1970, p. 869)	Measure on SRTMGL1 based on outcrop description
Bakili	13.933	40.569	-106	+ 6 / - 6	Barberi & Varet (1970, p. 869)	Measure on SRTMGL1 based on outcrop description
Bakili	13.924	40.569	-105	+ 6 / - 6	Barberi & Varet (1970, p. 869)	Measure on SRTMGL1 based on outcrop description
Southernmost	13.292	40.508	53	+ 7 / - 8	Brinckmann & Kürsten (1971)	Measure on SRTMGL1 based on outcrop description
Dahlak	15.633	40.029	5	+ 5 / - 2	Conforto et al. (1975)	Original paper
Ito Achil	14.792	40.735	12	+ 5 / - 2	Holwerda and Hutchinson (1968)	Original paper

Supplementary Table 3: List of outcrops mentioned in the literature.

SUPPLEMENTARY METHODS

Supplementary Method 1: GPS measurement

The Leica Viva GS10 was used either as conventional GPS or as differential GPS with a control points made available by Allana Potash Company. Because of the limited range of radio communication between the reference station and the mobile GPS, 18 new ground control points were measured to cover the area. An uncertainty of ± 1 m was considered for the differential GPS measurements. Used as a conventional GPS and after test on known control points in Switzerland, the uncertainty was considered as ± 4 m.

The Garmin GPSMAP 66i was used in a mode allowing a continuous measurement of 30 s. When possible, these measures were repeated after 30 to 60 min. Following tests on known control points in Switzerland, the uncertainty was considered as ± 8 m.

One outcrop (CO27) was measured by drone survey. A Wingtra drone was used with differential GPS using a Trimble R2 GNSS receiver as base station. This base was run for >12 h to obtain a reliable base positioning. The 3D model of the outcrop was then processed in Agisoft software, with a precision of 3 cm. The elevations of paleo shorelines observed in the field were then measured based on this 3D model. The total measurement uncertainty was considered as ± 1 m.

Supplementary Method 2: Geological mapping

The geological map (fig. 2) is a simplified version of the geological map of the Afar Depression (Rime *et al.*, 2022). It

was constructed based on existing maps, datings and descriptions available in the literature (Frazier, 1970; Brinckmann & Kürsten, 1970; Barberi *et al.*, 1971; Geological Survey of Ethiopia, 1971, 1978; Barberi *et al.*, 1972; Drury *et al.*, 1994; Duffield *et al.*, 1997; Bunter *et al.*, 1998; Carbone *et al.*, 1998; Abbate *et al.*, 2004; Bruggemann *et al.*, 2004; Buffler *et al.*, 2010; Sani *et al.*, 2017; Sembroni *et al.*, 2017; Le Gall *et al.*, 2018; Hagos *et al.*, 2020). These data were complemented by field mapping, including the western part of the depression from 13.6°N to 14.55°N , river cuts across the Belekiya Mountains (western margin of the depression), the region of Lake Afdera and the Sodonta Plain (40 km SSE of Afdera), the surroundings of the Tat'Ali range, the foot of the Nabro volcanic range, as well as along the roads linking Mekele to both Afdera and Dallol. Multispectral Landsat 8 imagery and other publicly-available aerial imagery from different providers (Google Earth, Bing Maps) and SRTM 1 arc-second Digital Elevation Model were also used. Furthermore, topographic analysis of rivers were used on the western margin to help determining the position of active faults. Manual picking of knickpoints was done on 130 individual rivers of the Belekiya Mountains (the western margin) between 13.8°N and 14.6°N where pre-rift rocks are in close proximity with the reefal terraces to complement field and remote sensing analysis to better constrain the position of faults (fig. 2). This analysis was done using the *Topographic Analysis Kit* TAK (Forte & Whipple, 2019).

Supplementary Method 3: Elevation, uplift, uplift rate

The elevation of the paleo shorelines ($h_{\text{PaleoShore}}$) was measured in the field

with different techniques described above. Its standard uncertainty is calculated as:

$$\sigma h_{paleoShore} = \sqrt{\sigma h_{sedi}^2 + \sigma h_{measure}^2}$$

where $\sigma h_{measure}$ represents the uncertainty related to the measurement method and σh_{sedi} represents the sedimentological uncertainty of the determination of the shoreline. This σh_{sedi} was determined in the field, outcrop by outcrop and only apply to the positive uncertainty as corals never develop above sea level.

In case of measurement based on the DEM, the $\sigma h_{measure}$ was calculated in a conservative way as:

$$\sigma h_{measure} = \sqrt{\sigma h_{abs}^2 + \sigma h_{pos}^2}$$

With σh_{abs} being the absolute altitude uncertainty of SRTM data of 6 m such as determined in the literature (Rodríguez *et al.*, 2006; Elkhachy, 2018). To account for degraded uncertainty in steep terrain and positional accuracy, the term σh_{pos} was added and calculated as the difference between the measured value and the minimum and maximum values of the DEM in a radius of 40 m (the DEM resolution being 30 m). This term is therefore different for positive and negative uncertainty.

The uplift (u) and its uncertainty was calculated as:

$$u \pm \sigma u = h_{paleoShore} - h_{rsl} \pm \sqrt{\sigma h_{rsl}^2 + \sigma h_{paleoShore}^2}$$

Where $h_{paleoShore}$ is the measured elevation of the paleo shoreline. The relative sea-level (h_{rsl}) is taken from Grant *et al.* (2014) As their values are relative to the nearby Hanish Sill, no

correction for glacio-hydro-isostatic contributions was necessary. According to Grant *et al.*, the relative sea-level during MIS5e was 5 ± 4 m. As it is unknown to which exact peak the MIS 7 sediments correspond, we used an average age of 219 ± 23 ka. MIS 7 a and e were at -12 m and MIS 7c was at -27 m. We therefore used an average sea-level value of -19 ± 11 m.

The uplift between two stages and its standard uncertainty are calculated as:

$$\Delta u \pm \sigma \Delta u = u_2 - u_1 \pm \sqrt{\sigma u_2^2 + \sigma u_1^2}$$

The tectonic uplift rate is calculated as:

$$\delta u = \frac{u}{\Delta t}$$

Where Δt is the duration of the considered interval.

The standard uncertainty of the tectonic uplift rate is calculated as:

$$\sigma \delta u = \sqrt{\left(\frac{1}{\Delta t}\right)^2 \sigma \Delta u^2 + \left(\frac{u}{\Delta t^2}\right)^2 \sigma \Delta t^2}$$

With

$$\sigma \Delta t = \sqrt{\sigma t_2^2 + \sigma t_1^2}$$

Where $\sigma \Delta t$ is the uncertainty on the time interval. Between MIS 5e and Recent times, Δt is considered to be 128 ± 2 ka (Grant *et al.*, 2014). Again, due to the uncertainty of MIS 7 sediments, we consider Δt between MIS 7 and Recent times to be 219 ± 23 ka, corresponding to the average between MIS 7a and e and an uncertainty containing all peaks. Δt between MIS 7 and MIS 5e is 91 ± 23 ka.

For the scenarios shown on fig.8, we assumed a $\sigma h_{paleoShore}$ over the gateway of 5 m and a σh_{rsl} of ± 4 m. MIS 7a, was

considered as 197 ± 2 ka, MIS 7c as 215 ± 2 ka and MIS 7e as 240 ± 2 ka (Grant *et al.*, 2014).

Supplementary Method 4: Amphila gateway elevation

The past elevation of the present-day Amphila gateway elevation (h_{paleoG}) is calculated as:

$$h_{paleoG} = h_{presentG} - u$$

Where $h_{presentG}$ is the present gateway elevation (16 m.a.s.l.) and u , the uplift defined above.

The standard uncertainty is calculated as:

$$\sigma h_{paleoG} = \sigma u = \sqrt{\sigma h_{paleoShore}^2 + \sigma h_{rsl}^2}$$

Where σh_{rsl} is defined above. For this area, the $\sigma h_{paleoShore}$ has been estimated at ± 5 m.

The future elevation of the gateway is extrapolated as:

$$h_G(t) = \left[\frac{h_{presentG} - h_{MIS5G}}{t_{MIS5}} \right] t + h_{presentG}$$

With h_{MIS5G} and t_{MIS5} the gateway elevation during MIS 5e and the age of MIS5e (127 ka) respectively. Its uncertainty is extrapolated as:

$$\sigma h_G(t) = \frac{\sigma h_{paleoG}}{t_{MIS5}} t$$

Supplementary Method 5: Zula gateway elevation

The time before the Zula gateway will reach present sea level ($t_{h=0}$) is calculated as:

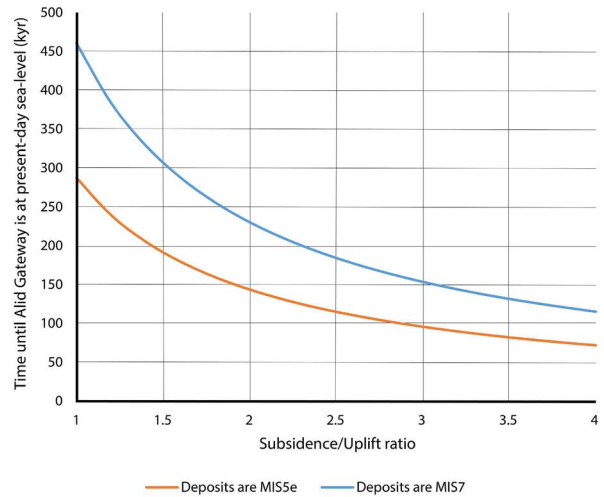
$$t_{h=0} = \frac{h_{presentG}}{\delta u R_{s/u}}$$

With:

$$\delta u = \frac{h_{paleoShore} - h_{rsl}}{\Delta t}$$

Whereas δu is the uplift rate and $R_{s/u}$ is the subsidence/uplift ratio. The present gateway elevation is 215 m and the outcrop elevation is at 100 m. $t_{h=0}$ is then plotted against the subsidence/uplift ratio. As the age of the highest carbonate deposits is unknown, two lines were calculated, assuming ages of MIS5e and MIS7e, representing both sides of the possible spectrum.

Subsidence/Uplift ratio estimations are ranging from 1.1 to 3.5 based on calculations in the Corinth rift (McNeill & Collier, 2004; de Gelder *et al.*, 2019). The timing that the Zula Gateway will reach present-day sea level ranges between 80 and 420 kyr.



Supplementary figure 3: Graph representing the time until the Alid Gateway is at present-day sea-level as a function of the subsidence/uplift ratio for scenarios where deposits spotted on the SE shoulder of the Buia graben are either MIS5e or MIS7.

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IV LATE PLEISTOCENE TO HOLOCENE EVAPORITES IN THE DANAKIL DEPRESSION: A MODERN SALT GIANT ANALOGUE?

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ABSTRACT

Salt giants are unique features of the geological record but remain poorly understood due to the absence of modern analogues. The origin of ancient evaporites and their highly variable sedimentation rates are often debated and ambiguous. The Danakil rift basin in northern Afar features several hundreds of meters of evaporites with deposition still going on today, and as such represents an unrecognized modern salt giant analogue. This study focuses on the multi-proxy analysis of a 625 m deep core in the central part of the basin. The core record, dominated by halite with subordinate clastic sediments, reveals for the first time the Upper Pleistocene to Recent geological, environmental and climatic history of the basin. Sediments witness restricted conditions of the last marine incursion (~ MIS 5e) followed by a hypersaline stage leading to the near-desiccation of the basin with the deposition of thick halite and potash deposits. Subsequent recycling of marginal halite deposits by meteoric waters in lacustrine and salt pan environments significantly increased the evaporite thickness in the subsiding central part of the basin. These findings have implications for the understanding of older fossil salt giants. They show that thick evaporite deposits can form in a very short period by evaporation of meteoric and seawater with only minor deep hydrothermal influences at basin scale.

IV.1 INTRODUCTION

Salt giants are thick and extensive evaporite deposits (100s to 1000s m thick and >100'000 km² according to the definition of Rodriguez *et al.* (2018) and are present in many sedimentary basins across the world (Hübscher *et al.*, 2007; Warren, 2010). These deposits are still poorly understood, mainly due to the absence of modern analogues. In particular, the mechanisms of formation of these evaporites, the hydrological and environmental conditions of the basin, and the surprisingly high sedimentation rates remain ambiguous and debated (e.g. Hsü, 1972; Hübscher *et al.*, 2007; Ryan, 2009; Scribano *et al.*, 2017). The deposition of salt giants might have had profound implications on global seawater salinity variations, thermohaline oceanic circulation, climate and life evolution (Hay *et al.*, 2006; Shields & Mills, 2020).

The Danakil evaporites are considerably smaller compared to classical salt giants. The Pleistocene Danakil Sea had a surface of approximately 9000 km², which is more than two orders of magnitude smaller than the Messinian basins of the Mediterranean (Warren, 2010). However, their significant thickness, marine origin and occurrence in an active rift setting mimic some aspects of ancient salt giants, e.g. the Carboniferous to Permian evaporites of the Barents Sea (Hassaan *et al.*, 2020), the Permo-Triassic Zechstein evaporites (Soto *et al.*, 2017), the Triassic to Jurassic evaporites of the northern Atlantic (Holser *et al.*, 1988; Decalf & Heyn, 2023), the Jurassic Louann Salt of the Gulf of Mexico (Hudec *et al.*, 2013), the Aptian salt of the southern Atlantic (Kukla *et al.*, 2018; Rodriguez *et al.*, 2018), or the Miocene Red Sea evaporites

(Hughes & Beydoun, 1992; Orszag-Sperber *et al.*, 1998; Smith & Santamarina, 2022). Processes related to the formation of the Danakil evaporites might therefore closely resemble those that formed past salt giants. Due to a more precise age control combined with an actualistic and process-based approach, the Recent Danakil Evaporites form an excellent analogue for their older and larger counterparts, offering new keys to understand these enigmatic deposits.

The Danakil evaporites are poorly studied as harsh environmental conditions, complex logistics and safety concerns largely prevented detailed field studies in the past. However, during the last decade, field studies have shed a new light on the sedimentological history of the depression and, in particular, on the Middle to Late Pleistocene Red Sea marine incursions (Atnafu *et al.*, 2015; Jaramillo-Vogel *et al.*, 2019, 2023). In the framework of these studies, data and unique core material from industrial companies were made available for scientific studies. Two cores in the basin were studied using a multiproxy approach including macro- and microfacies analysis, component analysis, the study of elemental and mineralogical content, organic matter analysis and absolute datings.

The main aim of the present study is to better understand (1) the timing and mechanisms of formation of the Danakil evaporites, (2) the paleoenvironmental and hydrological evolution of the basin and (3) the paleoclimatological conditions of the area. The latter is largely unknown in the Danakil Depression but of great importance for understanding the factors driving the out-of-Africa migration of *Homo sapiens* (Beyin, 2006, 2021; Abbate

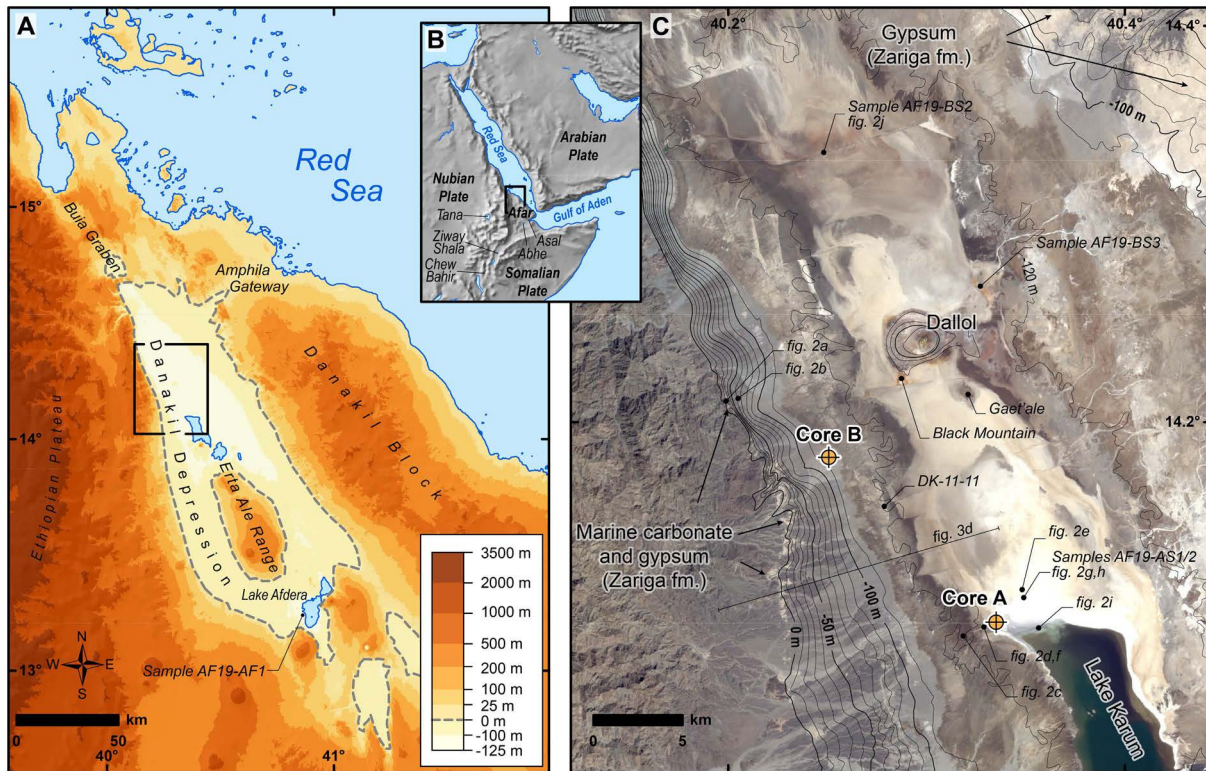


Figure 1: Maps of the study area. A) Topographic map of the Danakil Depression. B) Red Sea & Gulf of Aden region. Lake Abhe, Tana and Asai and Chew Bahir basin are indicated. C) Landsat 8 image of the salt plain with contours of the topography below sea level from SRTM. Main features and the localization of cores, samples and pictures are indicated.

& Sagri, 2012; Tierney *et al.*, 2017; Hill *et al.*, 2022). Finally, these results will help to understand how the sedimentary architecture of rift basins is shaped, in particular in the presence of thick evaporites.

IV.2 GEOLOGICAL SETTING

IV.2.1 The Danakil Depression

The Danakil evaporites are deposited in the Danakil Depression (Fig. 1A), a rift valley reaching elevations of 125 m below sea level (Fig. 1A). It is part of the Afar Depression, forming a triple junction between the Nubian, Arabian and Somali plates (Fig. 1B). The Danakil Block (Fig. 1A) is rotating counter-clockwise relative to Nubia, causing rifting in the Danakil Depression

(McClusky *et al.*, 2010; Viltres *et al.*, 2020; Rime *et al.*, 2023); chapter II, this volume). In the Danakil Depression, rifting is thought to represent an advanced stage of continental breakup with the Erta Ale volcanic range possibly marking crustal separation (Barberi & Varet, 1970; Barberi *et al.*, 1970, 1972; Bastow & Keir, 2011; Keir *et al.*, 2013; Le Gall *et al.*, 2018). The rotation of the Danakil block initiated approx. 10 Ma ago (Eagles *et al.*, 2002; McClusky *et al.*, 2010; Reilinger & McClusky, 2011; Rime *et al.*, 2023). The earliest continental syn-rift sediments of the Danakil Depression are Late Miocene to Pliocene Red Bed Series (Brinckmann & Kürsten, 1971; Le Gall *et al.*, 2018; Rime *et al.*, 2023; chapter II, this volume). During the Middle to Late Pleistocene, Red Sea incursions flooded the depression,

leading to the formation of fringing coralgall reefs topped by gypsum at the basin margins and forming the Zariga Formation (Faure & Roubet, 1968; Holwerda & Hutchinson, 1968; Lalou et al., 1970; Bonatti et al., 1971; Brinckmann & Kürsten, 1971; Atnafu et al., 2015; Jaramillo-Vogel et al., 2019, 2023). Field investigations showed at least 4 generations of coralgall terrasses (chapter III, this volume) and Jaramillo-Vogel et al. (2019) and Foubert et al. (in prep.) evidenced that the two most recent incursions date back to Marine Isotopic Stage (MIS) 5e (~129 ka) and MIS7 (~200 ka). Subsequent basin closure led to the desiccation of the basin and the deposition of extensive evaporite deposits.

IV.2.2 Present-day sedimentation

Active evaporite deposition is still taking place in the depression. At the foot of the escarpment, fluvial channels are eroding the basement and older sediments (Fig. 2A). Recent sheetfloods caused by seasonal sporadic precipitation events (Fig. 2B) result in the deposition of conglomerates with up to meter-sized boulders. Grain size decreases distally, with silt and mud-sized particles at the border of the lower part of the depression, forming a mud flat (Fig. 2C). The facies then gradually evolve into a saline pan characterized by large halite polygons (Fig. 2D). Artisanal mining of this halite is

known since centuries (Munzinger, 1869) and was still ongoing in 2020. Halite is formed each year during flooding of the salt plain by ephemeral and very shallow lakes (López-García *et al.*, 2020). Halite rafts and hopper crystals (Fig. 2E, F) form at the water-air interface by evaporation and sink to the lake floor. On the lake floor, halite is growing further as syntaxial overgrowths on settled rafts and hoppers, forming chevron halite beds and crusts (Fig. 2G). During subsequent desiccation, halite polygons develop within the halite crusts with secondary growth of halite into open desiccation cracks and fractures forming tepees (Fig. 2D). With time, halite crusts and darker laminae representing impurities due to fluvial or eolian input form the uppermost sediment fill of the basin (Fig. 2H). During dry periods, the water table is very close to the surface and Lake Karum is fed by saline springs (Fig. 2I). These waters drain from the escarpment. They have low salinities when flowing through pre-rift basement and sediments at the escarpment (Kebede *et al.*, 2008; Meaza *et al.*, 2019) but the measured salinity increases drastically in the subsurface when entering the depression (Ralph M. Parsons Company, 1967), indicating dissolution of ancient evaporites.

IV.2.3 Lithostratigraphy

The subsurface geology of the basin has mainly been described by the potash

Figure 2 ► : Present-day sedimentary facies in the Danakil Depression. The approximate position of each picture is presented in Fig. 1C. **A)** Fluvial channel cutting through older sediments. The salt plain and Dallol volcano can be recognized in the far distance. **B)** Sheetflood event transporting clastic material into the basin **C)** saline mudflat. **D)** Salt flat with polygonal and tepee structures. **E)** Halite rafts forming in a karstic pool at the surface of the salt plain. **F)** Recent rafts deposited on the salt flat during the last flooding. **G)** Euhedral halite crystals. Note that these crystals were found approx. 5-10 cm below the surface. **H)** Uppermost layer of the salt flat. White arrows point towards chevron halite crystals. **I)** Cold spring feeding lake Karum. **J)** Enigmatic, red-coloured pond on the salt flat.



Name(s)	Lithology	Interpretation/ Remark
Clastic overburden	Mainly consists of alluvial gravels, fine siliciclastic material, intercalated beds of gypsum and displacive halite	This unit is often poorly described as it was almost never cored. It was described to thicken towards the western margin of the depression. While some cores have a sharp contact with the underlying halite, others show a more gradual transition (Holwerda & Hutchinson, 1968).
Upper halite unit / upper rock salt / younger halite	Composed of medium to coarse-grained, bottom-nucleated poorly bedded halite with microkarst textures, variable anhydrite and clastic content, the latter increasing upwards	Warren (2016) interprets this unit as corresponding to a freshening upward cycle developed in a saline pan with meteoric water influence.
Houston formation	Divided in four subunits: (1) the Kainite Member formed by laminated kainite and halite; (2) the Intermediate Member composed of mixed and varying mineralogies often featuring a lower carnallite bed, a bischofite bed and an upper carnallite bed ; (3) the Sylvite Member , rich in sylvite, polyhalite, anhydrite and halite; (4) the Marker Beds formed by bedded halite with wavy anhydrite seams and clays.	Holwerda & Hutchinson (1968) interpret the Houston fm. as the primary deposition of a lower kainite bed overlain by a carnallite-rich bed that was leached in its upper part, removing $MgCl_2$ to form the Sylvite Member. Warren (2016) also interprets the Kainite Member as a primary, subaqueous, depositional unit formed by the evaporation of seawater. Conversely, the Intermediate Member is interpreted to have experienced significant alteration and early diagenesis by shallow phreatic brines, possibly formed during the instalment of a freshened brine lake or seaway. The Sylvite Member is also interpreted as a recrystallized unit, possibly of a carnallite precursor, containing barren zones of halite indicating the inflow of fresher waters.
Lower halite unit /lower rock salt / old salt	Mainly formed by halite and occasional seams of anhydrite	Warren (2016) interprets this unit as a shallowing-upwards sequence deposited in a subaqueous marine-fed basin.
Basal clastic unit	Clastic material ranging from clays to conglomerates	Poorly described as core material is scarce
Deeper	Largely unknown	Only few wells were drilled below the basal clastic unit but Holwerda & Hutchinson (1968) and Pudovskis et al. (2012) mention a possible 'deep' potash layer at 930 m in the central part of the basin and at 532 m on the western margin.

Table 1: Summary of the descriptions of the basin stratigraphy published in the scientific literature or internal reports of potash companies. Industrial nomenclature after Holwerda & Hutchinson (1968), Rauche & van der Klauw (2011), Pudovskis et al. (2012) and Warren (2016).

Figure 3 ► : Correlation between outcrop and well data of the Danakil Basin. **A)** Conceptual sketch of Late Pleistocene sediments belonging to the Zariga Formation (after Jaramillo-Vogel et al., 2019). **B)** and **C)** Lithological column of Cores B and A (the width of the column represents the competence of the lithological units). **D)** Topographic profile through the depression with the position of cores and outcrops from SRTM data. The position of A, B and C is presented in Fig. 1C. Note that the scale of A) is approx. 10x that of B) and C).



industry (e.g. Rauche & van der Klauw, 2011; Pudovskis et al., 2012) and in the scientific literature by Holwerda & Hutchinson (1968), Warren (2016), Bastow et al. (2018) and Bekele & Schmerold (2020). A summary of these works is presented in Table 1. The industrial units are also presented in Fig. 3C.

Based on geophysical data, Behle et al. (1975) estimate the subsurface sediment thickness south of Dallol to be between 2.2 and 3.5 km, the former being more plausible. These results were confirmed by Makris & Ginzburg (1987). Conversely, Tazieff (1969) and the C.N.R - C.N.R.S. Afar Team (1973) mention unpublished data showing thicknesses of >4 km and 5.5 km respectively. The deepest wells of the depression have 975 m of “virtually pure, bedded halite” (Holwerda & Hutchinson, 1968), confirming the significant evaporitic infill.

While it is clear that the uppermost part of the salt plain represents present-day sedimentation (López-García et al., 2020), the age and timespan of the underlying evaporites is largely unknown and debated. Based on faunal evidence, Holwerda & Hutchinson (1968) propose a mid or late Quaternary age. Brinckmann & Kürsten (1971) mention Pliocene to Quaternary ages. Hutchinson & Engels (1970) reports K/Ar datings on sylvite with ages ranging between 76 and 88 ka $\pm$ 15%. Based on pers. comm. with R.W. Hutchinson, the C.N.R - C.N.R.S. Afar Team (1973) report an additional age of 125 ka. However, the diagenetic nature of the sylvite deposits in the Danakil Depression (see below) and the known low reliability of this dating technique on sylvite (Hurley, 1966; Brookins et al., 1980) strongly question these results.

Due to their similarities with Red Sea evaporites, some authors even hypothesized that the lower evaporites might be Miocene in age (Talbot, 2008; López-García et al., 2020).

The stratigraphy of the basin is complicated in its central part by the Dallol volcano (see Fig. 1C) and surrounding hot springs (Holwerda & Hutchinson, 1968; Talbot, 2008; Nobile et al., 2012; Master, 2016; Warren, 2016; Cavalazzi et al., 2019; Kotopoulou et al., 2019; López-García et al., 2020; Otálora et al., 2022). The Dallol volcano is a volcanic edifice with a ~2.4 km deep magma chamber that witnessed a notable dyke intrusion in 2004 (Nobile et al., 2012). Intrusions are believed to have caused dehydration and hydrolysis of evaporite minerals, the doming of the edifice and hydrothermal activity at the surface (Otálora et al., 2022). Talbot (2008) proposes that the dome is mainly formed by marine salts “with minor local additions of hydrothermal salts”. The *Black Mountain*, situated SW of Dallol (Fig. 1C), is a hot spring with MgCl₂-rich fluids crystallizing bischofite and carnallite (Holwerda & Hutchinson, 1968; Kotopoulou et al., 2019; López-García et al., 2020). One of its ponds, *Black Lake*, is probably the most saline body in the world (>70%, Belilla et al., 2019). Similarly, *Gaet’ale* Pond (also called *Yellow Lake*) is extremely saline, dominated by CaCl₂ and MgCl₂ salts (Pérez & Chebude, 2017; Kotopoulou et al., 2019) and producing carnallite deposits (López-García et al., 2020).

IV.3 MATERIAL AND METHODS

Two cores were studied and sampled, Core A (drilled by BHP Billiton, industry name: DAN0003D, 14.0986N, 40.3349E,

120 m.b.s.l., Fig. 1C, 3C) and Core B (drilled by Allana Potash Corp., industry name: DK-11-14, 14.182N, 40.2506E, 116 m.b.s.l., Fig. 1C, 3B). Core A is the only available deep (625m) well in the basin cored throughout its entire length and reaching sediments below the Lower Halite Unit. This core is situated in the central part of the basin and unpublished seismic reflection data (Pudovskis *et al.*, 2012) do not show any unconformity, indicating a continuous sedimentary record. For those reasons, Core A was identified as the reference core within the framework of this study. In total, 277 bulk rock samples were taken from this core for sedimentary petrography, component analyses (microfauna), geochemistry (TOC, XRF), dating (^{14}C) and mineralogical analyses (XRD) resulting in more than 600 measurements (Table S4). Core B is 155 m long, reaches the *basal clastic unit* and is cored along its entire length except for the upper 39 m. Core B was drilled in a more marginal position of the basin (Fig. 3D), west of the major fault system forming the central part of the Danakil graben (Bastow *et al.*, 2018). 13 samples of this core were analysed for microfauna.

5 additional evaporite samples were taken at the surface of the depression (Fig. 1A, C, Table S4) for elemental and mineralogical analysis.

IV.3.1 Facies analysis

Facies analysis of Core A is based on the macroscopic description of full and half core sections (Fig. 4), unpublished geological descriptions by BHPB in Pudovskis *et al.* (2012), and the petrographic study of core slabs and thin sections (Fig. 5). Halite and other consolidated sediments were cut dry from half core sections as 5 to 10 cm slabs with

a circular saw mounted with a diamond blade. For thin section preparation, the slabs were further cut with a dry diamond wireline saw. Petrographic thin sections were prepared at the Department of Geosciences of the University of Fribourg using a diamond polishing lubricated with petrol to avoid dissolution.

The facies of Core B was described based on visual inspection of half and full core sections, combined with the analyses of pictures. In the case of non-recovery, downhole logging data reported by (Rauche & van der Klauw, 2011) were used to refine facies description.

Non-evaporitic sediments are mainly described according to grainsize (clay, silt, sand and gravel). Evaporites are described based on their main mineralogy (sulphates, potash and halite) and halite texture (clear, medium to coarse-grained bottom-growth halite; milky, cloudy chevron halite and laminated, fine-grained halite). Note that the term potash is used here in its broadest sense, including all high-concentration evaporite minerals and not only K-rich varieties.

IV.3.2 Component analysis

Samples of clastic sediments were sieved wet at 1mm, 125 μm , 63 μm and 32 μm (51 samples from Core A and 13 from Core B). The 63 μm and 125 μm fractions were hand-picked for qualitative microfaunal analyses. Relevant specimens were photographed and identified using standard optical microscopy (Nikon SMZ18) and scanning electron microscopy (Thermo Fischer SEM FEIXL30SFEG, Department of Geosciences, University of Fribourg).

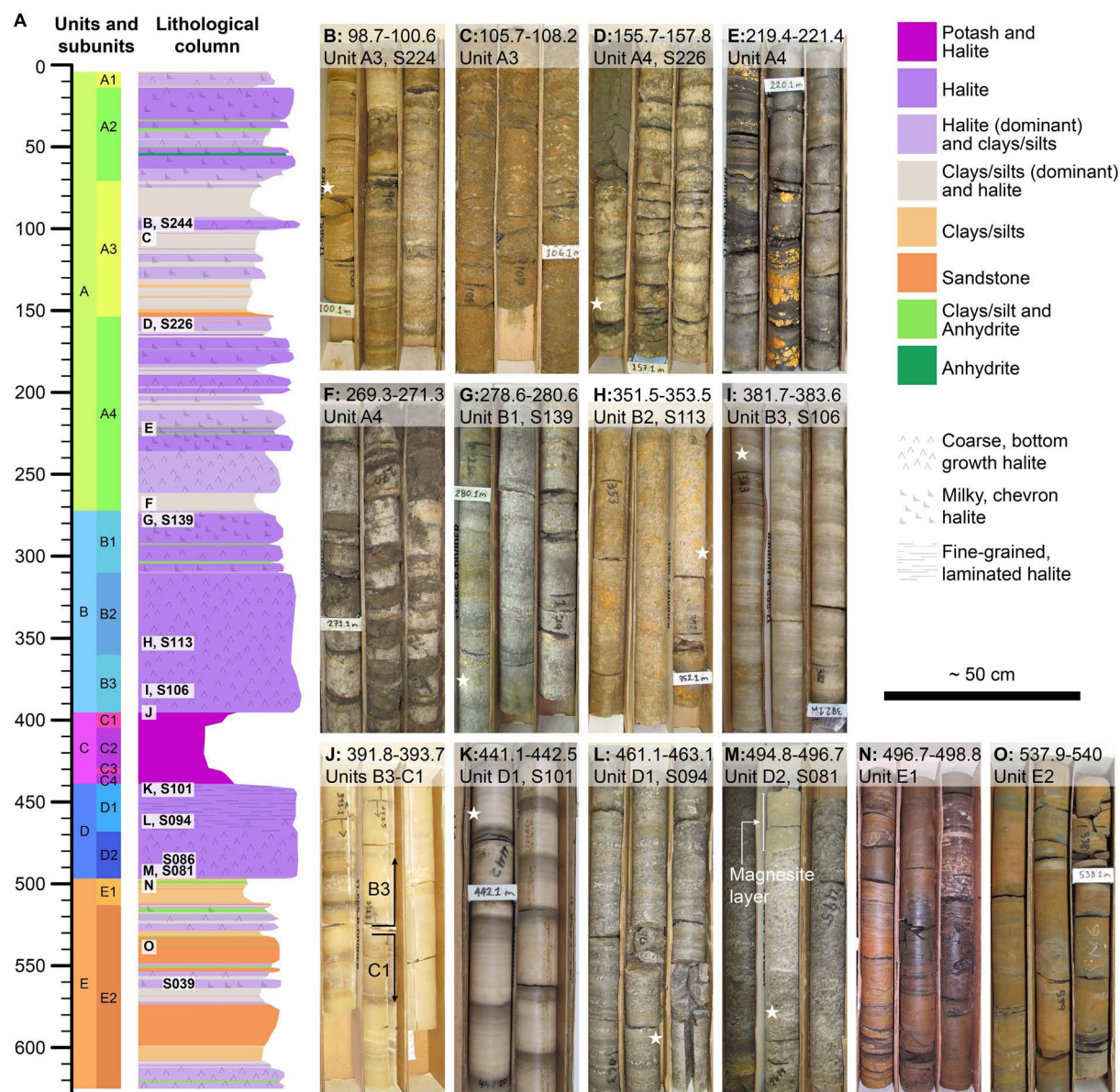
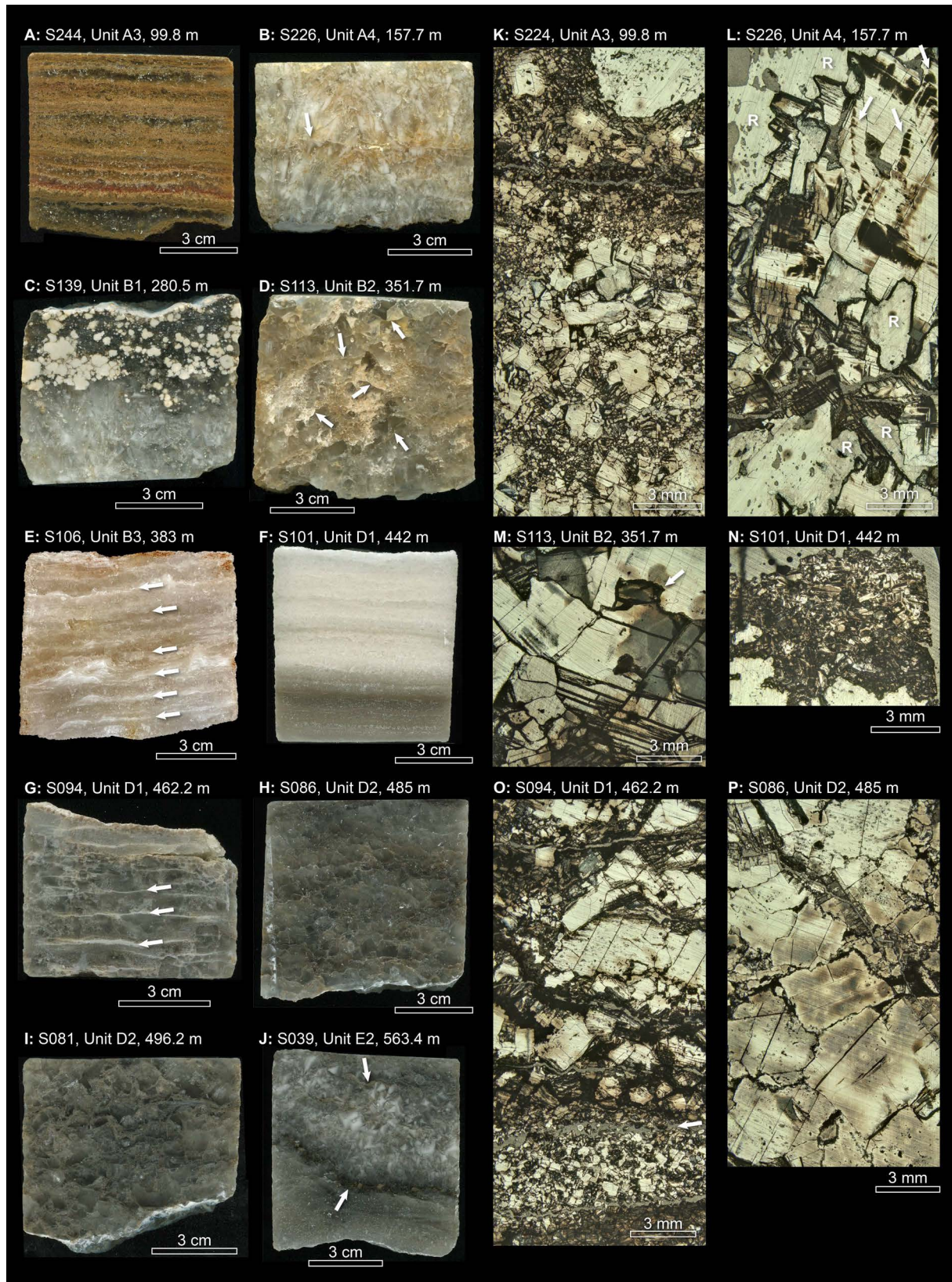


Figure 4: Lithology of Core A. A) Units, subunits and lithological column. B-O) core pictures. Depth, unit and sample numbers are indicated. White stars represent the sample position. B) Laminated medium to fine-grained halite. C) Clays and silts with forams and intra-sediment halite. D) Alternation of cloudy halite and clastic sediments. E) Clays and anhydrite between two halite layers. F) Silt and halite alternations. G) Cloudy halite alternating with anhydrite layers. H) Coarse halite. I) Laminated but coarse halite, with dissolution laminae. J) Conformable transition between unit C1 (carnallite, banded) and B3 (halite). K) Fine-grained, laminated halite. L) Alternation of fine-grained, laminated halite and coarse halite. M) Coarse halite forming the lowermost part of unit B. The greenish magnesite layer is visible. N) Clayey unit forming the uppermost part of unit A. O) Sand layer.

Figure 5 ► : Macro (A-J) and microfacies (K-P) of evaporites. Sample number, unit and depth are indicated. A), K) Fine to medium-grained laminated fabric. B), C), J), L) Cloudy, milky chevron halite. Arrows on B) and J) indicate dissolution surfaces. Arrows on L) show chevron-forming inclusions. "R" show recrystallized halite in cavities. D), E), M) Coarse-grained, bottom growth halite with dissolution surfaces (arrows). F), N) Fine-grained, laminated halite. G), O) Laminated alternations between coarse- and fine-grained halite. The white arrows represent polyhalite layers measured in XRD but removed during thin section preparation, H), I), P) Coarse-grained, bottom growth halite with dissolution features around the individual crystals. J) Cloudy, milky chevron halite with dissolution surfaces (white arrows).



IV.3.3 Geochemistry

In total, 179 samples were analysed for elemental content by wavelength dispersive X-ray fluorescence (WD-XRF) using a PANalytical Zetium XRF spectrometer (Department of Geosciences, University of Fribourg) (see Table S4). Both clastic and evaporite (halite) samples were prepared as pressed pellets (12 g of sample + 3g of wax). Evaporite samples were first prepared according to Herrmann *et al.* (1973) and Moretto (1988). Samples were crushed in ethanol, generously rinsed with ethanol and filtered to avoid fluid inclusion signals. Clastic samples were analysed for major and trace elements using the semi-quantitative Omnia and quantitative ProTrace calibrations and applications developed by Malvern PANalytical, respectively. Major elements, Cl and Br in halite samples were quantitatively measured using a specific calibration running on the Super-Q software and based on 22 in-house prepared pressed pellet standards composed of a mixture of halite, KBr, anhydrite and silicate. Measurement of a short series of trace elements (Sr, Rb, Pb, Zn, Cu, Mn, Ti, Ba) was performed using the ProTrace calibration. All results are given in wt.%.

Bromine content in halite is a proxy for brine parenthood and ion concentration (Boeke, 1908; Valyashko, 1956; Raup & Hite, 1978). As an incompatible element in halite, its abundance in crystals will rise with increasing ion concentration of the brine (e.g. McCaffrey *et al.*, 1987). Conversely, the recycling of marine evaporites by meteoric water will produce halite with lower bromine content (Holser, 1979; Warren, 2016).

There is a long controversy about the Bromine content of the first halite precipitating from marine water and thus, the potential use of Bromine as a proxy for distinguishing marine and meteoric parent fluids. While the theoretical minimum Br content should be 60-75 ppm in marine waters (Holser, 1966; Herrmann, 1972; Herrmann *et al.*, 1973), it has been shown that Br partition coefficient varies according to crystal growth rates, chemical composition of the brines (especially MgCl_2 content) and, to a small degree, temperature (Braitsch & Herrmann, 1963; Holser, 1966; Herrmann, 1972, 1980; Herrmann *et al.*, 1973).

Generally, the threshold values of the Br-content between marine and meteoric fluids varies between 40 and 75 ppm (e.g. Valyashko, 1956; Kuhn, 1968; Herrmann *et al.*, 1973; Kühn & Hsü, 1974; Holser, 1979; Wilgus & Holser, 1984; Holland, 1984; McCaffrey *et al.*, 1987; Garcia-Veigas *et al.*, 1995; Raup & Hite, 1996; Schreiber & Tabakh, 2000; Cendón *et al.*, 2004; Brennan *et al.*, 2013; Warren, 2016; Blamey & Brand, 2019; Ercan *et al.*, 2019; Karakaya *et al.*, 2019).

As pointed out by Hardie (1984), Taberner *et al.* (2000), Babel & Schreiber (2014) and Warren (2016), the identification of marine vs. meteoric parent brines should therefore not only rely on bromine content but also on additional stratigraphic, sedimentological, mineralogical, and geochemical evidence.

Additional geochemical analyses from BHPB and Allana Potash Corp. were used for refining the facies descriptions of the potash interval of Core A and B (Fig. S1 and S2). Geochemical analyses on core A were performed by BHPB using a hand-

held XRF (Innov-X Delta model used in Mining Plus mode, Pudovskis *et al.*, 2012). Geochemical analyses on Core B were performed by Kali-Umwelttechnik GmbH (Sondershausen, Germany) for Allana Potash Corp. by flame emission spectrometry (K^+ , Na^+ , Mg^{2+} , Ca^{2+}) and ion chromatography (Cl^- , SO_4^{2-}) and were reported by Rauche & van der Klauw (2011).

Organic matter analyses of 25 samples from the clastic fraction were performed with Technologies Vinci Rock-Eval 6 at the Institute of Earth Sciences of the University of Lausanne. Total organic carbon (TOC, wt%), hydrogen index (HI, mg HC/g TOC, HC = hydrocarbons) and oxygen index (OI, mg CO_2 /g TOC) were measured (see Table S4). HI and OI give an indication on the origin of the organic matter (e.g. Van Krevelen, 1993).

IV.3.4 Mineralogy

In total, 186 mineralogical analyses on bulk samples (hand-crushed in agate mortar) were performed with the Rigaku Ultima IV X-ray diffractometer (XRD) at the Department of Geosciences, University of Fribourg. The samples were measured from 15 to 70° 2 θ at 1.3°/min (step size 0.02°) with an X-ray source (Cu-anode, LFF) operating at 40kW, 40A and a D/teX Ultra detector. The results were analysed with the Rigaku PDXL2 software, using the Rietveld method for the quantification of non-evaporitic samples. In Fig. 6, phyllosilicates, as well as Ca-carbonates (calcite and high-Mg calcite) are grouped. The quantities are normalized to 100%, excluding evaporite minerals.

IV.3.5 Radiocarbon dating

Radiocarbon dating was performed on small wood fragments, as well as on foraminifera, ostracods and other shell fragments. Organic matter such as wood was treated to remove contamination by carbonates and humic acids. This was done by washing samples with an acid and a base at 60 °C for 1 hr each step (0.5 M HCl, 0.1 M NaOH, 0.5 M HCl, washes with MilliQ water in between) (Hajdas, 2008).

Dry sample material was weighed within aluminum cups for combustion in the Elemental Analyzer. The shells were dissolved in concentrated phosphoric acid. The CO_2 was analyzed using a Gas Ion Source (GIS) at the MICADAS system ETH (Synal *et al.*, 2007). Radiocarbon ages were calibrated using OxCal 4.4 and the IntCal13 calibration curve (Reimer *et al.*, 2013). Foraminifera and marine shell fragments analyzed in 2022 were calibrated using the IntCal20 (Reimer *et al.*, 2020) (see Fig. S3). The ages of the foraminifera and marine shells were close to the limit of radiocarbon dating. No reservoir correction was applied as it is unknown.

IV.4 RESULTS

IV.4.1 Sedimentary petrography

IV.4.1.1 Core A

Core A is composed of ~67% of evaporites (~58% halite, ~7% potash and ~2% anhydrite). Sulphates are only present as anhydrite and not as gypsum.

Besides evaporites, 33% of the core is composed of siliciclastic and carbonate sediments. Their grain-size ranges from clay to sand. Dominant mineralogical

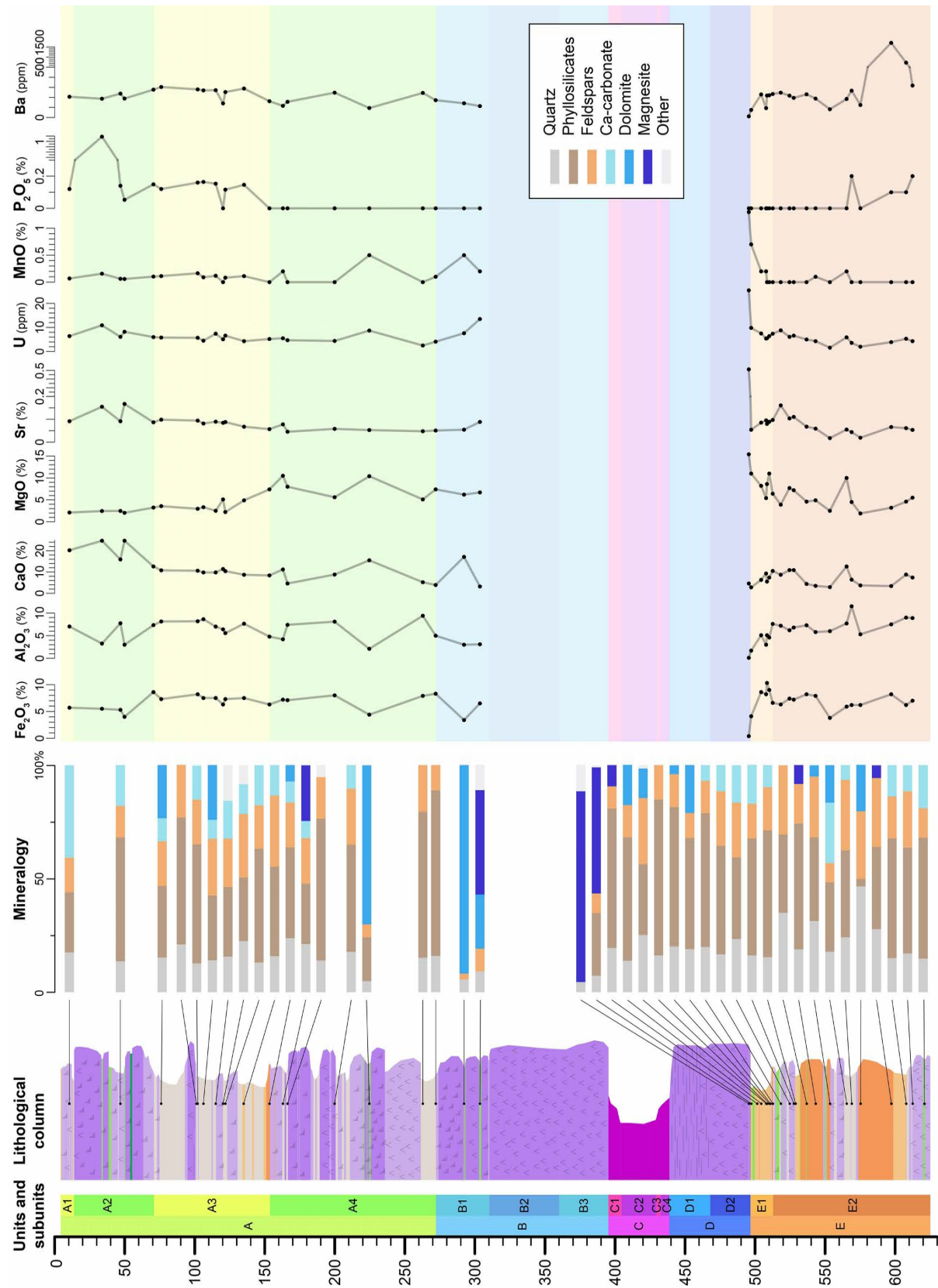


Figure 6: Mineralogical and elemental content of bulk non-evaporitic samples of Core A. Mineralogy from XRD analysis, normalization excluding evaporite minerals. XRF data show Fe_2O_3 , Al_2O_3 , CaO, MgO, Sr, U, MnO, P_2O_5 and Ba. Note the non-linear scale for some of these elements.

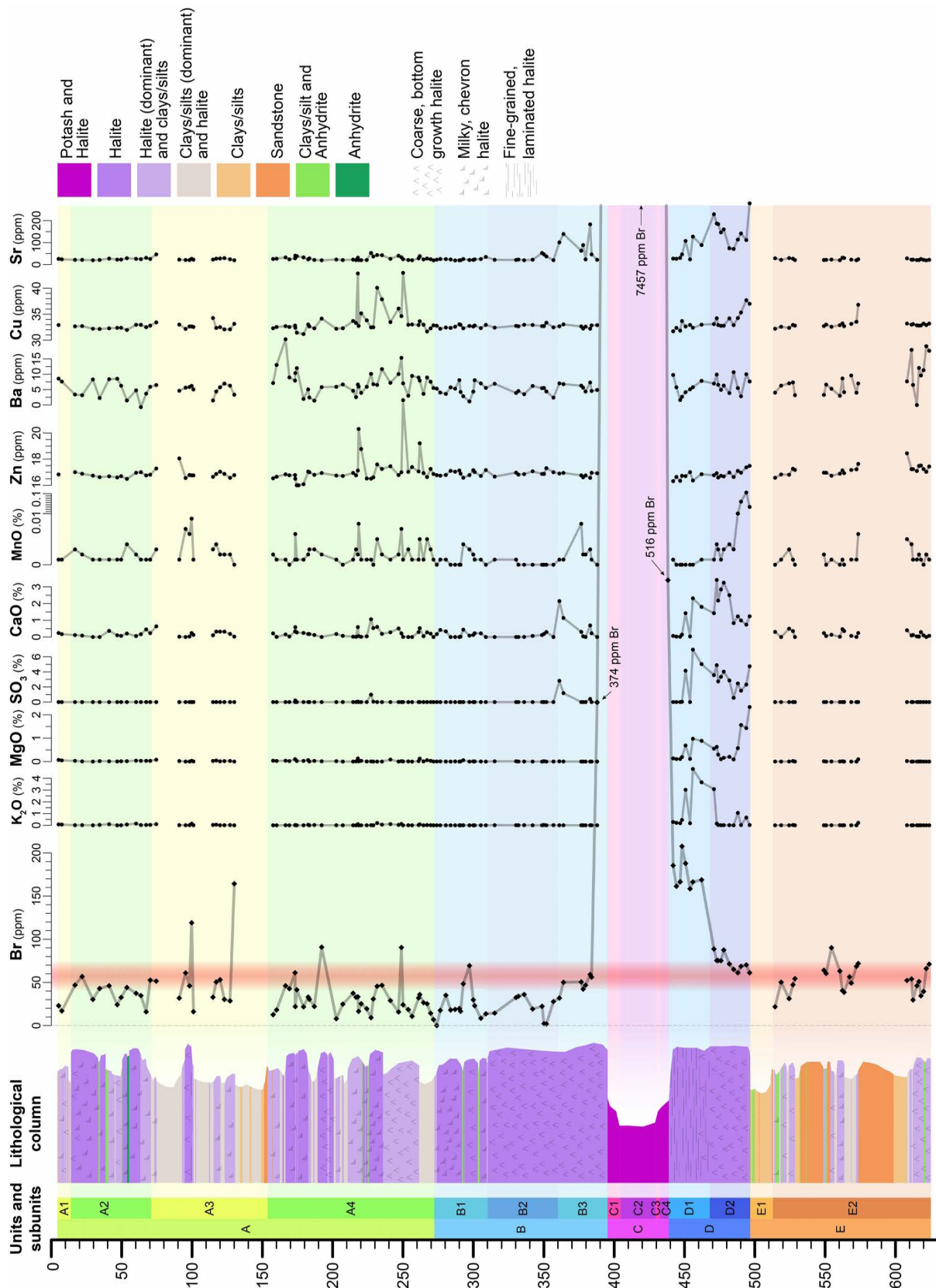


Figure 7: XRF analysis of halite of Core A. Br, K₂O, MgO, SO₃, CaO, MnO (note the non-linear scale), Zn, Ba, Cu and Sr are presented. The red line for the Br values indicates the commonly accepted range for the limit between meteoric (low) and marine (high) parent fluids (see text for discussion).

assemblages in the non-evaporitic samples are phyllosilicates, quartz and feldspars, while carbonates are composed of Ca-carbonates, dolomite and magnesite. These assemblages mainly reflect the composition of sedimentary grains and/or the bioclastic content. Some layers are more enriched in dolomite and magnesite representing chemical precipitates and/or diagenetic minerals. Component analysis revealed the presence of foraminifera, diatoms and shell fragments. Overall, the siliciclastic fraction dominates the non-evaporite part of the core, which will be called “clastics” herein after.

The core has been divided in 5 units and 15 subunits (Fig. 3, 4). Unit E (625.1 – 496.7 m) is dominated by clastics. Subunit E2 mainly shows the alternation between halite and clastics (with variable amounts of calcite). Clastic units are clay to sand-sized and mainly composed of quartz and phyllosilicates with minor amounts of carbonates and feldspars (Fig. 6). Minor anhydrite layers are present. Subunit E1 (Fig. 4N) is formed by silty to clay-rich material, mainly composed of phyllosilicates. Dolomite and magnesite (pure Mg-carbonate) are dominantly present at the top of the subunit.

Unit D (496.7 – 438.83 m) is exclusively composed of evaporites. While dominated by halite, XRF and XRD data (Fig. 7) indicate the presence of anhydrite, polyhalite and possibly kieserite in unit D. Subunit D2 shows clear, coarse-grained bottom growth halite (Fig. 4M, 5H, I). At the base of the unit, XRD data shows a 10 cm interval (Fig. 4M) of magnesite, anhydrite and huntite. Subunit D1 shows the transition from alternations of fine- and coarse-grained halite at the base (Fig. 4L, 5G, O) to fine grained, laminated halite (Fig. 4K,

5F) with cubic crystals (Fig. 5N) in the upper part of the subunit.

Based on facies analyses and portable XRF data, Pudovskis et al. (2012) recognize four mineralogical assemblages in unit C (438.83 – 395.3 m, see Fig. S1). Subunit C4 shows high sylvite content (with K₂O reaching 50%) along with carnallite and kainite. Subunit C3 shows mainly kainite. Subunit C2 shows bischofite with kainite bands. Subunit C1 shows a mix of carnallite, polyhalite, kainite and possibly kieserite. Facies observations and descriptions by the authors confirm this interpretation.

Unit B (395.3 – 272.3 m) is almost entirely composed of halite, except for the presence of minor anhydrite in subunit B3. The facies in subunit B3 changes from medium-grained, clear halite at the base to coarse clear halite with frequent dissolution horizons, resulting in clear laminations (Fig. 4I, 5E). At the top of the subunit, these dissolution surfaces disappear again, only showing (very) coarse clear halite. Subunit B2 is exclusively composed of coarse clear halite (Fig. 4H, 5D). Subunit B1 mainly shows a cloudy chevron texture with some intercalations of anhydrite (Fig. 4G, 5C) and minor intervals enriched in dolomite and magnesite (Fig. 6).

Unit A (272.3 – 4.3 m) is mainly composed of an alternation of halite and clastic material (Fig. 4). Subunit A4 shows an alternation of clastic and evaporitic material (figs. 4A, D, E, F). Halite is either forms enterolithic textures or is present as clear, coarse, bottom-growth halite or as cloudy, chevron halite with dissolution surfaces and recrystallized cavities (figs. 5B, L), indicating subaerial exposure. Some clastic intervals show chicken-wire

anhydrite (Fig. 4E), sometimes associated with dolomite (Fig. 6). Subunit A3 is dominated by clastic material (Fig. 4C). The base of this subunit is marked by sandstone but the rest of the subunit is composed of silt to clay material. Two intervals show a higher halite content (figs. 4A, B, 5A, K). The facies are either cloudy, with the alternation of fine-grained and coarser halite (figs. 5A, K) or clear coarse-grained halite. Subunit A2 presents two intervals nearly devoid of clastic material (14-39 m, 50-63 m) separating a more clastic-rich interval (39-50 m). Most of the halite present a cloudy/milky texture. Subunit A1 is formed by clastics with frequent enterolithic halite crystals or halite beds.

A small amount (ca. 400 individuals) of benthic foraminifera and ostracoda were found in the clastic intervals of the core, especially within subunits E2, A3 and at the base of A2 (Fig. 3C). A total of 15 benthic foraminifera species, were identified (see Fig. 8, 9). Subunit E2 is characterized by the presence of *Ammonia veneta* (over 50% relative abundance), *Quinqueloculina semiluna*, *Triloculina oblonga*, *Brizalina striatula*, *Bulimina elongata*, and *Rosalina globularis* (figs. 8 and 9). Subunit A3 and the base of A2 are more diverse with a total number of 11 identified species. *Ammonia veneta* is also the dominant species, whilst minor species are essentially represented by *Quinqueloculina* spp (*Q. semiluna* and *Q. viennensis*), *Elphidium* spp (*E. williamsoni* and *E. advenum*), *Bulimina elongata*, *Bolivina pseudoplicata*, *Nonion fabum* and *Haynesina depressula* (figs. 8 and 9).

The 4 identified ostracod species belong to the genera *Cyprideis* (both in E2 and A) and *Loxoconcha* (A2) (Fig. 10).

IV.4.1.2 Core B

Core B has been divided into 5 units (see Fig. 3B and S2). Unit ε (155 – 125 m) is composed of clastic material with intervals of clay/silt-sized clastics and intervals of conglomerates with cobble-sized clasts. Some layers of anhydrite are also present. At 154 m, *Ammonia veneta* and *Bolivina pseudoplicata* were found, along with diatoms. Unit δ (125 – 67 m) is composed of halite. The facies shows clear, coarse halite at the base, changing to alternations of fine- and coarse-grained halite and to laminated, fine-grained halite at the top. Rauche & van der Klauw (2011) describe unit γ (67 – 55 m) as mainly consisting of anhydrite with polyhalite and minor amounts of kieserite, carnallite, kainite, tachyhydrite and bischofite, as well as two layers of sylvite, at the top and bottom of the unit. Facies observations by the authors compared with elemental analyses confirm this (see also Fig. S2). Unit β (55 – 37 m) is composed of coarse clear halite. No core sections were recovered above 39 m depth but downhole logging and descriptions by Rauche & van der Klauw (2011) indicate a sharp lithological change at 37 m, marking the lower boundary of unit α (37 – 5.5 m) which is composed of clastic material.

IV.4.2 Geochemistry

IV.4.2.1 Core A

The Br content of halite samples (Fig. 7) ranges between 20 and 90 ppm in unit E. These values increase from 61 to 207 ppm in unit D, reach 7457 ppm in unit C, before decreasing to values below 60 ppm in units B and A. A few outliers show isolated high values in units B1, A4, and A3. K₂O and MgO values are low except in unit D. SO₃ and CaO show low values

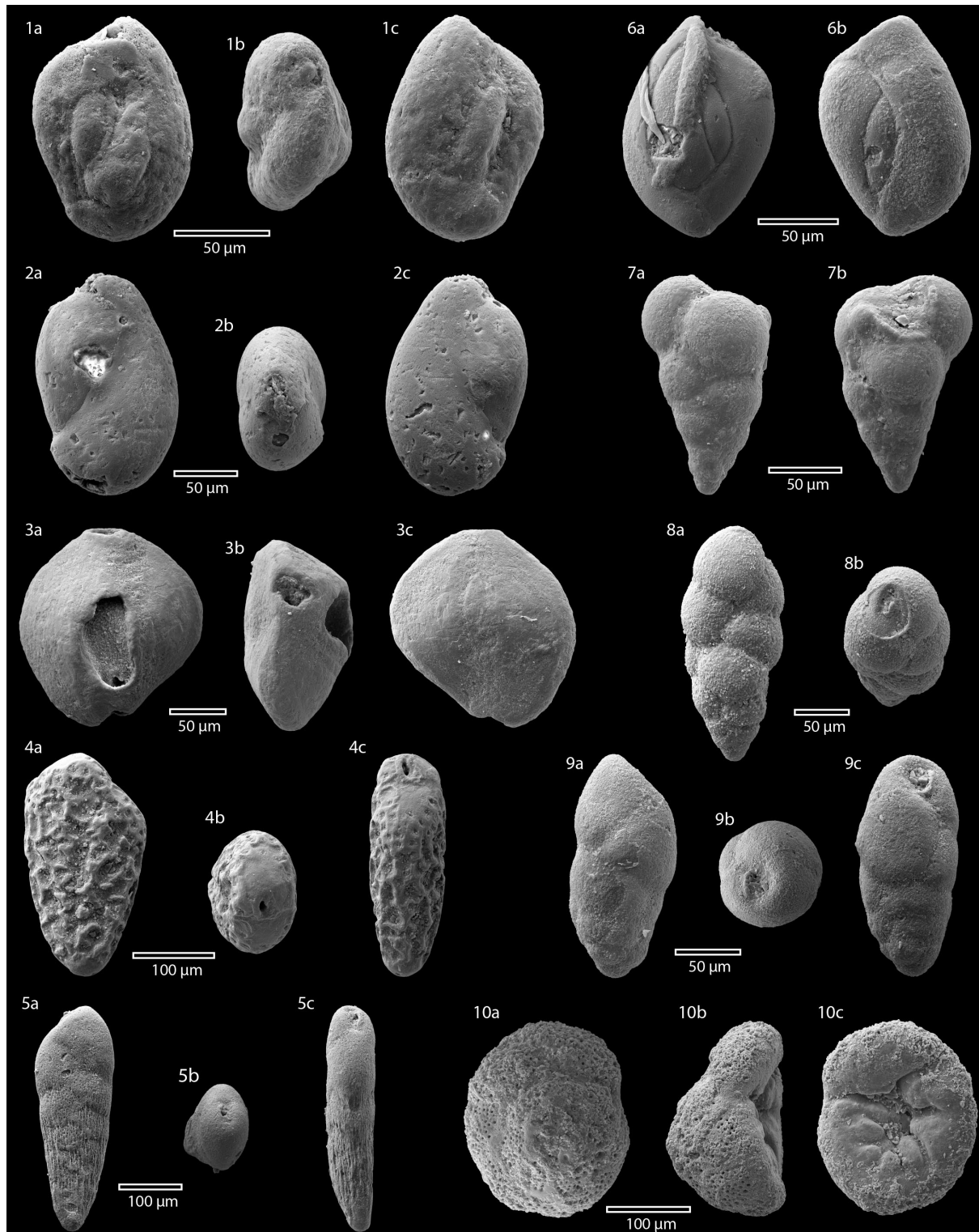


Figure 8: SEM images of foraminifera found in Core A. 1a-c) S240 (106.2 m, A3) *Quinqueloculina semiluna*; 2a-c) S005 (620.7m, E2) *Triloculina oblonga* (or *Miliolinella* sp.); 3a-c) S233 (121.9m, A3) *Miliolid* sp.; 4a-c) S238 (115m, A3) *Bolivina pseudoplicata*; 5a-c) S005 (620.7m, E2) *Brizalina striatula*; 6a-b) S238 (115m, A3) *Quinqueloculina viennensis*; 7a-b) S240 (106.2 m, A3) *Guembelitria* sp.; 8a-b) S251 (70.5m, A2), *Bulimina elongata* (morphotype 2); 9a-c) S038 (565.3m, E2) *Bulimina elongata* (morphotype 1); 10a-c) S032 (567.7m, E2) *Rosalina globularis*.

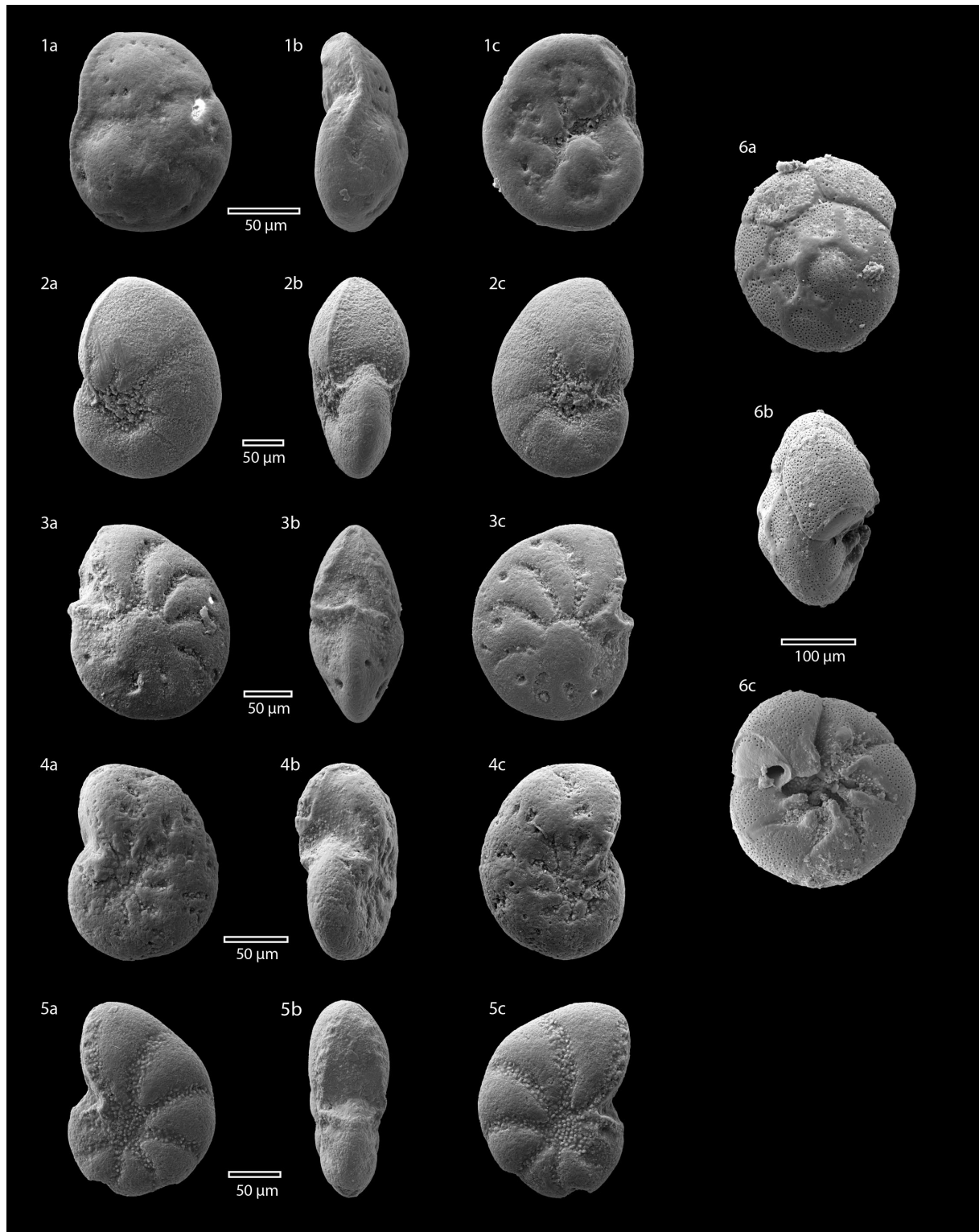


Figure 9: SEM images of foraminifera found in Core A. 1a-c) S238 (115m, A3) *Rosalina bradyi*; 2a-c) S240 (106.2 m, A3) *Nonion fabum*; 3a-c) S229 (135.2m, A3) *Elphidium advenum*; 4a-c) S233 (121.9m, A3) *Elphidium williamsoni*; 5a-c) S238 (115m, A3) *Haynesina depressula*; 6a-c) S006 (620.5m, E2) *Ammonia veneta*.

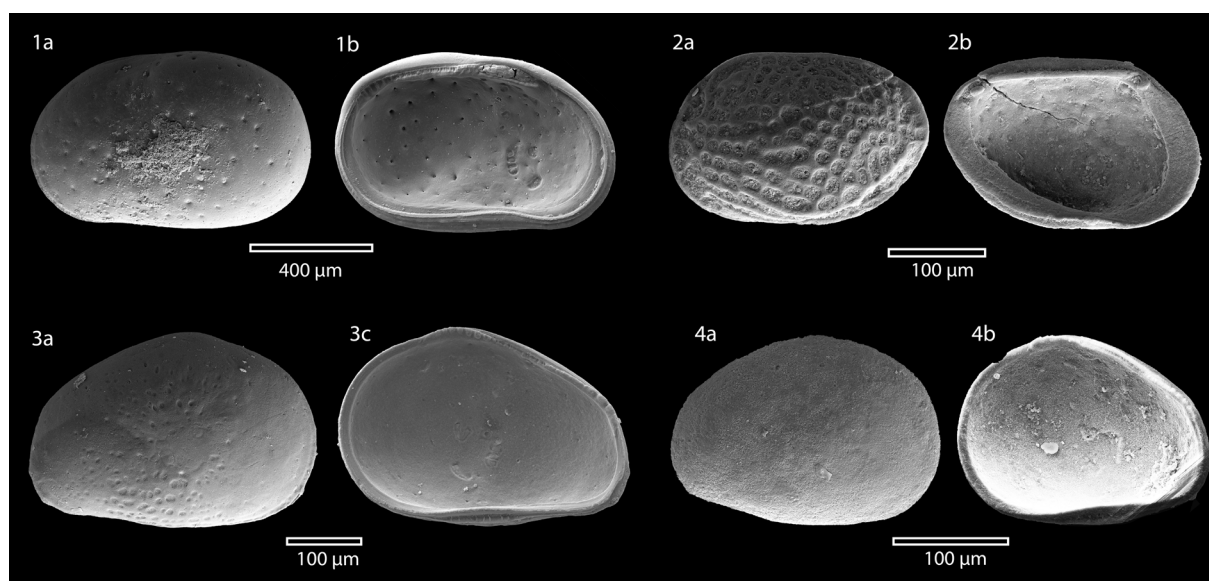


Figure 10: SEM images of ostracods found in Core A. 1a/b) S006 (620.5 m, E2) *Cyprideis torosa* cf. *pertorosa* (Gramann, 1971), female left valve. **2a/b)**: S229 (135.2 m, A3) *Loxoconcha* sp., juvenile left valve. **3a/b)**: S238 (115 m, A3) and **4a/b)**: S275 (10.4, A1) *Cyprideis* sp., juvenile left valves.

except in units D and B3. MnO, Zn, Ba and Cu show low values, but the signal is scattered in A4. MnO furthermore shows a peak (up to 1050 ppm) at the base of unit D. Sr only shows elevated values in units D and B3.

The Al_2O_3 and Fe_2O_3 contents of clastic sediments vary according to variations of phyllosilicates and feldspars (Fig. 6). The CaO content reflects the presence of Ca-carbonates and dolomite shown by XRD. MgO correlates with dolomite and magnesite content. Sr, U and MnO show a single strong peak at the base of unit D. P_2O_5 has increased values at the base of E2 and in units A3 and A2. Ba shows high values at the base of E2.

Organic matter analyses (Fig. 11) show that samples from units A, B and E2 mainly belong to type III or a mixture of type III and type II organic matter, with particularly high O index values. Samples of unit E1 evidence a higher H

index, representing mainly type I (sapropelic) organic matter.

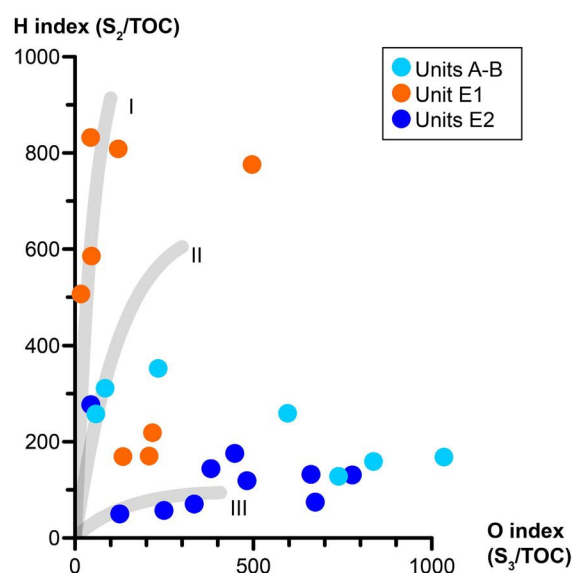


Figure 11: Van Krevelen diagram of non-evaporitic samples of Core A. It shows the ratio between the H index (S_2/TOC) and O index (S_3/TOC). Grey lines indicate the different types of kerogen. Most samples show low H index (type II and type III kerogen), except in subunit E1 with mainly type I kerogen.

IV.4.2.2 Core B

Elemental data (Fig. S2) show a peak in potassium (and to a lesser degree magnesium) at the base and top of unit γ . In-between these two beds, a strong increase in Ca and SO₄ and a decrease in Na and Cl is noted.

IV.4.2.3 Modern salt plain

Samples of Recent evaporites taken from the salt plain have a low Br content (34-47 ppm) for samples AF19-AF1, AF19-AS1/2, compatible with crystallization from saline continental waters of Lakes Karum and Afdera (position in Fig. 1). Conversely, samples AF19-BS2/3 taken near small ponds (including one with a peculiar red colour, Fig. 2J) indicate elevated values of 116 and 216 ppm. These samples also show elevated values of MgO (up to 0.8%) and K₂O (up to 3%), but normal values of Zn, Ba, Cu and Sr compared to other samples. These values are compatible with the XRD identification of sylvite in AF19-BS3.

IV.4.3 Datings

The results of radiocarbon datings are presented on Table S4 and Fig. S3. Two benthic foraminifera samples at 106.2 and 121.9 m were too small to be dated. Two benthic foraminifera samples at 620.5 and 620.7 were too old for radiocarbon dating.

IV.5 DISCUSSION

IV.5.1 Stratigraphy, sedimentation rates and correlation

Radiocarbon dating show that subunits A3 to A1 are Late Pleistocene to

Holocene in age, representing MIS 3 to MIS 1 (Table S4, Fig. 3). The age reversal at 10 m can be explained by the fact that wood is probably transported to the basin from the margin. As no absolute datings could be obtained from the lower part of the core, the chronostratigraphy relies on the correlation with marginal outcrops (Fig. 3).

Based on outcrops of the Zariga Fm. at the basin margin, Jaramillo-Vogel *et al.* (2019) showed that the last flooding of the depression happened during the MIS 5e sea-level highstand. Moreover, the rapid sea-level drop after MIS5e highstand (Grant *et al.*, 2014) indicates that any connection to the Red Sea was cut shortly afterwards. The potash unit and the lower halite unit (units D and C) therefore probably represent this last desiccation stage just after the MIS 5e Red Sea level highstand.

Cycles between fine- and coarse-grained halite or between halite and shale/anhydrite are known to represent annual cycles in the Dead Sea and Messinian evaporites (Manzi *et al.*, 2012; Sirota *et al.*, 2017). Cyclicities are also visible in the Danakil evaporites. Assuming that they also represent annual cycles, the sedimentation rates and duration of individual units can be discussed. At the base of subunit D1, cycles between fine- and coarse-grained halite can be observed. Most of them are between 1 and 6 cm thick with an average of 3 cm (Fig. 5G and 4L). This would indicate that sedimentation rates are comparable to what has been observed in the Dead Sea (Lensky *et al.*, 2005; Sirota *et al.*, 2021) and slightly lower than the 15 cm/yr observed in the Messinian Realmonte salt mine (Manzi *et al.*, 2012). Cycles in the uppermost 9 m of D1 (Fig. 4K) can be up to 25 cm thick (average ~12

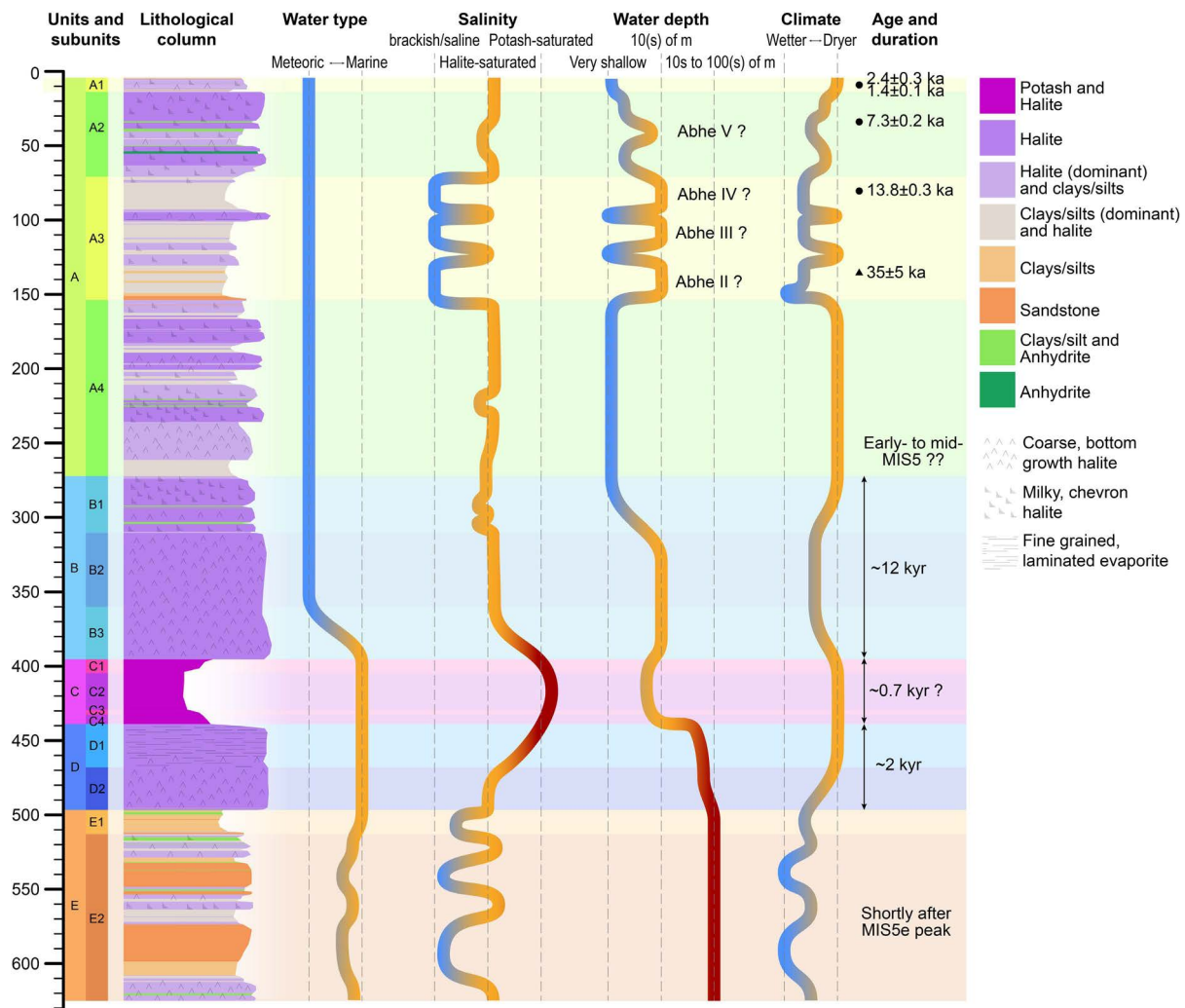


Figure 12: Paleoenvironmental (salinity, water depth and meteoric/marine waters) and paleoclimatic (wet/dry) interpretation of Core A. Age and duration are indicated.

cm), indicating very high sedimentation rates just before the deposition of potash. Assuming an average deposition rate of 3 cm/yr for the lower 49 m of unit D and 12 cm/yr for the upper 9 m, the whole unit would have deposited in less than 2 ka. Subunits C1, C2 and C3 present irregular cycles from 1 to 30 cm thick with an average of 6.5 cm. Assuming that these cycles are annual and assuming a similar sedimentation rate for C4, unit C (with an overall thickness of 44 m) might have deposited in less than 0.7 ka. If unit B was deposited at an average rate of 1 cm/yr as suggested by dissolution events (Fig. 5D, E, M) and knowing that unit B

has a thickness of 123 m, the whole unit could have been deposited in 12 kyr. Assuming that these deposition rates are correct, the transition between units B and A probably happened ~15 kyr after the basin became restricted, i.e. during early- to mid-MIS5.

The different units of Core A can tentatively be correlated with Core B and the outcrops of the Zariga Fm. The foraminifera-bearing, clastic-rich unit E can be correlated with the clastic unit ϵ of Core B (Fig. 3). Subunit E2 is correlated to a unit characterized by gypsum layers, stromatolites and aragonite crusts

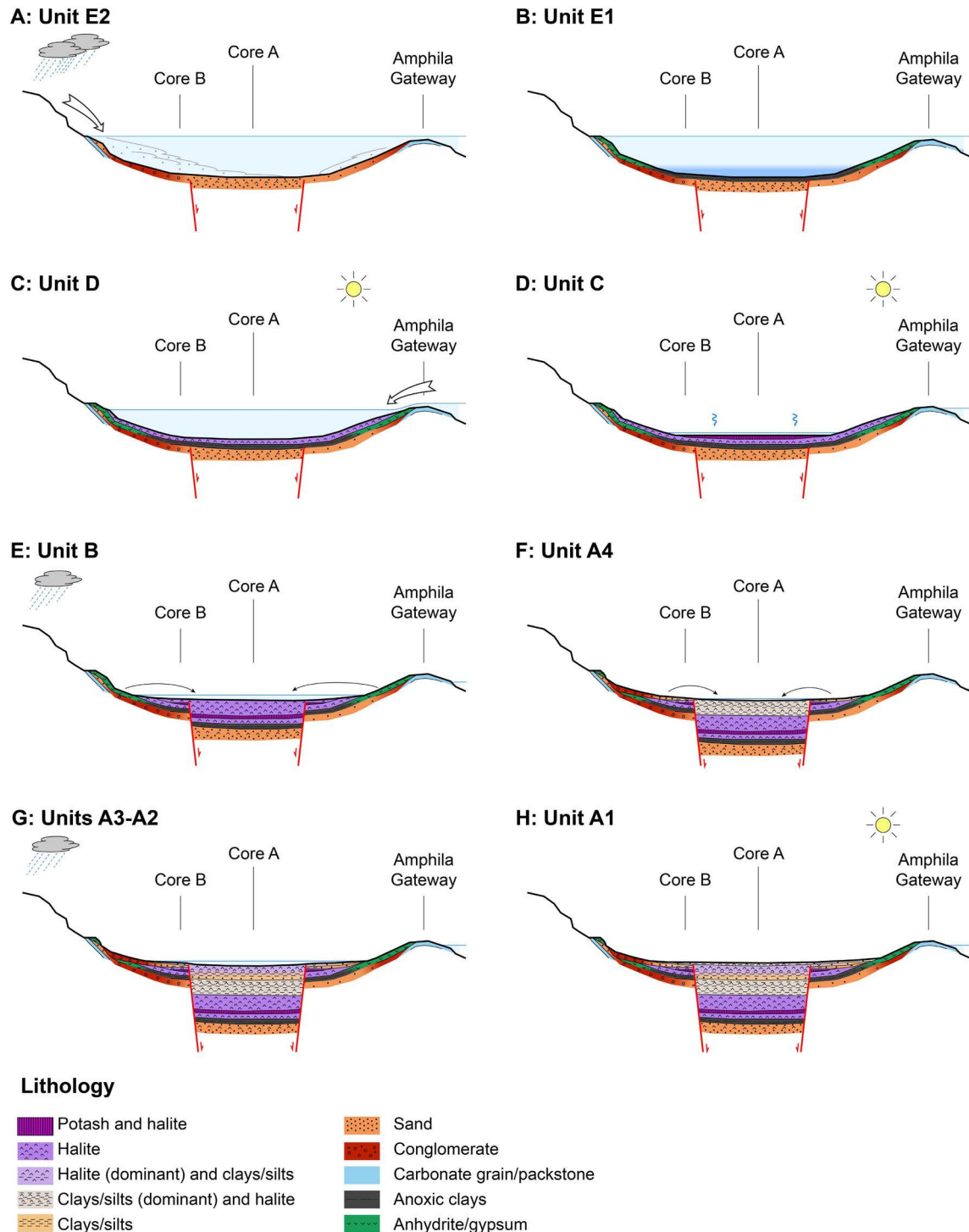


Figure 13: Depositional model for the Danakil Basin. Sketch illustrating paleo-environmental conditions since the deposition of unit E2. Please note that the cartoon is not to scale. The tectonic structure is simplified and hypothetical. **A)** Restricted marine stage shortly after MIS5e peak. A wet climate results in increased fluvial clastic input into the basin. **B)** The basin becomes more restricted and stratified with the deposition of gypsum on the margins and anoxic clays in the central part. **C)** Evaporation stage. Halite starts to precipitate under arid conditions. The basin remains flooded under a certain depth of marine brine supplied through the Amphila Gateway. **D)** Hyperconcentration stage. The connexion with the Red Sea is essentially cut, the brine depth decreases and potash minerals precipitate in the central part of the basin. **E)** Saline lake stage. Meteoric water dissolves the evaporites on the margins and recycles them in the central part of the basin. **F)** Saline pan stage. Dry climatic conditions induce an environment comparable to the present-day situation. **G)** Late lacustrine stage. A wet climate triggers the establishment of a brackish/saline lake that becomes hypersaline in its late phase. **H)** Modern saline pan.

topping the open-marine limestones at the outcrops of the Zariga Fm. (Jaramillo-Vogel et al., 2019, Negga *et al.*, in prep., Fig. 3A). In the Holocene marine evaporite basin of the Sabkha el Melah, dense borehole coverage allowed to correlate a magnesite-rich unit in the central part of the basin to gypsum-rich units on the margin (Perthuisot, 1971; Busson & Perthuisot, 1977). It is therefore interpreted that subunit E1 correlates to the deposition of gypsum at the marginal deposits of the Zariga Fm. (Fig. 1C, 3A). Unit D is correlated with unit δ ; which shows similar halite facies variations and a similar thickness. The potash-bearing units C and γ are also correlated. Units B and β show the same facies change from coarse-grained, laminated to coarse halite; although this unit is much thinner in Core B (16 m vs. 85 m). The correlation of the limit between units α and β to Core A is unclear. It is tentatively correlated with the limit between subunits B2 and B1 based on the appearance of clastic material and a change in the halite facies.

IV.5.2 Paleo-environmental and climatic variability

IV.5.2.1 Restricted marine stage

In subunit E2, benthic foraminifera (dominated by *Ammonia veneta*), ostracods (*Cyprideis*), and other shell fragments represent dominantly euryhaline species. For example, the species *Cyprideis torosa*, is a coastal and lagoonal euryhaline species, typical for brackish conditions, and occurring frequently in inland saline lakes (e.g. Wouters, 2002). The *pertorosa* subsp. has only been described in lacustrine deposits of supposedly variable salinities of the Danakil Depression (Gramann, 1971).

Since the specimens from core A did not show any signs of carapace nodding, layers containing *Cyprideis* were probably close to normal marine salinities (the same is generally applicable to *Loxoconcha* species). *C. torosa* has previously been observed associated with *Ammonia tepida* and *Q. semiluna* in early Holocene continental lake deposits in northern Saudi Arabia (Pint *et al.*, 2017). High levels of P_2O_5 and Ba, in particular at the base of the subunit, suggest high productivity.

Thick clastic intervals, including coarser sand-sized beds (Fig. 3O) and conglomerates in Core B are interpreted as delta systems and imply increased fluvial input indicating a very wet/humid climate, only interrupted by short dryer periods causing an increase in salinity (Fig. 12, 13A). The presence of halite and anhydrite layers indicates that the conditions were indeed temporarily hypersaline.

The Br content of halite alone does not allow to unambiguously distinguish between marine and continental waters as parent fluid for halite precipitation. However, the values are overall situated within the 40-75 ppm range, which would be very high for a purely continental brine body. It is therefore interpreted as a marine water body mixed with high meteoric water input. This interpretation is strengthened by the marine nature of the overlying units D and C (see below) and the correlation with outcrops (see above).

The wet environment during this period can be linked to the humid climate of eastern Africa and southern Arabia during MIS5e until approximately 120 ka (Fleitmann *et al.*, 2011; Rosenberg *et al.*, 2011; Tierney *et al.*, 2017; Lamb *et al.*,

2018; Nicholson *et al.*, 2020; Schaebitz *et al.*, 2021; Foerster *et al.*, 2022).

In subunit E1 and a small layer at the base of D2, the presence of dolomite (Vasconcelos & McKenzie, 1997), magnesite (Busson & Perthuisot, 1977), sapropelic organic matter (Tzevahirtzian *et al.*, 2022) along with the absence of fauna and gypsum (Busson & Perthuisot, 1977; de Lange & Krijgsman, 2010; Roveri *et al.*, 2014) suggest that this marine-influenced restricted basin developed into a stratified water body with anoxic conditions in the hypolimnion (Fig. 13B). Sulphate consumption by remineralization of organic matter or sulphate reducing bacteria may undersaturate the water with respect to gypsum deposition in the central part of the basin. The presence of gypsum on top of the marine deposits at the margin, which can be correlated with subunit E1 (Fig. 3), indicates that the water level was at or near maximum sea-level (Fig. 12, 13B).

IV.5.2.2 Evaporation stage

The high (>50ppm) Br content in halite and the high Sr content of unit D indicate the dominant presence of marine fluids (Fig. 13C). The increasing trend in Br content evidences increasing salinity, supported by micro- and macrofacies. The presence of coarse clear halite crystals (Fig. 5P) in subunit D2 is generally believed to represent bottom-growth halite precipitating from moderately supersaturated fluids (Dellwig, 1955; Li *et al.*, 1996; Kiro *et al.*, 2016; Warren, 2016; Sirota *et al.*, 2017). The inclusion-free halite textures often represent relatively stable chemical conditions, most likely reached under more than a few decimetres of water or might represent syndepositional/early

diagenesis recrystallizations, late burial diagenesis recrystallizations or recrystallizations related to deformation (Shearman, 1970; Lowenstein & Hardie, 1985; Schlöder & Urai, 2005; Salvany *et al.*, 2007; Lugli *et al.*, 2015). Samples from unit D show no signs of recrystallization and are thus interpreted as primary. However, irregular grain boundaries could indicate short-termed subaquatic dissolution events.

The lower part of subunit D1, characterized by the alternation of fine- and coarse-grained halite, probably represents very similar conditions to the present-day Dead Sea with high halite supersaturation in winter (deposition of halite cumulates) and stratified water with moderate supersaturation during summer (precipitation of bottom-growth halite, Sirota *et al.*, 2017). The fine grained, laminated halite in the upper part of D1 formed at the air-water interface or just below the thermocline as small cubes or hopper crystals (see Fig. 2D) in highly supersaturated fluids before settling at the bottom (Lowenstein & Hardie, 1985; Schreiber & Tabakh, 2000; Arnon *et al.*, 2016; Kiro *et al.*, 2016; Sirota *et al.*, 2017). XRF and XRD data indicate the presence of polyhalite, a mineral that has been described either as an early diagenetic replacement of anhydrite in marine environments, or as a primary mineral (Holser, 1966; Perthuisot, 1971). This mineral probably forms the white layers visible in Fig. 5G, on top of fine-grained halite intervals, and may represent a higher supersaturation, thus favouring the hypothesis of primary precipitation. It is unknown whether the kieserite found in the unit is primary or was precipitated during the deposition of unit C within open pore space. As unit δ of Core B

shows the same thickness and the same facies succession (Fig. 3), it is interpreted that the basin was flooded under a significant brine depth (10s to 100 of m, Fig. 12, 13C).

The absence of clastic material and the high and increasing salinity indicate warm and arid climatic conditions, in sharp contrast with the previous unit. This change can tentatively be correlated with the sharp transition to drier conditions observed after MIS5e highstand in southern Arabia, the Gulf of Aden and the Chew Bahir basin (Fleitmann *et al.*, 2011; Rosenberg *et al.*, 2011; Tierney *et al.*, 2017; Nicholson *et al.*, 2020; Schaebitz *et al.*, 2021; Foerster *et al.*, 2022).

IV.5.2.3 Hyperconcentration stage

This succession in unit C is comparable to the one presented in Table 1, except that the sylvite-rich interval is situated at the base of the unit and not at the top. Holwerda & Hutchinson (1968) and Warren (2016) showed that this sylvite layer is probably secondary, formed by the leaching of carnallite.

Overall, unit C is interpreted to represent an episode during which highly supersaturated brines were present in the Danakil Basin during very dry climatic conditions. According to Harvie *et al.* (1980) and Hardie (1990), bischofite crystallizes from high salinity brines, with an evaporative concentration factor well above 200. The mineral assemblage (rich in MgSO_4 , poor in Ca, devoid of typical meteoric minerals like mirabilite, thenardite, trona and pirssonite; Hardie, 1984, 1990, 1991; Spencer, 2000) and the high Br content indicate a marine parent brine and exclude a continental or deep hydrothermal origin of the brines. The

development of brines within the basin suggests that the Red Sea was disconnected from the depression.

The potassium-rich unit γ of Core B suggests that the more marginal position of the basin was at least partly flooded during the deposition of unit C, indicating that the brine level was not extremely shallow but could reach a few (tens?) of meters (Fig. 12, 13E).

Overall, units E, D and C show a concentrating upward trend as well as a shallowing upward trend. The climatic conditions during the deposition of unit C must have been very dry. Yet, the dry period that led to the deposition of units D and C was probably very short-lived (< ~3 kyr).

IV.5.2.4 Saline lake stage

The transition between unit C and unit B is sharp but shows no discontinuity or clastic interval, indicating the absence of emersion (and thus of complete desiccation). The rapidly decreasing Br content (from 374 to 42 ppm, Fig. 7) of subunit B3 implies a fast freshening upward trend. The presence of relatively high Br values, anhydrite and Sr still suggests a marine influence probably resulting from the leftover brines from the hyperconcentration stage being diluted by meteoric waters.

The coarse clear halite of subunits B3 and B2 indicates direct precipitation on the basin floor. Thin sections (fig 5D, E, M) show dissolution surfaces but no cavities, probably indicating short-lived subaquatic dissolution events. The absence of mosaic textures or fluid and solid inclusions purged to grain boundaries, as well as the presence of synchronous growth structures, indicate that the halite is primary (Hardie *et al.*,

1985; Lowenstein & Hardie, 1985; Schreiber & Tabakh, 2000).

Br content in subunit B2 is consistently very low and other elements show exceptionally constant content (Fig. 7), interpreted to represent a perennial and relatively deep meteoric lake under stable chemical conditions, only interrupted by frequent, possibly thermal, dissolution events. The presence of a similar δ unit in Core B evidences that the water body probably flooded a large part of the basin with a water depth reaching probably tens of metres (Fig. 12, 13E).

The change of facies to cloudy, milky texture of chevron halite present in subunit B1 is caused by the presence of fluid inclusions and is believed to indicate short-term variations in temperature, salinity, saturation and growth rate, most likely within a very shallow (a few decimetres) brine pan (Li *et al.*, 1996; Schubel & Lowenstein, 1997; Schreiber & Tabakh, 2000; Salvany *et al.*, 2007; Kiro *et al.*, 2016; Warren, 2016). This change of microfacies and the appearance of minor clastic intervals, as well as dolomite and magnesite layers, suggest a shallowing upward trend compared to subunit B2 (Fig. 12).

Overall, unit B indicates deposition in a saline continental lake with a shallowing upward towards subunit B1. The large size of this lake indicates relatively wetter climatic conditions than during the deposition of unit C. This period of relative humidity shortly after MIS 5e peak is not observed in the records of southern Arabia, the Chew Bahir basin and the Gulf of Aden (Tierney *et al.*, 2017; Nicholson *et al.*, 2020; Schaebitz *et al.*, 2021; Foerster *et al.*, 2022). Lake Tana, however, kept high

lake levels following the MIS 5e peak (Lamb *et al.*, 2018), possibly indicating a local climatic signal.

IV.5.2.5 Saline pan stage

The increased abundance of clastic intervals, the low Br content and the abundance of cloudy halite textures suggests that subunit A4 was deposited in an episodically inundated continental salt pan, showing very shallow ephemeral water bodies, similar to the present-day environment (Fig. 12, 13F). This indicates a further shallowing upward trend from B1 and dry climatic conditions. Varying minor elements like MnO, Zn, Ba and Cu indicate fluctuating brine chemistry in the ephemeral shallow water bodies.

Two samples show higher bromine content. These samples are outliers and do not reflect the general trend, suggesting a short-term and spatially restricted process. Similarly, a very high Br content is measured in some of the Recent evaporites of the salt plain, sampled near isolated ponds (AF19-BS2 and AF19-BS3 in Fig. 1C. see also Fig. 2J and table S4). Those ponds prove that spatially and probably temporally confined halite precipitation can lead to locally very high Br content values in a continental setting. Three interpretations can explain these high values. First, the subsurface seepage of marine waters through the Amphila gateway or even through fractures of the Buia graben (Fig. 1a) could result in the precipitation of Br-rich halite. Secondly, while the meteoric recycling of marine halite (containing ~60-200 ppm of Br) produces halite low in Br content, the leaching of potash minerals with very high (> 1000 ppm) Br content can lead to the formation of Br-rich halite, especially

if the volume of water is small (Ayora *et al.*, 1994; Taberner *et al.*, 2000). Finally, a hydrothermal origin of the water may explain the presence of Br-rich halite. Indeed, Vengosh *et al.* (2002) and Nahnybida *et al.* (2009) demonstrated that non-marine (magmatic) hydrothermal water (today produced in the vicinity of the Dallol volcano: Cavalazzi *et al.*, 2019; Kotopoulou *et al.*, 2019; López-García *et al.*, 2020; Otálora *et al.*, 2022) can have a higher Br content than normal meteoric water. Based on the data and observations, all three interpretations are plausible, and it is probably the combination of processes in different parts of the basin that trigger the local and confined precipitation of continental Br-rich halite.

IV.5.2.6 Late lacustrine stage

The clastic-dominated clay-rich subunit A3 indicates an environmental setting influenced by fresher water. The benthic foraminiferal assemblage composition, together with the observed ostracods and shell fragments, indicate that the study site was situated within shallow waters, possibly brackish. The dominant species composing the foraminiferal assemblage from subunit A3, such as *Ammonia* spp., *E. williamsoni*, *E. advenum*, *Q. semiluna* and *H. depressula*, are commonly found in coastal intertidal environments and estuaries (Horton *et al.*, 1999; Debenay & Guillou, 2002; Debenay *et al.*, 2006; Abu-Zied & Hariri, 2016; Mojtahid *et al.*, 2016).

The presence of foraminifera in this interval may indicate transport of propagules or adult individuals by birds from the Red Sea. Even during the MIS 2 lowstand, the Red Sea was situated less than 200 km away. Avian transport of

benthic foraminifera over long distances (up to several thousands of km) has been abundantly documented, noticeably for the same shallow-water genera found within unit A2 (*Ammonia* spp., *Quinqueloculina* spp., *Rosalina* spp., and *Bolivina* spp.) (Almogi-Labin *et al.*, 1992; Patterson *et al.*, 1997; Boudreau *et al.*, 2001; Abu-Zied *et al.*, 2007; Wennrich *et al.*, 2007; Giordana *et al.*, 2011; Riedel *et al.*, 2011; Pint *et al.*, 2017). The absence of foraminiferal test anomalies, which have been linked to the formation of tests under hypersaline conditions (Pint *et al.*, 2017), and the relatively high diversity, point towards a favourable environment, with a brackish to normal marine salinity, whilst elevated P₂O₅ and Ba values indicate high productivity. The presence of *Guembelitria* sp. (extinct since Cenozoic times) suggests a clastic origin from the Mesozoic sediments abundantly outcropping on the margins of the basin (Rime *et al.*, 2022). Subunit A3 was thus deposited within a brackish lacustrine environment and under wet climatic conditions (Fig. 12, 13G). Three intervals of halite are interpreted to represent a short increase in salinity, probably linked to short dryer climatic conditions and a decrease in water depth, as suggested by the milky halite facies. The Br content of these halite layers is relatively low, evidencing a meteoric origin. The two outliers with higher values are interpreted to represent a similar process as the outliers of subunit A4 discussed above.

The transition to the halite-rich subunit A2 indicates a salinity increase. The lack of clastic material in the two halite intervals dominated by cloudy textures suggests the presence of shallow hypersaline lakes. The relatively constant MnO, Zn, Cu and Sr values in

halite (Fig. 7) point to a relatively homogenous brine body, whereas the meteoritic origin of halite is demonstrated by its low Br content. The more clastic interval between the two halite layers may represent a wetter/fresher period with increased siliciclastic input (Fig. 12).

This wetter lacustrine stage starting before 35 ka contrasts with other records in eastern Africa and Arabia that place the onset of the African Humid Period (AHP) at approximately 15 ka and link its onset to an interval of “profound aridity” (Rosenberg *et al.*, 2011; Shanahan *et al.*, 2015; Tierney *et al.*, 2017; Foerster *et al.*, 2018, 2022; Schaebitz *et al.*, 2021). Yet, the different lake level highstands (Fig. 12) can be correlated with the different stages of Lake Abhe as defined by (Gasse, 1975, 1977; Gasse & Street, 1978; Khalidi *et al.*, 2020; Mologni *et al.*, 2021) and other lakes around the Afar Depression (Fig. 1, 12). The first highstand (~150-130 m) dated at 35 ± 5 ka correlates to the Abhe II period (which possibly ended around 30 ka; (Gasse & Street, 1978), and can be linked with the wet climatic conditions described in lake Tana around 35-40 ka (Lamb *et al.*, 2018). The second highstand (~115-100 m) correlates to the Abhe III period (~30-17 ka, Gasse & Street, 1978), whilst the third highstand (~90-60 m) may match the Abhe IV period (ending around 8 ka, Gasse & Street, 1978; Mologni *et al.*, 2021) that was also identified in Lake Ziway-Shala (Gillespie *et al.*, 1983). The last lacustrine highstand (subunit A2), dated at 7.3 ka, corresponds to the Abhe V stage, (Gasse & Street, 1978; Mologni *et al.*, 2021), as well as highstands identified in lake Asal (Gasse & Fontes, 1989), lake Afdera (Gasse, 1974) and possibly Ziway-Shala lake basin (Gillespie *et al.*, 1983). Abhe IV

and V periods correspond to the African Humid Period identified in eastern Africa and Arabia (Rosenberg *et al.*, 2011; Tierney *et al.*, 2017; Foerster *et al.*, 2018, 2022; Schaebitz *et al.*, 2021).

IV.5.2.7 Modern saline pan

The alternated succession of clastic and halite intervals in subunit A1 mimics what is depositing today in the Danakil Depression (Fig. 2) and represents similar environmental conditions, being a classical saline pan environment (Fig. 12, 13H). Hot and arid conditions are predominant, which matches other Late Holocene records of eastern Africa (Gasse, 1974; Gillespie *et al.*, 1983; Gasse & Fontes, 1989; Foerster *et al.*, 2018, 2022; Mologni *et al.*, 2021; Schaebitz *et al.*, 2021).

IV.5.3 Evaporite formation in the Danakil Depression: an analogue for salt giant formation in the past

The depositional model indicates that unit D formed due to the direct evaporation of seawater. The significant thickness of evaporites suggest that the connection between the basin and the Red Sea was not closed by one rapid and distinct event (which would have produced only a few meters of halite) but that the basin was continuously fed by the Red Sea by shallow spill-over. The deposition of that unit must thus have taken place shortly after the 128 ka sea-level highstand (Grant *et al.*, 2014) and before significant sea-level fall that would have cut any connection with the Red Sea. At that point, the basin reached a near equilibrium point where the influx of seawater from the Red Sea equilibrated both (1) the losses by evaporation, (2) potential (but unlikely

due to the shallow gateway) reflux, and (3) freshening of the brine by meteoric waters and precipitations.

Assuming the average deposition rate of 3 cm/yr for unit D and closed theoretical models suggesting that 13.7 m of halite can be deposited from 1000 m of evaporated seawater (Usiglio, 1849), the 58 m of halite deposited in unit D would represent the evaporation of a 4.2 km water column in less than 2000 years. Assuming a 9000 km² basin, steady-state conditions and no reflux, such evaporation implies an average influx of ~625 m³/s of seawater from the Red Sea.

Conversely, unit B was mainly formed by meteoric water influx in a perennial lake. As the salinity of water from the escarpment is not high enough to precipitate halite, the high salinity must be caused by the dissolution of older evaporites at the margin of the basin, as observed today. This assumption is supported by the much smaller thickness of unit B in Core B, situated in a more marginal setting and by the description of vertical dissolution cavities in a more marginal well (DK-11-11, see Fig. 1C, Warren, 2016). However, the smaller thickness can be also explained by other mechanisms : 1) The upper part of the unit was dissolved by younger fluvial deposits, 2) this part of the basin was flooded for a shorter amount of time, 3) This part of the basin was flooded but situated above the epilimnion and *halite focussing* mechanism (Sirota *et al.*, 2018; Kirkham *et al.*, 2020) concentrated the deposition in the central part of the basin, 4) Accommodation space was smaller. Whatever the reason for reduced halite precipitation at the margin of the basin, it should be highlighted that halite deposition was concentrated in the central part of the basin. If unit D was

deposited in a brine body nearing sea-level, a lot of marine halite was deposited up to high elevations around the margin of the depression and was available for recycling by meteoric waters into the central part of the basin. These waters might also have partly recycled older evaporites deposited on the margins and uplifted by tectonic movements (chapter III, this volume) or by shallow hydrothermal circulation.

This study highlights the importance of accommodation space and recycling of marine evaporites by meteoric waters for the formation of the great thickness of evaporite deposits. Evaporite recycling by meteoric waters has also been described in the Lorca Basin in Spain (Taberner *et al.*, 2000). The greater thickness of units A and B in Core A compared to the equivalent deposits of Core B stress the importance of accommodation space for the deposition of thick evaporite deposits. As the potash units of cores A and B deposited at nearly the same elevation in shallow waters, a subsidence of ~370 m occurred since the deposition of the potash, which is comparable with the estimations of Bastow *et al.* (2018). This indicates that the basin was never deeper than ~200 metres. A plausible age of ~120 ka for the deposition of unit C would imply a subsidence rate of ~2.9 mm/yr.

Several authors argued against a solar evaporation model for salt giants (Scribano *et al.*, 2017; Hovland *et al.*, 2018a; b; Debure *et al.*, 2019; Qin *et al.*, 2023), proposing that one single desiccation cycle of 1000 m of water could only produce ~15 m of evaporites and that too many cycles would be needed to produce km-thick deposits. These models do not take full measure of a possible protracted supply of seawater during the evaporation stage. The example of the

Danakil Depression shows that, in a basin accommodating only 100-200 m of water, under the right conditions (arid and warm climate, continuous supply of seawater through a gateway, reduced or absent reflux and sufficient accommodation space), a single, short (~3 kyr) cycle of direct evaporation of seawater may produce ~60 m of halite and ~40 m of potash, whilst the following periods of meteoric reworking can add more than 400 m of evaporite-dominated sediments in less than 130 kyr. If a similar process occurred during the three previous flooding events, as evidenced by marginal outcrops in the Danakil Depression (Foubert *et al.*, in prep.; chapter III, this volume), km-thick evaporite deposits might have been produced in less than 500 kyr.

It has been suggested that salt giants are produced by deep hydrothermal processes, including serpentinization of the mantle (Scribano *et al.*, 2017; Hovland *et al.*, 2018a; b; Debure *et al.*, 2019; Qin *et al.*, 2023). Hovland *et al.* (2018b) and Johnsen *et al.* (2021) used the Danakil evaporites as an example of this process. The present study shows that no deep hydrothermal processes are needed to explain the deposition of the Danakil evaporites. Solar evaporation can satisfactorily explain the upper 625 m of evaporite-dominated sediments of the Danakil Depression. Only a minor amount of halite was possibly mobilized by deep hydrothermal fluids, explaining the isolated high bromine content in halite. Moreover, according to the model of Otálora *et al.* (2022) and the present study, these fluids gained salinity by the dissolution of older, solar precipitated evaporites and reprecipitated by solar evaporation at the surface. This is compatible with the results of Talbot

(2008) describing only “minor local additions of hydrothermal salts” on the Dallol dome. At basin-scale, the deep hydrothermal influence is therefore minimal being only localized in specific spots within the vicinity of the Dallol volcano or other hot springs.

IV.6 CONCLUSIONS

Based on the multiproxy analysis of two sediment cores, the Pleistocene to Holocene evaporitic sedimentary record of the Danakil Basin has been unravelled precisely for the first time with the following conclusions:

- 1) The uppermost 625 m of basin fill of the Danakil Depression are mainly formed by halite and subordinate clastic sediments. This basin might therefore be considered as a Recent salt giant analogue.
- 2) The record can be divided into 5 different units. The lowermost (E) shows clastic and evaporitic sedimentation in a restricted marine basin during wet climatic conditions. In contrast, Unit D corresponds to the evaporation of seawater in an arid climate and represents the last connection with the Red Sea, whereas Unit C is characterized by the precipitation of potash minerals in a highly saturated brine of marine origin. Unit B depicts the refilling of the basin by a meteoric lake recycling older marine evaporites. Unit A represents and the deposition of evaporites and clastic sediments in a continental salt pan and a brackish/saline lake.
- 3) The record covering a period from MIS5e until present times proves that a significant thickness of evaporites

(>500 m) may accumulate in a very short amount of time (< 130 ka).

- 4) More than 100 m of evaporites were deposited by the direct evaporation of seawater during a very short time window (<3 ka) under an arid and warm climate (units C and D). The recycling of marine evaporites from the margins towards the centre of the basin, together with subsidence, played a major role after the isolation of the basin (units A and B) and explains the important total thickness.
- 5) Deep hydrothermal processes only played a very minor role in the deposition of the Late Pleistocene to Holocene evaporites of the Danakil Basin.
- 6) The Danakil Depression experienced a humid climate directly after the MIS5e sea-level highstand. After a short period of extreme aridity, followed by a lacustrine stage, dry conditions prevailed in the basin. A wetter, lacustrine phase started again before 35 ka. Present-day arid conditions were established during the Late Holocene.

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AUTHOR CONTRIBUTIONS

- VR: Conceptualization, Methodology, Investigation, Data Curation, Writing – Original Draft, Visualization
- HN: Investigation, Writing – Review & Editing
- RF: Methodology, Investigation, Writing – Original Draft, Visualization
- AEK: Methodology, Investigation, Writing – Review & Editing
- CP: Investigation, Writing – Review & Editing
- JCS: Investigation, Writing – Review & Editing
- IH: Methodology, Investigation, Writing – Review & Editing
- TA: Methodology, Investigation, Writing – Review & Editing
- BA: Writing – Review & Editing, Project administration, Funding acquisition
- TK: Conceptualization, Writing – Review & Editing Supervision, Project administration, Funding acquisition
- AF: Conceptualization, Methodology, Investigation, Validation, Data Curation, Writing – Original Draft, Supervision, Project administration, Funding acquisition

DATA AVAILABILITY

The newly produced data that supports the findings of this study are available in the supplementary material

of this article. Data from industrial companies and the former Ethiopian Ministry of Mines are not shared because of confidentiality reasons.

SUPPORTING INFORMATION

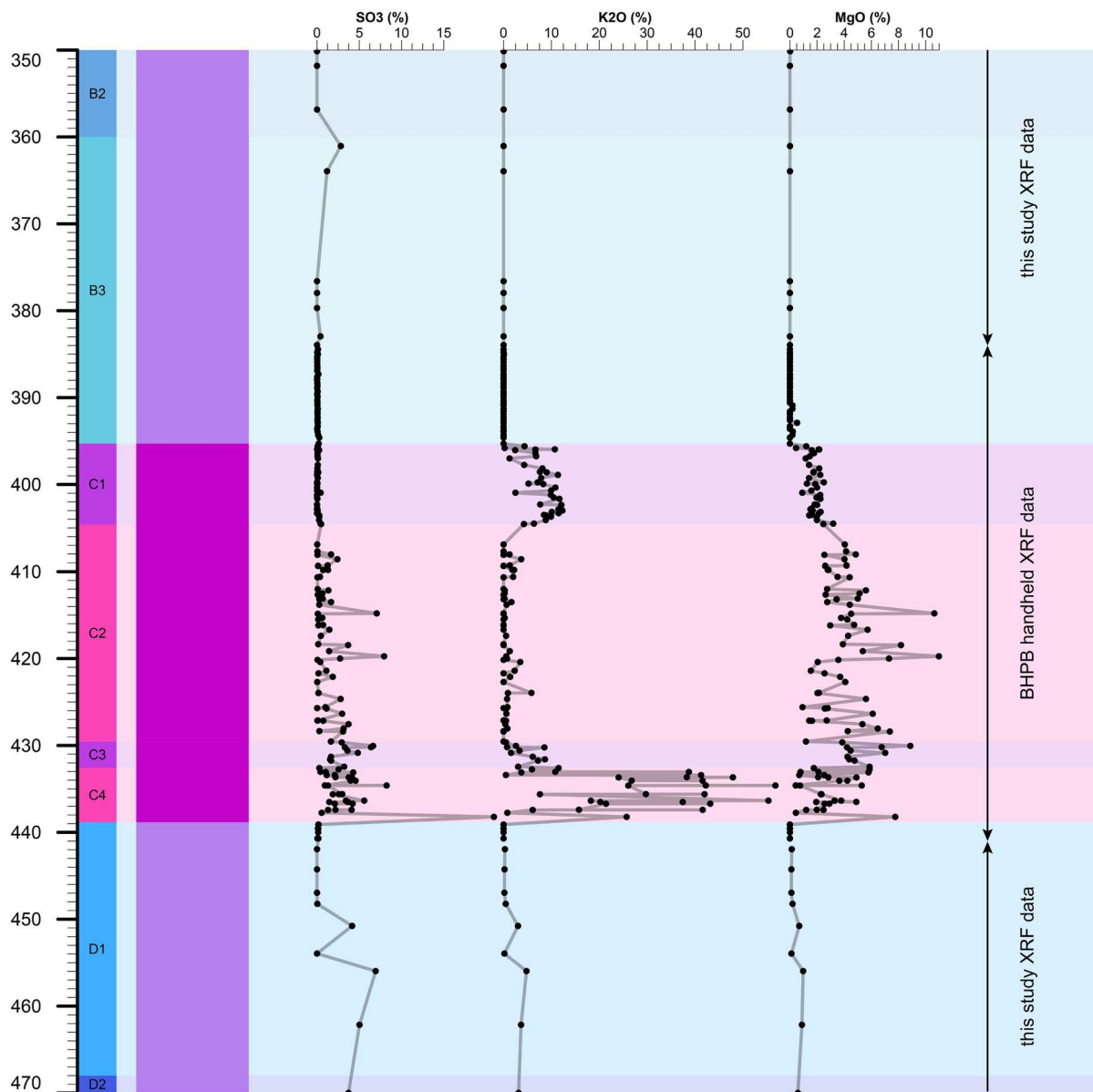


Figure S1: Detail of the potash interval of Core A. XRF elemental composition (SO₄, K₂O, MgO) of the main potash interval provided by BHPB is indicated. Data from BHPB are complemented by data from the present study.

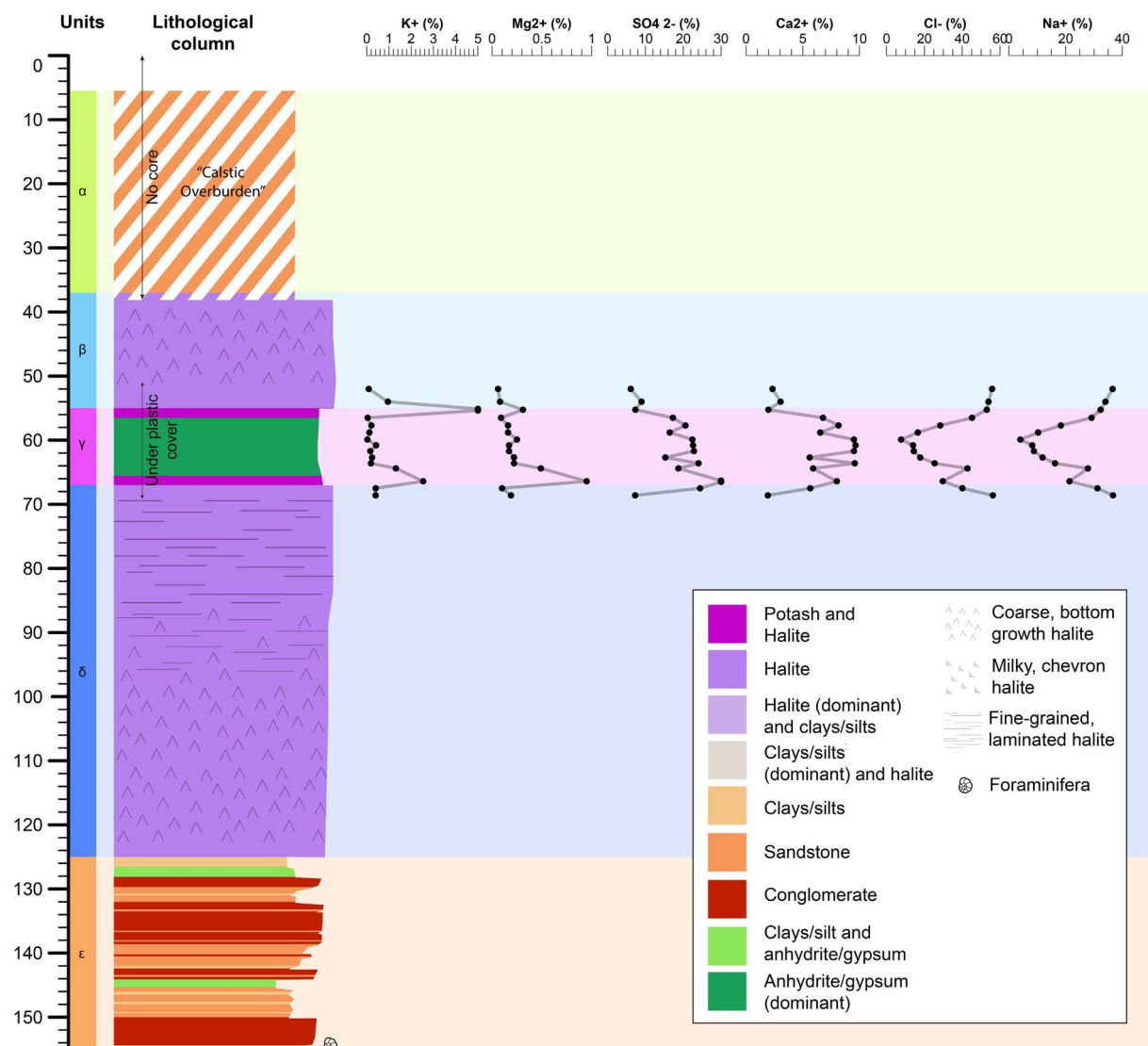
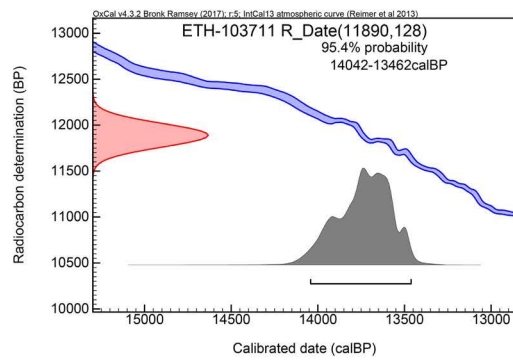
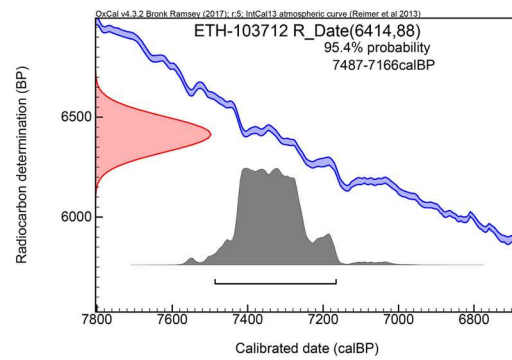
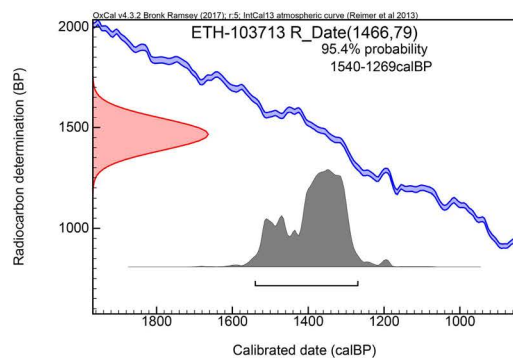
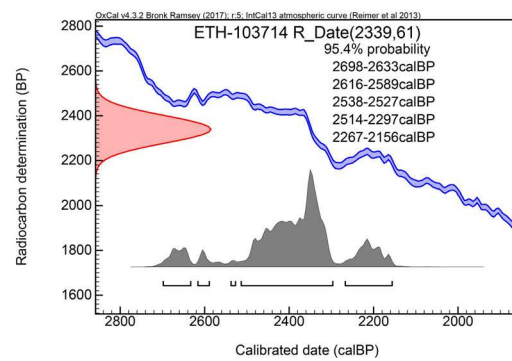
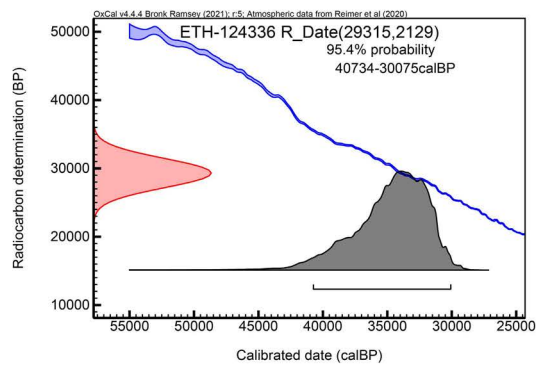


Figure S2: Lithological column of Core B. Elemental composition (K^+ , Mg^{2+} , SO_4^{2-} , Ca^{2+} , Cl^- and Na^+ from Rauche & van der Klauw, 2011) for the potash-rich intervals are indicated.

S251 (Wood), 70.5 m, Unit A2**S269 (Wood), 33.8 m, Unit A2****S274 (Wood), 10.6 m, Unit A1****S275 (Wood), 10.4 m, Unit A1****S229 (Shell fragments), 135.2 m, Unit A3****Figure S3: Results of radiocarbon dating on wood and shell fragments.****Table S4:** List of samples, depth of units and results of laboratory analysis (XRF, XRD, rockeval, radiocarbon) for Core A. Available in the numerical version of this thesis or, once published, on the website of the publisher.

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V NOVEL BATHYMETRY OF LAKE AFDERA REVEALS FAULT STRUCTURES AND VOLCANO-TECTONIC FEATURES OF AN INCIPIENT TRANSFORM ZONE (AFAR, ETHIOPIA)

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ABSTRACT

Lake Afdera is a hypersaline endorheic lake situated at 112 m below sea-level in the Danakil Depression. The Danakil Depression is located in the northern part of the Ethiopian Afar and features an advanced stage of continental rifting. The remoteness and inhospitable environment explain the limited scientific research and knowledge about this lake. Bathymetric data were acquired during 2 weeks expeditions in January/February 2016 and 2017 using an easily deployable echosounder system mounted on an inflatable motorized boat. This study presents the first complete bathymetric map of the lake Afdera. Bathymetric results show that the lake has an average depth of 20.9 m and a total volume of 2.4 km³. The maximum measured depth is 80 m, making Lake Afdera the deepest known lake in Afar and the lowest elevation of the Danakil Depression. Comparison with historical reports shows that the lake level did not fluctuate significantly during the last 50 years. Two distinct tectonic basins to the north and the south are recognized. Faults of different orientations control the morphology of the northern basin. In contrast, the southern basin is affected by volcano-tectonic processes, unveiling a large submerged caldera. Comparison between the orientation of faults throughout the lake with the regional fault pattern indicates that the lake is part of two transfer zones: the major Alayta–Afdera Transfer Zone and the smaller Erta Ale–Tat’Ali Transfer Zone. The interaction between these Transfer Zones and the rift axis forms the equivalent of a developing nodal basin which explains the lake’s position as the deepest point of the depression. This study provides evidence for the development of an incipient transform fault on the floor of the Afar depression.

V.1 INTRODUCTION

Lake Afdera is a hypersaline endorheic lake (Martini, 1969) located in the southern part of the Danakil Depression, in the Afar region, in northern Ethiopia (Fig. 1). Afdera is a name initially given by Afari people to the largest nearby active volcano (“Afe” meaning “mouth” and “Deda” standing for “large”, or the large mouth of the volcano). The lake is also sometimes called “Lake Afrera”, “Lake Giulietti” or “Egogi Bad”. Sedimentological (Gasse, 1974; Gasse and Street, 1978) and ecological data (Chiozzi et al., 2018)

confirm that the lake has been persistent for at least the last 10 ka, although the lake level has been up to 50 m higher than today during the Mid-Holocene (Gasse, 1974; Schaegis et al., 2017).

V.1.1 First explorers

Western exploration of the ‘Danakil Country’ and the Afdera region was initially launched in 1867 by Munzinger, the first explorer to describe the depression from a geomorphological perspective (Munzinger, 1869). The expeditions of 1875, 1881 and 1884 led by Munzinger, Giulietti and Bianchi were unsuccessful as none of the explorers

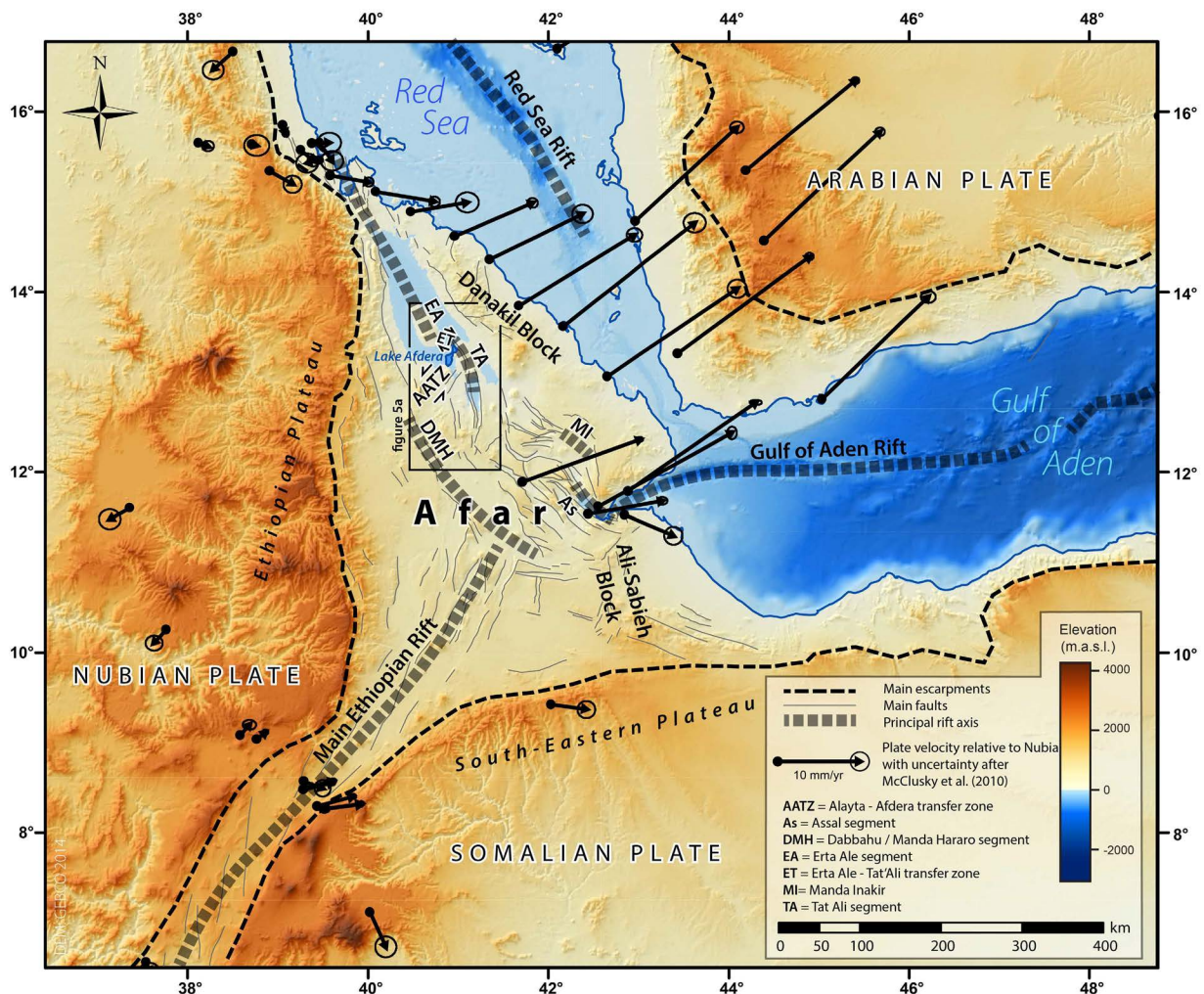


Figure 1: Elevation map of the Afar triple junction (East Africa), defined by the divergence of the Arabian, Nubian and Somalian plates as well as the propagation of the Red Sea, Gulf of Aden and Main Ethiopian Rifts. Black arrows show direction and velocity of the different plates (McClusky et al., 2010), relative to the Nubian plate. The black rectangle represents Fig. 5a.

survived (Nesbitt, 1935). The objective was to cross the Danakil desert from the Ethiopian plateau - ‘Abyssinia’- to the Red Sea. The next generation of explorers was more successful. Thus, Vinassa De Regny in 1919, Nesbitt in 1928 and Franchetti in 1929 all managed to cross the Danakil desert and collect geographical and geological data (Lupi, 2009). On January 26th, 1920, Vinassa de Regny was the first explorer to see the lake, which he named “Afrera” – most likely a misspelling from the volcano Afdera in its surroundings (Vinassa de Regny, 1931). Despite having visited the region a decade later, Nesbitt and Franchetti were the first to publish their results on the Danakil Depression in 1929 and 1930, respectively. Nesbitt called the lake “Egogi Bad”, a name given by his local guide. Franchetti named it lake Giulietti, as a tribute to Giuseppe-Maria Giulietti after discovering the remains of the explorer and his team near Beylul (Nesbitt, 1935; Taillibert, 1996). Franchetti also called the island in the middle of the lake Afdera – the lowest proven island in the world – after himself for being the first known person setting foot on it. Franchetti extensively explored the lake, and his name was assigned to one of the two endemic fishes of the lake Afdera: *Danakilia franchettii* (Vinciguerra, 1931; Getahun, 2001).

V.1.2 Geological setting

The Afdera lake is situated in a tectonically and volcanically active rift valley - the Danakil Depression - positioned between the Ethiopian plateau to the west and the Danakil block to the east (Fig. 1). This region is currently experiencing crustal stretching in an extensional environment representing the transition between continental rifting

and seafloor spreading (Bastow and Keir, 2011; Keir et al., 2013; Daniels et al., 2014; Bastow et al., 2018). The Afar depression opened up due to the separation of the Arabian plate from the Nubian plate following the Oligocene eruption of trap basalts related to the Afar plume (Bosworth et al., 2005). The eruption of the Ethio-Yemen traps 30 Ma ago was followed by the breakup of the Afro-Arabian plate and the formation of incipient proto – East African and Red Sea rift fractures (Baker et al., 1972; Hofmann et al., 1997; George et al., 1998). Simultaneously, the Gulf of Aden ridge propagated in west-southwest direction towards the present-day Afar region as stop-and-go propagation series (Manighetti et al., 1997; Audin et al., 2004). This kinematic process eventually isolated two rigid tectonic blocks known as the Danakil and Ali-Sabieh blocks which rotated in opposite direction (counterclockwise and clockwise respectively), opening the Afar Depression by a 'saloon-door' mechanism (Manighetti et al., 2001; Kidane, 2015). This anticlockwise rotation of the Danakil block formed the Danakil Depression. The individual rift segments are well developed in the Afar depression, consistent with the tectonic features observed in Mid-Oceanic ridge settings (Bridges et al., 2012; Dumont et al., 2018). However, the evidence for a mature and well-individualized transform fault in the Afar depression has not yet been documented. The Makarrasou Transfer Zone between the Assal and Manda Inakir rift segments; the Gamarri-Allol Transfer Zone between two aforementioned segments and the Manda Hararo rift segment in central Afar; the Alayta-Afdera Transfer Zone and the Erta Ale- Tat’Ali Transfer Zone (Fig. 1) are all characterized by normal

faults with orientations oblique to the rift axis, without forming a well-developed and defined transform fault zone.

Lake Afdera is situated in an area influenced by two transfer zones linking different rift segments. The Alayta – Afdera Transfer Zone links Erta Ale and Tat'Ali rift segments in the NE and the Dabbahu Manda–Hararo in the SW (Tazieff et al., 1972; Barberi et al., 1974; Barberi and Varet, 1977; Stab et al., 2016; Gardo and Varet, 2018; Varet, 2018; Le Gall et al., 2018; Ayalew et al., 2019; Zwaan et al., 2020; see also discussion in section 4.2) and the Erta Ale – Tat'Ali Transfer Zone links Erta Ale and the Tat'Ali rift segments (Illsley-Kemp et al., 2018; La Rosa et al., 2019, 2021). The lake is bounded by two volcanic ranges, the Erta Ale range in the NW and the Tat'Ali range in the E, of which the Borwali volcano directly overlooks the lake. Both volcanic ranges are considered as recent (<0.6 Ma; Barberi et al., 1972; Kidane et al., 2003; Lahitte et al., 2003; Stab et al., 2016). In the S, the lake is dominated by the Afdera volcano. Towards the W and the N, the lake opens up to relatively flat lava plains covered by a thin layer of Holocene to Recent lacustrine (Gasse, 1974) or aeolian sediments punctuated by hyaloclastite rings and volcanic cones (Tazieff et al., 1969). The lake features a small island (Franchetti Island or *Āfrerā Desēt*) in its southern basin. It is formed by volcanic rocks covered by a thin veneer of lake sediments and unique gypsum microbialites (Foubert et al., 2019).

Because of its isolation in the Danakil Depression and difficulties to access the lake, only a few studies have been carried out (Bonatti et al., 2017; Martini, 1969; Giofiantini et al., 1973). Bathymetric soundings have been attempted in the

1970's (Bonatti et al., 2017), but the precise lake floor mapping has never been performed, despite geological interest for this region. The bathymetry of lake Afdera and its comparison with outcropping structures is essential for the understanding of the regional structural architecture. Moreover, mapping Lake Afdera is valuable for economic and social reasons. The local population in the village of Afdera and its surroundings are economically dependent on the lake's salt production (Locatelli-Rossi, 1998; Bekele, 2016). The salt is extracted since the late 1950's by pumping salty lake water into salt evaporation ponds (Locatelli-Rossi, 1998, Suppl. Fig. 1). It is not clear whether hydrothermal springs considered as inflow compensate for the natural and artificial evaporation of the lake, but quantifying the lake volume will help to evaluate the stability of the lake Afdera and its fragile ecosystem, as well as the local salt exploitation. This study aims to estimate the lake volume and architecture by bathymetric mapping using an echosounder system mounted on a Rigid Inflatable Boat (RIB). Additionally, this study explores the feasibility of developing portable and easily deployable solutions for bathymetric mapping in challenging environments.

V.2 MATERIAL AND METHODS

The setup of an echosounder system on a mobile RIB has been essential to the success of the mapping survey (Fig. 2) and can be applied to the mapping of other lake systems or water bodies with limited access.

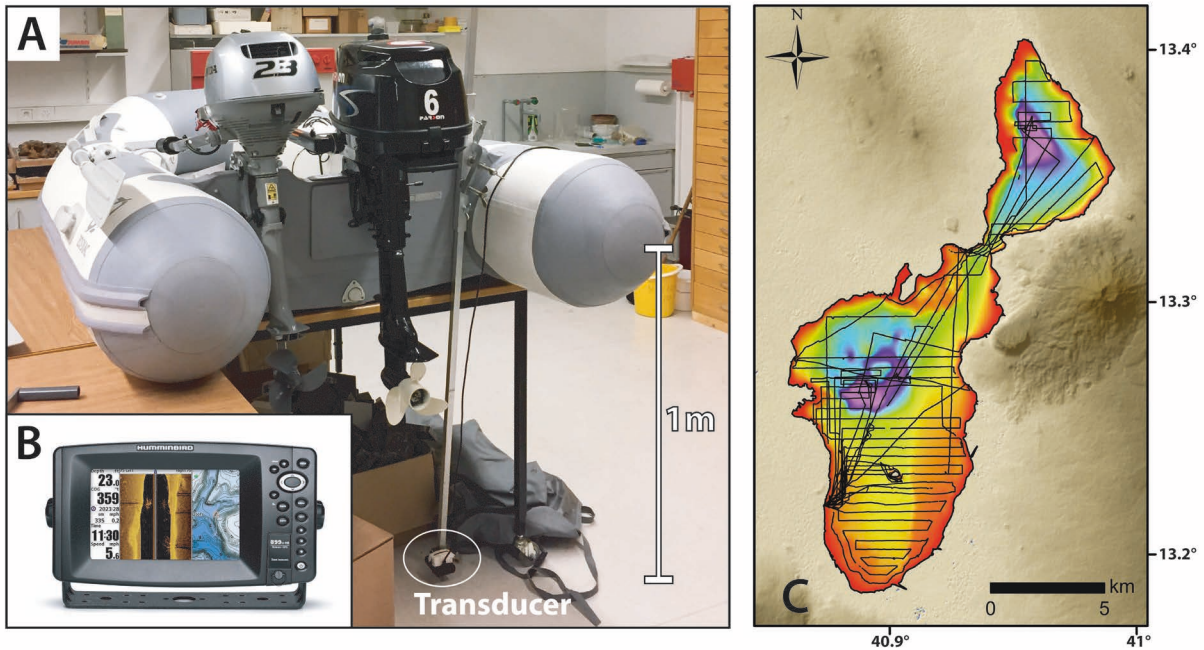


Figure 2: Material used during the bathymetric mapping. A. Rigid Inflatable Boat (Zodiac Cadet 270 Aero) with 2 engines (Parson 6HP water-cooled and Honda 2.3 HP air-cooled engines). The sonar transducer is attached to a perch plunged at 1 m below lake level. B. Echosounder *Humminbird 899ci hd si*. C. Map of the multibeam survey tracklines acquired during the 2 surveys in January 2016 and January 2017.

V.2.1 Bathymetric survey

Two-weeks expeditions were performed in January 2016 and 2017 to map the entire lake Afdera. About 500 km of multibeam survey tracklines (Fig. 2C) composed of 242'000 data points spaced on average by 2.1 m, with each point being measured every second at an average speed of 8 km/h. The first mission in 2016 lasted six days and allowed to examine the feasibility of the project and improve technical difficulties. During the second mission in 2017, seven full days were dedicated to complete the bathymetric mapping of Lake Afdera. Weather and logistical issues reduced the operational time to five full working days. Along the lakeshore, only a few locations rendered the deployment of the RIB system feasible with the most accessible location being *Aligenda hot spring* at the southwestern shoreline of the lake, close to the Afdera village. Therefore, all the survey tracks started at that point. The

survey tracks were planned with 400m distance between each track to obtain the most optimal spatial resolution taking into account the limited operational time.

The sonar system has been mounted on a small Rigid Inflatable Boat (RIB) system, resp. a Zodiac Cadet 270 Aero (Fig. 2A). The Zodiac Cadet is 2.7m in length, has an internal surface of about 1m², and can carry up to three passengers for a maximum load of 420kg. For safety reasons, the boat was powered with two thermic off-board engines, resp. a *Parsun F6ABMS* 6 horsepower (water cooled) as main engine and a *Honda* 2.3 horsepower air-cooled engine as backup.

The echosounder used was the mobile depth/fishfinder *899ci hd si system* from *Humminbird* (Fig. 2B). This echo sonar uses a dual-beam transducer type (XHS-9-HDSI-180-T) capable of giving instant temperature and side imaging, having a maximum working depth of 500 m and target separation of 63.5 mm. The

echosounder operated at 83 kHz with a swath width of 60 degrees. During acquisition, the echosounder has been set to ‘marine mode’ with a fixed conductivity setting of 55mS/cm. The system integrates a 10Hz internal GPS receiver built into the head unit with live chart plotting capabilities. To power the system, two packs of 12 V and 13.5 Ah Li-ion batteries - two *H2B182-B* from *HY-LINE* plugged in series - were used. Data were temporally stored on a regular SDHC memory card. In 2016, only the air-cooled engine was used, and the echosounder transponder was attached on the right tail of the rugged boat. Consequently, the sonar signal was severely affected by the engine vibrations and swash. Moreover, heavy weather conditions resulted in depth anomalies removed during processing. In 2017, the transponder was attached to an adjustable perch plunged at 1 meter below lake level and a water-cooled engine used to reduce vibrations.

V.2.2 Data processing

Sonar data were processed using ArcGIS software. Artefacts, depth aberrations and faulty readings (less than 1% of the total signal) were manually extracted and removed. Depths have been homogenized between the two missions where the probe was installed at different depths. Sonar data were depth-corrected through the calculated in-situ sound velocity of water samples collected at three different water depths and measured temperature and conductivity values (Table 1). The correction factor of 1.0721 was determined to correct the sonar data to the lake Afdera high conductivity/salinity setting, with an accuracy of $\pm 0.35\%$. The map was constructed using the “*topo to raster*”

algorithm and the *ANUDEM* elevation interpolation procedure (Hutchinson et al., 2011).

V.2.3 Fault analysis

Onshore faults have been mapped using satellite imagery (ESRI) and DEM (SRTM, 1 Arc-Second Global Elevation data) in a zone extending from Erta Ale in the north to Manda Harraro in the south. These faults are basically the same as already identified by Varet et al. (1975), Illsley-Kemp et al. (2018) and La Rosa et al. (2019, 2021). The lineaments of an area of 60 x 60 km around the lake were analysed and data plotted in a rose diagram. The orientation of faults larger than 500 m (97% of the set) were split into straight 500 m segments to obtain a fault-length weighted result. Smaller faults and/or segments were considered as single segments.

V.3 RESULTS

The main results of the bathymetry are shown in Figure 3. All data (raw bathymetry datapoints, interpreted bathymetry raster, interpreted immersed and outcropping faults) are available in a data repository (Schaegis et al., 2021). The lake covers an area of 117 km² with an average depth of 20.9 m. The total volume represents 2.4 km³ of water. The lake is divided into two distinct parts: the southern basin, being the largest with 87 km² surface coverage and containing 1.5 km³ of water, and the northern basin covering 30 km² with a volume of 0.9 km³. The southern part of the basin in the South is mostly flat and shallow with a depth ranging from 5 m to 15 m. The Franchetti Island represents a 300m diameter emerged land surface with abrupt topography. Further north, a

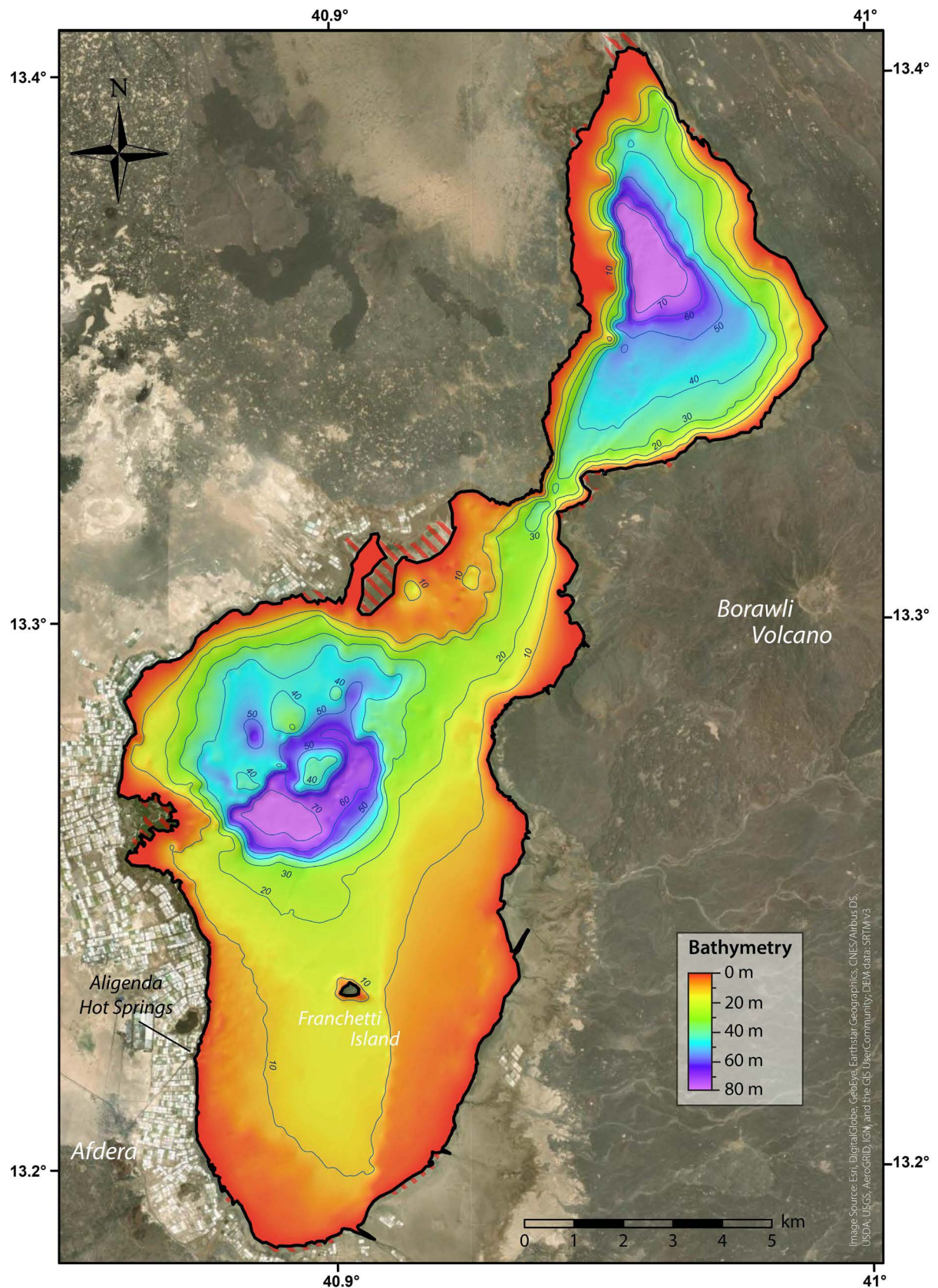


Figure 3: Bathymetric map of Lake Afdera with contour lines indicating depth in meters below lake level. The deepest point of the northern basin is 80 m below the lake level surface and the deepest point of the southern basin is 76 m below lake level surface. The lake surface is at 112 m below sea level. The red dashed surfaces indicate area periodically flooded by a few tens of cm of water.

depression (from 55 m to 71 m) is located within a 5 km diameter circular structure with a noticeable elevation (40 m) in its central part. The northern basin of the lake displays sharp bathymetric changes and a triangular shape with the edges oriented N-S; NNW-SSE; SW-NE. The western edge (N-S axis) is the steepest. The sinuosity of this escarpment is an artefact related to the interpolation algorithm applied between the multibeam survey tracks. Despite being disconnected, both basins exhibit maximum depths of 76 m and 80 m below lake level, or -188 m asl. and -192 m asl. for the southern and northern basin, respectively.

V.4 DISCUSSION

V.4.1 Elevation and dimensions of the lake

The elevation of the lake Afdera water level is today at -112 m asl covering a total surface of 117 km² (SRTM, 1 arc-second DEM). With a minimum elevation of -192 m *asl.*, Lake Afdera is the deepest lake and one of the lowest topographies in the Afar, after Lake Assal in Djibouti, at *ca.* -195 m *asl.* (Langguth and Pouchan, 1975). It is also the deepest known point of the Danakil Depression as the depths of lakes Bakili and Karum remain unknown.

The fluctuation in dimensions and elevation of the lake since the 1960's expeditions remain unclear, as data are incoherent between the reporting authors. The surface of the lake was estimated by Martini (1969) at 70 km² and as such cited by other authors (Gonfiantini et al. (1973) and Bonatti et al. (2017)). The first estimation by Martini (1969) was probably only

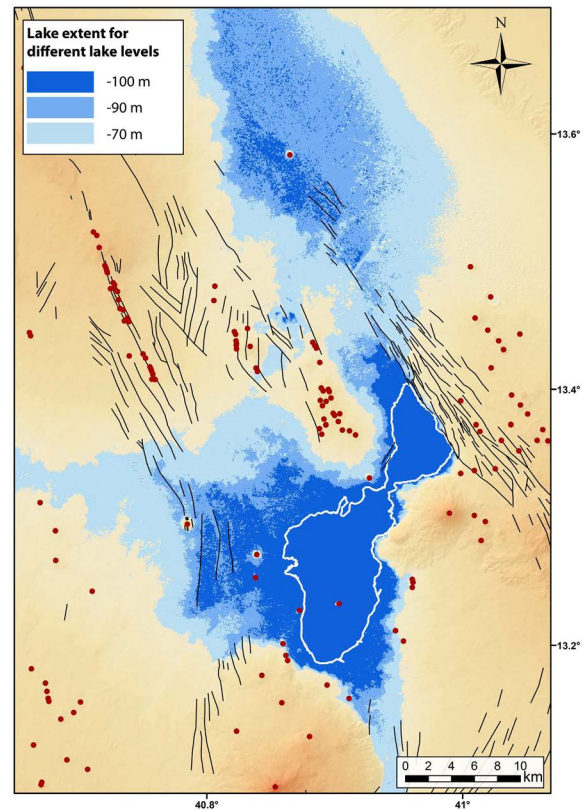


Figure 4: Map showing the potential extent of Lake Afdera assuming lake level elevation of - 100 m asl, - 90 m asl and - 70 m asl.

referring to the southern basin, the only part he mapped and studied, and was mistakenly cited as the total lake surface by the other authors.

The elevation of the lake level has been estimated differently by several authors. Martini (1969) only mentioned that it is situated below sea level. Gonfiantini et al. (1973) estimated the altitude at -70 m asl. Gasse (1974), Gasse et al. (1974) and Gasse & Street (1978) estimated the lake level at -80, -82 and -80 m asl., respectively. Finally, Bonatti et al. (2017) estimated the elevation of the lake in 1969 between -90 and -100 m asl. Based on these estimations and the present day measured elevation of -112 m asl., Bonatti et al. (2017) conclude that “lake’s level appears to have dropped by over 10 m during the last ~50 years”. Due to the flat topography, a higher lake level,

even only at -100 m asl., would cover an area twice as big as today's and would completely change its geometry (Fig. 4). The shoreline drawn by Martini, (1969) and other geological map data (Brinckmann and Kürsten, 1970, CNRS/CNR, 1971) fit exactly the current shape of the lake (Suppl. Fig. 2). Furthermore, local inhabitants never noticed any lake level variation (pers. comm. with local chairman, 2020). This fact is corroborated by the small variations, but the overall constant lake extent observed by remote sensing data collected between 1998 and 2016 (max 6.2% expansion and 4.8% shrinkage relative to the average, Nanis and Aly, 2020). To quantify these minor changes, images of the shore of a period of relatively high water (January 2016) were compared with other images of a period of relatively low water (January 2019, Suppl. Fig. 3). It shows a lake-level change of 25 ± 10 cm. Analysis of Landsat 8 images show that at the same time, the lake surface increased by 4.1 km², i.e. 3.5%. Assuming a linear correlation between lake surface and lake level change, a maximum variation of 11% reported by Nanis and Aly (2020) would correspond to a lake-level change of 80 ± 30 cm. This corresponds to a total lake volume change of 4 ± 2 %, which is surprisingly small. We therefore conclude that the lake level and water volume remained relatively stable for at least the last 50 years, despite evaporation in this arid environment and intensive artificial pumping. The mistakes in the estimation of lake elevation during the previous century are indeed caused by imprecise maps or high uncertainties related to the barometric measurement of elevation. Thus, we consider the conclusion of Bonatti et al., (2017) about lake level drop as incorrect.

Lake Afdera is the second largest lake in the Afar after lake Abbé in Djibouti. Lake Abbé covers an area of 320 km², but its volume is not well documented. According to Karimi et al. (2015), lake Abbé has a maximum recorded depth of 36m, but its water level varies greatly depending on the influx through the largest river in the Afar: the Awash River. Indeed, during dry periods, the lake Abbé level can decrease by 5m (Karimi et al., 2015). On the contrary, no major rivers feed Lake Afdera. The only identified water sources are hot springs (Martini, 1969; Bonatti et al., 2017; Foubert et al., 2019). Lake Afdera is probably the largest stable lake in Afar. Its stability could potentially be explained by largely hidden water reservoirs in the surrounding faulted subsurface (Gasse et al., 1974).

V.4.2 Structural lineaments

Fault mapping around Lake Afdera shows four main structural orientations (Fig. 5C) under the same regional NE-SW stress field, which are interpreted to be either part of the main NNW orientation of the Red Sea rift axis of the Danakil Depression or part of the two active transfer zones in the area.

The first one is the sinistral Alayta – Afdera Transfer Zone which links the main Dabbahu / Manda Hararo rift segment to the Erta Ale and the Tat'Ali rift segments. Although this transfer zone has been mentioned by several authors, no detailed description of its nature and fault orientation has been made. This first recognition of the *en échelon* structure of the Alayta and Tat'Ali range was mentioned by Tazieff et al. (1972). The transversal nature of the faults south of Afdera is discussed by Barberi et al. (1974) who also suggest

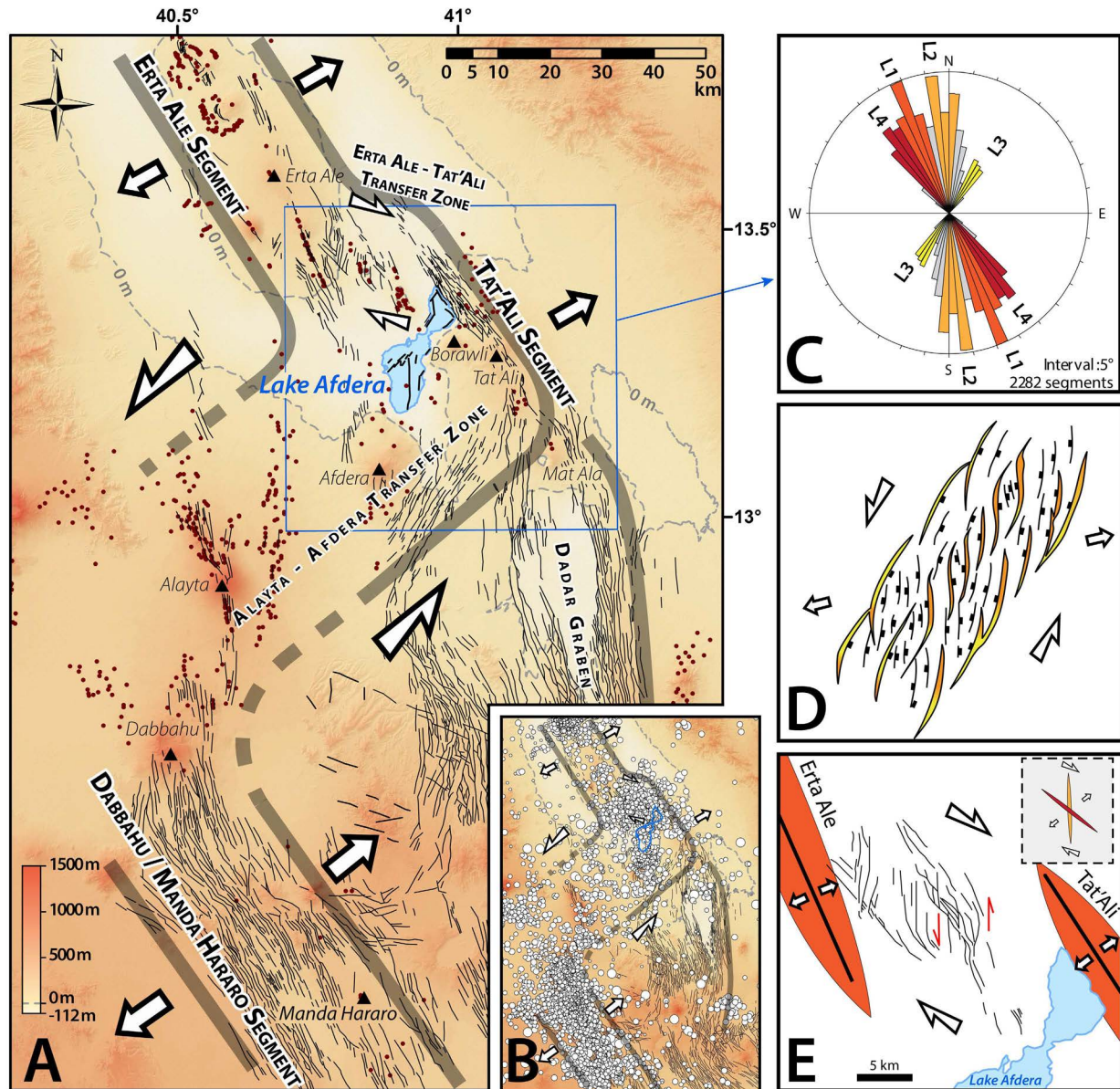


Figure 5: Structural analysis of the larger Afdera region. **A:** Mapping of tectonic lineaments (in black) and of volcanic cones or vents (in brown). **B:** Earthquake compilation in the study area by Zwaan et al. (2020) including data from Keir et al. (2006), Belachew et al. (2011) and Illsely-Kemp et al. (2018). **C:** Length-weighted rose diagram showing the orientation of the tectonic lineaments around lake Afdera in the area represented by the blue square on A. **D:** Results of analogue modeling of oblique rifting by McClay et al. (2002). The colors represent the orientations shown on panel B. Comparison between the Alayta – Afdera Transfer Zone and the model of McClay et al. (2002) shows striking similarities in fault orientation. This transtensional system is hence interpreted to have formed the L2 and L3 fault orientations. **E:** Structural interpretation of the Erta Ale – Tat’Ali Transfer Zone redrawn after La Rosa et al. (2019, 2021). It shows that this transfer zone formed L2 and L4 orientations.

that Lake Afdera might be part of the same transverse structure. Barberi & Varet (1977) then proposed that these faults represented a sinistral transform boundary. This concept of transfer zone between Dabbahu / Manda Hararo rift segment and the Erta Ale / Tat’Ali rift

segment has also been assumed in several recent publications (Stab et al., 2016; Gardo and Varet, 2018; Le Gall et al., 2018; Varet, 2018; Ayalew et al., 2019; Zwaan et al., 2020). It is also supported by earthquake data and seismic moment release (Zwaan et al., 2020, Fig. 5B). A

comparison between the fault pattern of the Alayta-Afdera Transfer Zone and analogue modelling of oblique rifting (McClay et al., 2002, Fig 5C) shows striking similarities, thus providing a conceptual model for the fracture pattern development. The Alayta-Afdera Transfer Zone would thus constitute a broad area of diffuse deformation and important magmatism forming a sinistral linkage zone between the Dabbahu / Manda Hararo rift segment and the Erta Ale / Tat'Ali rift segments.

The second transfer zone is the smaller dextral Erta Ale – Tat'Ali Transfer Zone, which links the Erta Ale and Tat'Ali rift segments and that has been investigated based on seismicity, InSAR, structural geology and described in great details by Illsley-Kemp et al. (2018) and La Rosa et al. (2019, 2021).

The first structural group mapped around the lake (L1, Fig. 5C) is oriented NNW-SSE and corresponds to the Red Sea rift orientation. It constitutes the main rift-axis orientation and is identical to the directions observed on the Erta'Ale range, the Tat'Ali and between Dabbahu and Manda Hararo. The Alayta – Afdera Transfer Zone is interpreted to have generated the L2 (N-S) and L3 (NE-SW) structural trends. The L2 (N-S) orientation is present west of Lake Afdera, on the Afdera volcano and around the Alayta, both as faults and volcano alignments. L3 (NE-SW) is mainly found south of Tat'Ali and Afdera volcano.

As demonstrated by Illsley-Kemp et al. (2018) and La Rosa et al. (2019, 2021), the Erta Ale – Tat'Ali Transfer Zone formed the L4 (NW-SE) orientation but also the L2 (N-S), which is shared with the Alayta – Afdera transfer system. The L2 orientation is hence common for the

two transfer systems. The four identified main structural orientations are consistent with the overall NE-SW opening kinematics of the Danakil depression. Lake Afdera is therefore situated in a unique position at the intersection of two transfer zones and the rift axis, possibly marking the position of an incipient transform fault (Illsley-Kemp et al., 2018).

V.4.3 Underwater topography

The bathymetry of the lake gives insights into the lake floor geomorphology. The slope map (Fig. 6A) indicates steep elongated features in the northern basin that are interpreted as faults. Their orientation coincides with the structural trends identified around the lake. The L4, L2 and L3 structural trends are all expressed in this northern basin. Hence, this basin is interpreted as a triangular tectonic graben caused by the intersection of three different fault orientations. The southern basin is marked by a clear L2 fault lined up along its length axis. The 5 km-wide circular feature representing the deepest part of this basin is interpreted as a submarine caldera. This caldera is structured by L3 and L2 faults. The elevation in the central part of the caldera is interpreted as a resurgent scoria cone (Fig. 6D). Its shape and size mirror the volcanic cone forming the Franchetti Island. This morphology is similar to the nearby Tat'Ali volcano which shows a 3 km wide caldera structured by L1 to L2 faults and similarly featuring a nested volcanic cone. Many other volcanoes in the Afar depression or the Main Ethiopian Rift feature comparable tectono-volcanic morphologies (caldera structured by faults and with nested volcanic cones, e.g. Hunt et al., 2019). The arcuate features

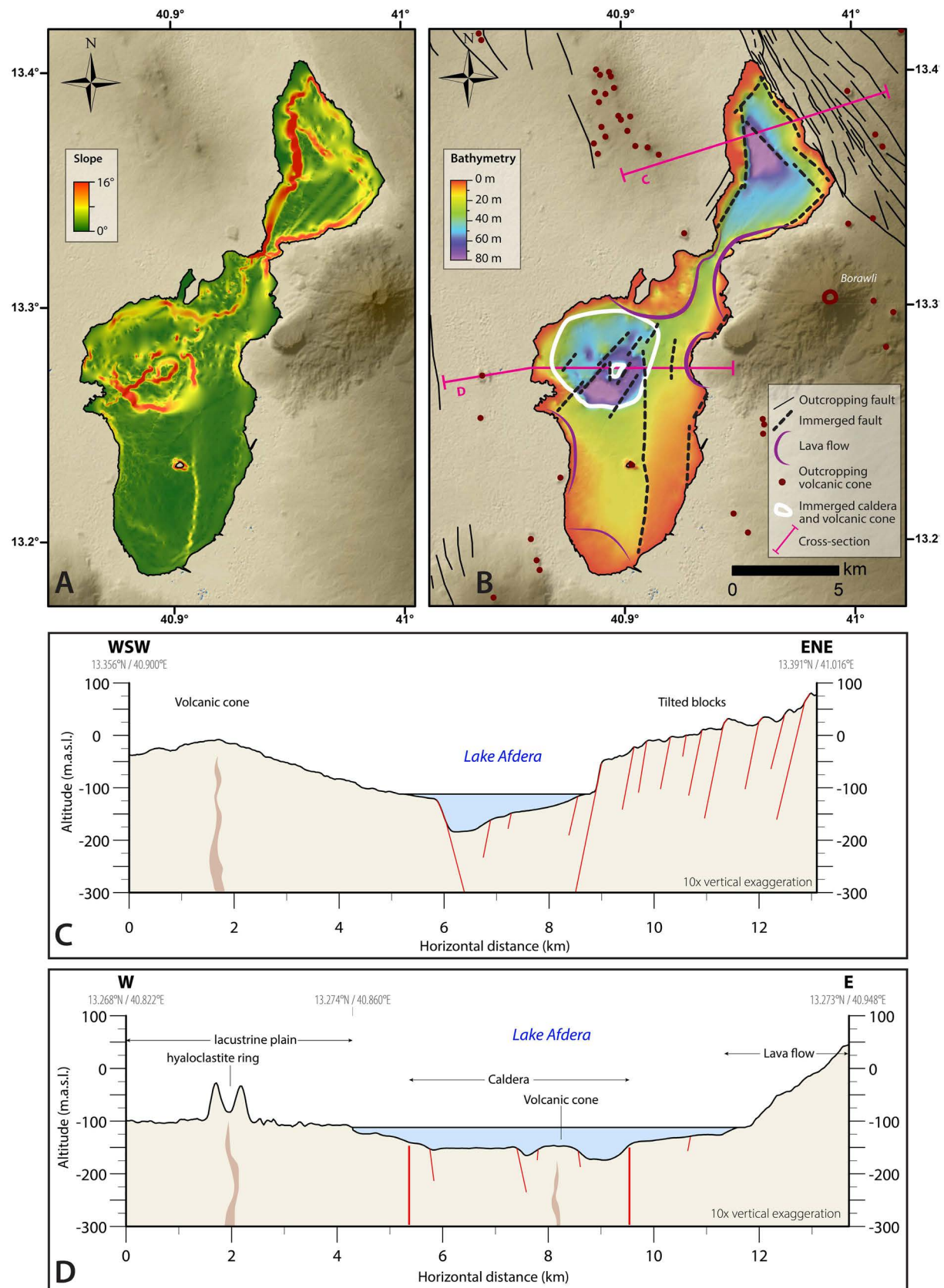


Figure 6: Interpretation of the lake morphology. Slope map (A) allows to identify several faults as well as a volcanic edifice in the lake (B). C and D show cross-sections of the northern and southern basin. The northern basin shows a graben morphology while the southern basin indicates a faulted caldera with a nested volcanic cone.

at the foot of the Erta Ale range, the Borawli and Afdera volcanos are interpreted to be lava flows. They are situated just offshore of lava flows either observed in the field or visible on satellite images.

The presence of a caldera in a depression is less common. However, several calderas in the area are found on nearly flat shield volcanoes (e.g. Erta Ale, Mat Ala, unnamed volcanoes at 12.09N, 40.89E; 12.165N, 40.927E) and several volcanoes are situated at very low elevations (e.g. Catherine volcano [13.912N, 40.449E, -123 m], Dallol volcano [14.240N, 40.299E, -92 m, no lava outcropping]). The combination of both is thus not surprising. We therefore interpret that this caldera is now situated at lower elevation because of the lower magma production compared to the neighbouring volcanic ranges (Erta Ale Range, Tat'Ali Range and Afdera Volcano).

Overall, these interpretations imply that both basins are genetically different. The northern basin is a tectonic influenced basin, while the southern one is controlled by volcano-tectonic morphologies. However, both form in response to the same regional stress field and at the intersection of fault sets of different orientations forming small-scale pull-apart basins.

V.4.4 Lake Afdera in the Danakil rift system

The geomorphological structures interpreted in Lake Afdera can be directly related to other rift structures in the Afar Rift System. The large caldera inferred in the southern basin is situated on the axis of the Erta-Ale rift segment, featuring numerous recent volcanoes.

The fault pattern with respect to the regional stress field indicates a strong influence of transverse structures in the local tectonics. Two transfer zones are linking the en echelon arranged rift axes segments. Lake Afdera is mainly structured by the major Alayta – Afdera Transfer Zone as shown by the predominance of L2 and L3 orientations (see Fig. 5A). The smaller Erta Ale – Tat'Ali Transfer Zone probably only influences the northern basin with L2 and L4 orientations. Regionally, the transverse nature of the zone is also expressed by petrological characteristics typical of a transverse structure (Barberi et al., 1974).

As discussed above, both basins formed mainly due to the interaction between the rift axis and the Alayta – Afdera Transfer Zone. Like Bonatti et al. (2017), we consider this lake and chiefly its northern basin to be the equivalent of the oceanic “nodal basins”, deep basins forming at slow-spreading ridge-transform intersections (RTI) (Sleep and Biehler, 1970; Fox and Gallo, 1984; Karson et al., 1984; Goud and Karson, 1985; Pockalny et al., 1988; Thiede et al., 1990). The triangular shape of the northern basin is typical for this structure. It can hence be considered as an incipient transform zone (see also Illsley-Kemp et al., 2018). The position of the lake Afdera as the deepest point of the Danakil Depression is therefore not fortuity but the result of the interaction between the rift axis and transverse structures. In contrast to their oceanic counterparts, faults of the Alayta – Afdera Transfer Zone are not perpendicular but oblique to the rift axis and, as such, the dominant kinematics probably involve oblique movement along the exposed faults. The zone acts as a

transtensional relay expressed in continental crust, with a broader area of strain partitioning than known from pure, more narrow oceanic transform faults affecting thinner oceanic lithosphere.

V.5 CONCLUSIONS

This study unveils the first complete bathymetric map of the lake Afdera with an area of 117 km², a volume of 2.4 km³, an average depth of 20.9 m and a maximal measured depth of 80 m below lake level. The lake Afdera is therefore the largest and deepest endorheic stable lake in Afar, hosting the lowest measured topography in the Danakil Depression. This study provides a present state of knowledge of the lake Afdera from its discovery in 1920 to the different attempts to estimate its depth. We evidence that the lake water volume did not fluctuate by more than a few percent during the last 50 years, despite artificial pumping and climatic evaporation. This could potentially be explained by large hidden water reservoirs in the surrounding faulted subsurface.

The bathymetric data indicate that the morphology of the northern basin is controlled by tectonics with faults oriented in three different directions, forming a triangular graben structure. The southern basin is interpreted to be influenced by volcano-tectonic processes with a large submerged caldera structured by faults and featuring a major caldera and associated resurgent volcanic cone. The fault orientations throughout the lake indicate that the lake is influenced by two transfer zones, the Alayta – Afdera Transfer Zone and the Erta Ale – Tat'Ali Transfer Zone. The former probably having a stronger

influence on the lake structure than the latter. The interaction between the Alayta – Afdera Transfer Zone and the rift axis in the northern part of the lake Afdera forms the equivalent of an oceanic nodal basins and explains the very low elevations encountered. Lake Afdera represents as such an incipient transform fault zone on the floor of the Afar depression.

Finally, this study attests that deploying portable and easy bathymetric mapping systems in remote areas result in the acquisition of valuable datasets in challenging environments.

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AUTHOR CONTRIBUTIONS

- J-CS: Concept / Design / Material / Data Collection (bathymetric survey) / Analysis and interpretation / Literature research / Writing.
- VR: Data Collection (lineament mapping) / Data Processing (bathymetric survey) / Analysis and interpretation / Figures / Literature research / Writing.
- TK: Supervision / Critical review / Interpretation.
- JM: Critical review / Interpretation.
- EG: Data Collection (bathymetric survey) / Critical review.
- BA: Critical review.

- AF: Supervision / Concept / Design / Critical review / Interpretation.

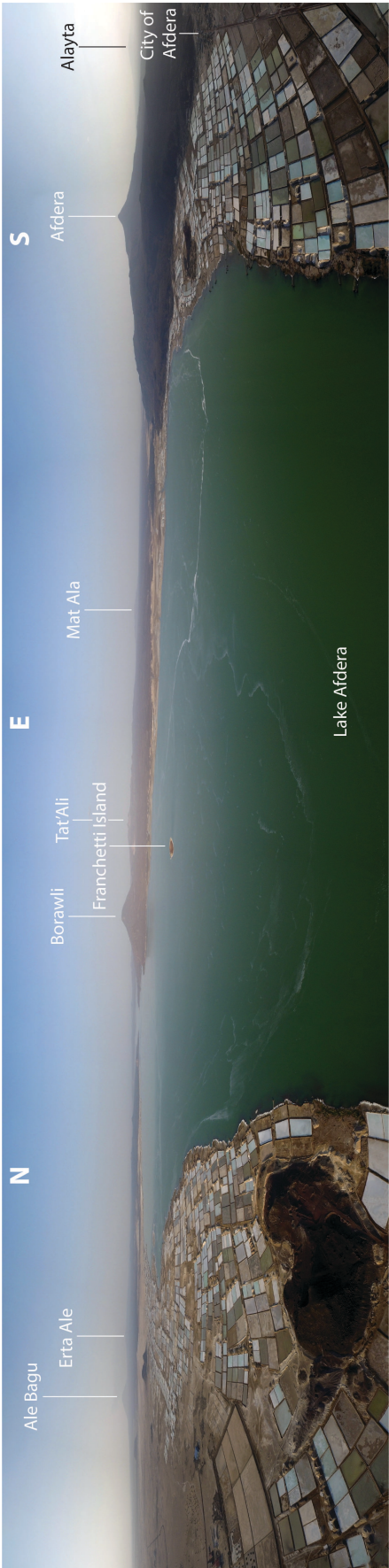
CONFLICT OF INTEREST

The authors declare that the research was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

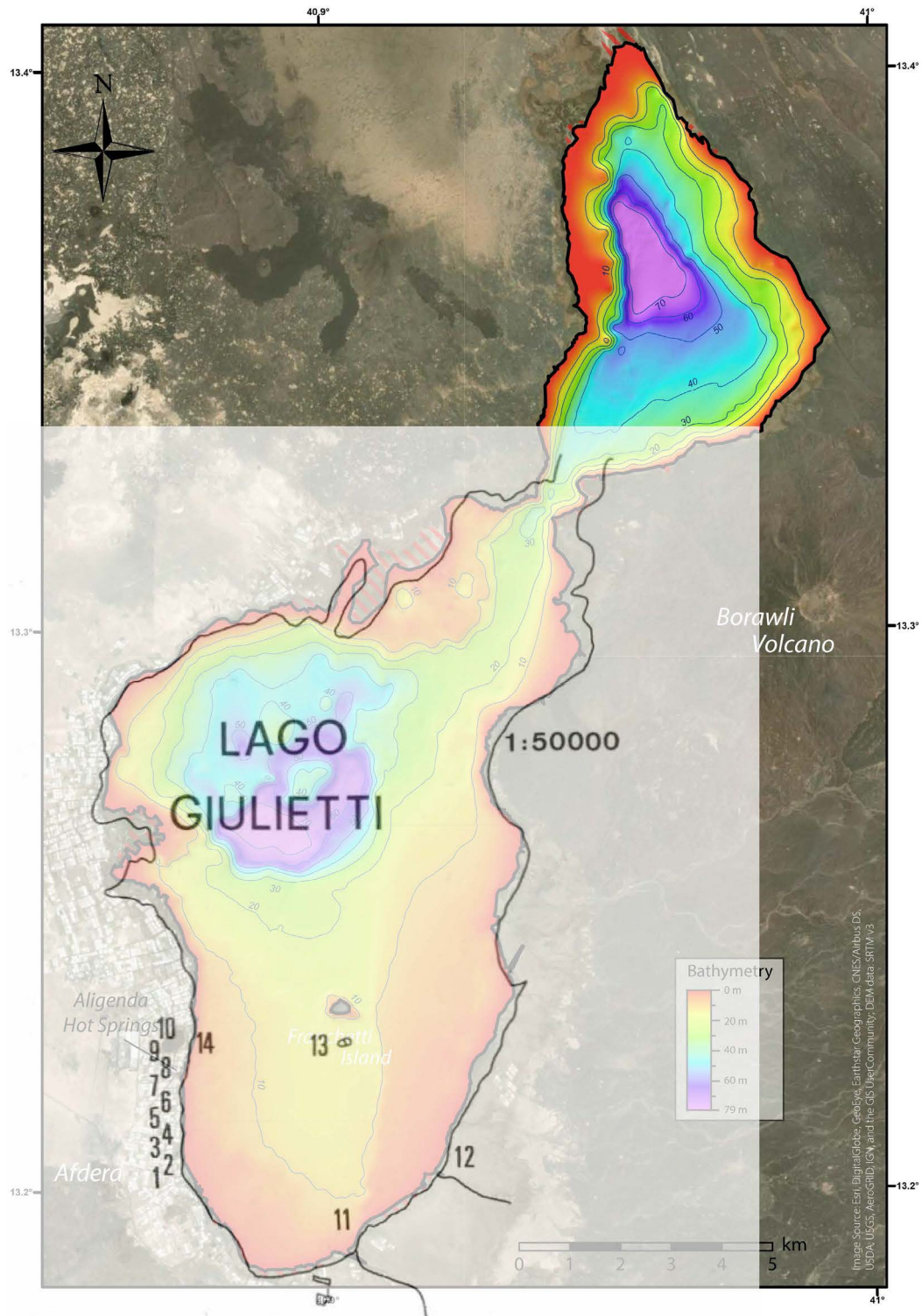
DATA AVAILABILITY

All raw data are available in the following data repository: ‘Schaegis, J.-C., Rime, V., Kidane, T., Mosar, J., Gebru, E. F., Atnafu, B., & Foubert, A. (2021). Bathymetric and structural data of Lake Afdera and surroundings, Afar, Ethiopia. PANGAEA. <https://doi.org/https://doi.org/10.1594/PANGAEA.935008> ‘

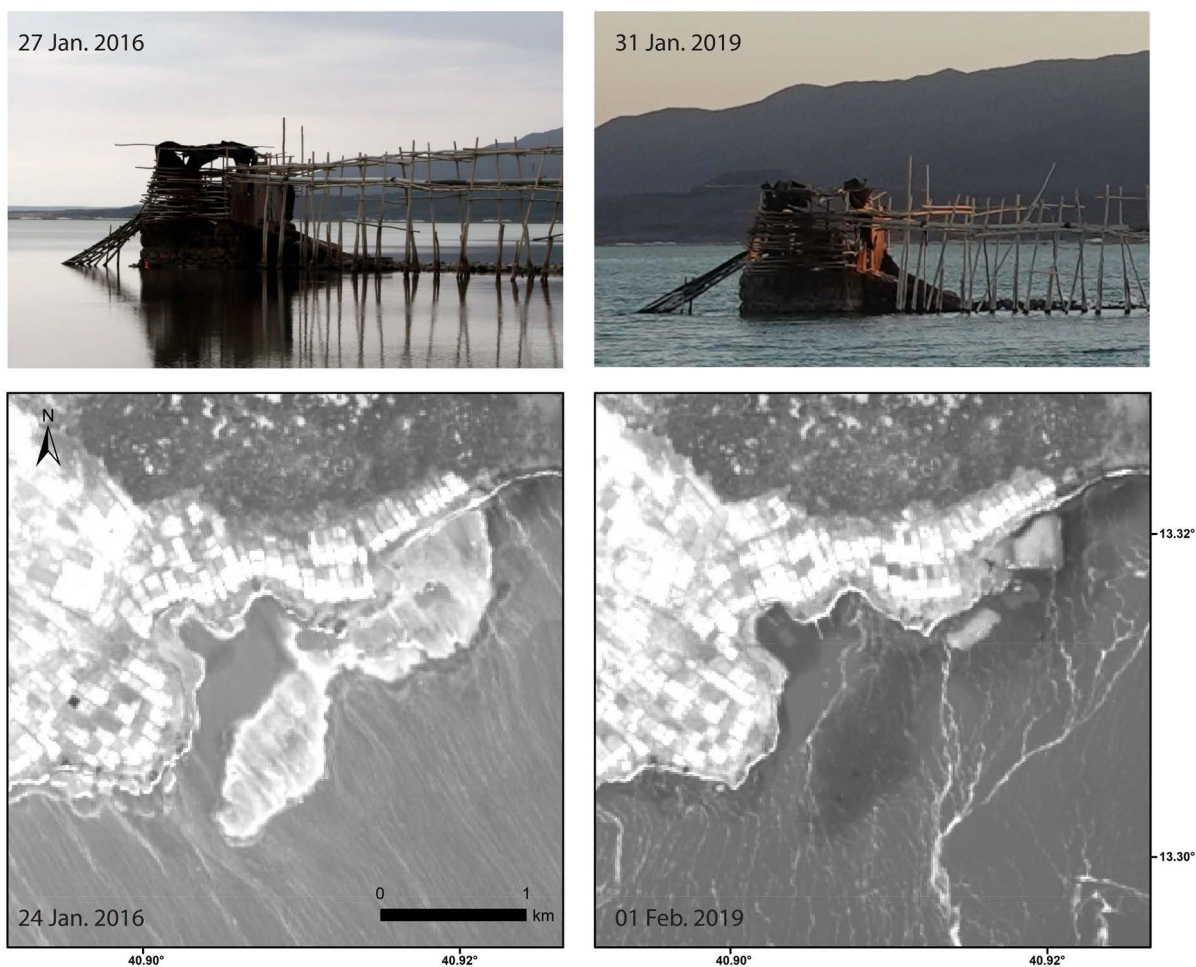
SUPPLEMENTARY MATERIAL



Supplementary Figure 1: Panoramic drone view of the Lake Afdera. Approximate position: 13.220N, 40.872E. Salt evaporation ponds appear in light colors around the lake.



Supplementary Figure 2: The comparison between the map of Martini (1969) as transparent overlay and present-day lake Afdera show no noticeable differences. It invalidates the hypothesis suggesting that the lake level was at significantly different elevation during the 60's.



Supplementary Figure 3: Comparison between a period of relatively low waters (January 2016) and a period of relatively high waters (January 2019). The upper images show that the lake level changed by approx. 25 ± 10 cm (small red mark on the upper left image). The lower Landsat 8 (Band 8) images show the related lake extent variation on the northern part of the southern basin. Over the whole lake, it corresponds to a surface change of 4.1 km².

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VI DISCUSSION, CONCLUSION AND OUTLOOK

VI.1 DISCUSSION

VI.1.1 Basin dynamic evolution of the Danakil Depression

The major aim of this study was to better understand the basin dynamics of the Danakil Depression with a particular focus on the structure, the evolution, the rates and the factors controlling these processes. The study sheds light on these aspects at different temporal and spatial scales.

The origin of the rifting in the Afar and Danakil depressions is situated several thousand kilometres north and initiated more than 50 million years ago. The subduction of the Neo-Tethys was driving the African continent towards Eurasia. During the Oligocene, the Arabian continent began separating from the African continent, the latter slowing down its northward movement (section II.2 and references therein). The localisation of this process in the present-day Afar region was probably related to the eruption of the flood basalts during the early Oligocene weakening the crust (section II.2 and references therein) and to the presence of several inherited structures of different nature and age meeting in the Afar (section II.5.3). After a long period of rifting in the Red Sea, tectonic reorganization in the Mid to Late Miocene led to a major rift jump from the Red Sea to the Afar Depression and to the counter-clockwise rotation of the Danakil

Block. This situation prevailed until now and shapes the entire Danakil Depression. After several million years of continental rifting dominated by mechanical extension, the importance of magmatic processes increased in the last half million years, in particular in the southern part of the Depression. This increase was probably related to a fast thinning of the crust and possibly indicate continental breakup in the southern part of the Depression (Bastow & Keir, 2011; Keir *et al.*, 2013). This thinning led to the subsidence of the Danakil Depression below present-day sea level and the creation of important accommodation space. During the Mid to Late Pleistocene, climate-induced eustatic sea-level variations combined with tectonic uplift of the rift shoulders and subsidence of the basin in the order of tens to hundreds of metres led to several cycles of flooding of the basin from the Red Sea (chapter III). These incursions allowed for the formation of coralgal reefs and associated marine deposits around the depression. The last incursion (MIS 5e, i.e. ~130 ka) was very short. The very shallow connection to the Red Sea and a rapid change to more arid climatic conditions led to the rapid (possibly <3 ka) deposition of >100 m of marine evaporites (chapter IV). After the complete isolation of the Depression after MIS5e sea-level peak, recycling of evaporites from its margins by meteoric waters and the creation of important accommodation space by tectonic

processes led to the deposition of more than 350 m of continental evaporites in the central part of the basin (chapter IV). Climatic conditions played a crucial role in shaping the Late Pleistocene to Holocene sedimentary architecture of the basin.

The rapid creation of accommodation space in the basin is caused by several factors. Stretching and thinning of the crust (see section II.6.1 and appendix II.A) causes first order subsidence. In the northern part of the Danakil Depression, normal faulting in the upper crust forms a fast-subsiding graben or half graben structure (chapter III, section II.7.1 and Bastow *et al.*, 2018). The load of the Erta Ale range probably plays an important role in the subsidence of the southern part of the Depression (Corti *et al.*, 2015; Le Gall *et al.*, 2018, section III.4.1). Lake Afdera is an example of the dual impact of tectonics and volcanic processes on accommodation space (chapter V). The relative distance of the Danakil Depression to the Afar hotspot also plays a role in its important subsidence. The Central Afar domain situated on the hotspot experiences magma-compensated thinning and more important dynamic topography. As such, it never subsided below sea level even though it experienced significantly more extension (section II.7.1).

Beside tectonics and volcanic controls, accommodation space in the Danakil Depression is also controlled by changes in base level, either sea or lake level. In this isolated basin, these changes are much more sensitive to small variations in the system than in an open ocean. Eustatic sea-level changes of a few meters can lead to a complete flooding or desiccation of the basin and a base-level variation in the order of 100 meters

(chapters III and IV). Similarly, lake level is very sensitive to climate and can change drastically within very short period of time (chapter IV). Finally, sedimentation rate plays an important role in the modification of accommodation space. Chapter IV shows that sedimentation rates can vary by at least two orders of magnitude in the basin depending on the hydrological and climatic conditions (see also below).

In the near geological future, anthropogenic sea-level rise will probably lead to a new flooding of the depression. Subsidence of the Buia Graben will potentially lead to a continuous connection with the Red Sea, even during glacial periods, within 200 ky to 1 My (chapter III), ending the current phase of fluctuation.

The present study therefore allows to better understand the structure and the evolution of the Danakil Depression. Relatively good timing constrains allowed to quantify relatively precisely the rates of several tectonic, sedimentological, hydrological, and climatic processes and to make some predictions for the future. Finally, the factors controlling these geological processes are better identified.

VI.1.2 On the complexity of geological processes in a rift basin

The results of this thesis highlight the **complexity** of geological processes in a rift basin. Chapter II shows that plate tectonics in the Afar region is complex. The Danakil Block is not acting as a single rigid block and its rotation only occurred >15 My after rifting initiation. The movement of the Ali Sabieh Block, Bia Anot fault and south-eastern Afar are

also complex and not yet clearly understood. Section II.7.1 shows that the Central Afar and the Danakil Depression represent two distinct types of rift with contrasting geological history. Chapter III shows that not only eustatic sea-level variation plays a role in flooding, but also tectonic movement of the rift shoulder and, to a lesser degree, sedimentation and volcanism. It is also shown that the position of the gateway might be different in the future than in the past. Chapter IV highlights that previous stratigraphic descriptions of the basin fill were too much simplified. The same chapter shows that the formation of thick evaporite deposits is a complex process involving the interaction between distinct parent fluids in different environmental settings and under changing climatic conditions. The climatic history of the Danakil Depression is also complex with very local climatic variations, e.g. a wet period probably started earlier in the northern Afar compared to other parts of Eastern Africa (chapter IV). Finally, it is proposed that Lake Afdera was not just the result of faulting but was also probably hiding a caldera (chapter V), evidencing the complexity in topography resulting from both tectonic and volcanic processes. Additionally, the constant volume of water in Lake Afdera for at least 60 years is a complexity *per se* as the opposite would be expected by the intense evaporation rates reported in the basin. This remains an unsolved mystery.

The present thesis also emphasizes the **spatial and temporal variability** of structures and processes in the Afar Depression. An example is the complex sedimentary architecture of the Danakil

Depression with lithologies varying at very short spatial and temporal scales, ranging from boulder-sized fluvial conglomerate to sandstone, claystone, evaporitic halite and gypsum, marine carbonates, lacustrine finely laminated clays and eolian sands. Chapter IV shows that significant changes in paleo-environments occurred in the last ~130 My. Chapter II reveals that the structure of the Danakil Depression is different than the one of the Central Afar and that both evolved differently through time. At smaller scale, chapter III displays dramatic changes in uplift rates on outcrops situated only a few hundred of meters away. These variations in uplift rate can be observed both perpendicular and parallel to the rift axis, owing to the intense and complex faulting and to the differential isostatic loading of the Danakil Depression.

This study also highlights the **non-linearity of geological processes**. In chapter IV, it is shown that evaporites can precipitate at rates up to 25 cm/yr while quieter periods (deposition of subunit A4) resulted in sedimentation rates less than 0.2 cm/yr. At larger scale, the rifting history in the Afar is ~27 My old but in the Danakil basin, more than 600 m of sediments were accumulated in less than 130 ka, i.e. less than 0.5 % of the whole rifting history³. By extrapolating those sedimentation rates towards past and future floodings, several kilometres of sediments might be produced in less than two million of years. Chapter III furthermore shows that the rate of tectonic movements are not constant. The measured uplift rate is systematically

³ or less than 1.2 % since rifting was focussed in the Danakil Depression (~11 Ma)

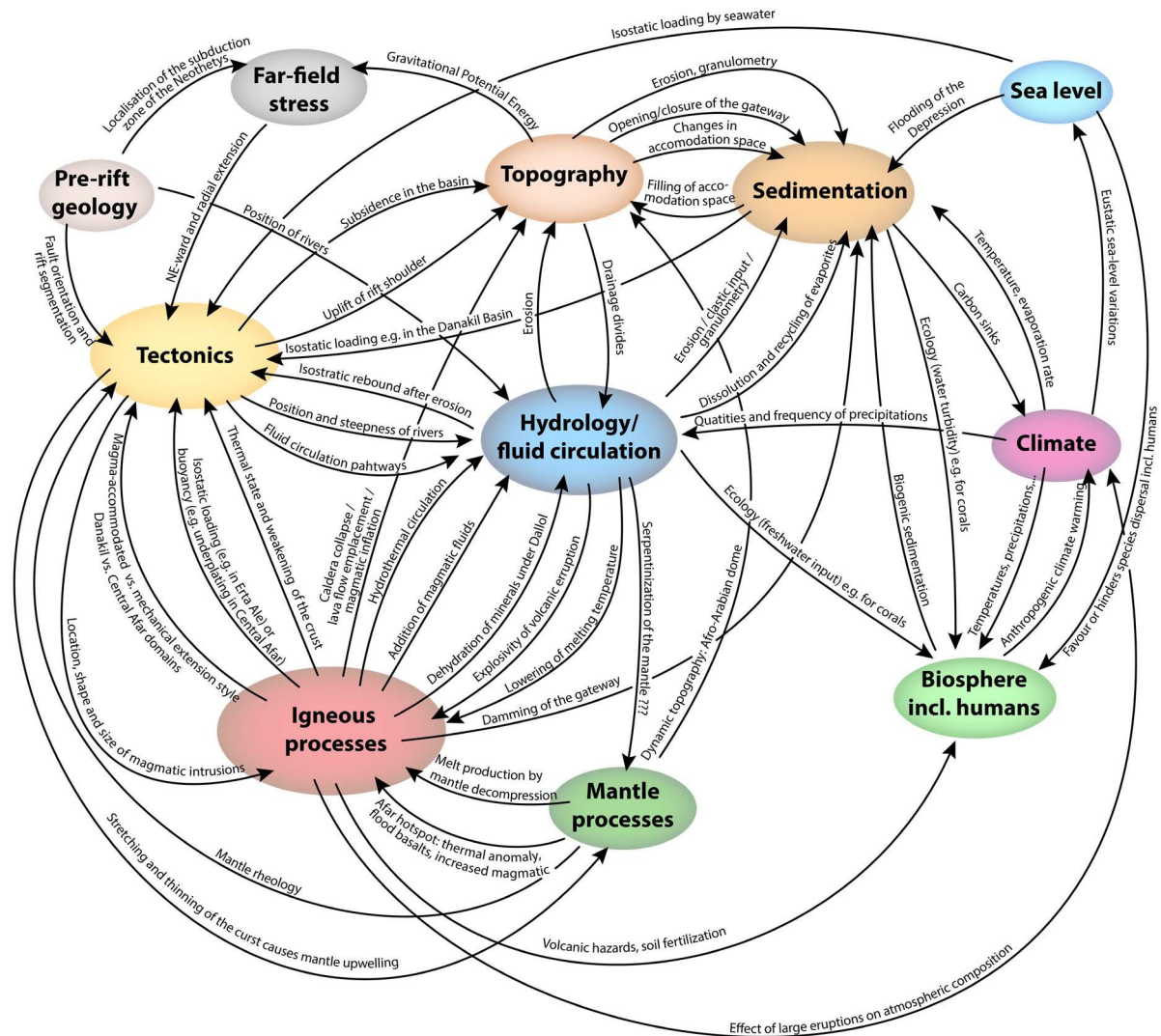


Figure 1: Schematic illustration of the complex relation and interaction between different processes controlling and driving the basin dynamic evolution of the Danakil Depression.

higher between MIS 7 and 5e than between MIS 5e and now. These examples show that the geological history and, in this case, rifting is defined by tipping points that might have a rapid and profound influence on the geological structure of an area, but also on its geography, potentially impacting oceanic circulation, climate and species dispersal.

The complexity, spatial and temporal variation as well as non-linearity of geological processes in rift settings are caused by the fact that multiple processes (e.g. tectonic, magmatism,

sedimentology, climate, biosphere, see fig. 1) are closely related, interacting and influencing each other with multiple feedback mechanisms. For example (fig. 1), anthropogenic greenhouse gases emissions induce climate change which increases sea-level. This will result in flooding of the Danakil Depression, increasing accommodation and sedimentation and, as such, increased isostatic loading that may trigger volcanic eruptions which could have again an influence on climate and on life. The thorough understanding of such

complex interactions requires careful scientific work, taking a multiproxy and/or multidisciplinary approach at different scales. All this relying on solid field mapping, detailed outcrop descriptions, well-defined integrative stratigraphic frameworks, rigorous sample collection and data analyses. Only then can rift and passive margin development be better understood.

VI.2 CONCLUSION

The present study aimed to better understand the basin dynamic evolution of the Danakil Depression. A multi-scale and multidisciplinary approach is used, ranging from tectonic plate scale to microscopic scale.

A holistic and integrative review of the geological history of the Afar Depression and the Afro-Arabian Rift System is compiled. It allows to propose new geological maps of the Afar Depression and of the southern Red Sea & western Gulf of Aden region. This review shows that geological events are diachronous throughout the region. A comprehensive kinematic model was developed, illustrating the evolution of the region. In contradiction to most of the published literature, it is shown that rifting in the Afar is propagating northward and not southward. It is also proposed that the Danakil Depression and the Central Afar represent two different domains with distinct structures and geological histories, with the Central Afar being much more influenced by the Afar Hotspot. As such, the Danakil Depression and the Central Afar represent two different types of rift and not two different stages of evolution of the same type of rift. The Danakil Depression is probably the equivalent of a magma-rich

margin and the Central Afar possibly represent the initial stage of an oceanic plateau.

Zooming in the Pleistocene and Holocene history of the Danakil Depression, the mechanisms of flooding of the depression are characterized by the measurement, in the field and by remote sensing, of the elevation of coralgal reef terraces. These data allow to describe recent vertical tectonic uplift/subsidence in and around the Danakil Depression. It is shown that the flooding of the Depression is controlled by eustatic sea-level variations and uplift of the rift shoulders. Potential future flooding of the Depression caused by anthropogenic sea-level rise (on short term) and continuous rifting (on medium term) is discussed.

The desiccation stage of the most Recent (MIS 5e, i.e. ~130 ka) marine incursion is further investigated by the study of a 625 m deep core drilled in the central part of the basin. This core highlights the thick evaporitic infill of the basin which can be considered as a *salt giant* analogue. The analysis of these deposits allows us to better understand the mechanisms and rates of evaporite formation in the Danakil Basin and elsewhere. This study furthermore constrains the Late Pleistocene to Holocene paleoenvironment and paleoclimatic evolution of the Danakil Depression.

Finally, the origin and nature of Lake Afdera is investigated, being a very Recent hypersaline lake in the Danakil Depression. The analysis of bathymetric data shows that it represents the lowest known elevation of the Danakil Depression and evidence a hidden caldera in the southern basin. The low elevation is explained by its position at

the intersection of the rift axis and two transfer zones. The volume of the lake and the variations of this value over the last 50 years could be quantified for the first time. This has important implications for the local population who use this lake for salt production.

VI.3 OUTLOOK

By investigating the Danakil Depression at multiple scales, the present work tried to shed light on some aspects of its basin dynamics. However, many questions remain, and new questions and hypothesis are formulated. Some of these questions and potential future research pathways are listed below. Please note that this list is non-exhaustive.

VI.3.1 Hydrology and hydrogeology in the Danakil Depression

While the structure of Lake Afdera is now better characterized (chapter V), important questions remain on its hydrology. Chapter V shows that the level and volume of the lake didn't vary much since at least 60 years. This comes as a surprise considering the high evaporation rate and the anthropogenic pumping of water for salt production. The water chemistry of the lake is another puzzle. Spring water composition is not similar to lake water composition. Several episodes of precipitation of different mineral phases are recorded around the lake and ongoing studies at the University of Fribourg will improve the understanding of this system. However, these studies will only partially shed light on the further understanding of the hydro(geo)logy of the whole basin. Other lakes (Karum, Bakili), as well as

the pathways and chemical evolution of groundwater remain largely unknown. The knowledge of the hydrology and hydrogeology in this basin, which is crucial for the local population, could be relatively easily improved by large-scale tracing experiments. The drilling of shallow wells could give further insights into the chemical evolution of water across the margin and the basin. It would also be beneficial to conduct these studies on the eastern margin of the Depression. As mentioned in chapter IV, the hypothesis of seawater seepage has been proposed to explain the unusually high bromine content in some isolated ponds of water. Such studies could however not explain deep hydrothermal fluid circulation, like the ones surfacing at Dallol, Gaet'Ale or Black Mountain. These would need further elemental and isotopic analysis.

VI.3.2 Late Pleistocene and Holocene climatic history of eastern Africa

Chapter IV brought some new insights into the paleoclimate of the Danakil Basin. While only shortly discussed, more detailed records are necessary in the Afar and should be compared and integrated with other records in eastern Africa and Arabia (Gasse & Street, 1978; Gillespie *et al.*, 1983; Gasse & Fontes, 1989; Hailemichael *et al.*, 2002; Fleitmann *et al.*, 2011; Tierney *et al.*, 2017; Lamb *et al.*, 2018; Oodendijk *et al.*, 2020; Schaebitz *et al.*, 2021; Foerster *et al.*, 2022) to understand the paleoclimatic variability during the Late Pleistocene and Holocene. Those results should be linked with the migration of *Homo Sapiens* out of Africa, trying to understand how climate influenced the movement of our

ancestors (Beyin, 2006, 2021; Abbate & Sagri, 2012; Tierney *et al.*, 2017).

VI.3.3 Flooding history of the Danakil Depression

The SNF-funded SERENA project results significantly improved the understanding of the Mid Pleistocene to Holocene sedimentary architecture of the Danakil basin linked with the flooding and desiccation cycles. As such, a re-evaluation of the current lithostratigraphy proposed more than 50 years ago by Brinckmann & Kürsten (1971) should be considered. This work started already and should be further finalized.

To constrain better the mechanisms of opening and closure of the basin, field studies on the eastern margin, in particular in the region of the Amphila Gateway, are necessary. Precise field observations, datings and GNSS measurement could give precious insights into this narrow connection between the Red Sea and the former Danakil Sea, including paleodepth, tectonic movement, as well as the extent of erosion and sedimentation during and after the marine incursion(s).

Several other key areas of the Danakil Depression would benefit from more detailed fieldwork and further datings for the understanding of tectonic movements in Danakil Depression, paleoenvironment and reef ecology. These include the Sodonta plain and the Harak graben. Preliminary observations within the framework of project SERENA indicate these deposits are lacustrine and not related to the marine flooding of the depression but Varet & Gardo (2020) mention the presence of coral reefs in the area. More detailed fieldwork could

further clarify this question and GNSS measurement could give very interesting indications on the rates of tectonic movements (uplift/subsidence) in this rift segment. Another key area would be the surroundings of Lake Bakili where Barberi & Varet (1970) mentioned the presence of reefs (see also chapter III). These reefs probably corresponds to the lowest outcrops of marine and paleoshoreline deposits in the Depression and a precise measurement of their elevation could give a minimum value for the subsidence rate in the basin. A final key outcrop would be the southernmost outcrop of the western margin (13.29° N, 40.51° E). This outcrop has been briefly described by Brinckmann & Kürsten (1971) but its age remains unclear. Furthermore, it presents a very large spatial extent, unusually important erosional features, and is situated at a surprisingly high elevation (chapter III). All these outcrops would represent different environmental conditions compared to the western and eastern margins and the sedimentological and paleontological study of these sites would surely bring new information on paleoecology and adaptation of the marine fauna to these special conditions.

The samples of Core A (chapter IV) should be further investigated to better constrain the Late Pleistocene and Holocene history of the basin which will help in the better understanding of the formation of salt giants. In particular, small scale annual cycles or yet unrecognized longer cyclicity should be looked at. New techniques could be used like fluid inclusions, as well as isotopic techniques like strontium isotopes (e.g. Denison *et al.*, 1998; Schreiber & Tabakh, 2000; Babel & Schreiber, 2014; Blamey & Brand, 2019), boron isotopes (Swihart *et*

al., 1986; Vengosh *et al.*, 1992; Xiao *et al.*, 2013), chlorine isotopes (Eggenkamp *et al.*, 1995; Eastoe *et al.*, 2007) and magnesium isotopes (Xia *et al.*, 2020). Dating evaporites is challenging and should be further explored to constraint the stratigraphy and rate of deposition. Datings of halite deposits have been attempted with beryllium, chlorine and uranium/thorium isotopes (Peng *et al.*, 1978; Phillips *et al.*, 1983, 1993; Jannik *et al.*, 1991; Luo & Ku, 1991; Belmaker *et al.*, 2013; Han *et al.*, 2014) and could be considered for dating the Danakil evaporites.

A significant leap in the understanding of the flooding history of the basin would definitely come with a deeper drilling of the Depression. While the last open-marine stage and the following desiccation are known from two wells (chapter IV) and outcrops on the margin of the Depression (Jaramillo-Vogel *et al.*, 2019; Endeshaw *et al.*, in prep.; Foubert *et al.*, in prep.; Negga *et al.*, in prep.), the effect of floodings stage in the central part of the basin remains to be unveiled. Was it a fast, catastrophic event or rather gradual? What are the type of sediments deposited and/or eroded during this flooding phase? Was the basin rapidly reaching open-marine conditions or did the basin first experience a long phase of basin restriction? What were the paleoclimatic conditions? Deep drilling will unveil a long and continuous paleoenvironmental, paleoclimatic and paleotectonic history that cannot be tackled by the study of the marginal outcrops. It should be mentioned that such a drilling proposal has been prepared. Seismic data and additional core and downhole logging datasets are currently studied at the University of Fribourg to constrain

further the flooding history of the depression.

Most scientific challenges mentioned above will be further tackled in the follow-up project of SERENA, the SNF-funded project CONNECT ‘Paleo-connectivity and Paleo-hydrological change in an initial rift basin’, and in the ICDP (International Continental Drilling Programme) Drilling Proposal ADD-ON. (ADD-ON: Afar Dallol Drilling: Onset of sedimentary processes in an active rift basin).

VI.3.4 Evaporites and salt giants

Chapter IV showed that the understanding of the formation of large evaporite deposits is poorly known and still controversial. This is particularly caused by the scarcity of modern analogues. As such, the few that are known, such as the Danakil Depression and the Sabkah el Mela (Perthuisot, 1971; Busson & Perthuisot, 1977) should be studied in much more detail, focussing in particular on the chronology and rates of processes which are difficult to study in ancient evaporite deposits. These basins should then be compared with older salt giants.

The study of evaporites is also in urgent need of new field or lab experiments. The studies of for example (Usiglio, 1849a; b) are outdated and should be repeated with modern analytical techniques, taking a quantitative approach to constrain better factors controlling evaporite deposition.

VI.3.5 Early geological history of the Danakil basin

While the SERENA project conducted detailed studies on the Recent flooding

history of the basin, much less is known from its earlier geological history. The Red Bed Series (or Danakil fm) has only been studied by Brinckmann & Kürsten (1971) and Le Gall *et al.* (2018). Further exploration, facies descriptions, datings and correlation at basin scale based on outcrops but also subsurface data could unveil a much longer environmental, climatic and tectonic history.

A first step towards the better understanding of the early basin history could come from the dating of lava samples collected during the SERENA 2020 expedition in boulder-sized conglomerate forming the syn-rift deposits above the pre-rift sediments in the northern part of the Danakil Depression. These conglomerates show growth-fold structures and are thus probably indicative of early basin formation. In chapter II, it has been postulated that rift initiation was younger in the northern part of the Danakil Depression. The dating of these samples could give quantitative answers to this hypothesis.

A second step could target prominent features, e.g. Mt Barali (13.46°N, 40.32°E) or the ridges at 13.62°N, 40.22°E and 13.21°N, 40.32°E. that can be used as correlative markers across the basin by subsurface seismic datasets. If these features can be dated, the understanding of its chronology could be significantly improved.

VI.3.6 Tectonic structure of the Danakil Depression

The present study shows that the tectonic structure of the Danakil Depression has a strong impact on its sedimentary architecture. This structure is furthermore a precious example of the

evolution of magma-rich margins. However, the knowledge of the architecture and evolution of this structure, as well as the link between tectonics and fluid circulation/hydrothermal systems remains poor.

Basic fieldwork, mapping and the detailed geological cross-sections could help better understanding the structure of the marginal parts of the Depression. Another step would be the geological mapping of the margins in order to unveil its spatial evolution. As suggested in section II, the spatial evolution of the rift from N to S also probably represents the temporal evolution of a magma-rich margin. Chapter II and appendices A and B show that this region is particularly well suited for remote sensing but the resolution of satellite data are too limited for such a detailed mapping. A relatively simple aerial survey with a small plane covering the whole western margin with vertical and oblique pictures, completed by targeted ground-truthing would unveil great details of its evolution.

For the central part of the Depression, the tectonic structure can probably only be better understood with geophysical methods. A first step is the study of existing seismic datasets currently ongoing at the University of Fribourg. A more complex and cost intensive step would be the acquisition of new seismic lines crossing the basin and reaching the eastern margin. The latter remains largely unknown. Another approach could be the acquisition of airborne gravimetric (and potentially magnetic) data that could unveil underground structures.

An important achievement for the understanding of the crustal structure of

the Danakil Depression would furthermore be the deployment of a dense network of seismic stations in the Danakil region, surrounding plateaus, Red Sea coast and islands. This *relatively* cheap approach would allow the better understanding of earthquakes in the region and unveil major active crustal faults, stress regime, magma emplacement and pathways. It would also provide a tool for the prediction of volcanic eruptions and for mapping seismic hazards. Moreover, seismic tomography could reveal the crustal structure, moho depth, presence of melt in the crust, mantle structure and flow. Possibly, the combination of this network with more distant stations may unveil a potential link between the Danakil Depression and the Afar hotspot and/or the African Superplume.

Numerical modelling could be performed to study the impact of isostatic processes caused by the rapid emplacement of volcanic edifices and the deposition of large volumes of sediment.

Finally, the Danakil Depression should be compared to other rifts and passive margins around the world where more data are available.

VI.3.7 Tectonic structure and basin history of the Afar Depression

Chapter II highlights many unknowns in the understanding of the Afar Depression. First of all, the new geological map of the Afar depression (appendix A) could be improved by further field studies and datings in the regions indicated on the uncertainty map, i.e. in particular the western margin of the Afar, the Awra Plain and the Assab region. One of the major unsolved

questions concerns the movement of the Ali Sabieh block and in particular the existence, position and motion of the Bia Anot fault (see section II.5.1.2). Fieldwork, remote sensing, mapping and paleomagnetic studies would greatly improve the understanding of this crucial part of the Afar. The south-eastern Afar is also one of the regions of the Afar where neither the structure, nor the geological history is well understood and could be improved by field and geophysical studies.

The links between the Afar Rift and the Red Sea Rift, both in the north and in the south, is one of the most puzzling questions of the structure of the Afar region. In the north, the structure and strain partitioning between Mt Ghedem, the Gulf of Zula and the Howlakil Bay, all showing important seismic activity (and a very low Bouguer anomaly for the Howlakil Bay), remain unclear (chapter II). In the south, the localisation of the plate boundary and the mechanisms of strain accommodation are unclear. In both areas, seismic surveys and tomography, such as discussed for the Danakil Depression, as well as InSAR may provide the most interesting results. Field surveys and potentially numerical and analogue modelling would furthermore improve the understanding of these enigmatic links.

The question of strain accommodation is also critical in Central Afar, which is one of the few places on Earth, with Iceland, where transient rifting processes can be observed. As such, GNSS and InSAR surveys, as well as analogue and numerical modelling efforts that are already ongoing should continue (Kogan *et al.*, 2012; Pagli *et al.*, 2014; Doubre *et al.*, 2017; Moore *et al.*, 2021; Maestrelli *et al.*, 2022; Sangha *et al.*, 2022).

VI.3.8 Rifting history and passive margin architecture of the Red Sea and Gulf of Aden

At a larger scale, the understanding of rifting history and structure of the Red Sea and Gulf of Aden is also fragmented. The age, let alone the limits of oceanic crust in the Red Sea, is not even known or commonly accepted (see e.g. Izzeldin, 1987; Almalki *et al.*, 2015; Augustin *et al.*, 2021; and section II.4.3.1). The sedimentary history of the Red Sea basin is also poorly known (section II. 4.3.2). Improved access to industrial data (Savoyat *et al.*, 1989; Hughes & Beydoun, 1992; Mitchell *et al.*, 1992) or the acquisition of new seismic data and deep scientific drilling could greatly improve our understanding of the structure of the Red Sea, the paleoenvironmental and paleoclimatic conditions, the history of evaporite deposition, the cause of the isolation and the origin of the first marine flooding of the Red Sea.

The knowledge of the age and structure of the northern Somalian and Yemenite coasts could furthermore be greatly improved by field and geophysical surveys. Finally, the pre-Oligocene history of the whole Afar region should be further investigated. Section II.5.3 explored the possibility of structural inheritance having an influence on rift localisation and segmentation. These structures should be better studied and comparison with other examples around the world, as well as numerical and analogue modelling should be able to better elucidate their impact on rifting processes.

VI.3.9 Plate tectonics and rifting processes

Potential future studies on plate tectonics and rifting processes are infinite but a few that are linked to the present study are listed here. One particularly interesting aspect is the link between tectonic, magmatic and surface processes in shaping rifts and passive margins. Section II.7.3 hinted at some links between the presence of hotspot-related magmatism, rift jumps and delayed breakup. Comparison with other examples around the Globe would help to identify more clearly these links and numerical modelling would help better understand the reasons for these effects. Another aspect is the understanding of the driving forces in such complex rift systems. Section II.2 showed that this is not yet completely understood. Finally, a very interesting question is how the architecture of these rifts and passive margins influences future subduction initiation, subduction processes and orogens.

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APPENDIX A: GEOLOGICAL MAP OF THE AFAR DEPRESSION

Published as:

Rime, V., Foubert, A., Atnafu, B. and Kidane, T. (2022) Geological map of the Afar Depression. Zenodo. <https://doi.org/10.5281/zenodo.7351643>

CONTENT

This dataset contains a geological map of the Afar Depression (1:1'000'000), including all the associated data.

This dataset includes:

- The map in JPEG, PDF, and GeoTIFF format
- The shapefiles of the map
- One document listing all sources used for the compilation of this map (*Source_Material_Afar_Depression.pdf*, presented on table II.1, chapter II of the present volume)
- One document describing the uncertainty of the map (*Reliability_geol_map.pdf*, presented hereafter)
- One dating database compiling published datings from the literature in .txt and shapefile formats. This database has been reviewed and mistakes in coordinates were corrected (see comments column). The precision of the position is also evaluated (1 = poor - possibly up to more than 10km uncertainty; 2 = average - no precise GPS measurement of

the coordinates but good localisation on map or description of the outcrop; 3 = very good - most of the time GPS measurement or very precise descriptions/map).

This database is an additional complement to the paper 'Rime, V., Foubert, A., Ruch, J. & Kidane, T. (2023) Tectonostratigraphic evolution and significance of the Afar Depression, *Earth-Science Reviews*, 244, 104519, <https://doi.org/10.1016/j.earscirev.2023.104519>'.

REFERENCES

The map was developed by compiling a large number of published maps, descriptions, datings and other studies, complemented by fieldwork and remote sensing. All sources and references are mentioned in the *Source_Material_Afar_Depression* (presented on table 1, chapter II of the present volume). Material and methods of mapping have been described in detail within the paper.

CITATION

When using the data, please cite the data as 'Rime, V., Foubert, A., Atnafu, B.



& Kidane, T. (2022) Geological map of the Afar Depression. Zenodo, <https://doi.org/10.5281/zenodo.7351643> and refer to the accompanying paper as 'Rime, V., Foubert, A., Ruch, J. & Kidane, T. (2023) Tectonostratigraphic evolution and significance of the Afar Depression, *Earth-Science Reviews*, 244, 104519, <https://doi.org/10.1016/j.earscirev.2023.104519>'

The map and additional data are given without any guarantee of correctness. Any use of these are under the user's full responsibility. The authors decline any responsibility.

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DISCUSSION ON THE UNCERTAINTY OF THE GEOLOGICAL MAP OF THE AFAR DEPRESSION

The map and the following descriptions are classified in four uncertainty categories:

- **Low uncertainty** (dark green): mapping is based on qualitatively and quantitatively good scientific data.
- **Medium uncertainty** (light green): mapping is based on lower amount of data but good extrapolation or remote sensing possibilities.
- **High uncertainty** (orange): small amount of data available but acceptable remote sensing possibilities or good amount of data but limited remote sensing possibilities, making the mapping more interpretative.
- **Very high uncertainty** (red): absent / very sparse / contradicting / low reliability data with limited remote sensing capabilities, making the mapping fairly interpretative.

Note that this document does not list all sources used to produce the map. The full list is available in the document 'Source_Material_Afar_Depression' and in the accompanying paper.

Uncertainty description

01: Eritrean Margin - Medium uncertainty

Several studies (Drury *et al.*, 1994; Ghebreab, 1998, 1999; Sagri *et al.*, 1998;

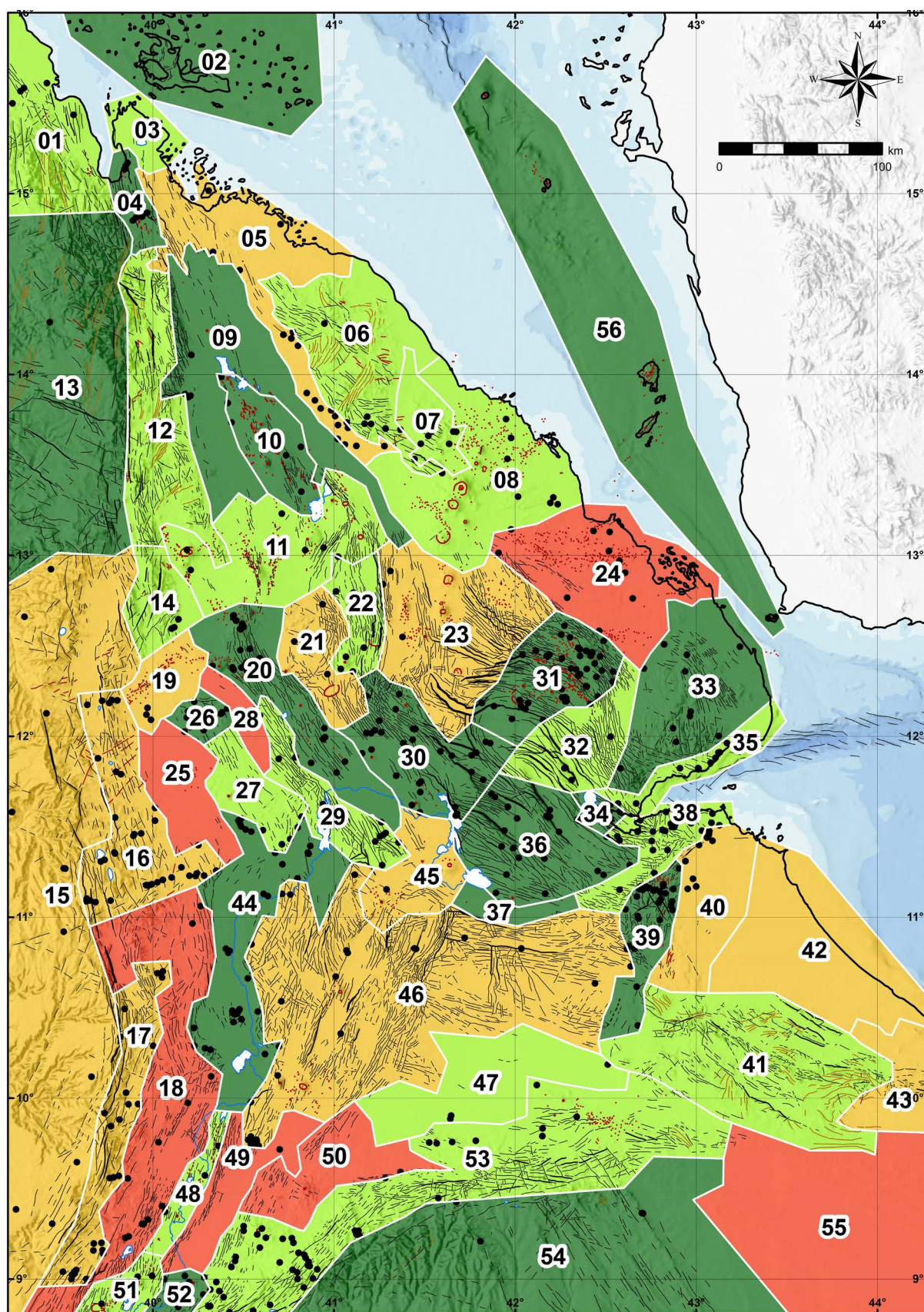


Figure 2: Uncertainty map of the Geological map of the Afar Depression. Each number corresponds to zones discussed below. Black dots indicate datings from the database. Colours according to the description above.

Ghebreab & Talbot, 2000; Teklay *et al.*, 2005; Shoshani *et al.*, 2006; Antonielli *et al.*, 2009) and good extrapolation possibilities by remote sensing result in a good reliability except for the relatively recent lava flows W of Mt. Ghedem which are of unknown age. Overall, the uncertainty is rated as medium.

02: Dahlak Islands – Low uncertainty

Sedimentological and structural studies (Angelucci *et al.*, 1975; Conforto *et al.*, 1976; Carbone *et al.*, 1998) and very good remote sensing capabilities result a low uncertainty.

03: Buri Peninsula - Medium uncertainty

Geological maps (Frazier, 1970; Ghebreab, 1999) and a detailed description (Buffler *et al.*, 2010) result a medium uncertainty.

04: Buia Graben – Low uncertainty

Detailed studies on the Alid volcano (Duffield *et al.*, 1997; Lowenstern *et al.*, 2006), the Dandiero basin (Abbate *et al.*, 2004; Sani *et al.*, 2017) and the Abdur Site (Walter *et al.*, 2000; Bruggemann *et al.*, 2004; Buffler *et al.*, 2010) allow a low uncertainty.

05: North-Eastern Danakil Margin – high uncertainty

Only few datings are available for the magmatic units in this area. The ones of Frazier (1970) are given without any detail or uncertainty and the ones of Brinckmann and Kürsten (1971) have huge uncertainties. Park (2010) mentions one Pliocene dating without further details. Mapped ages are therefore interpretative. Frazier (1970) gives good indication on the nature of the pre-rift outcrops along the coast. The Pleistocene

marine sediments found along the Red Sea coast are described by Bunter *et al.* (1998) and Holwerda and Hutchinson (1968), however the precise extent of these outcrops are difficult to identify. The overall uncertainty is rated as high.

06: Danakil Alps – medium uncertainty

Although no map covers the whole Danakil Alps, the maps and descriptions of Brinckmann and Kürsten (1970), Bunter *et al.* (1998), Sagri *et al.* (1998) combined with good remote sensing conditions provides a powerful tool to extrapolate the data presented in the literature. The overall uncertainty is rated as medium.

07: Dioita Massif – Medium uncertainty

The ages of the units are well described (mainly in Civetta *et al.*, 1975a, but also by (Bannert *et al.*, 1970; Brinckmann & Kürsten, 1971; De Fino *et al.*, 1978), allowing a medium uncertainty. However, the exact extent of the Dioita Granite is not perfectly clear.

08: Nabro Range – Medium uncertainty

The Nabro range has been relatively well studied, considering its remote location. Wiart and Oppenheimer (2005) and Wiart *et al.* (2000) established a robust relative chronology but no absolute datings are available, except for the historical flows mapped by Wiart and Oppenheimer (2000). The mapping of the Bidu Volcanic Complex therefore follows the age interpretation of Varet *et al.* (1975). The older magmatic products found in the SE part of this area (08 on fig. 2) are well constrained by datings, maps and descriptions of Brinckmann

and Kürsten (1971), and Civetta *et al.* (1975b, 1975a). Only the oldest rocks outcropping in the centre of the area are of unknown age. We interpret them as being of the same age as the ones forming the Dioita Massif. The overall uncertainty is rated as medium.

09: Danakil Depression - Low uncertainty

The Danakil depression was mapped based on several years of fieldwork exploration by the authors, the detailed maps of Brinckmann and Kürsten (1970) and several descriptions available in the literature (Holwerda & Hutchinson, 1968; Bannert *et al.*, 1970; Barberi *et al.*, 1970; Brinckmann & Kürsten, 1971; Le Gall *et al.*, 2018; López-García *et al.*, 2020; Mège *et al.*, 2021). While further work in this area would definitely be beneficial for the detailed understanding of the basin evolution, the map can be considered to have a low uncertainty at this scale.

10: Erta Ale Range –Low uncertainty

The Erta Ale Range is one of the most studied region of the Afar depression (e.g. Barberi and Varet, 1972; Watts *et al.*, 2020). While very few absolute datings have been done on the lava deposits of this range, its age younger than 0.6 Ma is widely accepted, giving a low uncertainty.

11: Southern Danakil Depression – Medium uncertainty

While datings are quite rare (Barberi *et al.*, 1972, 1975), remote sensing confirms the young age of the lavas and allows good extrapolation, giving a medium uncertainty.

12: Western Margin of the Danakil Depression – Medium uncertainty

This region is well covered by maps (Brinckmann & Kürsten, 1970; Geological Survey of Ethiopia, 1971, 1978a) and geological studies (Hagos *et al.*, 2016, 2020; Sembroni *et al.*, 2017). These sources are sometimes contradictory. Direct field observations by the authors and good remote sensing nevertheless allow a medium uncertainty.

13: Northwestern Plateaux and Mekele Outlier –Low uncertainty

This area has been studied since a long time (e.g. Beyth, 1972; Geological Survey of Ethiopia, 1971) and Recent research (Tesfamichael *et al.*, 2010; Hagos *et al.*, 2017, 2020; Sembroni *et al.*, 2017) complement these works resulting in a low uncertainty.

14: Limmo Massif – Medium uncertainty

The Limmo intrusive is well described (Black *et al.*, 1972). Mapping of the rest of this area is based on good remote sensing conditions that allows an overall medium uncertainty.

15: Western Plateau – High uncertainty

The age of the Plateau basalt is not well constrained. Several authors worked on these units and although sparsely distributed, a reasonable amount of absolute datings and maps are available (Geological Survey of Ethiopia, 1978b; Mohr & Zanettin, 1988; Drury *et al.*, 1994; Hofmann *et al.*, 1997; Chernet *et al.*, 1998; Ukstins *et al.*, 2002; Coulié *et al.*, 2003; Kieffer *et al.*, 2004; Wolfenden *et al.*, 2004, 2005; Geological Survey of

Ethiopia, 2009b; a, 2010a, 2011; Stab *et al.*, 2016; Feyissa *et al.*, 2017; Ayalew *et al.*, 2021). Several datings are provided without exact coordinates or precise maps. The major concern is that this region was mainly mapped based on petrological criteria not necessarily representing a good chronostratigraphy, as noted by (Mohr & Zanettin, 1988; Rooney, 2017). Different sources provide different interpretations in terms of litho- and chronostratigraphy. The lack of tectonic overprint allow a good extrapolation of the results of published studies. Mapping is therefore mainly based on absolute datings, extrapolated based on morphological criteria (e.g. some large flows can be traced in canyons based on DEM data), rather than on lithostratigraphy. The overall uncertainty is rated as high.

16: Central Western Margin – High uncertainty

Although many studies (Kunz *et al.*, 1975; Coulié, 2001; Ukstins *et al.*, 2002; Coulié *et al.*, 2003; Wolfenden *et al.*, 2005; Rooney *et al.*, 2013; Stab *et al.*, 2016; Ayalew *et al.*, 2019) along the Dessie-Chifra and Dessie- Mille roads and further north resulted into numerous absolute datings, it is very difficult to map the extent of each unit, partly due to the complex tectonic overprint and the presence of vegetation complicating the use of remote sensing data. Overall, the uncertainty is rate as high.

17: Upper south-western Margin – High uncertainty

While absolute datings and detailed tectonic analysis have been published on this margin (Chernet *et al.*, 1998; Ukstins *et al.*, 2002; Wolfenden *et al.*, 2004, 2005), the tectonic complexity and the difficulty

in using remote sensing data due to the vegetation cover renders the mapping difficult and fairly interpretative, resulting in a high uncertainty.

18: Lower south-western Margin – Very high uncertainty

This area lacks sufficient datings and descriptions. The data (WoldeGabriel *et al.*, 1992b; Chernet *et al.*, 1998; WoldeGabriel *et al.*, 2013; Feyissa *et al.*, 2017) are sparse, not very precise and sometimes poorly georeferenced. The use of remote sensing data is feasible, albeit fairly interpretative, resulting in very high uncertainty. On the southernmost part, the maps, Ar/Ar datings of Wolfenden *et al.* (2004) and K/Ar dating of Feyissa *et al.* (2017) are in contradiction with the map of Abebe *et al.* (2005) . Last are indicating the presence of Pleistocene "Chefe Donsa pyroclastics". We followed the interpretation of former authors as they include absolute datings but very high uncertainty remains.

19: Dabbayra Volcanic Field and N Awra plain – High uncertainty

No studies were published for the Dabbayra Volcanic Field. However, the very recent nature of some of the lavas are evident from remote sensing data. Other outcrops were mapped by extrapolating data from the northern Sullu Adu area (i.e. for Pliocene rocks). In the Northern Awra Plain, some volcanic edifices were dated (Coulié, 2001; Stab *et al.*, 2016) but the sediments surrounding them were never studied in detail. The overall uncertainty is rated as high.

20: Dabbahu – Tendaho – Low uncertainty

The numerous works and datings published for this range (Barberi *et al.*,

1972; Lahitte *et al.*, 2003; Ferguson *et al.*, 2013; Field *et al.*, 2013; Medynski *et al.*, 2015), along with excellent remote sensing possibilities results in a low uncertainty.

21: Afara Dara and Eastern Manda Hararo – High uncertainty

Mapping of the remote Afara Dara is based on one dating by Barberi *et al.* (1972) and the descriptions of Varet (2018). The presence of only a few datings in the surrounding Quaternary lavas leaves room for interpretation, resulting in a high uncertainty.

22: Dadar Graben – Medium uncertainty

Numerous datings in the area (Barberi *et al.*, 1972; Lahitte *et al.*, 2003) allows a good constrain on the age range but the variety in lava types and the intense faulting complicate the interpolation of the stratigraphic units. Field observations by the authors complement the published data. The overall uncertainty is rated as medium.

23: Northern central Afar – High uncertainty

This area has barely been explored. No absolute dating is available. Good remote sensing conditions allow the extrapolation from the adjacent regions and refinement of the map of Varet *et al.* (1975). The overall uncertainty is rated as high.

24: Assab Region – Very high uncertainty

This region is very difficult to map as it has probably never been explored further inland. Barberi *et al.* (1975) published only a few questionable and poorly-georeferenced datings. Some

datings were published by (Civetta *et al.*, 1975b) but the mapping remains largely interpretative. The Recent unit of the Assab range is mapped much smaller than Barberi *et al.* (1971) and Varet *et al.* (1975) to conform the datings of (Civetta *et al.*, 1975b) and remote sensing indicating older than recent lavas. The overall uncertainty is rated as very high.

25: Awra Plain – Very high uncertainty

No sedimentary units have ever been mapped, neither studied in this region. The mapping is an extrapolation of the units mapped in the (far) Mille river basin, last being reasonably well constrained. The mapping in this zone should be therefore be considered more relative than absolute and the the uncertainty is considered as very high.

26: Northern Sullu Adu – Low uncertainty

Thanks to the extensive work of (Stab *et al.*, 2016), this key area is very well constrained in terms of litho- and chronostratigraphy, resulting in a low uncertainty.

27: Southern Sullu Adu – Medium uncertainty

While this horst has not been studied in detail, interpolations and extrapolations from stratigraphic units mapped in adjacent regions (Kunz *et al.*, 1975; Coulié, 2001; Kidane *et al.*, 2003; Deino *et al.*, 2010; Stab *et al.*, 2016; Alene *et al.*, 2017; Saylor *et al.*, 2019; Alemseged *et al.*, 2020) results in a medium uncertainty.

28: East of Sullu Adu – Very high uncertainty

No data exist for this area. Mapping is based on the interpretation of remote sensing data and the ages of surrounding magmatic units, resulting in a very high uncertainty.

29: West Tendaho – Medium uncertainty

Some datings (Barberi *et al.*, 1975; Kidane *et al.*, 2003; Lahitte *et al.*, 2003) and good remote sensing conditions result in a medium uncertainty.

30: Eastern Tendaho – Low uncertainty

Dense dating network (Kidane *et al.*, 2003; Lahitte *et al.*, 2003 among others) and good remote sensing possibilities result in a low uncertainty.

31: Siyyarou to Moussa Ali – Low uncertainty

Dense dating network (Lahitte *et al.*, 2003; Manighetti *et al.*, 1998 among others) and good remote sensing possibilities result in a low uncertainty.

32: Northern Djibouti – Medium uncertainty

While data are sparse, good remote sensing possibilities and the map of Le Gall *et al.* (2015) result in a medium uncertainty.

33: North-eastern Djibouti – Low uncertainty

This area has been studied and mapped in detail (Barberi *et al.*, 1975; Le Gall *et al.*, 2015; Zumbo *et al.*, 1995 among others). Combined with very good remote sensing possibilities, the large

timespan for each mapped geological unit result in low uncertainty.

34: Asal – Low uncertainty

Many studies, maps and datings have been published for this area (Chessex *et al.*, 1975; Courtillot *et al.*, 1984; Zumbo *et al.*, 1995; Manighetti *et al.*, 1998; Daoud *et al.*, 2011; Le Gall *et al.*, 2015), resulting in a low uncertainty.

35: Tadjoura – Medium uncertainty

Several datings are available (Manighetti *et al.*, 1998; Daoud *et al.*, 2011), along with the map of Le Gall *et al.* (2015) but the spread in ages and the tectonic overprint complicates the interpretation in terms of the magmatic stratigraphic units. The overall uncertainty is rated as medium.

36: Western Djibouti – Low uncertainty

The large amount of datings (Barberi *et al.*, 1972, 1975; Courtillot *et al.*, 1984; Zumbo *et al.*, 1995), the mapping of Le Gall *et al.* (2015) and the low spread of the ages results in a low uncertainty.

37: Goba'Ad – low uncertainty

The mapping is based on the detailed map of Gasse *et al.* (1987), resulting in a low uncertainty.

38: Djibouti – Medium uncertainty

Many studies have been conducted in the area (e.g. Audin *et al.*, 2004; Barberi *et al.*, 1975; Chessex *et al.*, 1975; Courtillot *et al.*, 1984; Daoud *et al.*, 2011; Zumbo *et al.*, 1995) but its tectonic complexity and the juxtaposition of different units complicates the interpretation. The overall uncertainty is rated as medium.

39: Ali Sabieh Block – Low uncertainty

Numerous studies conducted on the Ali Sabieh Block, including datings and mapping (e.g. Barberi *et al.*, 1975; Black *et al.*, 1975; Chessex *et al.*, 1975; Gasse *et al.*, 1986; Geological Survey of Ethiopia, 2015; Le Gall *et al.*, 2015, 2010; Varet *et al.*, 1975; Zumbo *et al.*, 1995) and the very good remote sensing possibilities result in low uncertainty, even for the Ethiopian southern part. There is - however - one inconsistency between the map of Geological Survey of Ethiopia (2015) and other authors (Gasse *et al.*, 1986; Le Gall *et al.*, 2010, 2015) concerning the interpretation of the Mesozoic sandstones. The Geological Survey of Ethiopia (2015) places them stratigraphically below the limestones while the other sources state the opposite. Considering the detailed observations and descriptions of (Gasse *et al.*, 1986; Le Gall *et al.*, 2010), we place the sandstones above the limestones, and thus assign them a Cretaceous age.

40: Eastern Ali Sabieh – High uncertainty

While this area provides good remote sensing possibilities, the discrepancy between different authors (Barberi *et al.*, 1975; Zumbo *et al.*, 1995; Audin *et al.*, 2004; Daoud *et al.*, 2011) regarding the nomenclature (e.g. confusion between Somali and Dalha basalts) and datings, as well as the scarcity of the datings complicates the interpretation of the age of the different units. The overall uncertainty is rated as high.

41: Somalian margin – Medium uncertainty

Four geological maps (Geological Survey Somali Democratic Republic,

1957, 1970, 1971a; b) covering part of this area were used as input data for the mapping. However it is not known whether they are based on field survey or remote sensing data. Ali (2015, 2009) published more recent maps but at smaller scale. Despite the small amount of studies, good remote sensing possibilities allow to interpolate and extrapolate these maps with a medium uncertainty.

42: Western Guban Basin – High uncertainty

No study exists on the Quaternary sediments of the western Guban basin. Mapping only based on remote sensing and relationship with adjacent Miocene lavas. Without serious calibration through field data and without a good age control, the interpretation is fairly interpretative. The overall uncertainty is rated as high.

43: Mt Elmis – High uncertainty

The age of the volcanics around Mount Elmis is often cited as being "late Miocene" (Ali, 2009, 2015; Ali *et al.*, 2018) but no datings are available. This age is probably inferred from the analysis of the stratigraphy in seismic data and wells in the Guban basin (Ali, 2015). The overall uncertainty is rated as high.

44: Middle and lower Awash Valley – Low uncertainty

Numerous studies have been conducted in the middle and lower Awash Valley as well as in the Mille basin due to its anthropological significance (Taieb *et al.*, 1976; Aronson *et al.*, 1977; Tiercelin, 1986; White *et al.*, 1993; Clark *et al.*, 1994; Kalb, 1995; Renne *et al.*, 1999; WoldeGabriel *et al.*, 2001; Levin *et al.*, 2004; Quade *et al.*, 2004; Semaw *et al.*,

2005; Wynn *et al.*, 2008; DiMaggio *et al.*, 2009, 2015a; b; Deino *et al.*, 2010; Campisano, 2012; Haile-Selassie *et al.*, 2015; Saylor *et al.*, 2019; Braun *et al.*, 2019; Garelo, 2019; Alemseged *et al.*, 2020). The smaller-scale map by Kalb *et al.* (1982) allows to interpolate lithological units between the different studies. This allow a low uncertainty although interpretation based on remote sensing data is difficult.

45: Gabilema – High uncertainty

While no datings are available in this area, good remote sensing data, extrapolation of stratigraphic units from the neighbouring areas and the map of Varet *et al.* (1975) result in a high uncertainty.

46: Southern Afar – High uncertainty

This part of the Afar is not well studied. Most of the datings are from Barberi *et al.* (1975) but have high uncertainty and unprecise positions. Only a few additional datings are available (Black *et al.*, 1975; Kunz *et al.*, 1975; Lahitte *et al.*, 2003; Hujer *et al.*, 2015). However, published maps (Varet *et al.*, 1975; Geological Survey of Ethiopia, 2014, 2015) are valuable. The relative consistency across the datings allow to postulate the dominance of Pliocene lavas, even though early Quaternary lava might be present in the upper part of the succession. Towards the eastern part of the area, some lavas characterized by a more recent texture based on remote sensing data were mapped as Q_{2v}, in accordance with Varet *et al.* (1975). The overall uncertainty is rated as high.

47: Buren Plain – Medium uncertainty

Although no study describes the sediments of the Buren Plain, remote sensing gives solid indications (absence of erosion, traces of active sedimentation) on the Recent age of the sediments. The uncertainty is rated as medium.

48: NMER – Medium uncertainty

A few but well localized datings (Chernet *et al.*, 1998; Feyissa *et al.*, 2017) and good remote sensing data allow a medium uncertainty.

49: Eastern NMER – Very high uncertainty

No datings or detailed studies exist for the lavas and sediments outcropping in this region. Mapping is therefore very interpretative and as such, the uncertainty is rated as very high.

50: Afdem – Very high uncertainty

Few studies exist for this area. It was mapped based on the Dire Dawa geological map (Geological Survey of Ethiopia, 1985) and extrapolation from the neighbouring areas with remote sensing, resulting in a very high uncertainty.

51: Kone – Fentale – Medium uncertainty

The presence of a few absolute datings combined with existing maps (Williams *et al.*, 2004; Vogel *et al.*, 2006) and the extrapolation of mapping further south (Morbideilli *et al.*, 1975; Morton *et al.*, 1979; Siegburg *et al.*, 2018) resulted in a medium uncertainty.

52: Awash – Low uncertainty

Numerous datings by (Kunz *et al.*, 1975; Morbidelli *et al.*, 1975; Chernet *et al.*, 1998; Kidane *et al.*, 2010; Feyissa *et al.*, 2017) and the Nazret geological map (Geological Survey of Ethiopia, 1978b) allow a low uncertainty.

53: South Afar Margin – Medium uncertainty

Numerous datings (Black *et al.*, 1975; Kunz *et al.*, 1975; Tiercelin *et al.*, 1979; Berhe, 1986; WoldeGabriel *et al.*, 1992a; Chernet *et al.*, 1998; Audin *et al.*, 2004; Benito-Calvo *et al.*, 2014; Feyissa *et al.*, 2017) and the presence of good geological maps (Geological Survey of Ethiopia, 1978b, 1985, 2007, 2010b; Mège *et al.*, 2016) allow a good understanding of the margin. However, the lack of precision in the position of some datings and the complex tectonics still leave some room for interpretation, resulting in a medium uncertainty. Sickenberg and Schönfeld (1975) was used for the mapping of the Chorora Fm.

54: Ogaden basin – Low uncertainty

The relative simplicity of the geology and the good coverage by geological maps (Geological Survey of Ethiopia, 1978b, 1985, 2007, 2010b; Mège *et al.*, 2016) result a low uncertainty.

55: Somalian Plateau – Very high uncertainty

This region is poorly studied. The Somalian part is mainly based on Ali (2009). Questions raised on how to map large area covered by eluvium. The Harar geological map (Geological Survey of Ethiopia, 2010b) on the Ethiopian side describes a quaternary cover of eluvium and calcrete. This map also reports on the

presence of the Jesomma Formation without field evidence. Other maps (Geological Survey of Ethiopia, 1996; Ali, 2009; Mège *et al.*, 2016) focus on the underlying consolidated rock supposed to be the Jesomma Fm. To represent the fact that this area does not represent a recently active sedimentary basin, we opted to show the underlying lithology, even though their limits are not clear. The age of the Jesomma Fm. itself is debated. While Ali (2009) and Chernet *et al.* (2020) and the Harar geol map (Geological Survey of Ethiopia, 2010b) map it as Cretaceous, the Geological Map of Ethiopia (Geological Survey of Ethiopia, 1996) and Mège *et al.* (2016) indicates a Paleogene age. We mapped it as Cretaceous. Overall, the uncertainty is rated as very high.

56: Red Sea Island – Low uncertainty

The geology of the Perim Island is well described by Al-Mikhlaifi *et al.* (2018) and Mallick *et al.* (1990) and there is a general consensus about the recent age of the Hanish, Zubair and Jebel at Tair islands (Gass *et al.*, 1973; Xu *et al.*, 2015). The uncertainty is rated as low.

APPENDIX B: GEOLOGICAL MAP OF THE SOUTHERN RED SEA & WESTERN GULF OF ADEN REGION

Rime, V., Foubert, A., Atnafu, B. and Kidane, T. (2022) Geological map of the southern Red Sea & western Gulf of Aden region. Zenodo. <https://doi.org/10.5281/zenodo.7351765>

CONTENT

This dataset contains a geological map of the southern Red Sea & western Gulf of Aden region (1:3'100'000), including all the associated data.

- This dataset includes:
- The map in JPEG, PDF, and GeoTIFF format
- The shapefiles of the map
- One document listing all sources used for the compilation of this map
(*Source_Material_GmsRSwGoAr.pdf*, presented on table II.1, chapter II of the present volume)

This database is an additional complement to the paper 'Rime, V., Foubert, A., Ruch, J. & Kidane, T. (2023) Tectonostratigraphic evolution and significance of the Afar Depression, *Earth-Science Reviews*, 244, 104519, <https://doi.org/10.1016/j.earscirev.2023.104519>'.

REFERENCES

The map was developed by compiling a large number of published maps, descriptions, datings and other studies,

complemented by remote sensing. All sources and references are mentioned in the *Source_Material_GmsRSwGoAr*, (presented on table II.1, chapter II of the present volume). Material and methods of mapping have been described in detail within the paper.

CITATION

When using the data, please cite the data as 'Rime, V., Foubert, A., Atnafu, B. & Kidane, T. (2022) Geological map of the southern Red Sea & western Gulf of Aden region. Zenodo' and refer to the accompanying paper as 'Rime, V., Foubert, A., Ruch, J. & Kidane, T. (2023) Tectonostratigraphic evolution and significance of the Afar Depression, *Earth-Science Reviews*, 244, 104519, <https://doi.org/10.1016/j.earscirev.2023.104519>'.

The map and additional data are given without any guarantee of correctness. Any use of these are under the user's full responsibility. The authors decline any responsibility.

ACKNOWLEDGEMENTS

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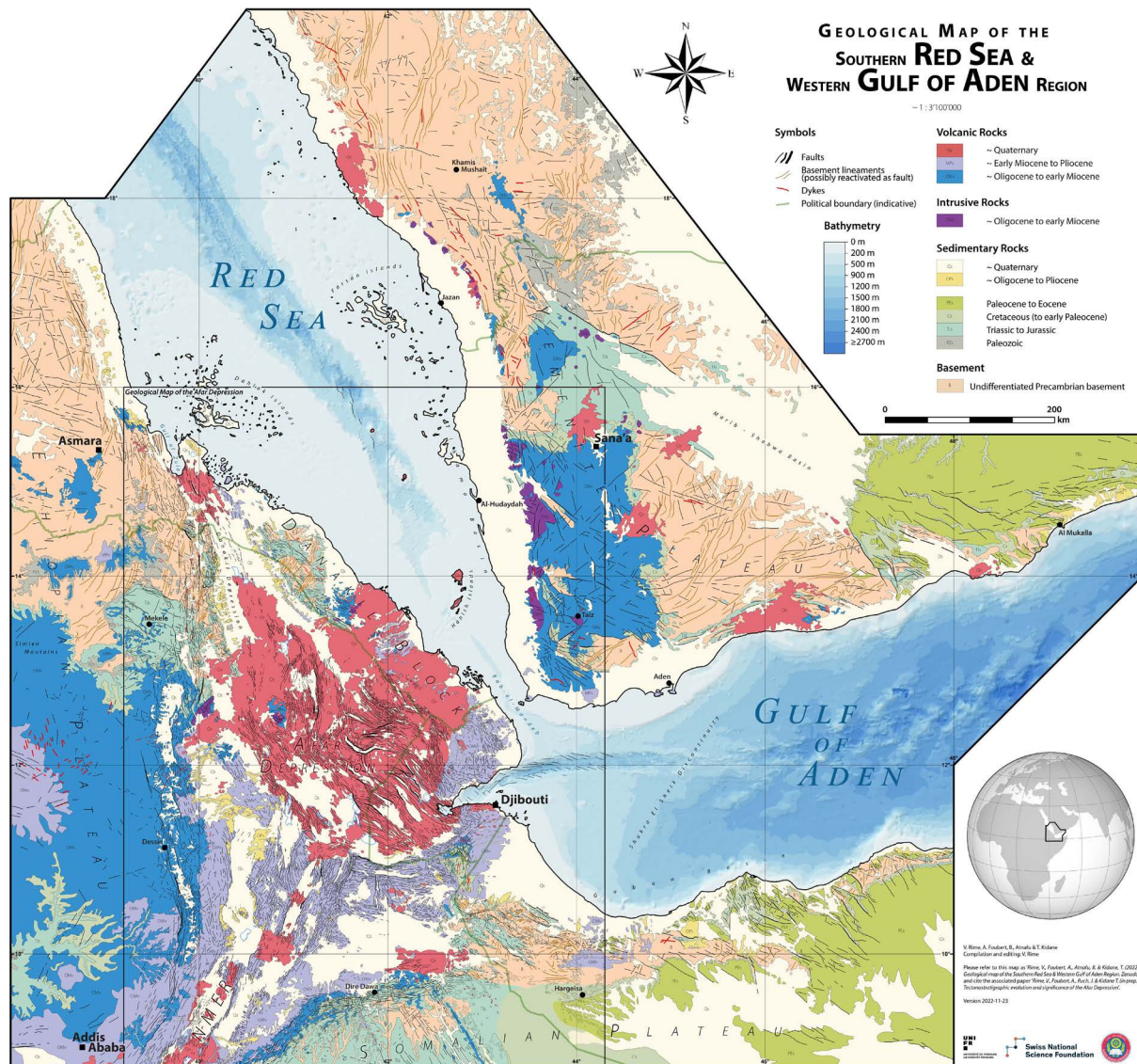


Figure 1: Geological map of the southern Red Sea & western Gulf of Aden region

of the Northern Afar 200021_163114). We are grateful to the University of Fribourg (Switzerland), the University of Addis Ababa (Ethiopia), the Ethiopian Ministry of Mines and Energy, the Ethiopian Geological Survey, Circum Minerals, former Allana Potash and Yara Dallol for their support. We particularly acknowledge Samuel Getachew that helped us to access some of the geological

maps. We thank David Jaramillo-Vogel, Jean-Charles Schaegis, Haileyesus Negga, Addis Endeshaw, Ermias Gebru, Eva de Boever, Juan-Carlos Braga, Pia Wyler, Xenia Haberditz, the Ethioder team as well as the regional and local administration of the Afar for their help and support during fieldwork.

APPENDIX C: OCEANIC ISOSTASY AS A TRIGGER FOR THE RIFT-TO-DRIFT TRANSITION: COMMENT

Published as:

Rime, V. and Foubert, A. (2022) Oceanic isostasy as a trigger for the rift-to-drift transition: COMMENT. *Geology*, **50**, e554. <https://doi.org/10.1130/G50625C.1>

Conder (2022) presents the possibility of oceanic isostatic loading being a key component for rift-to-drift transition and discusses this effect with real-world examples. While we recognize that the isostatic effect of oceanic loading might trigger asthenospheric flow and basal traction contributing to extensional forces, it is important to highlight that temporal or spatial coincidence of the inundation of a basin and the oceanic nature of the crust is not sufficient to show that the former caused the latter. This is especially applicable as the opposite causality is an established process: rifting and oceanic spreading will systematically form a depression into which seawater will eventually flow once connected to existing oceans (rifting causing flooding). Additional parameters are known to play an important role in the rift-to-drift transition such as magmatic activity and mechanical behavior of the crust (e.g., Sapin et al., 2021). These other factors are largely ignored by Conder. To justify the proposed hypothesis, the numerous examples of basins that did experience marine flooding but never reached oceanic spreading should be considered (e.g., Rockall, Porcupine, Viking,

Carpathian, Dniepr-Donets, Greater Caucasus). The Afro-Arabian Rift System is presented by Conder as an example for the temporal coincidence between flooding and rift-to-drift transition. Unfortunately, a precise review of the chronology evidences that this example does not support this coincidence. Rifting in the central Gulf of Aden (GoA) started during the early Oligocene (Fantozzi and Ali Kassim, 2002; Ali and Watts, 2016) and marine flooding occurred during midOligocene time (ca. 30–27 Ma; Hughes et al., 1991), more than 10 m.y. before ocean spreading (ca. 16 Ma; Fournier et al., 2010). In the Red Sea (RS), faulting initiated at ca. 29 Ma (Bohannon, 1986) but flooding started much earlier than 6 Ma such as mentioned by Conder. Hughes et al. (1991) showed that the first marine conditions were already established in the southern RS during the late Oligocene (ca. 27–23 Ma) and that widespread deep-water conditions prevailed since early Miocene time. At that time, even the Gulf of Suez (which never reached oceanic spreading) was flooded. The initiation of seafloor spreading in the central RS is not well constrained, but happened probably

between ca. 13 Ma (Augustin et al., 2021) and ca. 4.6 Ma (Schettino et al., 2016); i.e., long after the initial flooding. The timing of the events (rifting, flooding, and oceanic spreading) does not “differ dramatically” between the RS and the GoA, the former being only a few million years older than the latter for each of these events. This delay is coherent with the counter-clockwise rotation of the Arabian plate implying a propagation of rifting from the GoA to the RS and smaller extension within the RS (e.g., Joffe and Garfunkel, 1987). Considering the never-inundated continental East African Rift as an argument is not relevant considering that this system experienced significantly smaller extension (Corti, 2009). Concerning the northern Afar, flooding of the Danakil Depression occurred more than 80 km away from where magnetic stripes were observed. Any conclusion on causality is therefore stretched. The description of the South Atlantic history is also more complex, and rifting cannot be described as “continual” northward-directed “unzipping”. The breakup of the northern and southern parts occurred independently at 138 and 120 Ma, respectively, only connecting at 113–112 Ma in the Santos Basin (Heine et al., 2013).

The example of Iceland should also be considered with caution. The geophysical data presented (Conder, 2022, his figure 4) are, in our opinion, not conclusive. Free-Air data do show a very noticeable transition at the platform boundary, and magnetic anomaly data show that the atypical signal is not limited to Iceland but extends far offshore toward the west-northwest and east-southeast under the Greenland-Iceland-Faroe Ridge. Iceland is known to be far more complex than the

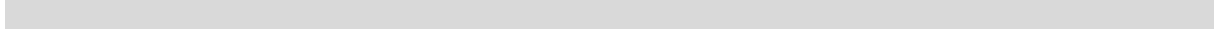
simple distinction between oceanic and continental domains (Foulger et al., 2020) and the effects of the Icelandic plume probably have a significantly stronger influence on the crustal architecture than inundation. Although Conder presents a plausible mechanism where flooding increases extensional forces during rifting, real-world examples mentioned in his paper give no evidence that the ocean plays an “integral” or “important role” in the rift-to-drift transition. Such a conclusion would need to consider other parameters like magmatic production, mechanical behavior of the crust, and the amount of extension, as well as counter-examples where flooding did not result in oceanic spreading.

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APPENDIX D: CURRICULUM VITAE



VALENTIN RIME

Postdoctoral researcher in Earth Sciences

EDUCATION & EMPLOYMENT

Postdoc FNS

04.2023-present

Department of Geosciences, University of Fribourg, Switzerland
Structural and sedimentological analysis of seismic data in the Danakil Depression

PhD in Earth Sciences

01.2018-03.2023

Department of Geosciences, University of Fribourg, Switzerland
Thesis: Multi-scale basin dynamic study of the Danakil Depression (Afar)

Master of Science in Earth Sciences

2015 – 2017

University of Fribourg, Switzerland
University of Bern, Switzerland
Option: Geology
Thesis: Tectonics of the Neuchâtel Jura Mountains

Bachelor of Science in Earth Sciences

2012 – 2015

University of Fribourg, Switzerland
Propaedeutics: physics, mathematics and informatics (12 ECTS each)
Complementary branch : physics (30 ECTS)
Thesis: Étude tectonique de la structure de Hermrigen

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Fribourg, Switzerland

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ABOUT ME

Birthdate 17.09.1991

Citizenship Swiss

Gender Male

Driving license Yes

LANGUAGES

French native speaker

English fluent (C1)

German advanced (B2)

Italian basics (A2)

REFeree

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University of Fribourg

Switzerland

RESEARCH FOCUS

Tectonics

- Rift and passive margin tectonics
- Fold-and-thrust belt tectonics
- Balancing and restoration
- Plate tectonics and geodynamics
- Tectonic geomorphology

Sedimentology

- Rift-related sediments architecture
- Evaporites and salt giants sedimentology
- Carbonate sedimentology
- Paleoenvironment and paleoclimatology

Multi-disciplinary

- Basin dynamics
- Interactions between tectonics, sedimentology, magmatism, sea-level and climate
- Paleogeographic/paleogeological reconstructions

Regional geology

- Afro-Arabian Rift System, Afar and Danakil Depression
- Jura fold-and-thrust belt
- Molasse Basin

TECHNICAL EXPERIENCE

Fieldwork

- Field exploration and mapping in remote areas (incl. 2 field expedition in northern Ethiopia)
- Field and well logging
- Drone survey

Labwork

- Thin section analysis
- X-Ray Fluorescence
- X-Ray Diffraction
- Sediment component analysis

Data analysis

- Geographic Information Systems (ArcGIS)
- Plate Tectonic reconstruction (GPlates)

- Remote sensing (satellite data)
- Seismic data analysis (Kingdom)
- Topographic analysis (TAK)
- Photogrammetry (Agisoft)
- Cross-section construction
- Forward tectonic modelling, balancing and 3D geological modelling (Move by Petroleum Experts)

Visualization

- Logging (Strater)
- Figure creation (Adobe Illustrator, Photoshop)
- Film editing (Adobe Premiere and Cyberlink PowerDirector)

TEACHING EXPERIENCE

Student supervision

Department of Geosciences, University of Fribourg, Switzerland

- Co-supervision of a BSc-thesis on Danakil evaporites (Xenia Haberditz) 2018-2019
- Co-supervision of a MSc-thesis on Danakil fluvial terraces (Bastien Bornet) 2020-2023

Assistant

2018-2022

Department of Geosciences, University of Fribourg, Switzerland

- Cartes et coupes I and II (BSc beginner and advanced cross-section construction practical)* 2018-2022
- Sédimentologie TP (BSc general sedimentology practical)* 2018-2021 (2023)
- Méthodes de sédimentologie TP (BSc 5-day sedimentology field- and lab work practical) 2018-2021
- Applied Sedimentology (MSc, 4h practical on sequence stratigraphy) 2021, 2023
- Dynamic Alps (MSc 5-days field excursion across the Alps) 2020
- Géologie Générale (BSc, 1x 2h course, occasional supervision of practical) 2019, 2022
- Histoire de la terre (BSc, occasion supervision of practical) 2022-2023
- Stage de terrain II (BSc 5-days mapping course in the Alps) 2019
- Stage de terrain I (BSc 3-days mapping course in the Jura) 2019, 2023
- Cartographie et cartomatique TP (BSc mapping and GIS practical) 2018

*Main course manager

Student Assistant

2015-2017

Department of Geosciences, University of Fribourg, Switzerland

- Cartes et coupes I and II (BSc beginner and advanced cross-section construction practical)
- Cartographie et cartomatique TP (BSc mapping and GIS practical)
- Introduction aux géosciences TP (BSc general geology practicals)
- Microscopie des magmatites (BSc magmatic rock microscopy practical)
- Excursion de géologie I and IIA (BSc 1-day field courses in Switzerland)
- Stage de terrain I (BSc 3-days mapping course in the Jura)
- Stage de terrain II (BSc 5-days mapping course in the Alps)

TRAININGS AND COURSES

FIELD COURSES

- Mesozoic-Cenozoic Dinaric foreland basins (06/2023). *IAS Conference Field Trip*
- General Alpine Geology in the Engadine (07/2022). *CUSO – ESPP doctoral school*.
- Evolution of Volcanism in the Central Antilles Arc from the Miocene to the Present Day (05/2022). *CUSO – ESPP doctoral school*.
- Sedimentological and Volcano-Tectonic Processes in the Afar, Ethiopia (01-02/2020). *CUSO – ESPP doctoral school*. Member of the organizing team
- Microbial carbonate reservoirs architecture: from depositional facies to seismic scale geometries in the Triassic of Dolomites (Italy) (09/2019). *IAS Conference Field Trip*
- The Messinian sediments of the Piedmont Basin: a record of the Messinian Salinity Crisis and of circulation of gas-rich fluids (09/2019). *IAS Conference Field Trip*
- Depositional and facies models: The bay of Mont Saint-Michel, France, (09/2019). *CUSO – ESPP doctoral school*

- Coral Reefs in the Maldives: Tropical Marine Ecology and Sedimentology (09/2018). *CUSO – ESPP doctoral school*
- Consequences of karstification, energy and water supply (05/2018). *CUSO – ESPP doctoral school*
- Bachelor and Master field courses in the Jura (CHE), Molasse basin (CHE), Alps (CHE and ITA), Aeolian Islands (ITA), Atlas (MAR), Crete (GRC) and Tenerife (ESP) (2012-2017). *University of Fribourg*

COURSES

- How to Write a Compelling Grant Proposal for Scientific Research (11/2022 – 2 days course) *CUSO doctoral school*
- “Flügel” Course on Carbonate Microfacies (04/2022 – 5 days course) *CUSO doctoral school*
- Kingdom Geological Data Loading V-ILT (07/2021 – 1-day course) *IHS Markit Training*
- Geodynamics, from plate tectonic theory to paleogeographic reconstructions (03/2021 – 4 days course) *CUSO doctoral school*
- Engage, Inform, Inspire: Bring your presentation skills to the next level (10-11/2019 – 2 days course) *CUSO doctoral school*
- Video Journalism for Researchers (06/2019 – 2 days course) *CUSO doctoral school*
- Visualising your Research: Creative Approaches and Using Images to Help Communicate your Work (04/2019 – 2 days course) *CUSO doctoral school*
- Kingdom Seismic Interpretation (10/2018 – 3 days course) *IHS Markit Training*

MEMBERSHIP

International Association of Sedimentologists
European Geosciences Union

Since 2019
Since 2020

AWARDS

CUSO ESPP days Best poster presentation award	2022
Basin Research Outstanding reviewers 2020-2021 award, https://doi.org/10.1111/bre.12657	2022
SNSF Scientific Image Competition First prize in the category "Locations and instruments", https://www.snf.ch/en/xcdCNJikRZXwYWC5/page/science-communication/image-competition	2021

ADMINISTRATIVE RESPONSIBILITIES

CUSO doctoral school

2020-2023

PhD representative for the Earth Surface Processes and Paleobiosphere section.

University of Fribourg

2020-2021

Representative of the Intermediate staff of the structural and selection commission (hiring process of a new professor).

SCIENTIFIC OUTREACH

NEWSPAPER AND MAGAZINE INTERVIEWS/CONTRIBUTIONS

- “Quand le désert raconte la mer”, *printed and online issue of Universitas 03* | 2021/2022, August 2022, <https://www.unifr.ch/universitas/fr/editions/2021-2022/die-wueste/>
- Photo essay about scientific expeditions in the Danakil Desert, *printed and online issue of Universitas 03* | 2021/2022, August 2022, <https://www.unifr.ch/universitas/fr/editions/2021-2022/die-wueste/quand-le-desert-raconte-la-mer.html>
- “Photographier la naissance d’un océan”, *online article in Alma & Georges webmagazine*, 14.06.2021, <https://www.unifr.ch/alma-georges/articles/2021/photographier-la-naissance-dun-ocean>
- “Un explorateur tout terrain”, *online article in La Liberté*, 31.07.2021, <https://www.laliberte.ch/news/regions/un-explorateur-tout-terrain-613351>
- “La naissance d’un océan en Ethiopie”, *online video clip by La Liberté*, 31.07.2021, <https://www.laliberte.ch/news/regions/la-naissance-d-un-ocean-en-ethiopie-613819>
- “Un explorateur dans le désert africain”, *full-page printed article in La Liberté*, 02.08.2021
- “Un doctorant au-dessus du volcan”, *insert in L’Illustré*, 11.08.2021

FILM AND CONFERENCES

- “Il était une fois... la mer dans le désert.” *Self-made film about our scientific expeditions in Ethiopia*
 - Carnets de Voyages film festival (<https://www.carnetsdevoyages.ch>), Marly, Switzerland (20.11.2022)
 - Fachschaft event, University of Fribourg, Switzerland (24.11.2022)
 - Meeting of the *Société de Volcanologie Genève* (09.01.2023)
 - Presentation for the *Musée Valaisan des Bisses* (11.05.2023)
- “Comment la tectonique et l’eau ont façonné les paysages de notre planète” *Conference for “Le Nouveau Cercle” in Fribourg*, 11.04.2022

PEER REVIEWS

- Heida, H., Raad, F., Garcia-Castellanos, D., Jiménez-Munt, I., Maillard, A., & Lofi, J. (2022). Flexural-isostatic reconstruction of the Western Mediterranean during the Messinian Salinity Crisis: Implications for water level and basin connectivity. *Basin Research*, 34(1), 50-80, <https://doi.org/10.1111/bre.12610>

PUBLICATIONS

PEER-REVIEWED JOURNAL ARTICLES

First and Co-first* author publications:

Submitted:

- **Rime, V.**, Negga, H., Fentimen, R., El Korh, A., Pirkenseer, C., Schaegis, J.-C., Hajdas, I., Adatte, T., Atnafu, B., Kidane, T. & Foubert, A. (submitted) Origin of Late Pleistocene to Holocene evaporites in the Danakil Depression: a modern analogue for understanding salt giants in Earth history. *Submitted to Sedimentology*.

Published:

- **Rime, V.**, Foubert, A., Ruch, J. & Kidane, T. (2023) Tectonostratigraphic evolution of the Danakil Depression. *Earth Science Reviews*, 244, 104519, <https://doi.org/10.1016/j.earscirev.2023.104519>
- **Rime, V.** & Foubert, A. (2022) Oceanic isostasy as a trigger for the rift-to-drift transition: COMMENT. *Geology*, 50 (11), e554, <https://doi.org/10.1130/G50625C.1>
- Schaegis, J. C. *, **Rime, V. ***, Kidane, T., Mosar, J., Gebru, E. F., Atnafu, B., & Foubert, A. (2021). Novel bathymetry of Lake Afdera reveals fault structures and volcano-tectonic features of an incipient transform zone (Afar, Ethiopia). *Frontiers in Earth Science*, 9, 530, <https://doi.org/10.3389/feart.2021.706643>
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- **Rime, V.**, Mosar, J., Schori, M., Sommaruga, A. (2018) New possible age constrain on the deformation of the Swiss Neuchâtel Jura Mountains. *Oral presentation held at the 26th SwissSed Meeting, Fribourg, Switzerland.*
- **Rime, V.**, Mosar, J. (2017) Tectonic structure and dynamic of the Neuchâtel Jura Mountains: new insight from detailed mapping and forward modelling techniques. *Poster presented at the 15th Swiss Geoscience Meeting, Davos, Switzerland.*

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- Negga, H., **Rime, V.**, Hailu, A., Jaramillo-Vogel, D., Braga, J.-C., Atnafu, B., Kidane, T., Foubert, A. (2019) Coupling of Pleistocene sea level fluctuations with corallgal reef growth in the Danakil Depression (Afar, Ethiopia). *Poster presented at the 17th Swiss Geoscience Meeting, Fribourg, Switzerland.*
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- Endeshaw, A., Negga, H., Jaramillo-Vogel, D., **Rime, V.**, Atnafu, B., Braga, J.C., Kidane, T., Foubert, A. (2019) Paleo-ecology of Pleistocene corallgal reef terraces in the Danakil Basin (North Afar, Ethiopia). *Poster presented at the International Symposium on Fossil Cnidaria and Porifera, Modena, Italy.*
- Fentimen, R., Feenstra, E., Rüggeberg, A., Hall, E., **Rime, V.**, Vennemann, T., Rosso, A., Adatte, T., Van Rooij, D., Vogel, H., Schröder-Ritzaz, A., Krengel, T., Spezzaferri, S., Foubert, A. (2019) The influence of climatic variability on the long-term growth history of a Mediterranean cold-water coral mound (Melilla Mound Field). *Oral presentation held at the International Symposium on Fossil Cnidaria and Porifera, Modena, Italy*
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- Negga, H., Jaramillo-Vogel, D., Schaegis, J.C., **Rime, V.**, Perrochet, L., Wyler, P., Filfilu, E., Hailu, A., Braga, J.C., Atnafu, B., Kidane, T. and Foubert, A. (2018) Coralgall reefs in the Danakil depression (Northern Afar, Ethiopia). *Poster presented at the 26th SwissSed Meeting, Fribourg, Switzerland*
- Negga H., Jaramillo-Vogel D., **Rime V.**, Schaegis J.-C., Perrochet L., Wyler P., Filfilu E., Hailu A., Braga J.C., Balemwal A., Tesfaye K., Foubert A. (2018) Palaeo-environmental evolution of the Danakil Depression during the Pleistocene (Northern Afar, Danakil Depression), *Poster presented at the 16th Swiss Geoscience Meeting.*

NON-PEER-REVIEWED JOURNAL/MAGAZINE ARTICLE

- **Rime, V.** (2023) Expéditions scientifiques dans la Dépression de Danakil, *Bulletin de la Société de volcanologie Genève*, 219, 18-20
- Kohn, J., **Rime, V.** (2019) Hertzsprung-Russel-Diagramm von Sternhaufen, *Orion*, 2, 6-11

OTHER SKILLS (NON-ACADEMIC)

Mountain rescue

Secours Alpin Suisse, Station de secours de Bulle

- Mountain rescuer since 2013
- Secretary of the rescue station 2014-2021
- Head of training of the rescue station 2021-2022
- Intervention leader since 2020
- Up to date Basic Life Support certificate

Mountaineering instructor

Swiss Alpine Club, Section Moléson

- Ski touring instructor for teenagers since 2009
- Mountaineering instructor for teenagers since 2012
- Alpine specialist of the Swiss Army since 2012
- Activity manager for the OJ Moléson (2007 – 2018)
- Climbing training manager for the OJ Moléson (2013 – 2015)

Sports and other interests

- Mountaineering, incl. rock climbing, trad climbing, ice climbing, ski touring, alpinism, hiking
- Travelling (incl. Morocco, India, Kirghizstan, Turkey, Thailand, Norway, Iceland, Canada, Iran)
- Running, trail running
- Occasional: swimming, bike riding, kayaking
- Sailing (Offshore licence in progress)
- Helicopter flying
- Concerts and music festivals

Already published:

- Vol. 1 **HILLGÄRTNER, Heiko** (1999) The evolution of the French Jura platform during the Late Berriasian to Early Valanginian: controlling factors and timing (203 pp.).
- Vol. 2 **DUPRAZ, Christophe** (1999) Paléontologie, paléocéologie et évolution des faciès récifaux de l'Oxfordien Moyen-Supérieur (Jura suisse et français) (247 pp.).
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