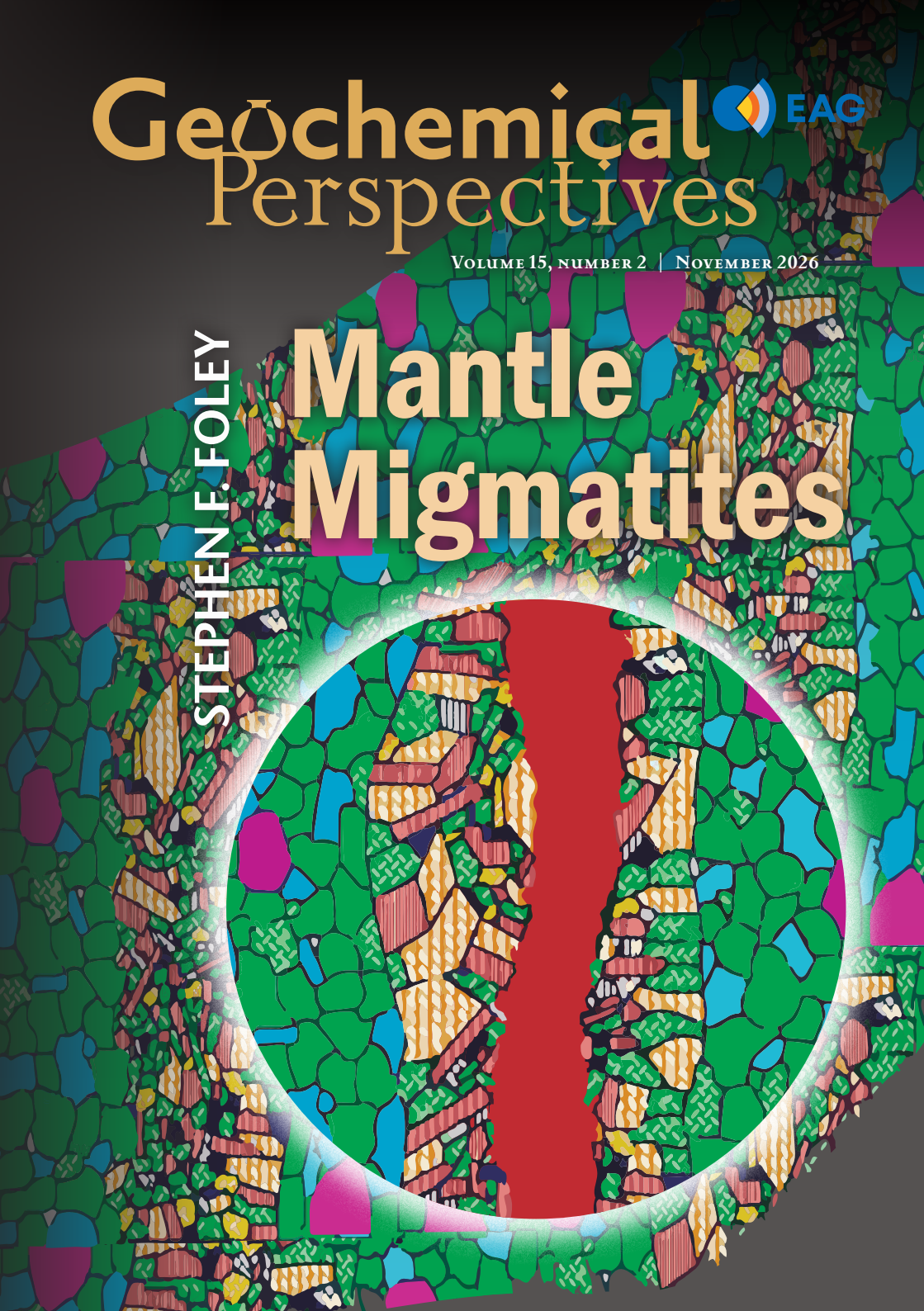
Geochemical Perspectives

EAG

Volume 15, number 2 | November 2026

Mantle Migmatites

Stephen F. Foley



Geochemical Perspectives

Geochemical Perspectives is an internationally peer-reviewed open access journal that publishes invited contributions from leading scientists in the field of geochemistry.

Each issue of *Geochemical Perspectives* features a single, in-depth article that explores the past, present, and future of a geochemical field – presented in a clear, accessible way by respected members of our community. The articles bring together important scientific discoveries in these areas with the human and, often, the technological stories behind their development, offering insights directly from leading scientists who contributed to shaping and advancing the field.

Geochemical Perspectives articles are designed to resonate with the entire geochemical community. Although revolving around the core of the science of the field in question, they offer expert, personal viewpoints on how a research area has evolved, where the field stands today, and where it may be headed next. The overarching goal of *Geochemical Perspectives* articles is to commit to paper for future generations the history of the field of geochemistry.

This article is distributed under Creative Commons Attribution Non-Commercial No-Derivatives 4.0 License (CC BY-NC-ND 4.0), anyone is free to share (copy and redistribute) the material in any medium or format under the condition that:

- appropriate credit is given to the original author and publication
- a mention is included if changes were made to the original material
- a link to the license is provided

However, the material cannot be adapted (remixed, transformed, built upon) or used for commercial purposes without prior permission from the author.

For information on *Geochemical Perspectives*, visit:
www.geochemicalperspectives.org.

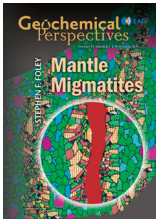
The publisher assumes no responsibility for any statement of fact or opinion expressed in the published material.

ISSN 2223-7755 (print)
ISSN 2224-2759 (online)
DOI 10.7185/geochempersp.15.2

Principal Editor for this issue
Klaus Mezger, University of Bern, Switzerland

Reviewers
Stephan Klemme, University of Münster, Germany
Sandro Conticelli, University of Florence, Italy

Cover Layout **Pouliot Guay Graphistes**
Typesetter **Michel Guay**
Printer **Deschamps impression**



About the cover
Illustration of mantle migmatite consisting of hydrous veins rich in brown minerals cutting peridotite dominated by green. Melting (red) is concentrated in the veins. Based loosely on original figure in Foley (1992).

EDITORIAL BOARD


Janne Blichert-Toft
ENS Lyon, France


Don Canfield
University of Southern Denmark, Denmark


Sarah Gleeson
University College Dublin and Research Ireland Centre for Applied Geosciences, Ireland


Klaus Mezger
University of Bern, Switzerland


Richard Pancost
University of Bristol, UK


Caroline Peacock
University of Leeds, UK


Helen Williams
University of Cambridge, UK


Editorial Manager
Marie-Aude Hulshoff
European Association of Geochemistry


Graphical Advisor
Robert Dennen
Geoscience Editing

CONTENTS

Foreword IV

Acknowledgements VIII

Mantle Migmatites 193

Abstract 193

1. Introduction and Rationale 196

2. Rungs on the Career Ladder 200
   2.1 Getting Started in Geology: Yorkshire and Southampton 200
   2.2 Newfoundland – First Geochemistry and First Ice-Polished Rocks 203
   2.3 Tasmania – First Experiments 206
   2.4 Max Planck Institute for Chemistry 209
   Text Box 2.1 – Choosing a Post-Doc Position 209
   Text Box 2.2 – Memories of Ted Ringwood 211
   2.5 Göttingen – University in the West 212

Text Box 2.3 – The Curious Concept of Habilitation – A Foreigner's Guide 215
2.6 Greifswald – University in the East 216
2.7 Mainz and Geocycles 218
Text Box 2.4 – German Universities and the Max Planck Society 219
2.8 Macquarie and the CCFS 221
Text Box 2.5 – Some Remarks on Gender and Equity in Germany and Australia 222
Text Box 2.6 – From GAU to Super-GAU 224

3. Primitive Mantle Derived Alkaline Rocks 226
3.1 A Short History of Mantle Derived Alkaline Rocks 228
3.2 Primitive and Primary Melt Types 231
3.3 Ruminations on Rock Classification 233
3.4 Kimberlites and Unrelated Rocks 234
3.5 Lamprophyres 236
Text Box 3.1 – The History of the Rock Name Aillikite 238
3.6 Types of Primitive Alkaline Melts 239

4. Mantle Rocks and Mantle Metasomatism 249
4.1 Timeline for Mantle Petrology and Mantle Metasomatism 250
4.2 A Brief History of Mantle Metasomatism 252
4.3 The Rock Record – Metasomatic Minerals and Rocks 257
4.3.1 Massif Peridotites 258
4.3.2 Megacrysts 259
4.3.3 Xenoliths 261
4.3.3.1 Fe-Ti-rich assemblages: the PIC and IRPS suites 261
4.3.3.2 The MARID and PKP suites 263
4.3.3.3 Mica clinopyroxenites 264
4.3.3.4 Amphiboles and the amphibole-apatite suite 266
4.3.3.5 Where is the carbonate? 267
4.3.3.6 Reaction zones 269

5. Experiments on Melting Peridotite 271
5.1 A Short History of Experimental Studies Relevant to Mantle Derived Alkaline Melts 271
Text Box 5.1 – The Debt to Peter Wyllie 274
5.2 Early Experiments in Simple Systems 275
5.3 Melting Experiments on Peridotites 277
5.3.1 Volatile-Free Peridotites 280
5.3.2 Peridotite with Water 284
5.3.3 Peridotite with Carbon Dioxide 286
5.3.4 Peridotite with Mixed H_2O and CO_2 287
5.3.5 Other Volatiles 291
5.3.6 Melting in Reduced Mantle 292
5.4 Melting of Dry Pyroxenites 293

5.5 Liquidus Experiments on Alkaline Rocks 294
5.6 Reaction Experiments 298

6. Experimental Melting of Mantle Neosome Assemblages 302
6.1 Melting Points of Hydrous Minerals 304
6.2 Constraints on the Choice of Neosome Assemblages for Experimental Melting Studies 306
6.3 Melting Points of Hydrous Ultramafic Assemblages. 308
6.4 Melt Compositions 310

7. Where Are We and Where Are We Going? 316
7.1 Comparison of Experimental Melts with Primitive Mantle Derived Melts 316
7.2 An Update on the Concept of Vein Melting 319
7.3 Where Are We Going? New and Missing Information 325
7.3.1 Melt/Rock Reaction Products 327
7.3.2 New Isotope Systems 328
7.3.3 Which Metasomatic Agents? 329
7.3.4 Accessory Phases and Neglected Tectonic Environments . . 331
7.3.5 Redox Effects on Melting of Mantle Migmatites 332
7.3.6 Kimberlite Models and the Origin of Carbon in Carbonated Silicate Melts 333

8. Outlook 337

References 342

Index 369

FOREWORD

My quest in this article is to review and integrate the many lines of evidence that indicate that the mantle sources of many mafic rocks are complex mixtures of more than one rock type, and that the traditional concept of the Earth's mantle as consisting of homogeneous peridotite has run its course. I choose the label "mantle migmatites" to emphasise the parallels with crustal migmatites, in which the volumetric minority of the rock – the veins, or neosomes – are derived by partial melting and resolidification of those partial melts. Once the mixed rock is formed, the neosomes have lower melting temperatures than the rest of the rock (paleosome or melanosome) and so would contribute substantially to the melt if the migmatite were to be re-melted.

Evidence that the mantle does not consist of homogeneous peridotite started to accumulate in the 1970s. This was the decade in which the concept of mantle metasomatism was introduced, and geochemists, petrologists and experimentalists began to get to grips with the effects of volatile components – which was usually expected to mean H_2O and CO_2 – on melt compositions and reactions with mantle rocks. Up until this time, volatiles carried an inscrutable air: there had not been much advance on the opinion from 50 years earlier that a volatile component *"does just what one may wish it to do"* (Bowen, 1928, cited by Burnham, 1979). Another 45 years later in 2024, mantle metasomatism generally remains a poorly defined solve-it-all concept that allows one to avoid being too specific.

The even vaguer term "mantle heterogeneity" also started to crop up in the late 1970s and was frequently interpreted to refer to the results of chemical changes which had no expression in modal mineralogy, or else it did not address the interplay between composition and mineralogy at all. Imprecise characterisations like this can be very enduring, and more than 40 years later many articles *discover* and describe mantle heterogeneity – even in article titles – as if this is a surprising and novel finding. The usage of the two labels *mantle metasomatism* and *mantle heterogeneity* over the last 50 years highlight that our understanding of melting processes in the mantle have become stuck and need a circuit breaker. Part of the problem here is that these terms are vague, since as Maxwell (1984) notes: *"Implicit, covert presuppositions are always much harder to challenge and resist than explicit assumptions".*

Here, I question the homogeneous mantle peridotite paradigm and seek to promote the hypothesis that *most* melting events in the upper mantle that lead to volcanism at the Earth's surface begin by melting of mixed source regions consisting of olivine-free or olivine-poor ultramafic rocks that occur as veins or dykes within peridotite. To be clear, I do not refer just to rare alkaline rocks which many of those who study the much more voluminous basaltic rocks might consider to be exotic and so worthy of neglect. I will concentrate on the origin of these alkaline rocks because they contain the components of the veins and dykes in their purest, least diluted form. I will show that most mafic rocks *must* contain the same components even if diluted almost beyond recognition.

The key to the mantle migmatite hypothesis is that the significance of olivine-free hydrous ultramafic rocks in the production of melts greatly outweighs their abundance in the mantle. This will be true wherever the melting point of the ultramafic veins is lower than that of the peridotite, meaning that melting events will begin in the veins. This revamping of our understanding of melt generation also carries economic ramifications because it revises the possible mechanisms by which rare metals are concentrated in igneous processes.

The mantle migmatite concept is an expansion of earlier concepts of melting of mixed source regions that were required to explain the origin of potassic alkaline rocks (Foley, 1992a), a hypothesis that is receiving more attention recently. However, this has not generally been applied to more common rock types, even though mid-ocean ridge basalts have long since been shown to contain evidence for the involvement of recycled crust (Hirschmann and Stolper, 1996) and MORB sources themselves are required by their isotope signatures to be mixtures of diverse components (Salters and Dick, 2002). The chemical effects of hydrous ultramafic dyke rocks in the source can be recognised in alkali picrites (Ellam and Cox, 1991), even though picrites are generally considered to originate from high degrees of melting of peridotites. Meimechites are another type of strongly magnesian magma that may be enriched in alkalis (Sobolev *et al.*, 2009).

We have now reached a "tipping point" where not even abundant volcanic rocks can be explained without mixed source regions, but we find ourselves with very few experimental melting studies of mantle rocks comprising anything other than peridotite. A *Web of Science* search (7th June, 2026) finds almost 8,000 hits for mantle heterogeneity and more than 5,130 for mantle metasomatism, and 10 times as many experimental studies on peridotite as on all types of other ultramafic rocks put together.

This reminds me of another hobby horse of mine that I have often cited for students: the on-line search at Macquarie University currently reaches back to 1981, whereas it is earlier at the Australian National University (ANU). The time frame covered differs according to which university you work at, but the problem is the same: there is an arbitrary cut off so that many of the original works are not found. There is a danger that modern students accustomed to working on-line may stop there and not consider looking for older material. I will attempt to combat this trend here by mentioning older work from which later trends developed, but which may escape the widespread on-line search engines that seem to have become the lifeblood of science.

In keeping with the tradition of this journal, I write in a review style with early career researchers and graduate students in mind, but I include some anecdotes that may be of more interest to older readers. The timeline follows my own career path, which started with alkaline igneous rocks and passed later to mantle rocks and high pressure experiments. I often emphasise older, and in some cases more obscure, papers that attracted my interest and have been influential in the development of my own thinking over the years. In many universities, it is



Figure 0.1 The Earth from the Moon before Apollo.

unfortunately no longer possible to do what I very much enjoyed in my younger years, namely browsing shelves of books and journals in the university library without really knowing what I was looking for – books and journals that are now shut away in automatic retrieval systems for people who already know what they expect to find. This is how I first became aware of journals such as *Soviet Geology, Geologica Carpathica* and *Kristallinikum* whilst working on my MSc in Newfoundland. I also came across the image in Figure 0.1 in this way in a book in the basement of the Geology Department at the University of Tasmania during my PhD years, where reprints, old books and rocks were stored. Although I no longer know its source, I have used it in classes because it beautifully illustrates how our viewpoints and concepts are dependent on the state of knowledge at any one time. Ever since the Apollo 8 astronauts took a picture of the Earth while orbiting the Moon in 1968, everybody knows what the Earth looks like from a distance, but Figure 0.1, published several decades earlier, exhibits more knowledge of the Milky Way than of how the Earth appears from the outside. We should use this as a reminder that there is a lot we don't know and that our simple conceptions, including of the mantle, may be far from the real thing.

Stephen Foley

School of Natural Sciences,
Macquarie University, North Ryde,
New South Wales, Australia

Research School of Earth Sciences,
Australian National University,
Canberra, Australia

ACKNOWLEDGEMENTS

The thesis summarised in this article has been many years in the making during which I have benefitted greatly from discussions with many people, especially students and colleagues, during my time in Gottingen, Greifswald, Mainz and Sydney (Macquarie). Many of the 49 PhD students and 29 Masters' students that I have had the pleasure to supervise or co-supervise have contributed data and/or experiments that support this work. There are too many to name, added to which I might make a mistake and forget someone, but their input is much appreciated.

I am grateful to several people who read parts or all of the manuscript, namely Stephan Klemme, Lynton Jaques, Dejan Prelević, Isra Ezad, Dorrit Jacob and Sandro Conticelli. Klaus Mezger deserves an award for his extremely painstaking checks of the manuscript and I am indebted to Marie-Aude Hulshoff and Rob Dennen for their work on text and figures during production.

MANTLE MIGMATITES

ABSTRACT

Textbooks on igneous petrology explain a range of primitive basalts and more alkaline volcanic rocks such as basanites, nephelinites and melilitites as products of various degrees of partial melting of mantle peridotite – a supposedly uniform rock consisting of four minerals olivine, ortho- and clinopyroxene and either spinel or garnet, depending on depth. However, all is not well in igneous petrogenesis: many more alkaline rocks such as lamproites, lamprophyres, kimberlites and carbonatites do not fit this homogeneous peridotite mantle model. They require other ultramafic rocks or mixed rock source regions; however, they are volumetrically relatively insignificant and so the discrepancy is often ignored by most people who use the standard peridotite model. This simple peridotite model is widely used in neighbouring specialities that do not consider complex source regions.

The solution is that the mantle source regions of magmas are more complex, containing veins and dykes with abundant hydrous minerals that have accumulated over the eons. The results of episodic melt infusions are mantle

migmatites, the details of which have been hidden behind the shroud of vague labels such as mantle metasomatism and mantle heterogeneity that have resisted attempts to specify what they mean for decades.

Recently a role for mixed rock source regions containing pyroxene derived from deeply recycled eclogites has become accepted, but the need for hydrous and carbonate-bearing minerals in rocks distinct from peridotite has only recently emerged as a necessity. Recent experimental work shows that these hydrous ultramafic rocks melt before peridotite (including hydrous peridotite) in mixed sources under most geodynamic conditions and so will be components, however diluted, in almost all mantle derived melts. This includes voluminous types such as picrites and komatiites that are *mostly* derived from peridotites by further melting. The veins and dykes – the neosomes of mantle migmatites – are minor components of the mantle but nevertheless important because they always melt first and so cannot be ignored. Many voluminous basaltic rocks are thus a kind of geological palimpsest; they still bear evidence of events that are swamped by later melting of peridotite.

My career has overlapped to a large degree with the development of the mixed source concept and so this review tracks the history of relevant areas that I have worked on: mantle derived alkaline rocks, mantle petrology, mantle metasomatism, and experiments on peridotites and hydrous ultramafic rocks.

After a brief career history (Section 2) I review the different types of alkaline rocks that may originate by melting of hydrous ultramafic rocks and carbonates in the mantle (Section 3). They cannot be simple melting products of peridotite alone, not even if the peridotite is "metasomatised". I point out similarities and differences between these alkaline rocks for the non-specialist and to illuminate some of the semantic problems surrounding these rocks, some of which have changed their names more than once. This semantic confusion can bewilder the non-specialist, so it is not surprising that many would rather ignore them and retain the simple peridotite model.

Section 4 reviews the evidence for the variety of mantle source rocks, mostly from mantle xenoliths. Amphiboles and micas are common, usually with clinopyroxene. Other minerals such as the Ti oxides (ilmenite, rutile and perovskite), as well as apatite and carbonate may be abundant in mantle dykes despite usually being thought of as accessory phases.

The history of experiments at mantle pressures (Section 5) shows that the concept of mantle metasomatism followed the first experiments on peridotite remarkably quickly in the 1970s. This may be part of the reason that alkaline melt origins were sought simply by adding H_2O and CO_2 to peridotite melting experiments. Liquidus experiments on alkaline rocks were common, but failed to find the desired melting conditions of volatile-rich peridotites. Melting of volatile-rich peridotite never worked well as an explanation, but most were satisfied with *ad hoc* modifications, leaving only few people to continue working on explaining the strongly alkaline rocks.

Experiments on the melting of hydrous ultramafic rocks (Section 6) – the mantle migmatite neosomes – have mostly been published since 2010. The results we now have show that rocks with amphibole or phlogopite will usually melt to a substantial degree before peridotites begin to melt and so will dominate the early melt compositions. Comparing experimental melts to natural rocks (Section 7) shows that melting of amphibole-rich dykes produces sodic alkaline melts akin to nephelinites and melilitites, whereas mica melts more slowly, resulting in high K_2O/Na_2O in melts that resemble lamproites.

There is a huge variety of potential dyke rocks in the mantle, but few of these have yet been subjected to melting experiments. In the future, more experimental work should consider carbonate-rich assemblages, Ti oxides and sulfides, and rocks with both calcic amphibole and phlogopite. The mineralogy of reaction zones between infiltrating melts and peridotite wall rocks has come into focus, but the range of melts produced by re-melting of these reaction assemblages have not yet been studied in detail.

Using the analogy of thin section microscopy, it is time to switch lenses to a more powerful objective and appreciate changes in scale. For example, we need to move away from global averages such as *the* oxidation state of the mantle and from experiments on global average sediments that do not exist anywhere. Melting processes depend on conditions and rock compositions that vary on a small scale. Many melts emanate from an oxidised beneath the cratons in which carbonate melts can exist in places, but C is present only in reduced form at both shallower and deeper levels. Care must be taken when referring to low degree melting in mixed source regions as this depends on spatial scale.

Progress will rely on demystifying metasomatism and heterogeneity by being more specific about what they are supposed to mean and about the abundances of volatile species involved. Explanations of melt origins can be expressed as a continuum between melting of peridotite alone and melting of pure neosome assemblages. Between them is a transition zone where melts such as nephelinites, basanites, melilitites, and alkali basalts have alternative explanations: their sources may consist either of volatile-bearing peridotite or mixtures of hydrous dykes and peridotite. The hydrous and carbonate melts produced from mantle migmatite neosomes are important carriers of critical metals, making study of mantle migmatite melting process societally topical and important.

1. INTRODUCTION AND RATIONALE

The lesson of history is that our firmest convictions are not to be asserted dogmatically; in fact, they should be most suspect; they mark not our conquests but our limitations and our bounds.

Kline (1980)

Throughout my career I have been puzzled at the apparent disparity of expectations when it comes to conceptualisation of the lower crust and upper mantle. Mapping in coastal Labrador in 1980 for my MSc degree took me to the mid- to lower crust, where beautiful, extensive, ice polished outcrops showed me the complexity of crustal rocks from several kilometres down. Figure 1.1a is my favourite locality summing up the Archean crust, since it contains the main components of tonalitic gneisses and rafts of metabasaltic rocks welded together by late stage intrusions of granitic rocks. From field outcrops people expect that the lower half of the continental crust consists of a mixture of multiply deformed rocks with different lithologies and various origins, and that they are often migmatites (Fig. 1.1b).

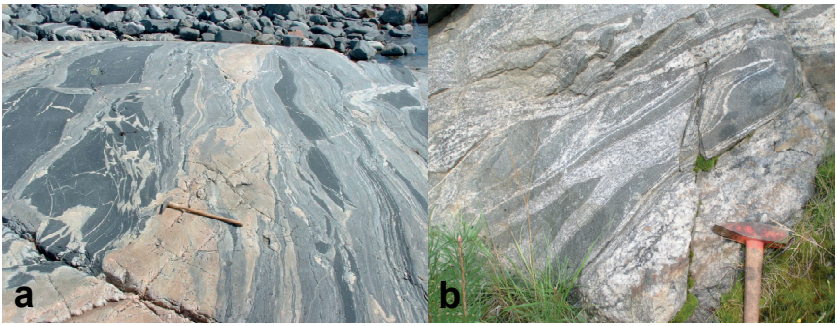


Figure 1.1 Lower crustal migmatites. **(a)** Rafts of metabasalt in tonalitic gneiss, both cut by granitic veins. Turnavik Island, Labrador. **(b)** Complex folds in late Archean tonalitic gneiss, near Iisalmi, central Finland.

My introduction to the mantle was mostly theoretical. I visited Labrador to look at, and sample, alkaline igneous dyke rocks formed by melting of the upper mantle. I had learned that this mantle was made up of peridotite consisting of just four minerals; olivine, orthopyroxene, clinopyroxene and one of garnet or spinel, depending on the depth. There had been some revelations in the preceding decade about the necessity to have water and carbon dioxide in the mantle, but these volatile components were generally assumed to be influencing a mantle composed of homogeneous peridotite.

There were two direct sources of information about mantle rocks: massif or Alpine peridotites and xenoliths in volcanic rocks. Xenoliths in kimberlites have been used to conclude that the mantle is made of peridotite for almost 100 years (Wagner, 1928), but they suffer the insurmountable problem that they are small, so that homogeneity on a large scale cannot be assessed. Large bodies of massif peridotite outcropping at the surface have more to say about large scale homogeneity. These were recognised in a famous paper by O'Hara and Mercy (1963) to be "*little altered fragments of the mantle*", although other possible origins were debated late into the 1960s so that Ernst (1965) conservatively considered a mantle origin for these peridotite bodies *"as being plausible"*. Massif peridotite bodies commonly display outcrops of tens of metres of apparently uniform peridotite, contrasting with the complex constitution of many lower crustal rocks (particularly continental ones).

Why should the crust/mantle boundary be a single line of such fundamental change? What is the nature of this boundary anyway? I thought back to an excursion to the Lizard Peninsula in Cornwall during my final year of undergraduate studies where I was introduced to the "petrological Moho" in the field. Southampton and Newfoundland geologists had recently reinterpreted the Lizard to be an ophiolite (featuring what is still one of my favourite article titles: *"A new tale for the Lizard"*; Strong *et al.*, 1975), competing with Dave Green's interpretation of the peridotite as a high temperature intrusion (Green, 1964). I recall Southampton University's Nic Badham standing on the shore at Coverack declaring he had one foot in the mantle (peridotite) and the other in the crust (gabbro) – an image that leaves a memorable impression on a student. But can you do that? As for mantle xenoliths, this too has a scale problem, as investigations at larger scales often find changes in the abundances of ultramafic to mafic rocks to be gradual over hundreds of metres (Dilek and Furnes, 2014). Also, it didn't even address the issue of the difference between the petrological and seismic Mohos, which is more of a problem beneath the continents. Thirty years after my visit to the Lizard one could still find diverse opinions on whether these versions of the Moho coincide or not. The impression this uncertainty may leave on a non-specialist could be that we are not even sure what the crust/mantle boundary is.

In massif peridotites, minor proportions of other rocks occur, particularly dykes of pyroxene-rich rocks, as in the bodies at Balmuccia, Lherz and Almklovdalen. However, this occurs on the small scale of centimetres to tens of metres that would not be recognised if we were to zoom out to the scale of kilometres that is appropriate for remote geophysical surveys.

Keeping the mantle simple as peridotite may be an expression of an oft quoted philosophical principle called Occam's Razor, which is named after the work of William of Ockham in the 1300s (I am not sure at what stage the spelling changed). As Bertrand Russell summarises: "*Occam is best known for a maxim which is not to be found in his works* (that's curious!)*... but has acquired the name of 'Occam's razor'*. This maxim says: *"Entities are not to be multiplied without necessity""*

(Russell, 1946, p. 462), or, in modern terms, "keep it as simple as you can". If geophysical measurements, made at a distance of tens to hundreds of kilometres, do not have the resolution to decipher lithological changes at a scale of tens of metres, then surely it is simplest to assume that these inhomogeneities are insignificant and that the mantle can be safely summarised as homogeneous peridotite. We cannot directly access much of the mantle, so why make it unnecessarily complicated? A related philosophical viewpoint with a dose of practicality is: "*The case for simplicity is pragmatic. It simply is easier to calculate with simpler theories*" (Newton-Smith, 1981).

Evidence that the mantle is more complex than homogeneous peridotite began to accumulate in the 1970s. Two general types of rocks were outlined; peridotites which were thought to be the main mantle constituent, and pyroxene-rich rocks that were interpreted as solidified melts that had not made it out of the mantle (Wilshire and Shervais ,1975). The pyroxenites were accepted to be derived from solidified melts once composite xenoliths were found in which the pyroxenites could be seen to intrude and/or surround the peridotites (Irving, 1980). This led to culinary analogues such as "marble cake" and "plum pudding" mantle (*e.g.,* Allegre and Turcotte, 1986). One could say that the mantle was already struggling with its homogeneity image at this time, but this simple concept has nevertheless proven resilient.

My purpose in this article is to expand the concept of migmatites to the upper mantle. This proposes that the Earth's mantle can no longer be realistically modelled as homogeneous peridotite from a petrological and geochemical viewpoint, and that it is counterproductive to continue to do so. I draw on the origins of mantle derived igneous rocks, particularly alkaline ones, and variations in mantle rocks, and try to fit these together. This interplay of igneous and mantle petrology and geochemistry mirrors the scope of work I have done with colleagues and students over much of my career. I am well aware that I am swimming against the tide implied by Occam's Razor and that *"the need for a different system will not be recognized unless we can be persuaded that our actual system does not work so well."* (Ayer, 1973). This is my purpose with this review.

I begin by summarising the evidence that most alkaline igneous rocks cannot be direct melting products of peridotite, but that they require other ultramafic rocks in their sources (Section 3). This requires staking out and characterising the variety of primitive, mantle derived melt types in order to assess which ultramafic rocks may be relevant.

This is followed by a short history of mantle metasomatism (Section 4), concluding that while most of the upper mantle is indeed peridotite, it is criss-crossed by dykes of other rocks to a depth of at least 160 km beneath the continents and probably much deeper. These may seem to be volumetrically insignificant and so are frequently overlooked, but where they contain hydrous minerals, they melt first and the hydrous minerals play the principal role in determining the compositions of melts.

In Section 5, I draw on high pressure experimental results to analyse our knowledge of the melt compositions that may be produced from peridotite. This characterises the rather fuzzy compositional limits of melts of peridotite, particularly with volatiles.

In Section 6 I characterise what we know about the range of melt compositions that could be produced when hydrous pyroxenite, glimmerite and hornblendite dykes re-melt. This section illuminates that, despite the much greater mineralogical variety of these dykes relative to peridotite, they have attracted only a fraction of the number of experimental studies that peridotites have enjoyed.

I next attempt to fit these three lines of evidence together to zero in more closely on the source assemblages and conditions of melting for a range of mantle derived igneous rocks (Section 7). This should help to focus and refine what we mean by expressions such as mantle heterogeneity and mantle metasomatism.

2. RUNGS ON THE CAREER LADDER

I have had a rather unusual geological education in having completed four degrees in four different countries. This contrasts with the careers of many academics and gives me a rich history of experiences in different university and research environments. I hope that some of this short series of recollections is entertaining, or at least illuminating.

2.1 Getting Started in Geology: Yorkshire and Southampton

Life has many events that I like to refer to as bifurcation points, as I find the bifurcation diagrams in chaos theory to be an excellent illustration of the concept. These may be conscious decisions that you make, or you may have no control over them, but they change the direction of your life like points in railway tracks beyond which life will never be the same again. Choosing whether to accept a new job or moving to a new town, city or country are examples.

I began my career in geology because of one such bifurcation point that was beyond my control: I started to study geology in the last two years of school only because of a timetable clash! I had to choose three subjects to study to A-level for my final two years of school in North Yorkshire, England, when I was 16. I had chosen biology, geography and chemistry, but I was informed that the timetable would not allow it, so I had to quickly find a replacement for one of them. The school was in a bit of disarray: it was just completing its transformation from a boy's Grammar School into a sixth form college – I had been in the final year of the Grammar School, which thus ceased to exist when I finished the fifth form. The new sixth form college was given the grand name of Prior Pursglove College in honour of the founder of the grammar school, thus appearing to co-opt its history and give the false impression that the College had been around since 1517 instead of 1974!

As the College started its business, the scope of subjects offered diversified massively from what the old school had offered, since all students would be limited to the age range 16–18 from now on. This included geology, which I chose to study for two reasons: firstly, it seemed similar to physical geography, which had always been my favourite subject. I very much enjoyed the outings this involved, recalling day trips to see glacial geomorphology in the Lake District and impressive limestones with sink holes, caves and pavement at Malham in the Yorkshire Dales. Secondly, I had heard good reports about the teacher, Dr. Ernie Bradford, a small, active man shortly before retirement with a curious and, at times, amusing habit of unexpectedly emphasising particular words. The geography teachers seemed to idolise him. I did an O-level in Geology in the first

year and an A-level in the second in what proved to be my best A-level result. I was always puzzled why it was not possible to take those three subjects, as it didn't seem such a strange choice to me.

I went off to Southampton for my undergraduate studies, which was about the furthest university from home, which was very tempting to me at the time. Still 17, I was baited more by the leafy campus at Southampton and its attractive prospectus than any knowledge of the university. I often think of this in the discussions in recent years about how to recruit new students. Modern universities seem to assume that students (or their parents) look deeply into the reputation of the university. I didn't, but then as you can see (Fig. 2.1a) I was pretty green at the time; I was more interested in running and exploring the world than in taking studies seriously. At Southampton I was in a cohort of just 19 students (half of them in Fig. 2.1b), just three of whom went on to do post-graduate degrees and another three who moved directly into work as professional geologists.



Figure 2.1 **(a)** Just turned 18, from my student card when I enrolled in university studies at Southampton. **(b)** Group photo at South Stack lighthouse, Anglesey, in 2nd year. I am fifth from the left; fifth from the right is Jim Andrews.

In moving so far from home I was following my love of the exotic, but I soon realised that here in Southampton in the far south it was me that was the exotic one! My fellow students came from close by, from London, Cornwall and even south Wales, but none from more than 90 km north of Southampton. I had to concentrate on softening my northern accent as they claimed they couldn't understand me, and they were disappointed to hear that my father was neither a coal miner nor a sheep farmer, which would have suited their prejudices about northerners. I spent many of my weekends with the cross country club or at the athletics track and enjoyed visiting many universities (and fields and forests near them) in southern England at athletics events.

I studied at the old Geology Department on the main campus, not the Oceanographic Institute at the docks which everyone knows nowadays. Oceanography was already strong though, and I was intrigued by the tank models of the complex tides around the Isle of Wight, which showed why Southampton

has four high tides a day. I loved the field excursions and have fond memories of Wales (Fig. 2.1b) and the south coast of England as well as my independent mapping project at Ingleborough in North Yorkshire (Fig. 2.2), for which I got the prize for best map. My interest in sedimentary rocks at this time benefitted greatly from the teaching and mentorship of Derek Moore, from whom I heard my very first lecture at Southampton on deltas.

I was taught igneous petrology by Miss Dorothy Wisden using a textbook which had first appeared in 1891. Needless to say, it was a largely petrographic book. The geochemistry lecturer was Mike Cosgrove, who, unbeknown to me at the time, had studied potassic igneous rocks in Devon (Cosgrove,1972).



Figure 2.2 View of part of my independent mapping area at Ingleborough, northwest Yorkshire. The hilltop is the Carboniferous Millstone Grit, below which the deltaic Yoredale Beds lie above extensive limestones. There was only one igneous rock in this area, a lamprophyre dyke.

In my final year I needed to choose a lab project, which turned out to be my first brush with the Earth's mantle. Southampton's structural geologist was Jim Andrews, who had described the first kimberlites in Greenland (Andrews, 1969). After this discovery, Jim had attended the first *International Kimberlite Conference* at Cape Town in 1973 and had collected xenoliths in South Africa and Lesotho. Six peridotite and two granulite xenoliths from Matsoku became the subject of my lab project. I don't remember much about them now, other than that the peridotites contained big garnets and chrome diopsides, which was enough to spawn my embryonic interest in mantle rocks.

Nic Badham was the economic geologist – a chirpy and self confident type who left for industry soon after I finished at Southampton. He had done his PhD in Alberta and told many stories of Canada, and the idea of fieldwork in the barren and frozen north appealed to me greatly. He had been in the same undergraduate class as John Malpas at Oxford, who was now at the Memorial University of Newfoundland. This is how the connection for my MSc in Newfoundland arose. I learned of my final exam results in Julian Pearce's office at the Open University whilst visiting John to plan my MSc project; Julian was away on sabbatical at the time and John, also on sabbatical, had moved into it.

My introduction to Newfoundland students came when I was invited to join their field trip to Cornwall at Easter (1979) which Nic Badham led. It was on this excursion that I visited the Lizard Peninsula.

2.2 Newfoundland – First Geochemistry and First Ice-Polished Rocks

August 1979 took me to the Memorial University of Newfoundland (MUN) in St. John's, which was warm when I arrived, but this soon changed and there was ice on ponds on 1st October during my first field trip to the centre of the island. The weather took some getting used to; I had never experienced gales and fog at the same time before! The winters were long for a young Brit from close to the Gulf Stream: I recall vying for space on the rocks between ice floes on a field trip to New World Island (central Newfoundland) the next May, and shovelling snow from the athletics track so that we could hold the first meet of the season on 31st May!

I found Newfoundland to be very welcoming, but most people there say they are half Irish so maybe my name helped? Life was very laid back: cars on four lane highways would all slow to a stop to let pedestrians across! It made me wonder how many Newfies are killed when visiting the mainland, expecting the same behaviour from drivers there. My favourite sport of track and field athletics was equally welcoming: imagine my excitement as a poor student on a squalorship when flights to Ottawa and Montreal were paid for me to take part in national meetings! I was Meet Manager of the Canadian National Junior Championships when I had just turned 23 – this would have been unthinkable in Germany, where one has to serve for decades before being thought mature enough for such a role.

Newfoundland also saw the beginning of my lengthy informal and unintended course in intercultural studies. During my 3 degrees in 3 countries, I shared flats or houses with 38 people from 12 different countries. Apart from two boys from Hong Kong and Greece in Southampton, living closely with foreigners began here in Newfoundland. Canadians and Americans were not so surprising, but there were just as many mainland Chinese: this no longer seems worthy of mention, but in 1980, this was the first wave from China to be allowed to venture abroad after the end of the Cultural Revolution. I shared flats with four over two years and recall heated political squabbles between my flatmates and "overseas Chinese" who arrogantly nurtured their own opinions. My love of dumplings started here; homemade ones spiced with ginger and spring onions!

At work, I was to be supervised by John Malpas and Ken Collerson on a project on kimberlite dykes in Labrador (Fig. 2.3), but Ken fled to Australia when I had been there just 4 months (I like to think this wasn't related to my arrival). I had gone expecting to do a PhD, but I soon found that the rules wouldn't allow it – I had to do an MSc first because of the worthless nature of my English BSc

– I had not done a year long Honours thesis! This was the first event that led me to eventually formulate *Foley's Law* about the destiny of students abroad: "*When you move as a student to a new country, the people there believe their system is better than where you came from, regardless of where you move to or from*". This law was cemented in place in Tasmania where I was told that my 2 year Canadian MSc had about the same value as an Australian BSc Honours thesis!

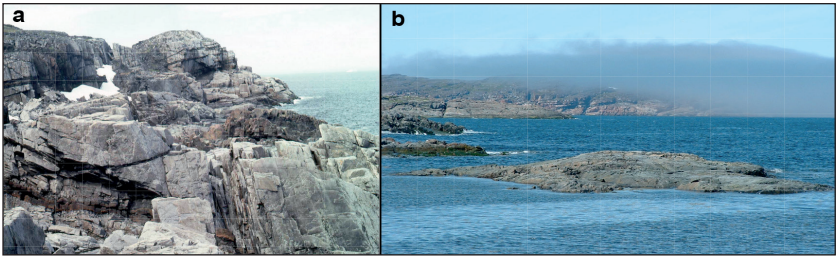


Figure 2.3 **(a)** Criss-crossing dykes of various ages (1375–125 Ma) at Cape Makkovik, Labrador. **(b)** Rapidly incoming fog, which was cause for alarm when boating between islands in 1980, but no longer in 2003 after the advent of the "go to" function in GPS!

The facilities at the Geology Department at MUN struck me as more diverse and modern than at Southampton, although methods seem archaic when looking back from 2025. Rare earth elements were analysed by X-ray fluorescence on papers after a week's wet chemistry and the data was stored on flimsy micro-cassettes (much smaller than music cassettes, in case you know what those are) with tapes that would often get screwed up into balls, needing 20–30 minutes to disentangle them. I did appreciate the hands on software for the plotter, however, and spent enjoyable hours designing and programming new shapes and sizes for symbols to plot geochemical results. In my final semester at MUN – it must have been November or December 1981 – I attended a seminar on the new embryonic method of ICP-MS which our geochemist Brian Fryer had arranged. There was a long discussion afterwards which gave me the impression that this method had no future, as nobody seemed to believe the interferences would *ever* be sorted out! I returned to Newfoundland in the early 1990s to use the Laser-ICP-MS that was pioneered particularly by Henry Longerich. It was Henry who reminded me at this time that my MSc thesis had been the first MUN thesis written on a computer! I had done a course on programming that Henry taught, and he was interested in my progress because he had advised me when I used the computer to compile track and field statistics (I was responsible for compiling National Junior lists for the Canadian Track and Field Association). I wrote the thesis in parts on the university's main frame (there were no personal computers yet) but ran into a last minute panic when I couldn't find a high quality printer: the thesis was not going to be accepted by dot matrix printer on

green and white lined paper which my drafts had been printed on! I eventually found a new daisy wheel printer in the Department of Education which they let me use!

I was left alone to organise fieldwork in Labrador, which I managed with the help of fellow graduate students who put me in contact with a mining company that was prospecting for uranium on the Labrador coast. The field season was plagued by local attitudes that I did not have the background to be prepared for. I thought I had it all organised when my field assistant dropped out at what seemed like the last minute, preferring to make money playing with his band instead. When we got to Makkovik in Labrador (Fig. 2.4) my reservation of a boat to work on the islands turned out to be worthless: the salmon were running early and the boats were needed for the short fishing season, so none were available for me. There was no airport on the Labrador coast at the time, so we were supported by the mining company and flown by helicopter to islands first. On returning to the mainland coast, I had a horrendous reaction to insect bites and my hands swelled to twice their usual thickness. I was flown out to hospital in Northwest River to deflate. On my return to St. John's, my supervisor John had heard of these events and joked "*I didn't expect to see you back… ever*!" I was lucky to have such a caring and empathetic advisor as John Malpas. He would often lend his truck to students, and when my father died just 3 months after my arrival, he said I needed to be home and drove me straight to the airport, where he bought me a transatlantic ticket at the check-in desk, not caring if he would ever get the money back.



Figure 2.4 The town of Makkovik on the central Labrador coast in typical summer weather. In 1980 it was reachable only by sea or seaplane.

2.3 Tasmania – First Experiments

Newfoundland gave me the appetite to see more of the world, and I started to look towards Australia. Just two years earlier, in 1979, kimberlites with diamonds had been found in Western Australia (Jaques *et al.*, 1984), so I wrote to three universities asking if anyone would supervise me to work on these new rocks towards my PhD. All replied that they would gladly accept me if I bring my own money, but one – some character called Ringwood at the ANU – told me I should ask Dave Green in Tasmania. So I did!

In the final weeks of 1981 I was hoping I would receive positive news about a scholarship in Tasmania, but arranged a back up plan of doing isotope geochemistry on the Labrador dykes at MUN with Brian Fryer. One of my best Christmas presents ever was a telex (some of you might have to look up what that is) from Dave Green on 24th December telling me that I had been granted a University of Tasmania scholarship! Now I just had to get a visa and find the money for a flight. I reserved a flight, which was difficult, because the travel agency couldn't do anything with the concept Tasmania so that the first ticket I was offered was to Dar-es-Salaam! By the way, PhD student Sebastian Tappe finally did the backup isotope project 20 years later in Mainz.

I arrived in Tasmania at Easter 1982, and my four and a half years there (prolonged by a five month stay in Antarctica) was a very happy time. I was treated to a field trip to central Tasmania in the first week: used to the expansive, ice-polished outcrops of coastal Canada, I had to get used to people giving rock names to what I viewed as orange sand and brown sand!

1982 was the year of the Gordon Dam demonstrations, which eventually prevented the construction of a hydroelectric dam on the Franklin River in western Tasmania. Figure 2.5 gives you an idea of the country in western Tasmania. Historically, this now seems to have been a tipping point for conservationism. Although demonstrations to prevent a huge hydroelectric dam now seem the obvious thing to do, it was hard to push it through in Tasmania in the 1980s.

Here my multicultural living continued: at work, I was mixed with students from Australia, Indonesia, USA, New Zealand, Iran, West Germany and China. At home, my circle of friends came from all over Africa in addition to Tasmanians. As an introverted European, I must have looked out of place at parties with many of the University's African students in our house. As well as Australians, I had flat mates from Nigeria, Tanzania and Zambia, the third of whom, Francis Tumeo, was a mature student who had written speeches for President Kenneth Kaunda. I learned a lot about African ways from discussions with Veddy Ndyetabura, with whom I enjoyed the successes of Tanzanian athletes at that year's Commonwealth Games. This period also introduced me to mental health issues, as three of my many flat mates suffered intense headaches, breakdowns or vivid nightmares that landed them in hospital. They were later attributed to exam anxiety, acute homesickness and child abuse (*e.g.*, Grandma

sprinkled broken glass in his sandwiches!), which were invaluable experiences for me when I had to deal with my own cPTSD issues in my fifties caused by events in my early to mid-teens[1].



Figure 2.5 Trevor Falloon and Tony Crawford fishing for their (and my) supper on a field trip to Macquarie Harbour, western Tasmania, ca. 1985.

When I arrived in Tasmania, Dave Green gave me a choice of possible projects, which I later always tried to do for my own students. One of these was to revisit the Lizard peridotite, the subject of his own PhD, to revisit his interpretation, which had been questioned by a group of geologists from Southampton and Newfoundland (Strong *et al.*, 1975). I guess I was seen as ideal for this task, as I now had degrees from both offending institutions. However, I chose to study the Gaussberg olivine leucitite, with the understanding that I could move on to experiments on the Western Australian lamproites. Dave was intrigued to know how these compositions could be derived from mantle peridotite! I note here that he still assigned olivine leucitites to peridotite sources many years after I thought I had demonstrated that they cannot be partial melts of four phase peridotite (see Section 5). Ah well, that's part of the motivation for this article. I wasn't the first or last student he tried to interest in the Lizard project.

The PhD project in Tasmania was largely experimental, which I began after dropping a project on the compositions of melt inclusions in the Gaussberg rocks, since I couldn't make any sense of the results after several months. My lab

1. During my recovery period I came to appreciate that completing my first degree at Southampton was a much greater achievement than the MSc or PhD!

book records my first high temperature experiments under the auspices of Keith Harris in October 1982, the results of which appeared in my second paper (Foley, 1985). My first high pressure experiment was performed on 2nd December 1983.

I was one of several of Dave's PhD students, including Klaus Nickel, Wayne Taylor, Scott Kuehner, Trevor Falloon, Margaret Wallace, Steve Eggins, Nic Odling, John Adam and Ai Yang. Many of these have, at the time or later, produced landmark papers in petrology: those directly from the Tasmanian laboratory include the first determination of the reduced mantle solidus (Taylor and Green, 1988), the discovery of the stability region for carbonatite melts in the mantle (Wallace and Green, 1988), some of the most widely applied geothermometers and geobarometers for peridotites and eclogites, and detailed studies of mid-ocean ridge basalt genesis. Imagine my surprise and vexation when I saw that the history of petrology by Davis Young (Young, 2003) mentions David H. Green only briefly as an assistant to Ted Ringwood (p. 513-514) and doesn't mention that this proud Tasmanian was ever in Tasmania at all! The achievements of his excellent research group are missing completely. He is also missing from Yoder's (1993) time table of petrology, who notes "*it is often said that history is the distorted view of a single observer, and this timetable is no exception*." I couldn't agree more, Mr Yoder! No wonder many Australians complain their efforts are often forgotten or ignored by American scientists. For more balanced view, you can refer to the memoir of his life by Jaques *et al.* (2025).

A major positive factor in my scientific training were the weekly lunchtime group meetings – another concept I later used with my own students. Dave was extremely busy at national level and so, given the isolation of Tasmania, was often away in Canberra. He had already been Vice-President of the Australian Academy of Science by this time and was a member of the Australian Science and Technology Council for the Prime Minister and Cabinet, as well as on other advisory boards. The group meetings, often led by post-doc David Ellis when Dave was away, were the most effective catch up for him on the varied student projects when he was in town, but for me they were excellent training in explaining and communicating my results to others. Despite the isolation, we had international visitors, including Alan Thompson from ETH Zürich, Alok Gupta from Roorkee, and Igor Ryabchikov from Moscow, who were active contributors to these meetings. Tasmania was my "*time of intoxicating intellectual excitement*" as Lewens (2016) wrote about Karl Popper in Vienna.

Also leading the way at the time was the University of Tasmania's Central Science Laboratory, a forerunner of many in Australia, which housed analytical facilities from all across the sciences. I doubt if I would have been introduced to FTIR spectroscopy if the facilities had been dispersed as in most universities at the time.

Tasmania was much more isolated than it is now. Apart from the difficulty of reaching it in the first place (funds kindly provided by the I.C.I. Educational Trust in London), there were no visits home or from relatives or friends. I spent much of my time on the Domain (a hill in Hobart) at the athletics track, or on runs that would be partly along beaches[2].

Journals came by sea mail and so we were constantly about five months behind Europe and North America: it was always exciting to have someone return from an international conference and report on recent publications that we hadn't seen or heard of yet! I got to the Big Island (mainland) several times, but more often because of athletics than geology. I recall Dave Green's non-plussed look when he met me at Melbourne airport on one of his many trips to Canberra! I went to two national conferences in Australia (Sydney and Adelaide) and two international meetings thankfully came to us: my Tasmania years closed with the *4th International Kimberlite Conference* in Perth in 1986, where I was astonished to be appointed a session chairman whilst still a PhD student. I must have made a good impression somewhere.

2.4 Max Planck Institute for Chemistry

When looking for a post-doc position, I applied in Canada and the UK (Edinburgh) before deciding on the Max Planck Institute for Chemistry in Mainz, West Germany (Text Box 2.1). I was tempted by the belt apparatus, which enabled experiments to 60 kbar instead of the 40 kbar limit of piston cylinder instruments, and probably also again by the lure of a new country. I arrived in Mainz to a major change in living environment and in outlook. I thought I was well seasoned at moving to new countries, but this time I had also moved to a new language, which was harder than I expected. I had been told before moving there that it wouldn't be a problem because everyone there speaks English, but this is not what I found: the Max Planck Institute that these people were referring to seemed to be the Geochemistry Division, whereas I was in Cosmochemistry, which seemed to consist mostly of Austrians! For the foreign ear, the Austrian accent renders the place into the "Mox Blonck Institut" für "Kosmokimee", which is what I had to get used to. They *could* speak English, but they mostly *didn't*, and the technical and workshop staff couldn't. The only sensible way forward was a two month intensive German course, which helped me immensely in the long run.

Text Box 2.1 – Choosing a Post-Doc Position

History records that I moved to the Max Planck Institute for Chemistry in Germany when I finished my PhD, but there were several bifurcations in the route that got me there. I had an offer to go to Canada, which I decided to turn down before hearing

2. Those of you who live in big cities can imagine the joy of living in Sandy Bay.

for sure that the Max Planck option was going to work. Feeling somewhat isolated in Tasmania and unsure what to decide, I was advised against this Canadian job by an international visitor to Tasmania (Alok Gupta).

I also received an offer of a two year NERC Research Fellowship to work at Edinburgh, which arrived just after I had received confirmation from Mainz. A copy of the letter I wrote to Edinburgh records the reasons for my preference at the time: *"the opportunity to work with the belt apparatus up to 60 kbar"* and *"to live and work in another country of which I have no experience"*. However, to turn down a NERC Fellowship was more unusual than I had realised at the time: some years later when visiting the UK to prepare other applications (which were not successful), I heard *"Oh this is the guy who turns down NERC Fellowships"*. I never did get back to Britain.

The Max Planck Institute for Chemistry in Mainz had an excellent reputation – even my mother had heard of it (!) because my father had spoken of it (he was an industrial organic chemist). She said he would have been very proud.

Cosmochemistry was a whole new field for me and seminars and the whole Earth scale of discussions during coffee breaks were very illuminating. The Director, Heinrich Wänke, was known for nodding off in the front row during seminars and waking to ask a very astute first question. My outlook on science also benefitted greatly from talking to the many international big name visitors such as isotope specialist Alan Zindler (Florida), Ted Ringwood (see Text Box 2.2), eminent palaeomagnetist Keith Runcorn, George Wetherill (Carnegie Institution of Washington; some will know the name from a Pb isotope plot) and lunar experimentalist Dave Walker (Lamont-Doherty Observatory), not to mention others who visited the Geochemistry division. These chats mostly took place during coffee breaks above the stairs in the entrance hall. Mostly at the other side of the car park, but partly in the same building as me, were the geochemists, including Al Hofmann, Nick Arndt, the Steve's Goldstein and Galer, Roberta Rudnick, Bill McDonough, Klaus Mezger and Uli Christensen, although it seemed strange to refer to Uli, a geophysicist who later led the Max Planck Institute for Solar System Research, as a geochemist. More about the team and facilities in cosmochemistry can be gleaned from the *Geochemical Perspectives* issue by Herbert Palme (2018), whose Figure 4.1 shows an aerial picture of the buildings I mention here.

The Mainz belt apparatus was the phenomenal achievement of Gerhard Brey and Rudi Weber. It now stands in the University of Frankfurt, where Gerhard took it when he moved there in the 1990s. Various laboratories around Germany had played with belt apparatuses before – I saw various bits of defunct belt apparatus in different universities, including Göttingen – but as far as I am aware, the Brey-Weber version is *still* the only high temperature calibrated belt apparatus (Brey *et al.*, 1990). This allowed routine, long duration experiments[3] at 60 kbar with temperature control as good as the widespread piston cylinder

3. Well, long duration as long as Gerhard didn't lean on the red button marked *Notaus*.

apparatuses and much better than the multi-anvils of the day. It is a remarkable piece of equipment that deserves to have been copied more. I spent many happy hours running experiments (Fig. 2.6).

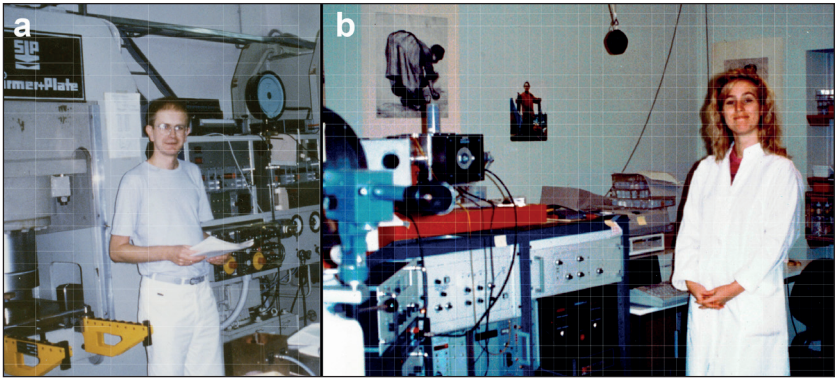


Figure 2.6 **(a)** In the lab of the belt apparatus at the Max Planck Institute for Chemistry, Cosmochemistry Division, in Mainz. **(b)** Dorrit with Emil Jagoutz's home built single collector mass spectrometer one floor above the experimental lab.

Mainz threw another bifurcation into my life, which rotavated my life views more than most of you would realise: it was here that I met my future wife Dorrit, at the time a chemistry lab assistant with Gerlind Dreibus in the Cosmochemistry Division before she (Dorrit, I mean) began to study mineralogy. I was very impressed when Dorrit stood in to run the Secretariat for a week when Heide Prager was away at a conference, answering the phone in any of three languages flawlessly. Our future became clear at Craig Smith's leaving party at the Cowboy Club in Gonsenheim, where we shared a bottle of wine when all the others were drinking beer.

Text Box 2.2 – Memories of Ted Ringwood

My new office in the Research School of Earth Sciences (RSES) of the Australian National University (ANU), where I sit to write this manuscript, is just five metres and two doors from the *Ringwood Room*, one of four seminar rooms named after former directors of the RSES. I had several brushes with the great Ted Ringwood without ever working with him. The first, noted above, was when I applied to do a PhD at the ANU and he re-directed me to Tasmania. Later, I applied for a post-doc position to work in his multi-anvil lab in 1986, but not being Japanese, I had poor cards.

I had several longer chats with him when he was a regular visitor to the Max Planck Institute for Chemistry in Mainz; there weren't that many people around who would write on such all encompassing subjects as the formation, origin and composition of the Earth, Moon and terrestrial planets, and so they tended to congregate in Mainz. Bill McDonough was also stationed in Mainz in the Geochemistry Division at the

time. One occasion was a rather Austrian Christmas party in the Cosmochemistry Division's kitchen, where the two ex-pat Australians sat and discussed things nobody else could relate to, such as Australian Rules football (Ted was an avid Hawthorn supporter, in case you were wondering… or that means anything to you).

We met again at the *1992 Goldschmidt* conference in Reston, Virginia, where Ted had a room on the same floor of the hotel as Dorrit and I. My "breakfast exams" followed: the three of us met at breakfast each morning, where Ted grilled me on his chosen subject for the day. It was interesting to share ideas with such a famously audacious thinker, but I had to have my wits about me. There was never any doubt in Ted's mind that he knew the answers. On the subject of kimberlite origins, I remember him saying "*well, I think you're wrong there, Steve. As a matter of fact, I* know *you're wrong!*" On the last day we didn't see him at breakfast time but met him later in the day at the conference, where he smiled at Dorrit and asked: "*Did you have a nice breakfast?*". He was fully aware of what he was doing to me.

I initially had two years funding from the Max Planck Gesellschaft (Society), but I stayed for four, with the latter two funded by a project submitted to the German Science Foundation (Deutsche Forschungsgemeinschaft = DFG). However, the continuation in 1989 was rejected, introducing me for the first time to the capricious recommendations of grant proposal reviewers: a crystallographer thought I should do what *he* would have chosen to do (synthesis experiments on mineral end members) instead of what I had proposed to do (the role of titanate minerals in subduction processes), since he didn't consider island arc processes worth investigating! The rejection came just in time to threaten me with unemployment, causing the cancellation of our planned honeymoon.

I became increasingly exasperated at looking for a job (just as many early career researchers are these days; nothing has changed here!), and, with the UK seemingly no longer interested in me (see Text Box 2.1), I finally took the plunge into a new lake without knowing how deep it was – the German university system. I accepted a position at the University of Göttingen as assistant to Matthias Rosenhauer, who would become one of my most important mentors.

2.5 Göttingen – University in the West

In November 1990 I began what I expected would be a short foray into the German university system as Wissenschaftlicher Assistent. I eventually left Germany more than 22 years later! Life was very different in this parallel universe to the Max Planck Society, which is a federally funded pure research system. Initially, I seemed to be one of very few foreigners in German Earth Sciences (it is a very different picture today),

Göttingen is a charming small town in the centre of Germany with many old half-timbered buildings. I knew of the University of Göttingen from my interest in the philosophy and history of science as a long standing centre of mathematics. Carl Friedrich Gauss, Bernhard Riemann and David Hilbert were

all from Göttingen, and I learned here that Riemannian geometry, which I believe underlies relativity theory, was first laid out in Riemann's Habilitation thesis (see Text Box 2.3). Physics also has an illustrious past in Göttingen: it was on the wooded hills to the east of the town where Werner Heisenberg first met Niels Bohr in 1922 (Heisenberg, 1973). Göttingen was proud of having more Nobel laureates than any other city in Europe, but this didn't seem to be widely known: at one of the early European Geoscience Union conferences I was asked by a Cambridge scientist who saw I was from the University of Göttingen: "*Which city is that in*?"

The long academic history of the University of Göttingen was accompanied by some unusual academic traditions. All exams, whether undergraduate or PhD, consisted of oral exams in 3–4 subjects. After completion of the PhD exam at the Examinations Office in the centre of town, successful candidates are presented with a hat made by their fellow students, sat in a rickety cart and pulled to the Gänseliesel, a statue of a goose girl in the middle of town. The newly minted Doktor then has to climb the statue, present the Gänseliesel with flowers, and give her a kiss on the cheek (yes, the girls too; Fig. 2.7). The hurdles differ depending on the time of year: in summer there is a moat of water to cross, whereas in winter I know of a case where lips stuck to the Gänseliesel's cheek at –13 °C!



Figure 2.7 Climbing the Gänseliesel immediately after graduation, in this case Dr. rer. nat. Dorrit Jacob, later Director of the Research School of Earth Sciences at the Australian National University.

The structure of the Earth Science departments at Göttingen seemed to be leftovers of former magnificence and/or planning mistakes. Mineralogy alone was subdivided into four separate departments (why?) and in the Mineralogisch-Petrologisches Institut where I worked there were five professors but only one

Assistent (me), instead of the normal 1:1 ratio. Attitudes were old fashioned and seemed to be governed by the ghost of Helmut Winkler, a phenomenally famous metamorphic petrologist, who had retired about 15 years earlier and had run a large research centre. He was still haunting the place. The relicts of the Centre were evident in a large experimental lab which had had housed more than 80 hydrothermal bombs, *three* of which were now functional, two piston cylinder apparatuses, and various small bits of former belt apparatus that had never worked properly. It was Winkler who introduced experimental petrology to Germany. Three of the five professors at the time of my arrival were Winkler's former staff, and, probably mostly due to its now top heavy personnel structure, the place seemed pretty sleepy. This was a useful lesson in succession planning for later in my career!

I managed to complete my Habilitation in just three years (Text Box 2.3) and spent a lot of time correcting manuscripts for my German colleagues. However, more than once I received reviews of my own manuscript submissions which suggested that I should show my work to a native English speaker before resubmission! Had my command of the English language not improved as much as I thought since moving south from home at 18, or do some reviewers base their opinions on the author's address?

Field work during this period included visits to Siberia and the Alps, the latter due to the very independent PhD student Thomas Zack. I visited Siberia twice in 1989 and 1993, two years either side of the dissolution of the Soviet Union, and the contrast in the town of Aldan was remarkable. Aldan had vastly expanded markets and colour by 1993, but the logistics of fieldwork had become much more difficult, with the seemingly arbitrary wishes of *new businessmen* to negotiate. We flew into Inagli on a huge military helicopter in 1989 but were left to limp along some barely passable roads in 1993. The student field assistants on the 1993 trip were Konstantin Litasov and Vlad Malkovets, who have both gone on to notable careers.

In the Göttingen years, beginning in 1992, I rekindled my connections with the Memorial University of Newfoundland, where most of the forecast insoluble interferences with ICP-MS had been overcome and Henry Longerich, Simon Jackson and Brian Fryer had become pioneers in Laser-ICP-MS. This led to a number of very successful articles including the first application of Laser-ICP-MS to high pressure experiments (Jenner *et al.*, 1993).

It was in Göttingen that I started my 11 year stint as Chief Editor of Lithos. This took up much of my time and was a very different job to working with today's on-line submission systems: I had to deal with immense quantities of paper and felt to be constantly packing white courier envelopes. The journal boomed; when I stepped down in 2007, they replaced me with three editors!

Matthias Rosenhauer was my best mentor ever and his work behind the scenes made sure I was successful despite being like a wood splinter in the skin of the German university system. He died far too young in 1996 and I took his place as Acting Professor for three years before a replacement was found.

Why three years? That is how long it could take to replace a professor in the German university system for several reasons (see Text Box 2.3): (1) they didn't start looking for the first few months to save money, (2) a Professor from another university had to be included on the selection panel and the time table had to suit them, and (3) the panel had to submit a list of 3 names in order of preference to the State Minister of Education (the university did not appoint professors)[4], who would then, after a delay of his choosing (to save money), negotiate with the first on the list. If the first person was just feathering their nest at home, then it would proceed to Number 2, and so on.

Text Box 2.3 – The Curious Concept of Habilitation – A Foreigner's Guide

It is difficult to translate the concept of Habilitation for people in English speaking countries (and many others, for that matter). It is a qualification taken at least three years after the PhD and so is sometimes likened to a DSc but is in fact very different animal.

I had taken a job as Wissenschaftlicher Assistent, which is a temporary position assigned directly to a single professor, with responsibilities to teach whichever part of his teaching load he chooses (I say he because there was famously only one female professor in German Mineralogy at the time). The position was for three years but could be extended for another three as long as the Habilitation was completed in the first three. The Habilitation is a further research thesis based on research done since the PhD and is a qualification needed to apply for full professorships. There was discussion at the time as to whether I needed to do this, being a foreigner, but to do it seemed safer than to fall at the last hurdle on a technicality. The Habilitation process consists of a thesis and a public lecture on a subject unrelated to the thesis contents, because this is basically an exam to see if the candidate is capable of teaching as a professor. The lecture was to be 45 minutes and the tolerance seemed to be ±1 minute, which was the most important criterion for some listeners! Mine was about whether plate tectonics operated in the Archean, which I had to read up on for this lecture: this opened a new avenue of interest for me, eventually resulting in some of my best known work (*e.g.*, Foley *et al.*, 2002a).

For my Habilitation, I was more fortunate than many in having four years of research from the Max Planck Institute behind me. I pinned together several papers and visited the Dean, who had to approve the submission: he was concerned that it was in English and asked me to write a German summary (theses in English were not normal at the time). On my next visit, he recommended using *only* the German summary and leaving out all that English nonsense! Personally, he claimed not to mind, but there may be some of his other colleagues who might not approve! This is the story of my only full length German language paper; I had written it and so I submitted it to a journal. The resulting degree was the Venia Legendi in Mineralogy, which includes petrology and geochemistry, but *definitely not* geology, and so my first three degrees in geology were now essentially negated.

4. Some universities have modernised: the university president is now allowed to make professorial appointments!

Assistent positions and their continuation Privatdozent (after the Habilitation) are strictly time limited contract positions, so the pressure is on to find a professorship, which is a *permanent* position. The concept of the German professor, at least at the time, was one of near untouchability. I am reminded of Morris Kline's description of German mathematicians in the 1800s, which certainly gelled with the navel gazing views of some professors as late as the 1990s: "*The intuitions of great men are sounder than the deductive demonstrations of mediocraties*" (Kline, 1980). After I had moved to Australia, a leader of a university research office (not Macquarie) exclaimed on hearing I had been a German professor: "*Oh, so you were a god!*" But apart from a precious few who somehow thwart the system, it is impossible to become a professor at the university in which you currently work, which is the reason I did not even apply for the position I had covered for 3 years in Göttingen.

Prior to my Habilitation, I had already toured Germany giving lots of seminars, but now I would be part of the professorial circus which is a group of young wannabes who apply for professorships and meet periodically to give lectures whenever interviews take place. These are public lectures so everybody knows who is involved. Each time one would "get off the bus" and the others would continue on to the next stop. Most of my circus cohort eventually became established professors.

I spent *seven years* on this circuit before finally being appointed professor just a few months before hitting the immovable wall of unemployment. It was demoralising, perhaps because I had been included on lists at 2nd and 3rd place for my first two applications *before* my Habilitation. Most annoying were instrument-specific appointments, where they were looking for a professor to run an existing lab as if a technician, and twice being told that although I was the best candidate for a C3 position they would not appoint me because they considered me to be C4 material (C4 was the more senior of the two professorial levels). They had decided for me that I would move on after a year or two!

The legacy of the Göttingen years was very useful later; having been at all academic levels there I now thought I understood how the system works, which helped me greatly in Greifswald and Mainz later on.

Private life in the Göttingen area was wonderful: we lived in a rural area, initially in a place with a population of 17, and then renovating an old farmhouse, looking after horses, and then two boys of pre-school age. We would have loved to stay, but we couldn't – the system wouldn't allow it! You have to *move* to become a professor with a permanent position in Germany. I think the reason behind this is that most students had not moved at all until that stage in life, so the rule was to prevent universities becoming totally insular.

2.6 Greifswald – University in the East

Whilst at the Max Planck Institute in 1989 I experienced the fall of the Berlin wall. It was exciting to watch on the TV, but it really hit home two days later when we went into the centre of Mainz to find it full to the brim with light blue Trabant cars. The next year we moved to Göttingen, where our farmhouse was only

5 km from the former border between the two Germanies. Here we experienced how towns such as Heiligenstadt just east of the border were renovated almost completely within a few years after 1989.

In 2000 I moved as professor to Greifswald with my family, a small university in a small town on the Baltic coast known mostly, if at all, for medicine. Here, 11 years after the fall of the wall, the renovation wave had not yet arrived; buildings in Greifswald were still sporting bullet holes from the second world war. The Geology Department had just six professors and a handful of students each year – almost exclusively locals. I was the third Wessie (a common label for West Germans at the time), so now the scales were balanced at 3:3, with some of the Ossies (East Germans) not approving of this influx of foreign opinions and values. Previously, professors had always come from Berlin (the other two new Wessies were also from Berlin!) and it was difficult to avert their collective eyes from the Master City! Those with these opinions dressed in light grey, tone in tone with the walls of the dilapidated university building in which we were housed. The meeting of the two cultures is ably illustrated by the sculpture in Figure 2.8, which is in Duderstadt close to the border near Göttingen, showing that feelings of mismatch were felt by both sides.



Figure 2.8 The *Wiedervereinigungsbrunnen* in Duderstadt (by Karl-Henning Seeman, erected 1994) beautifully expresses the wishes and problems associated with the reunification of Germany in 1990. Westerners and Easterners mostly wanted to unite but ran into considerable cultural problems – the invisible wall – in doing so.

We spent much of our time clearing out rooms and corridors for renovations in our few years in Greifswald; I occupied four offices one after the other, averaging about 9 months in each. In the German Democratic Republic, you never threw anything away, broken or not (because you would not be able to replace it), and so the cellars and attics were full of the most impressive collections of materials

from decades ago. There was a huge attic room full of laboratory glassware, but no laboratories left in which it would be used; there were about 20 petrographic microscopes with mirrors instead of inbuilt light sources, and the best collection of crystal models I have ever seen! There were many displays of rock samples, all chipped with military precision into rectangular shapes of around 15 x 12 x 5 cm to fit in uniform boxes. In the cellar was the geology club, in which you could peruse ancient books in which staff members had written essays on how Sputnik had inspired their productivity.

Here I was responsible for teaching all of mineralogy, meaning petrology, geochemistry, and crystallography, which was very good for my general education. There were often depressingly few students though. Despite the distracting renovations and reorganisations, we managed to acquire the department's first Laser-ICP-MS system (Jacob, 2006), which had to be negotiated up the stairs to the top floor as there was no lift. We also produced what I am told were the university's first ever publications in both *Science* (Jacob *et al.*, 2000) and *Nature* (Foley *et al.*, 2002a), but I doubt that anyone there remembers that now. There were other positives at Greifswald, particularly the natural environment and kids' birthday parties on the beach, which was just 500m from home. But all in all, it was cut off from the international sphere and poorly accessible: a day's meeting at the DFG in Bonn for me was a three day outing because of the poor travel connections. Some friends that said they would visit us when the autobahn was finished; we left before it was!

The state of Mecklenburg-Vorpommern, in which Greifswald lies, was never one of the richest areas of Germany, and in 2002 it was announced that the university had to cut 158 academic positions – a considerable number in view of the fact that the whole university had only 7,000 students. I was present at a Faculty meeting where the Dean, physicist Klaus Fesser, announced that one of those to go would be mine, because I was moving to Mainz! This is a perfect illustration of how rumours and under the counter information ran the place, because at the time this public announcement was made, I had not been offered a job in Mainz! We were pleased to move back to Mainz, especially Dorrit, who, despite funding her own position with grants, had been told in no uncertain terms in her first year in Greifswald by the Head of Department that she was the "fifth wheel on the cart" and wouldn't receive Departmental support for anything.

2.7 Mainz and Geocycles

I returned to Mainz in 2003, but to the Johannes Gutenberg University (JGU) and not the Max Planck Institute for Chemistry. These were very different establishments, although housed in neighbouring buildings on the JGU campus. German university professors were like freelance researchers and teachers who all happened to have offices in the same building; there was often little common interest to knit them together. Professors had to cover a certain area of teaching

but were otherwise left to themselves and productivity differed accordingly. Some even redefined their own areas of responsibility to suit their research interests. The position of Department Head rotated on a two year cycle and had virtually no power; most viewed it as a task to be avoided (the title Geschäftsführender Leiter translates closer to business leader than to Head of Department; any translation you look up attributes too much power to it!). It was far removed from the modern pseudo-corporate set ups at Australian universities (see Section 2.8).

There were a few close contacts with Max Planck's Geochemistry Division, less so with Cosmochemistry, and stronger, but entirely separate, ones between the Max Planck Atmospheric Chemistry Division and JGU's Physics of the Atmosphere. I arrived just at the right time to take advantage of these links and to benefit from the Federal Government's *Excellence Initiative* even well before it started. The Excellence Initiative was invented to combat the impression that Germany's best universities didn't register in international university rankings, which were of growing importance to those in university leadership positions, but still not on the radar of most others in 2004.

The Federal Government, therefore, decided to begin a competition for special funding for the best 10 universities. This was slowed significantly by the fact that university funding was the responsibility of the States, of which there were 16, not 10, and so some were going to miss out! Whilst the argument about what to do about this took place, Mainz's State government (Rhineland-Palatinate), led by education minister Jurgen Zöllner, rushed to the front of the pack and invented their own funding initiative for State Excellence Clusters to pre-empt the national system. They called this *Wissen Schafft Zukunft*, which is one of the better plays on words I have seen used as a political slogan.

Text Box 2.4 – German Universities and the Max Planck Society

Max Planck Institutes are usually located on university campuses to promote research collaboration. This worked in Mainz with the links I note in Section 2.7 and between the Chemistry Department of the University of Mainz (JGU) and the Max Planck Institute for Polymer Science. But the two institutions have entirely different philosophies and structures, so that to make a joint appointment, the two incompatible systems need to be united in some way. When a Max Planck Division Director retires, the default expectation of the Max Planck Society (Max Planck Gesellschaft = MPG) is that the Division will be closed and a new one will be opened in a new, cutting edge, research area, not necessarily in the same place. Some may recall the public outrage at the closing of Mainz's Max Planck Geochemistry Division on Al Hofmann's retirement, which is logged in letters to *Nature* and the like. This is really a misunderstanding of how different the Max Planck system is to universities. Given that this happened shortly after the initiation of *Geocycles*, we tried to use a JGU professor position for the continuation, as Alfred Kröner had just retired from JGU's Geoscience Department. A joint appointment had worked for Stephan Borrmann between the Max Planck Institute and JGU's Physics of the Atmosphere a few years earlier. I met with the MPG's Science Section leader Robert

Schlögel to stake out the playing field and discuss the inconsistencies between the two systems; there was a definite let's make this work attitude on both sides, despite the incompatible systems.

A university appointment is a much more public affair. Professorships at German universities are numbered and retirement ages are fixed; a university cannot just invent a new one like in many other countries. This is the reason you will often see the retiring Professor named in job adverts; it is a specific position that they are trying to fill. This system can be an advantage for future planning, as the number and timing of future position availabilities can be seen years in advance. Disadvantages are that you can't invent an extra position, even if you have the money, and the ministry can block new appointments to save money (which is very frequently done). After a position is advertised, the number of interviewees (usually 5–7, occasionally as many as 10; this is the circus I describe in Text Box 2.3) and their names are public knowledge. In contrast, the MPG doesn't want to tolerate Number 2s because its philosophy is to take only the best, which Number 2 cannot possibly be. Head hunting is common in the MPG – also a no-no for universities. The only solution for our joint appointment was for the position to be advertised and the same person to be Number 1 for both MPG and University, as well as being good enough to pass the MPG's stringent standards. If Number 1 turns it down, the MPG system decrees that the rest of the university list is pointless; it is game over! Regrettably, after two entire rounds, we had to give it up and no joint appointment could be made, although the procedure did have the effect that two applicants were promoted to full professor elsewhere at universities that can react much more quickly than German ones! Thankfully, JGU and the Max Planck Institute for Chemistry worked together well on many other initiatives, including *Geocycles* and the interdisciplinary Max Planck Graduate Center (https://www.mpgc-mainz.de), in which PhD students have supervisors from both the JGU and MPG. This was a prototype for Centers of which there are now several around Germany (https://www.mpg.de/13833218/max-planck-graduate-center).

Using the links between the JGU and the Max Planck Institute, we teamed up with Mainz's Archaeological Museum (RGZM) and the University of Applied Sciences and submitted a proposal for an Earth System Science Centre named *Geocycles* (Fig. 2.9), which I led. After presenting this to the usual review panel and to state politicians including the Premier Kurt Beck, we were one of two successful Clusters. The other was immunology, but neither physics nor chemistry, which met with much gnashing of teeth in unsuccessful mouths, as they *knew* themselves to be the premier departments in the university! As one physics professor said to me: *"even a blind chicken sometimes finds a grain"*. The blind chickens could see the road ahead though; we motored happily through two more sets of green lights until the funding system changed in 2013.

I led the Centre with atmospheric chemist Jos Lelieveld (Max Planck Institute for Chemistry) and archaeologist Kurt Alt (Mainz University) for a very successful eight years. Unfortunately, we were not able to transform this into a national Centre despite having many more top quality researchers than successful applicants (now the blind chicken's teeth are gnashing). Different political things counted at the national level! This was a very enjoyable peak



Figure 2.9 One of the images designed (by Jochen Körner) to advertise *Geocycles*, representing the integration of geosphere, biosphere, atmosphere, hydrosphere and anthroposphere in the Earth system. This was designed years before I knew I would be moving to Sydney, chosen accidentally here to represent the anthroposphere!

of my career in organisational things: working more closely with university leadership than ever again (special mention to President Georg Krausch and Vice-President for Research Ulrich Förstermann), and meeting and discussing science organisation with state and national leaders. I found some truth in the opinion that professors work directly for the ministry: I was once shut away with several other JGU professors until midnight in Jürgen Zöllner's ministerial suite with the directive: *"don't come out until you have a plan I can use"*. We were sustained by snacks from his cupboard, but you wouldn't have wanted to be a vegetarian!

2.8 Macquarie and the CCFS

In 2013 I moved to Macquarie University in the northern suburbs of Sydney. When I joined Macquarie, its Earth Sciences star had risen mostly due to the efforts of Sue O'Reilly, and it was widely viewed as one of the leading Earth Science departments in Australia. It had the CCFS (Australian Research Council's *Centre of Excellence for Core to Crust Fluid Systems*); the only national Centre of Excellence in the Earth Sciences and one of only two across all subjects at Macquarie. It quickly became clear to me that Australian universities take much more notice of university rankings than German ones had done, and here the Earth scientists ranked far higher than any other subject in the Faculty of Science and Engineering in the subject rankings. It was one of Macquarie's few research flagships.

The CCFS was larger in geographic extent than *Geocycles* had been, with nodes in Western Australia, but it was much smaller in scientific scope. It was a Solid Earth centre with a definite lean towards the lithosphere and mineral deposits, as was everything in Australia at the time to tick the national benefit box. It was steered by an impressive international advisory board that included representatives from the mining industry, government organisations and a university president.

We moved to Macquarie when Dorrit was awarded a *Future Fellowship* by the Australian Research Council. This was one of eight awarded to Macquarie Earth scientists (four of them women) over 8 years! Another measure of the quality of the team was the award of three medals from the Australian Academy of Science between 2015 and 2017. A refreshing change from Germany was the gender balance amongst academic staff (see Text Box 2.5): one third of the staff at Macquarie were female. In Mainz I had been involved in the appointment of the *first ever* female professors in *two* departments!

Text Box 2.5 – Some Remarks on Gender and Equity in Germany and Australia

As noted in Section 2.4, Dorrit has been my partner since my post-doc days in Mainz. We have thus long experience of the "two body problem" that haunts many university scientists, and how it has varied with time and location.

We arrived at Macquarie to find that a third of the academic staff were female (recall from Text Box 2.3 that there was one female mineralogy professor in all of Germany in 1988!), and we were amazed to learn that we were the fourth husband and wife team in our department. Both would be unthinkable in the universities of many German states, even in the 2010s. I knew of several cases of partners who were both professors, but in different *States*, and we knew of cases where appointment negotiations had resulted in pairs being given positions at *different universities* in the same State (remember appointments were made by the State Minister). The idea that an appointee could negotiate a position at the same university for his/her spouse, as in common in many countries including Australia, was not possible in the German states I worked in, although I am told it did occur in at least one state. I suspect this was underlain by an undercurrent of terror that a married couple could become all powerful, which is presumably psychologically related to the above mentioned, dated concept of the all powerful professor (Text Box 2.3). It was also impossible for me to employ my partner, even on soft money, and so she had to find her own funding, which she did successfully, scratching together part-time positions on grants she wrote herself *for 25 years* including her PhD. Her first continuing position came when she was awarded a Heisenberg Professorship in 2012. In the Heisenberg Professorship system, five years' salary were paid by the granting body (DFG), after which the university had to promise to transfer the awardee to a *real* professor position. Since we left for Australia in the first year of this Heisenberg Professorship, Dorrit first drew a university-funded salary of any kind, anywhere, in 2017, following four years on her ARC Future Fellowship! I recall a national meeting organised by the DFG in 1989 to plan a new funding scheme, where the two invited French participants asked: "*where are the women?*"; only one German woman was

present in a room of about 50! Where a woman did acquire a position, she was frequently totally swamped by committee work as the *Frauenbeauftragte* – the token women's voice – which was usually ignored.

Part of the reason to move to Australia, despite the debacle with my promised position (see Text Box 2.6), was that Dorrit had followed me for long enough through three job changes (to Göttingen and Greifswald, then to Mainz): it was now time for me to take the back seat. This worked out well as she moved on to the Australian National University as Director of the Research School of Earth Sciences just before the cuts hit Macquarie. The downside to Macquarie's more flexible approach to employing women is that they were all just as easily removed when the cuts came!

The Geochemical Analysis Unit (GAU) had everything you need for geochemistry from rock to *in situ* trace element analysis, and funds seemed to flow for any recent technological development. Macquarie scientists had been pioneering in *in situ* Re-Os dating of sulfides, and *in situ* Hf isotopes in zircons, and supplied many labs around the world with the Laser-ICP-MS data reduction software *Glitter*.

The experimental facilities were older, so that I wondered if the building had been constructed around the presses: the answer, Trevor Green has told me, is yes for one, whereas the other came on its side through the door. We slowly acquired new presses; two rapid quench piston cylinders (Ezad *et al.*, 2023) and a 1000 ton multi-anvil from Voggenreiter (Fig. 2.10), although the latter spent far too long in a container at the edge of campus whilst the labs were renovated and extended.

Teaching was different to Mainz because I had much less influence in choosing what to teach. Curricula are more regimented in Australia and team teaching is common. The igneous petrology course was occupied and taught by others (but geochemists, not petrologists) so I was left with metamorphic petrology. I did very much enjoy the field excursions in much wilder places than even exist in Germany, with amusing moments when students suddenly realised they had no phone reception or internet for two weeks for the first time in their lives.

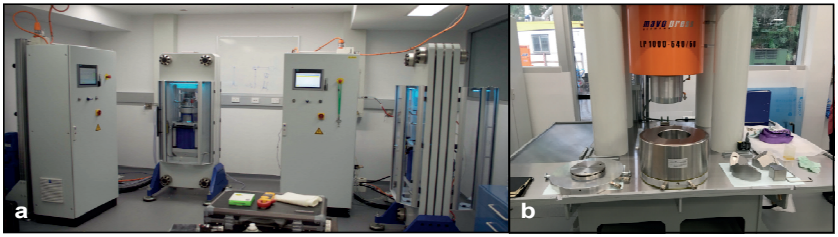


Figure 2.10 The Macquarie high pressure laboratory in 2021. **(a)** Two rapid quench piston cylinders, designed following a prototype developed by Stephan Buhre at *Geocycles*. **(b)** 1000 ton multi-anvil apparatus from Voggenreiter GmbH (Germany).

Text Box 2.6 – From GAU to Super-GAU

When I arrived at Macquarie, GAU stood for the *Geochemical Analysis Unit* where an excellent and extensive instrument park of geochemical equipment was housed and run. It was one of the main reasons that Dorrit and I decided to join Macquarie. The label GAU was somewhat amusing when we arrived from Germany, where Super-GAU is the vernacular for the worst possible disaster, based on *Größter Anzunehmender Unfall* for a nuclear meltdown. Little did I know that within a few years the one would become the other at Macquarie.

The bumpy ride for me started immediately when the new Dean refused to fund the continuing position promised to me by the previous Dean and I was left with only a four year contract covered by central funds that were supposed to be for my lab start up (meaning there was nothing left for that). The new Dean described my appointment to me as "*the worst possible outcome for Macquarie*"! Although the science was exciting and the avenues for collaboration numerous, I seemed to have arrived on a battlefield: in Mainz, *Geocycles* used the analogy for funding that "*the tide lifts all ships*" to reassure those suspicious of large grants and afraid of being left out. Here, the big ships were well afloat, but they were busy with broadside attacks on each other like in a 16th century sea battle. I can't help wondering if this was noticed at higher levels in the university and was a factor when the time for cuts came a few years later.

It is interesting to contrast the role of Deans and Department Heads in Germany and Australia, namely their impotence in Germany and power in Australia. There were three Deans in my first eight years at Macquarie, the first killed off my promised position which the university leadership had agreed to. The second tried to push the Earth Sciences in a new direction, apparently trying to counteract the power and independence that the CCFS had acquired. The third eliminated the Earth Sciences completely in 2021, despite it running a research flagship of the university. I had to learn to appreciate that the system in Australia is much more hierarchical than in Germany if those in charge wish it to be. I noted in Section 2.7 that I had worked closely with the University President and even State politicians in Mainz: after 11 years at Macquarie I can report that I exchanged more words with the Bundeskanzler in Germany than I did with my Vice Chancellor at Macquarie.

The GAU in 2014 was in effect run by the Centre of Excellence for Core to Crust Fluid Systems rather than the Department of Earth Sciences. This was a source of tension for the new Dean and Head of Department in 2015, who took control of it and drove out Norman Pearson who had managed the facility for many years. Funds were redirected away from the CCFS in a time of ill-advised expansionism: we were told to expect 6–8 new academic appointments in the next couple of years, only two of which happened. The plan for an extension to the experimental lab which I had negotiated (the new multi-anvil was too tall for the roof of the existing one) was stopped and replaced by a complete basement renovation at almost 20 times the cost, including three large wet labs that would never be used for their intended purpose. The multi-anvil was finally installed more than three years later after the renovation was completed.

Despite this turbulent period, the closure of the Earth Sciences in 2021 was really a result of national COVID lockdowns. Australia closed its borders, preventing foreign students from coming in, yet governments of the last 30 years had made sure that foreign students were the main source of income for universities! Starved of a third

of its income, Macquarie saved money by cutting any curriculum with small student numbers, whereby small meant classes of less than 50, so geology and geophysics were gone (as they would be in most places).

There are lessons here for those in other countries too: with flexibility comes insecurity. I was impressed when I arrived in Australia that expansion was easier than in Germany because you could invent new positions. But it was also possible to eliminate them: a professor is *permanent* in Germany but merely *continuing* in Australia. Academics (everywhere?) are constantly given the impression that their research records are of paramount importance, but these counted for *absolutely nothing* when the money ran out in COVID Australia. The poorly funded German system is much more resilient: Mainz twice had far fewer students than Macquarie when I was there, and yet it lives on.

The journey here and the story of Earth Sciences at Macquarie in the last few years may prompt some to wonder whether I regret the move from Mainz. Despite the turbulent events of the last few years, I do not. I joined a vibrant research department which expanded further to 22 academic staff in 2015, but after being merged into oblivion twice (see Text Box 2.6), I was one of just two remaining Earth scientists by April 2023. Macquarie was just the first: Australia has now seen at least three closures of Earth Science departments and mergers of others with geography, so that the ANU has one of few dedicated Earth Science schools left. The country lacks a national plan for universities and for subject areas within them. Each university makes its own decisions on what to cut in these difficult times, but the decisions show that they view themselves as businesses and only secondarily as educational institutions. Geoscientists are much better at national cooperation, with AuScope (https://www.auscope.org.au/) coordinating funding and access to geochemistry research hubs nationally. Perhaps it would be better if geoscientists would run the universities.

I now have an office in the Research School of Earth Sciences at the Australian National University. A real estate agent would say I have a view of Lake Burley-Griffin, meaning you can catch a glimpse of it if you squint through the trees. It is the perfect place to look back on a career and assemble the different experiences and interpretations that all lead towards mantle migmatites.

3. PRIMITIVE MANTLE DERIVED ALKALINE ROCKS

The purpose of this section is to outline the range of igneous rocks that may be derived principally or predominantly by the melting of mantle migmatite neosomes. Most primitive melts from the mantle are assumed to be partial melts of peridotite. In order to stake out which melts may owe their compositions, either partly or wholly, to melting of other rocks in the mantle, I review here those melt types that contain signatures in major or trace elements, or in isotopes, which indicate the contribution of rocks other than peridotite. This includes many alkaline rocks whose origin has been assigned at times to partial melting of peridotite under the influence of volatile components, particularly H_2O and CO_2. I outline a spectrum of primary melts of peridotite as delineated by high pressure experiments and then focus on the borderline cases that cannot be explained without volatiles present. Although many of these are strongly alkaline melts that are commonly associated with volatiles, such as kimberlites, carbonatites and lamproites, indications of hydrous ultramafic rocks in the source are present for other rocks that are normally assumed to have formed by relatively large degrees of melting of peridotite, but contain abundances of components such as TiO_2, K_2O or P_2O_5 too high to originate from peridotite alone. Examples are alkali picrites in South Africa, Greenland, eastern Siberia and Antarctica (Ellam and Cox, 1991; Litasov and Taniguchi, 2002; Larsen *et al.*, 2023) and some meimechites (Dimroth, 1970; Sobolev *et al.*, 2009). Melts such as these may begin as melts of neosomes that become diluted to various extents by continued melting to include surrounding peridotites. The undiluted initial melts of the mantle neosomes are most likely represented by alkaline rocks and, therefore, I concentrate on the primitive alkaline rocks here.

Although my undergraduate laboratory project had involved mantle xenoliths carried to the surface by the Matsoku kimberlite (Section 2), my introduction to alkaline rocks started in earnest with my move to Newfoundland to begin my MSc. This was shortly after the publication of an inspiring book that gave me just the background I needed to begin my study of alkaline rocks. This was *The Alkaline Rocks of Canada* by Currie (1976) and contained descriptions of all the known occurrences of alkaline rocks in Canada, including many fascinating ones that seem to have been ignored since, such as Castignon Lake in Quebec (Dimroth, 1970). I soon found, as does every newcomer to alkaline rocks, that the nomenclature is vast and confusing: *"although alkaline rocks comprise less than 1 per cent of igneous rocks, almost one third of all rock names are of alkaline rocks"* (Currie, 1976). I therefore offer Table 3.1 – a simplified glossary that will probably have experts screaming but should suffice to help the non-specialist. Added to this sheer volume of weird names comes the confusion that usage of many of them has changed with time: Rock (1987) gives an example where the same rock has been given five different names! I have grouped the entries in Table 3.1 according to possible primitive magma types rather than individual rock names.

Table 3.1 Simplistic summary of alkaline rock names for the non-specialist. Much longer definitions for some rock types are given in Le Maitre (2002).

Rock name	Definition, description and comments
Alkaline lamprophyre	Includes **camptonite** (Plg > Kfs), **sannaite** (Kfs > Plg), **monchiquite** (glassy). Ti-rich mafic phenocrysts typical of alkaline rocks. Some monchiquites may be chilled camptonites.
Calc-alkaline lamprophyre	Includes **minette** Kfs > Plg; Bi > Hb), **vogesite** (Kfs > Plg; Hb > Bi), **kersantite** (Plg > Kfs; Bi > Hb), **spessartite** (Plg > Kfs; Hb > Bi). Many associated with granitic rocks, may have diverse origins. Minettes are problematic – probably from at least two primary magmas, one related to lamproites.
Ultramafic lamprophyre	Includes **damtjernite**[5] (Ne, Kfs; Ol + Phl + Cpx), **alnöite** (Mel; Ol + Phl + Cpx), **aillikite** (Cb; Ol +Phl). Hydrous phenocrysts are mica in all three. Probably two primitive melt types.
Lamproite	Ultrapotassic, peralkaline rocks (low CaO, Al_2O_3 bulk rocks) with Ol, Phl phenocrysts in diverse groundmass including K richterite. May have leucite and/or sanidine.
Kamafugite	Acronym group name made up of **katungite**, **mafurite** and **ugandite**. Essential kalsilite in groundmass. IUGS terminology gives them very different names apparently denying an association.
Carbonatite	Igneous rock with >50 % modal carbonate. Usually has <15 % SiO_2 in bulk rock composition.
Kimberlite	Extremely MgO- and volatile-rich (H_2O and CO_2) rocks. Now restricted to the old Group I kimberlites; not rich in K_2O, extremely low Na_2O.
Orangeite	Previously Group II or micaceous kimberlites. Phenocrysts dominated by Phl and Ol. Some people now name them carbonate-rich lamproites.
Meimechite	Extremely MgO-rich lava type. Officially a TiO_2-rich (>1 wt. %) equivalent of komatiite (IUGS), but some are rich in alkalis.
Ouachitite	Defunct lamprophyre name: groundmass feldspathoids and carbonate.

Mineral abbreviations: Plg = plagioclase, Kfs = K feldspar, Ne = nepheline, Mel = melilite, Cb = carbonate, Bi = biotite, Hb = hornblende, Phl = phlogopite, Ol = olivine, Cpx = clinopyroxene.

5. Brøgger (1921) used the spelling damkjernite for damtjernite, which has been promulgated in several publications since, but as explained by Barth and Ramberg (1966) this was a misspelling of the place name.

This plethora of names is the despair of the petrologist or geochemist who doesn't look at alkaline rocks: common rocks such as basalts or rhyolites can be defined simply by a box on an X-Y plot, whereas rare ones such as kimberlites and lamproites have a page or more of text to define them in the IUGS system (Le Maitre, 2002). I have even witnessed exasperated laughter when using rare rock names during a seminar I gave on lamprophyres!

I begin with a short history of the alkaline rocks in an attempt to throw back the heavy curtains of nomenclature and remove some of the mystique surrounding their names. My goal is to assess the range of primary melts and not the range of obscure rock names. I then outline what we know of primary melts of peridotite in order to spotlight the types of melt that need other explanations. The range of alkaline rocks must express the variety of neosome assemblages: here I want to illuminate this variety to get a handle on the assemblages and their origins.

3.1 A Short History of Mantle Derived Alkaline Rocks

Figure 3.1 presents a timeline for the alkaline rocks that may illustrate the sources of some of the confusion for the non-specialist surrounding the many names. I am sure specialists and aficionados of specific rare alkaline rocks could (and will) take issue with many of my comments in the coming sections, but this does not concern me.

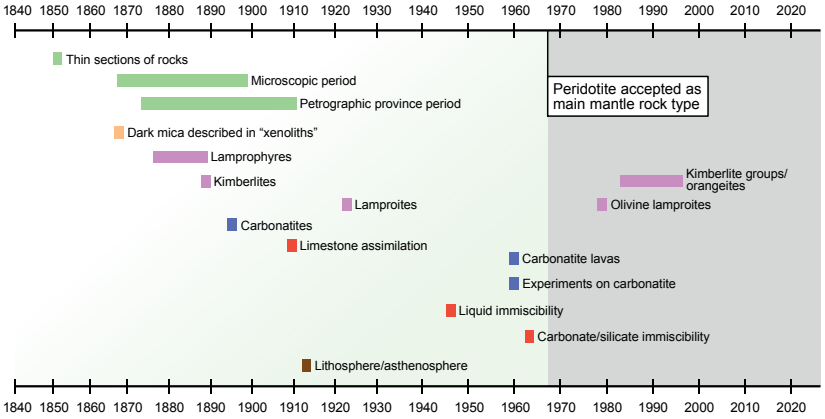


Figure 3.1 Timeline of investigations of alkaline rocks. The grey shaded area marks the period in which peridotite was accepted as the main mantle rock type (see Section 4).

The story of the history of the alkaline rocks begins at about the same time as the discovery of the first hydrous minerals in mantle rocks (Wolf, 1867), since the technology of making thin sections was spreading at this time. In his review of the development of petrography in the 19th century, Cross (1902) labelled the last third of that century as the *Microscopical Era*. In the near absence of chemical analyses, the microscope was the main instrument of investigation. Right from the beginning, the feldspathoids featured at the forefront, as Zirkel (1870), in what is recognised as the first textbook of mineralogy and petrography (Loewinson-Lessing, 1954; Young, 2003), described three groups of basaltic rocks: feldspar basalts, leucite basalts, and nepheline basalts. In these early works, the descriptions are not related to the abundance of the rocks, with alkaline rocks such as nephelinite often treated before basalts. The number of rock names grew rapidly, exacerbated by descriptions from individual localities studied in isolation. I wonder if there was a collector's effect here, with unusual rocks prized and described in detail. We can recognise it now as a sign of the times – cataloguing was a main activity in many branches of science. This new mass of rock names soon led to the concept of petrographical provinces which tried to link apparently similar rock types over thousands of kilometres. Adding to the proliferation of rock names was the largely European habit of using distinct names for rocks of a particular age. In his history of petrology, Young (2003, p. 181) claims that this died out when Rosenbusch's (1877, 1887) books went out of print, but I remember having to exorcise the concept of a melaphyre[6] out of students on my arrival in Greifswald in 2000.

The lamprophyres had a field decade from their first description in 1874 (Gümbel, 1874) with Rosenbusch (1887) listing about 40 publications about them. He was slow to catch on though, remarking in his 1877 book that, after examining Gümbel's thin sections, he could not see the justification for a lamprophyre group (Rosenbusch, 1877)[7]. Ten years later he listed many of the rock names still in use today (Table 3.1), having massively extended Gümbel's meaning of the word. One of these is alnöite, a melilite-rich ultramafic lamprophyre (UML) named after the island of Alnö in Sweden, in which the first carbonatites were described by Högbom (1895). Högbom referred to them as carbonate dykes, clearly recognising their igneous dyke form and noting that they can contain up to 20 % apatite (Högbom, 1895), now a well known feature of many carbonatites. Igneous carbonatite was recognised as such on a visit to the Fen complex in southern Norway by Brøgger and Goldschmidt (Brøgger, 1921), and the first carbonatite lavas were described from western Uganda by Holmes and Harwood (1932). Alkali carbonatite lavas were described from Oldoinyo Lengai, northern Tanzania, by Dawson (1962) – a locality which has, in my opinion, led to an over emphasis on sodic carbonatites in experiments and in petrogenetic models for

6. Melaphyre is (was) a basalt exclusively of Upper Palaeozoic age!

7. In a classic piece of correction of the past, in his 1898 book Rosenbusch attributed the naming of the lamprophyres to Gümbel (1879), thus post-dating his denial of 1877.

carbonatites. The reaction of the Oldoinyo Lengai sodic carbonatite lavas with the moist atmosphere led many authors to conclude that calcitic or dolomitic carbonatite lavas were erupted as alkali carbonatites and later altered.

The name kimberlite replaced "blue ground" in 1888, of interest only as the matrix of diamond (Lewis, 1888). I chase the evolution of confusion around kimberlite nomenclature into the 21st century below, but only in so far as relevance for the delineation of groups of magmas that may have different neosome inputs is concerned. Lamproites were known (without this name) as extremely K-rich rocks in Wyoming, USA, southeastern Spain and western Australia before Niggli invented the name lamproite for his own theoretical purposes (Niggli and Beger, 1923). These were petrological curiosities until the discovery of diamond-bearing olivine lamproites in 1979 (Jaques *et al.*, 1984), ironically in the Kimberley region of Western Australia, so that their relationships with kimberlites, the other diamond host rocks, had to be clarified.

The other events logged in Figure 3.1 concern the introduction of concepts for the origin of alkaline rocks or methods of investigation. The concept of silica undersaturation, which is common in alkaline rocks, dates from Shand (1913), about the same time as the hypothesis of limestone assimilation (Daly, 1910), which appeals to the desilication of originally SiO_2-richer parental melts. This, as well as the idea that carbonatites might be melted limestones, was largely abandoned following a series of experimental studies beginning with Wyllie and Tuttle (1960). I note as an aside here that the 1960s are often seen as the beginning of experiments on carbonate melts, although first experiments were done by Niggli (1916, 1919) more than 40 years earlier. There is often a lag of around 40 years before the general acceptance of a new concept, as for the peridotite mantle in the 1960s after it should have been clear since 1928 (Section 4). Occasional appeals to the generation of carbonatite by melting or assimilation of sedimentary limestone in the following 50 years appeared like cries in the wilderness. However, the melting of limestones during subduction to produce carbonatites has recently seen a revival (Liu *et al.*, 2015).

I chose to mark 1948 for immiscibility in magmas: this is in fact the date of its revival by Fenner (1948), who noted "*Petrologists in general seem at present to take immiscibility as impossible*". Immiscibility in simple silicate systems had been known much earlier (Greig, 1927) but dropped by most as an explanation for natural rocks after criticism by Bowen (1928). Immiscibility between silicate and carbonate melts as a mechanism for carbonatite generation was discovered by Koster van Groos and Wyllie (1963) and benefitted from a tailwind after the separation of carbonate melts from nephelinite melts appeared to explain the sodic carbonatites of Oldoinyo Lengai (Hamilton *et al.*, 1979).

3.2 Primitive and Primary Melt Types

A glance into any textbook on igneous petrology written in the last 60 years will find basalts and other mafic igneous rocks explained as the products of various degrees of partial melting of the upper mantle, which is supposed to consist, either explicitly or implicitly, of uniform peridotite. Since the 1960s, this has been bolstered by many high pressure experimental melting studies on peridotite that have delineated the origin of a wide variety of voluminous melt types, but have had difficulty in explaining rarer, more alkaline melt compositions (Section 5).

With decreasing degree of melting of peridotite at ≈100km komatiite melts at >30 % melting would give way to picrite, through olivine tholeiite at ≈15 % melting, to alkali basalt, basanite, nephelinite and finally melilitite at 5 % melting or less. The low degree *melts* are especially important here: I consider these further together with volatile-rich melts in Section 3.6. However, there are critical uncertainties in the composition of low degrees of melting of peridotite for two main reasons. Firstly, the interpretation of the results of melting experiments becomes very difficult when little melt is present, especially when they contain volatile components such as H_2O and CO_2. This prevents the formation of glass and results instead in a mass of tiny crystals (Section 5). This leads to large uncertainties in the melt compositions. Secondly, it is debated at which degree of melting the melt becomes extractable, a parameter that is not tested in experiments aimed at determining melt compositions. Theoretical models suppose that extremely low degrees of melt (<0.1 %) are mobile, particularly carbonate melts (McKenzie, 1989). However, these purely physical models neglect chemical dissolution/precipitation and deformation processes as well as the growth of crystallographic flat faces (Watson, 1982; Faul, 1997). Experiments appear to favour a minimum melt proportion of around 3 % for the extraction of silicate melts (Faul, 1997).

In terms of identifying primary melts amongst natural volcanic rocks, most textbooks agree on a few criteria for recognising primary, mantle derived melts. The term primary magma has most recently been summarised as "*the melt composition in equilibrium with its source at the time of extraction from that source*" (Foley and Pintér, 2018). This distinguishes a primary melt from a parental melt, which is the starting point for a fractionation series, and thus is frequently used for the most primitive member of a series of volcanic rocks that may not contain any primary melts. A primitive melt can have many of the same characteristics, but may have been modified since leaving its source, and so differs from a primary melt.

For our purposes here, I note that there is no requirement in this definition for the source to be in the mantle, let alone from peridotite. If we restrict our attention to mantle derived melts, then we immediately run into the problem that the accepted criteria for identifying a primary melt implicitly assume that the mantle source consists of peridotite! Four such criteria are often used and

none of them are foolproof, especially when it comes to considering mantle migmatite neosomes. A strong case must depend on positive evidence from at least two of these criteria.

The first criterion – the one on which most emphasis is usually placed – is that the Mg-value should be 70 or greater. This is the case for all but 2 of 59 melts in experiments on peridotite melting by Falloon and Green (1987, 1988) and Walter (1998). I define the Mg-value here as 100*Mg/(Mg + Fe); many do not use the factor 100, but I personally prefer to deal with numbers between 0 and 100 than a range of 0 to 1. The value of 70 derives from the distribution coefficient for Mg and Fe between olivine and melt. This distribution coefficient, usually denoted $K_D^{Fe/Mg}$, has a value of around 0.30, but may vary between 0.25 and 0.38, depending on melt composition and oxygen fugacity (Ford *et al.*, 1983). Since olivine is by far the most abundant mineral in peridotite, this value should apply approximately to peridotite too and has been shown to do so in many melting experiments on peridotite.

The second criterion for primary melts is that they should contain at least 300 ppm Ni. This is also based on equilibrium between olivine and melt for which the partition coefficient[8], denoted D_{Ni}, is around 10 but can also vary greatly (4 and 40 for basaltic systems). Here again, the modal dominance of olivine in peridotite determines that the bulk partition coefficient between peridotite and melt should be around 7, so that an average peridotite should give rise to a melt with approximately 300–350 ppm Ni.

The third criterion is obvious, namely that the occurrence of peridotite xenoliths shows that the melt that carried them has come straight from the mantle. However, peridotite xenoliths also occur in melt types such as phonolites (Irving and Price, 1981) which have been interpreted as melts that fractionated within the mantle, weakening considerably the use of this criterion in isolation.

The fourth criterion is high contents of Cr: this is more problematic than Ni because the bulk partition coefficient for Cr can vary greatly depending on whether garnet or spinel is present in the residual peridotite. The Cr_2O_3 content of spinel is much higher than that of garnet and varies depending on how much melt the peridotite has previously lost. Using the Cr_2O_3 content of melts as an indicator of primary melts is thus more useful for melts from the garnet stability field.

8. I follow the recommendations of Takahashi and Irvine (1981) and Beattie *et al.* (1993) that K_D should be used to designate a distribution coefficient, which is the distribution of *two* elements between two phases, and D for the partition coefficient, which is the distribution of *one* element between two phases. Considerable confusion can arise from the frequently heard vernacular usage of K_Ds for partition coefficients, which runs against these recommendations. An alternative distinction between K_D for major elements and K_d for trace elements also finds no consistent usage. I do not find it helpful that the Geochemical Earth Reference Model (GERM) website maintains a database entitled *Partition Coefficient (Kd) Database*, despite its originator being one of the authors of the Beattie *et al.* (1993) paper!

Primary melts of mantle neosomes need not obey these criteria, as olivine is generally not present. Where these occur as discrete dykes they are likely to vary in Mg-value as the melt fractionates during flow along a channel (Section 7). Since the first crystallising phases (phlogopite, amphibole, clinopyroxene) have broadly similar $K_D^{Fe/Mg}$ to olivine (Mengel and Green, 1989), the Mg-value of the melt will decrease as it flows and crystals growing from the edges of distal parts of the dyke will have lower Mg-values. The Mg-value may also decrease towards the centre of the dyke, so that the bulk Mg-value can be much lower than that of a peridotite (Best, 1970; Hearn, 1994). It has recently become clear that phlogopite and amphiboles may have large partition coefficients for first row transition elements, including Ni (Ezad *et al.*, 2024a, 2025). I will return to this in Section 8.

3.3 Ruminations on Rock Classification

No one disputes that there is an essential difference between day and night. Nevertheless, we have difficulty specifying exactly when the night begins and the day ends. An exact demarcation is not possible.

Dürr (1988)

Your body does not eliminate poisons by knowing their names.

Watts (1951)

When I was working on my review of ultrapotassic rocks (Foley *et al.*, 1987) there appeared to be no systematic groupings of the ultrapotassic rocks, nor even an accepted definition of what constitutes ultrapotassic. In the years following the publication of this paper I was approached regularly for discussions on rock classification, mostly by people with very animated opinions. The assumption that I am deeply interested in rock nomenclature has pursued me for decades; once you are classified as a classifier, there seems to be no escape. One could forgive an outsider for keeping away from the frequently changing terminology of alkaline rocks, viewing the vast literature as quicksand produced by the interminable bickering of alkaline rock specialists.

My reason for straying into this territory in this article is that concentrating too much on the use of correct names (which change with time anyway) can remove the focus from deciphering the origin of the melts. My primary interest in alkaline igneous rocks is to improve our understanding of how they are formed, and not what they are called. Copi (1978, p. 496) notes "*Classification is generally most important in the early or less developed stages of a science*"; one could portray this to reflect our current understanding of the petrogenesis of alkaline rocks!

The IUGS classification system is intended to give names to rocks, embodied in the title of Streckeisen's 1976 paper: "*To each plutonic rock its proper name*". James Lovelock put his finger on the difference with the following

comment (but about biologists, not geologists): "*I fancy that to biologists the living world is a vast set of book collections held in interconnected libraries. In this dream, the biologists are like competent librarians who devise the most intricate classification of every new library they discover, but never read the books*" (Lovelock, 1988, p. 20). The underline is mine: classification in terms of nomenclature is just a means to an end; what interests me is to open and read the books. It seemed logical that rock classification should follow the hierarchical rules of biology until the mid-nineteenth century. However, petrologists continued to follow this well into the 1900s, culminating in the terminological blind alley of Johannsen (1938) that divides rocks into numerous Families and Orders, many of which then carry the entry "no examples"!

The 1987 paper used the concept of a resemblance classification which I will extend here beyond ultrapotassic rocks to include other candidates for partial melts of either mantle neosomes or volatile-assisted peridotite. A resemblance classification depends on similarity to standard members and dissimilarity from standard non-members. In Figure 3.2 West Kimberley (Australia) and Gaussberg

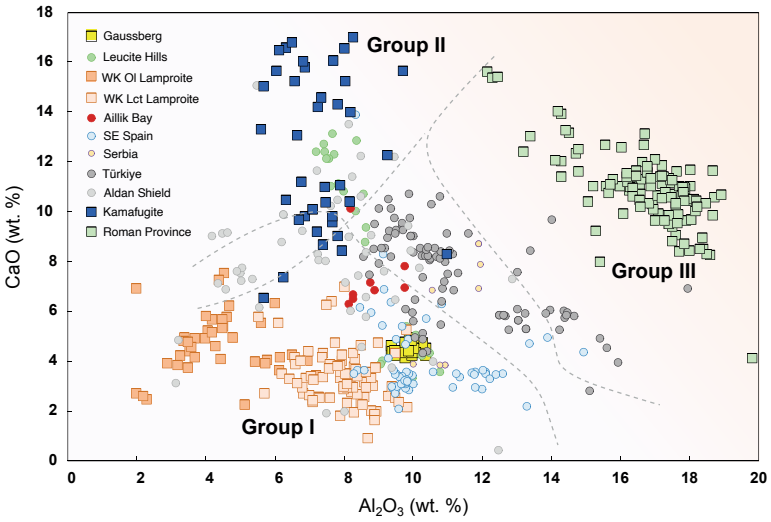


Figure 3.2 Revisit of the Al_2O_3 *vs.* CaO plot used to delimit ultrapotassic rock groups by Foley *et al.* (1987) with more recent data added. Standard members of groups are plotted as large squares (orange = West Kimberley lamproites [dark = olivine lamproites, light = leucite lamproites], yellow = Gaussberg). Standard members of the other two groups are also large squares (blue = Ugandan kamafugites; green = Roman Province leucitites). Smaller circles are ultrapotassic rocks from localities said to include lamproites (light blue = SE Spain; light green = Leucite Hills; orange = Serbia, grey = Turkey; dark red = Aillik Bay). Many newer data, mostly for orogenic lamproite localities, show transitional characteristics between the groups. Grey dotted lines are the original group delimiters from 1987.

(Antarctica) serve as the standard members of Group I (the lamproites); these standard members are plotted as the larger orange and yellow squares. The standard members for Group II (kamafugites of western Uganda) and Group III (Roman Province leucitites) are also plotted for reference.

Figure 3.2 includes reference lines, not strict boundaries, which Mitchell and Bergman (1991) considered to *"demonstrate the failure of the classification by… [establishing] a fourth group of transitional rocks"*. This seems central to Mitchell's philosophy on classification, for whom *"there can be no transitional magma types"* (Mitchell, 2020, p. 126). However, if the source regions consist of two or three source rock types, then different proportions of melts from these source rocks will logically lead to transitional melts. Others cite the original paper incorrectly as having defined three inked in regions with no transitional cases. This reminds me of the philosophy *"it is fine to draw lines, provided we do not mistake them for boundaries"* and that *"every boundary line is also a potential battle line"* (Wilber, 1988).

There is now a lot more data than was available in 1987 and there have been several experimental studies on the melting of different hydrous ultramafic rocks, so we are now in a much better position to compare positions on these diagrams with melts of different source mineral assemblages. This is discussed in Sections 6 and 7.

3.4 Kimberlites and Unrelated Rocks

The rock name kimberlite has a very chequered history. It originally meant little more than *"the matrix of the diamond"* (Lewis, 1888). However, its use expanded greatly, and it has had to shed many allegedly related rock types over the years, to a stage where it now refers to only a fraction of the rocks that were called kimberlite 60 years ago. The group definition has banished other magma types now considered to be ultramafic lamprophyres, carbonatites, or orangeites.

In my opinion, the kimberlite community itself has done the clarification of their petrogenesis a disservice by promulgating the expression "kimberlites and related rocks" for several decades. This phrase, and with it the unwarranted association of unrelated rock types, began, as far as I can ascertain, with Mitchell (1970) and appeared as book or section titles in the proceedings of several International Kimberlite Conferences, beginning with the 3rd in 1984 (Kornprobst, 1984)[9]. Although it may be legitimate to discuss the relationships, if any, between kimberlites and melilitites, lamproites, kamafugites, ultramafic

9. The 1984 conference proceedings also suffered from a publishers' aberration, common in the 1980s, in which books and some journals were printed in exceptionally ugly, inappropriately thin fonts that looked like they had come straight from a typewriter (which in some cases they had), often with inconsistent line spacing. This transferred almost all the onerous work of typesetting to the authors. Thankfully, this fashion only lasted about 10 years.

and alkaline lamprophyres, and even oceanic carbonatites, labelling them all as *related* to kimberlites may have led to confusion in some quarters. Mitchell's (1986) book on kimberlites has a chapter entitled *Kimberlites and Related Rocks* in which he distinguishes kimberlites from several types of rocks that are NOT related to them.

The separate origins of micaceous kimberlites were accepted after they were shown to have distinct Nd, Sr and Pb isotope characteristics (Smith, 1983) which saw them rebranded as Group II kimberlites. Mitchell (1995) then resurrected Wagner's (1928) name orangeite for the Group II kimberlites to prise them away from the kimberlites, but has dropped this recently, preferring to class the orangeites as a type of lamproite (Mitchell, 2020; see below).

The other major step in dismantling the kimberlite edifice was to recognise that kimberlites are not related to carbonatites. In one of the more important single contributions (a classic paper), Mitchell (1979) showed convincingly that kimberlites and carbonatites are not genetically related, and that the "central complex kimberlites" belong to the lamprophyres. One of the main locations of confusion was Alnö (Sweden): in a paper that I regard as holding the unofficial world record for self citation (23 of 27 references; 22 of them as single author), von Eckermann (1966) refers to ouachititic and carbonatitic kimberlites and claims that alnöitic magma grades into kimberlite. To confuse things further, Kresten *et al.* (1981) noted that most alnöites at Alnö do not contain melilite and so are not alnöites!

The kimberlites that remain after shedding all these other types, including orangeites, are generally accepted to be low degree melts of carbonated mantle peridotite. Many genetic models now assign kimberlite magmas to low degree melts deep in the asthenosphere that have picked up an extremely voluminous lithospheric cargo of xenoliths and xenocrysts (Giuliani *et al.*, 2023). This might make them the most likely of the rock groups considered in this article to be derived from much deeper than the lithosphere and not to involve any dykes or veins of material other than peridotite.

3.5 Lamprophyres

Lamprophyres are very diverse rocks, only some of which are relevant to discussions of origin from mantle migmatite components. I make a short survey of their history and types here in an effort to isolate the relevant ones. Table 3.1 contains a guide to the approximate meanings of the many names, which I hope will improve digestion.

Lamprophyres are usually accepted to be porphyritic dyke rocks with phenocrysts of hydrous minerals (mica or amphibole) in which felsic minerals are restricted to the groundmass. Most of the many names depend on the identities and proportions of hydrous phenocrysts and felsic minerals in the groundmass. Lamprophyres have been surrounded by controversy ever since

their inception, with some doubting that a separate group name is justified at all. This applies also to their origin, including whether many may simply be volatile-rich versions of more common magma types. Le Maitre (2002) described them as a *"diverse group of rocks that chemically cannot be separated easily from other normal igneous rocks"*.

The history of the name lamprophyre was tracked by P.J. Beger in a book usually cited as Niggli (1923)[10]. Gümbel (1874) coined the term lamprophyre for mica-rich diabase-like rocks[11] which occur as dykes and have dark mica phenocrysts and a groundmass dominated by plagioclase. Rosenbusch (1887) expanded its use into a group name for dyke rocks with either mica or amphibole phenocrysts and feldspar in the matrix. The restriction of lamprophyre to dyke rocks has been criticised because alkaline lamprophyre lavas and tuffs are known *e.g.,* Allan and Carmichael (1984). In terms of bulk composition, the predominance of mica or amphibole tends to reflect potassic or sodic affinity.

Nicholas Rock[12] was the most prominent lamprophyre scholar of the last half century, producing many summaries of four lamprophyre groups: alkaline lamprophyres, calc-alkaline (or shoshonitic) lamprophyres, ultramafic lamprophyres and lamproites (Rock, 1977, 1984, 1986, 1987, 1991). Nick was a prodigious worker with an astonishingly large vocabulary and a great interest in the statistics of rock compositions. He advocated the recognition of a lamprophyre clan, which broadened the scope of lamprophyres to include lamproites (Rock, 1981) and kimberlites (Rock, 1987), which are treated separately here. This was not endorsed by the IUGS and appeared to many like a hostile takeover; it fell into disuse after Nick's tragic death in 1993.

The threefold division into calc-alkaline, alkaline and ultramafic groups has survived (Table 3.1), although there is often confusion surrounding the coexistence of more than one group, implying gradations between them. For example, Rock (1977) notes that many sannaites have been misidentified: my own sannaites (Foley, 1984) are now referred to as damtjernites, which, in petrogenetic terms, assigns them to ultramafic instead of alkaline lamprophyres (Tappe *et al.,* 2005). The Sokli alkaline lamprophyres (Finland; Vartiainen *et al.,* 1978) are clearly ultramafic. A main reason for Nick Rock choosing to use aillikite instead of ouachitite (Text Box 3.1) was that ouachitite was originally used for alkaline lamprophyres (Rock, 1986).

---

10. This is another measure of the power of German (in this case Swiss) professors in days gone by (I hope; see also Text Box 3.3). The book – the origin of the term lamproite – is usually cited as Niggli (1923) even though P.J. Beger, a mere Privatdozent (≈ senior lecturer) wrote 2/3 of the book! I have corrected this impression in my reference list.
11. Remember that a diabase here is a German Palaeozoic basalt and not an American dolerite!
12. I have often wondered if there is something in their names that causes scientists to concentrate on rock classification, noting that Hubert de la Roche (1986) was also busy in this field at the time.

Text Box 3.1 – The History of the Rock Name Aillikite

Aillikite is now accepted as the name for ultramafic lamprophyres (UML) that lack feldspars and feldspathoids and are dominated by carbonate in the groundmass (Rock, 1986; Tappe *et al.*, 2005). But there was a rocky road to achieving this which hardly anyone knows about. I had gone to Newfoundland to study kimberlites in Labrador, but these were the central complex kimberlites that were later re-assigned to the lamprophyres.

At about this time, the IUGS Subcommission for the classification of igneous rocks had become active, eliminating many rock names from the era when petrographic descriptions resulted in a plethora of local names (see the 166 page glossary of Le Maitre, 2002). However, the suggestions for lamprophyres, carbonatites and melilitic rocks (Streckeisen, 1979) were not thought very successful, as gauged by a comment from Peter Kresten, a Swedish expert on what we now call ultramafic lamprophyres: "*I feel that the whole nomenclature of these rocks is a mess, which has not been improved by Streckeisen's suggested nomenclature*" (letter 28th April 1982).

In my M.Sc. thesis I did not refer to aillikites, but this changed with the write up for publication by early 1983. The rock name aillikite had previously been used by Kranck (1953) who added: "*they have in earlier reports been called aillikites, which name, however, should be regarded as strictly local, and not representing a new rock type.*" I begged to differ: this name seemed to beautifully fill a conspicuous gap in the table of lamprophyre names.

I sent the submitted manuscript to Nick Rock, knowing that he was working on a review of ultrabasic lamprophyres (older label for the UML; Rock, 1977). A satisfactory name for carbonate-rich lamprophyres was lacking, and Nick was planning to use oauchitite, but in reply to my letter was clearly yet to make a definite decision: "*I think there is much to be said for using 'aillikite'…. As you say, ouachitite is not without its problems and my main reason for using it is that both Von Eckermann and Kresten and Edelman used it for carbonate-rich melilite and feldspathoid-free lamprophyres…. I have changed my mind and think I will use 'damtjernite' for these carbonate-rich varieties*" (letter 28th July 1983). By mid-1984 I had convinced him to use aillikite (letter 13th June 1984), since the Aillik Bay rocks are the perfect standard members for this group, but he enquired "*What happened to your paper with Malpas for J Petrology? I have quoted it in this review on the assumption that it's coming out already!".* I had to explain: "*J Petrol turned it back as being too long (!) and… last I heard (about 2 months ago) it was in review [at CJES]. With a bit of luck it will be in press by now and it should certainly be out before the end of the year.*" (my letter, 24th June *1984!*). I add with some amusement now that this overlong contribution had 38 manuscript pages at a time when the Journal of Petrology commonly had papers of 50 *printed* pages or more!

My paper eventually appeared in the *December 1986* issue of *CJES* (Malpas *et al.*, 1986); this is the paper that was twice referred to as Foley and Malpas, 1986 in press (Rock, 1986, 1987). I apologised to Nick: "*You will see that it has been drastically shortened, which in itself is not a bad thing, except that I was not consulted on which bits to shorten. The description of aillikite has suffered, and is now next to non-existent, which is a shame since people may seek out this paper (they may even find it!) after reading your J. Petrology review. All mention of amphiboles is gone, plus a number of analyses you refer to.*" (my letter, 6th April 1987).

Nick's purpose with his 1986 paper was not just to review, but to simplify the terminology for the UML (now ultramafic). He summarised: "*much previous confusion seems to have arisen from the terminological gap now filled by aillikite*" (Rock, 1986, p. 163). Whilst the introduction of a new name at a time when many were being eliminated may seem tasteless, I note that Tappe *et al.* (2005) recommend just three names for UML, a considerable reduction from earlier times.

I turn now to the question of primitive lamprophyre melt compositions. Some examples of all three groups contain mantle xenoliths and so are mantle derived, although the calc-alkaline group probably also includes crustal melts related to granites. Alkaline lamprophyres may contain peridotite xenoliths (Ehrenburg, 1979) and ultramafic lamprophyres include hydrous pyroxenites at many localities, whereas lherzolites/dunites occur in alnöites (Foley *et al.*, 2006) but are rare in aillikites (Hutchinson and Frei, 2009; Röper *et al.*, 2025).

The four calc-alkaline lamprophyres in Table 3.1 have almost identical average compositions (Rock, 1984, 1991). They were assigned to two of seven lamprophyre magma types by Rock (1991) who thought them to be broadly lamproitic melts contaminated by crustal materials to different degrees. Minettes are a diverse group: some are related to granites, whereas others may be related to lamproites (Esperança and Holloway, 1987); they probably encompass both alkaline and calc-alkaline magma types.

The compositions of the alkaline lamprophyres are akin to volatile-rich alkali basalts, basanites and nephelinites. Camptonites are most like alkali basalts; some monchiquites are said to be chilled equivalents of camptonites, whereas others correspond to basanites and nephelinites (Rock, 1986).

Aillikites are carbonate-rich ultramafic lamprophyres (UML), which may grade into carbonatites within one dyke. They have no volatile-poor equivalent and are intermediate in composition between carbonatites and nephelinites; they fall into the experimentally determined miscibility gap. They are the most similar natural rocks to experimental melts of peridotite with CO_2 and H_2O at high pressures (Foley *et al.*, 2009). Rock (1991) claimed there were two distinct UML primary magmas, one aillikitic and the other basanitic to melilititic that gives rise to alnöites. Carbonatite complexes are said to have a complete gradation between alkaline and ultramafic lamprophyres (Rock, 1991); I suspect there may not be much difference between many alkaline and carbonate-poor ultramafic lamprophyres in terms of petrogenesis.

3.6 Types of Primitive Alkaline Melts

This section briefly reviews the range of mantle derived melts as background to later discussion of the source rocks they are potentially derived from. Figure 3.3 is an expanded primary melt spectrum that includes melts of anhydrous peridotite (green background), probable melts of hydrous mantle migmatite neosomes

(brown) and those which can be discussed either primarily as melts of peridotite with volatiles or as melts of mixed sources including neosomes (indicated by coloured dykes). Note that in this transitional band, melts labelled with the same names may have totally different origins. For example, nephelinites are purported to be partial melts of peridotite with water, and melilitites to be melts of peridotite with water and CO_2 (Section 5). However, we will see later that nephelinitic to melilititic melts may also be derived by melting of amphibole-rich ultramafic rocks in mantle migmatite neosomes: the melting reactions of rocks containing amphibole are dominated by melting of amphibole, logically leading to melt compositions similar to amphibole, which is Ne-normative when expressed in CIPW norm terms. The borders between the coloured bands should not be seen as solid: some picrites have alkali contents too high to be derived from peridotite alone, whereas aillikites are probably hybrid melts with a large contribution from peridotite (Tappe *et al.*, 2006). Since the melting points of neosome assemblages lie below that of peridotite (Section 6), melting will progress to include peridotite so that many melts usually attributed to peridotite alone are probably mixtures.

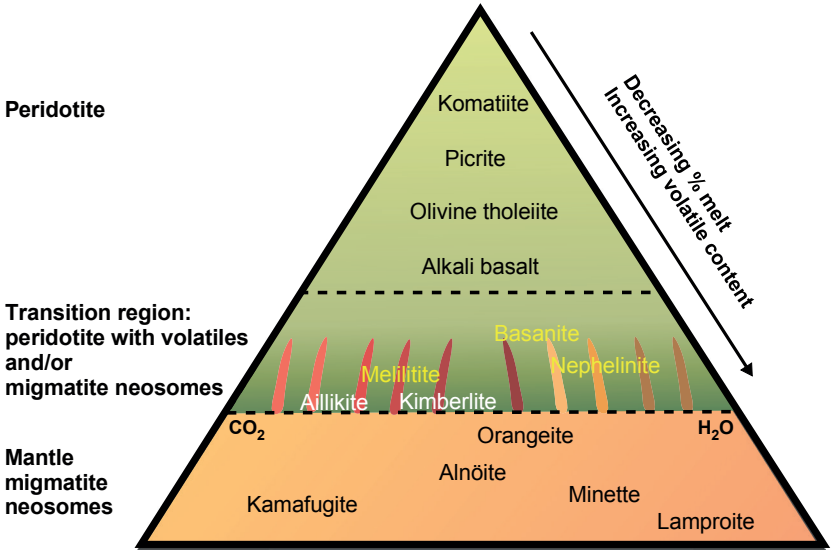


Figure 3.3 Spectrum of mantle derived melts for all mantle source rocks and volatile-bearing conditions. Melts in the green field above the dashed line can be produced from anhydrous peridotite. Melts in the brown field are predominantly from migmatite neosomes but may include peridotite melt components. Those in the transitional field have been explained as melts of peridotite in volatile-bearing conditions (darker green indicates higher volatile contents), but many may alternatively be principally melts of mantle migmatite neosomes. Melts requiring CO_2 at source are listed to the left and those requiring H_2O to the right.

The numerous melt types in the brown region of Figure 3.3 reflect the wide range of neosome assemblages. However, most experimental melting studies have concentrated on peridotite (Section 5), whereas only a small fraction of the same workload has been put into potential neosome rock types, so that assessment of their origins has only recently become possible.

Isotopic evidence for input from lithospheric mantle or recycled crustal components has accumulated for 50 years: examples for lamproites are shown in Figure 3.4, which contrasts them with basanites from oceanic areas that help to delineate the mantle array. Rocks with the lowest $^{143}Nd/^{144}Nd$ can be explained by their derivation from LREE-rich neosomes that have been stored in the lithosphere for 2,000 to 3,000 Myr, whereas $^{87}Sr/^{86}Sr$ is controlled principally by the isotopic ageing due to the high Rb/Sr of phlogopite in the source. $^{87}Sr/^{86}Sr$ values above 0.72 for Spanish lamproites are probably due to high amounts of phlogopite in the source rock, whereas the low $^{143}Nd/^{144}Nd$ lamproites of North America have $^{87}Sr/^{86}Sr$ values of <0.7061, allowing only small amounts of phlogopite (Foley and Ezad, 2024).

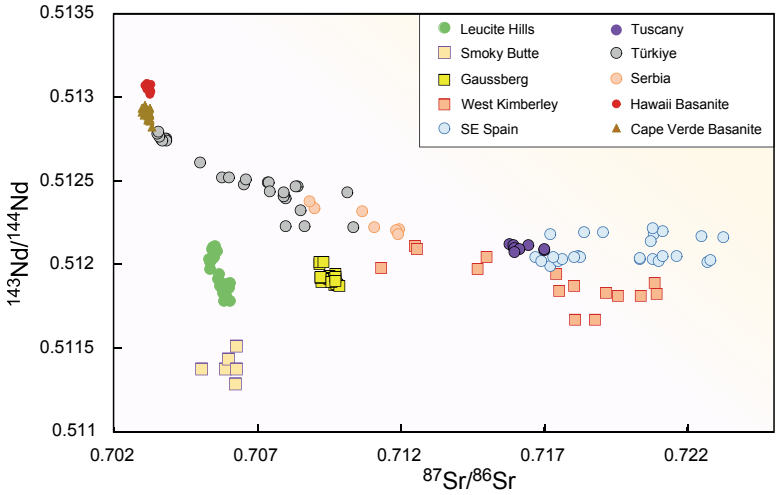


Figure 3.4 Nd-Sr isotope plot for selected lamproite localities compared to basanites from Hawaii and Cape Verde Islands. Most localities with high $^{87}Sr/^{86}Sr$ also have Pb isotopes indicating sedimentary contributions at source. The low $^{143}/^{144}Nd$ without significant $^{87}Sr/^{86}Sr$ enrichment for Leucite Hills and Smoky Butte indicates old enrichment without significant mica. Modified after Foley *et al.* (1987).

The weakness of the IUGS recommended Total Alkali – Silica (TAS) plot for many alkaline mantle derived melts is illustrated in Figure 3.5, indicating that we need other plots to investigate similarities and differences between primitive melt types. In these plots, most rock types are represented by several

symbols, which indicate average compositions for individual regions to assist legibility and prevent overcrowding of the plot by plotting all individual rock analyses. The IUGS boxes defining rock types are not helpful for about half of the symbols in Figure 3.5, as ultramafic lamprophyres, orangeites, kimberlites, melilitites, most nephelinites and kamafugites all plot in the single foidite field. This reiterates the comment of Currie (1976) that alkaline rocks are not so much rich in alkalis as poor in silica.

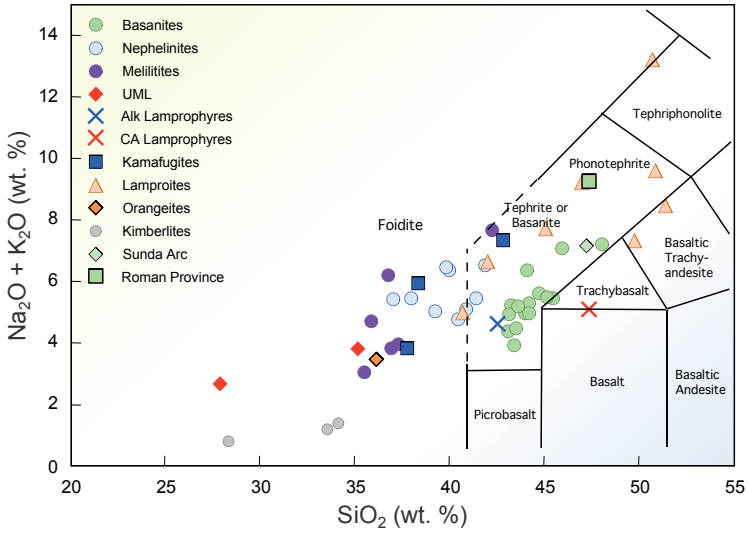


Figure 3.5 Total alkali-SiO_2 diagram showing mantle derived melts discussed here in terms of average compositions for legibility. The basanite => nephelinite => melilitite trend with decreasing SiO_2 may correspond to decreasing degree of melting of peridotite as well as increasing pressure and CO_2/H_2O at source. Many of these alkaline rocks have low SiO_2 contents lying outside in the IUGS fields for foidite. The multiple identical symbols refer to different localities, including oceanic ones for basanites, nephelinites and melilitites.

Furthermore, the TAS plot does not distinguish between sodic and potassic rock types since the two are simply added on the Y axis. For this reason, the K_2O *vs.* SiO_2 plot is often used (Fig. 3.6). This appears to distinguish a series of less potassic melts from kimberlites, through melilitites and nephelinites to basanites from a more potassic series including lamproites, kamafugites and arc volcanics. Apart from kimberlites, the former constitutes a sodic series with K_2O/Na_2O generally <0.6. The two series are distinguished clearly in Figure 3.7, which shows a dearth of primary alkaline melts with intermediate K_2O/Na_2O ratios. This duality of compositions is probably explained by the presence of phlogopite in the source of the potassic melts and calcic amphibole in the source of the sodic melts, as will become clear in Section 6.

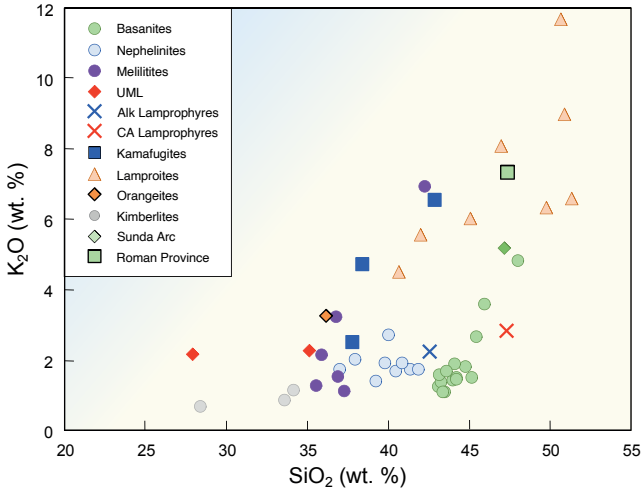


Figure 3.6 K_2O *vs.* SiO_2 plot for the same rock types and localities as in Figure 3.5. This plot differentiates between a sodic trend (for basanites, nephelinites and melilitites) and a potassic trend which diverges from around 35 wt. % SiO_2 and 2 wt. % K_2O. Note some basanites and melilitite localities have more potassic characteristics.

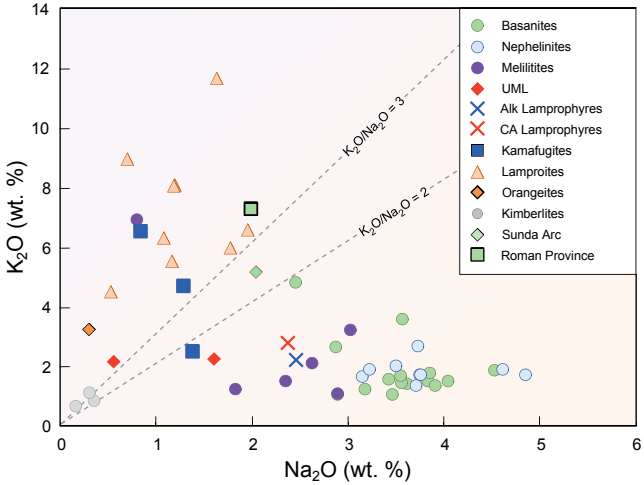


Figure 3.7 K_2O *vs.* Na_2O plot for the same rock series as in Figure 3.5 serving to de-shroud the sodic and potassic rock series that are not separated by the IUGS approved total alkali–silica plot. The sodic series are likely controlled by amphibole in the source and the potassic series by phlogopite in the source. Dotted lines are commonly used delimiters for ultrapotassic rocks.

Taken together, Figures 3.6 and 3.7 present several features of note for individual melt types. There are several basanite and melilitite localities that deviate from the overall sodic trend. The potassic melilitites are from Italy, which is also the region where the K_2O-richest kamafugites occur. The potassic basanites are from the Sunda Arc, Iran and, to a lesser extent, the western branch of the East African Rift. The average compositions for alkaline and calc-alkaline lamprophyres are close together, perhaps questioning their distinction. They clearly plot with the sodic rocks, although there is a danger that these averages may include melts with distinct origins. Kimberlites have high K_2O/Na_2O but very low total alkalis (0.77–1.38 wt. %), which is less than average picrite (2.1 wt. %), questioning the common habit of considering kimberlites to be alkaline rocks at all.

Figure 3.8 extends the CaO *vs.* Al_2O_3 plot used for ultrapotassic rocks (Fig. 3.2) to all the mantle derived alkaline rock types considered in Figure 3.5. This shows that the sodic series of basanites, nephelinites and melilitites occupy the high Al_2O_3 and CaO part of the diagram. The ultramafic lamprophyres occupy a similar position to melilitites, particularly the melilite-bearing alnöites. The higher CaO of ultramafic lamprophyres relative to kimberlites relates to their high content of carbonate.

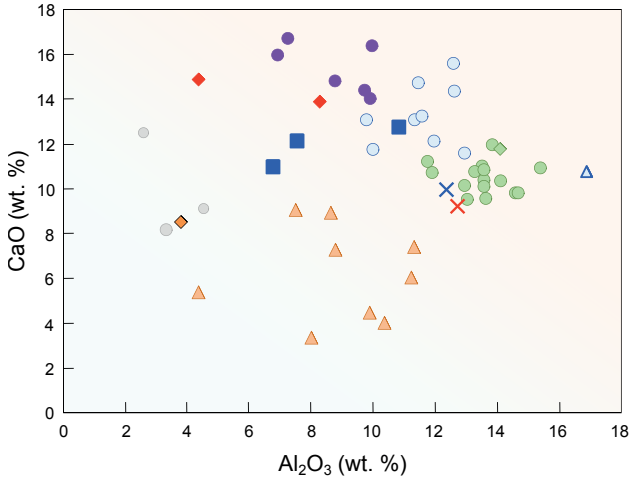


Figure 3.8 CaO *vs.* Al_2O_3 plot for the same rock series as in Figure 3.5. This plot was designed for ultrapotassic rocks (Fig. 3.2) and is used here also to sodic series. Symbols as in Figure 3.5.

The K_2O/Na_2O *vs.* SiO_2 plot (Fig. 3.9) was used by Foley *et al.* (2025) to identify mica-rich sources for melts with high K_2O/Na_2O, based on experimental melt compositions of hydrous ultramafic rocks. All melts with $K_2O/Na_2O > 1$ probably derive from phlogopite-bearing source rocks. The range in SiO_2 contents shows that a variety of phlogopite-bearing source rocks must be involved.

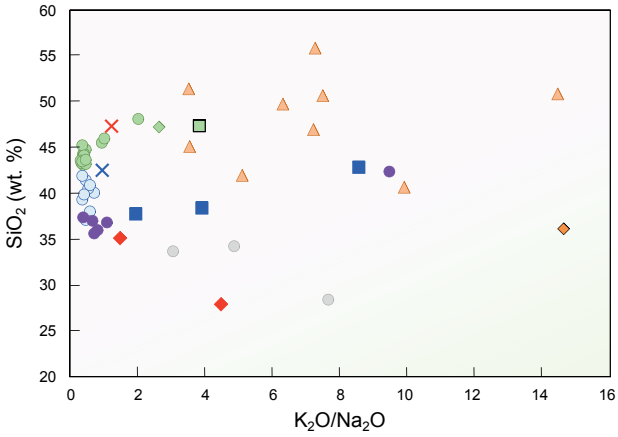


Figure 3.9 K/Na *vs.* SiO_2 plot highlighting the large range of compositions shown by potassic rocks. Melts with $K_2O/Na_2O > 1$ probably have phlogopite in the source. Symbols as in Figure 3.5.

Le Bas (1978) outlined two groups of nephelinites with distinct fractionation trends. Evolved members of the two series are similar but they originate from distinct starting points (Fig. 3.10), implying distinct parental magmas. He assigned Group I to an alkali-basalt-nephelinite association and Group 2 to a carbonatite-nephelinite association. This underlines once again that there may be multiple sources for melts hiding behind one rock name. The most CaO-rich of the nephelinite averages (Cameroon Line) plots very close to the melilitites, which are generally thought to require CO_2 in their sources (Brey, 1978), lending credence to Le Bas's hypothesis.

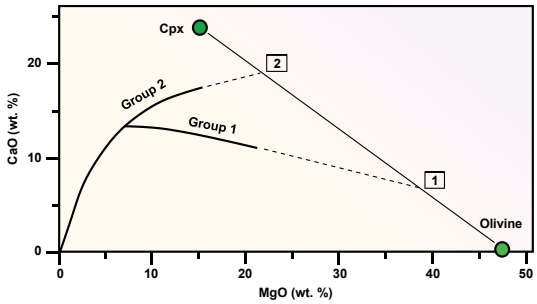


Figure 3.10 Two nephelinite groups related to distinct fractionation series identified by Le Bas (1978). Group 1 is characteristic of the alkali basalt–nephelinite association and is controlled by olivine fractionation. Group 2 melanephelinites are associated with carbonatites and controlled by clinopyroxene-rich fractionation assemblages.

Figure 3.11 shows the large variation in TiO_2 contents of primary mantle derived alkaline melts which exhibit strong regional variations. This was documented for lamproites by Foley and Wheller (1990) but is seen here to be equally true for kamafugites and to a lesser extent for more voluminous rock types including basanites and nephelinites. Many of the melts with less than 2 wt. % TiO_2 are from collisional environments, expressing the well known subduction signature: these include the orogenic lamproites as well as basanites from Iran and the Sunda Arc. The low TiO_2 kamafugite locality is the subduction-related Intra-Appenine province of Italy. This plot also brings out the strongest compositional difference between alkaline and calc-alkaline lamprophyres, in keeping with the Ti-richer compositions of clinopyroxene and phlogopite in alkaline lamprophyres (Rock, 1977, 1984).

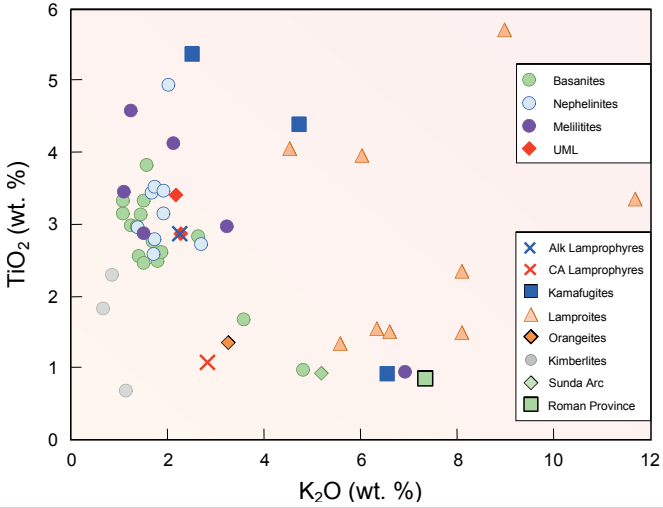


Figure 3.11 TiO_2 concentrations in mantle derived alkaline melts show a large range and are regionally characteristic, implying differences in Ti oxide mineral types and abundances in the sources.

Primary melts for the alkaline lamprophyres have been suggested to be volatile-rich equivalents of "dry" melts of peridotite corresponding to the alkali basalt to nephelinite range (Fig. 3.3). Camptonites may be the wet proxies of alkali basalts (Rock, 1977), whereas monchiquites "*correspond broadly to basanites and nephelinites*" (Rock, 1986). This does not gel with the opinion that many monchiquites are chilled heteromorphs of camptonites (Rock, 1977). Bongiolo *et al.* (2015) also considered Brazilian monchiquites to be derived from hydrated nephelinite magmas and attributed this melt to low degree melting of lherzolite. The average for calk-alkaline lamprophyres plots close to that for alkaline lamprophyres except for TiO_2 and SiO_2 (Figs. 3.5 and 3.11), which may suggest

related source compositions. A more thorough investigation of the differences and ranges would be opportune, given the data amassed since Rock's (1991) book, but is beyond the scope of this review.

Amongst the ultramafic lamprophyres, the amount of carbonate in the source appears to control the bulk compositions. As noted above, average alnöite is similar in composition to melilitites, whereas aillikites have lower SiO_2 due to grading towards carbonatite, even in a single dyke (Malpas *et al.*, 1986; Tappe *et al.*, 2006). Aillikite is the igneous rock closest in composition to partial melts of peridotite with both CO_2 and H_2O at depths of 120–180 km (Foley *et al.*, 2009) – more so than kimberlites (Foley *et al.*, 2019) – which may suggest an origin from peridotite in the uppermost asthenosphere. However, their isotopes require additional input from lithospheric rocks (Tappe *et al.*, 2006; Zozulya *et al.*, 2020).

It is doubtful whether any carbonatites are primary mantle derived melts (Schmidt *et al.*, 2024), but they are frequently discussed as metasomatic agents within the mantle and so may be present in mantle dykes. Experiments show that carbonatite melts are present just above the solidus of carbonated peridotite (Yaxley *et al.*, 2019; Section 5), but geothermal gradients at 120–200km are well above this solidus. Instead, higher degrees of melting correspond to aillikite melts (Foley *et al.*, 2009), which are then parental to carbonatites, either by fractionation at crustal levels (Tappe *et al.*, 2009) or by reaction with peridotite as they migrate at around 150 km depth (Chen *et al.*, 2025). The origin of aillikite at depths of 150–200 km is confirmed by the occurrence of rare garnet lherzolite xenoliths (Tappe *et al.*, 2009: Röper *et al.*, 2025).

I return now to the orangeites, which started life as lamprophyric or micaceous kimberlites, then morphed into Group II kimberlites before Mitchell (1995) resurrected the name orangeite. However, this has been dropped by some authors recently in favour of naming them carbonate-rich lamproites (Pearson *et al.*, 2019). I retain orangeite, agreeing with Tappe *et al.* (2022) that there are sufficient differences between orangeites and lamproites to distinguish them. Orangeites share high K_2O/Na_2O with lamproites but the total alkalis are very much lower (average $K_2O < 3.3$ wt. %; Ngwenya and Tappe, 2021). The position on Figure 3.8 is much closer to kimberlites than to lamproites. A further distinction that is important for their petrogenesis concerns the high CO_2 content, which is an order of magnitude higher, on average, than for most lamproites and shows a strong negative correlation with SiO_2, indicating a high influence of carbonate (Fig. 3.12). The average CO_2 content for the Kaapvaal orangeites is 5.1 wt. % (Ngwenya and Tappe, 2021), whereas those for lamproites from West Kimberley (0.43 wt. %), Gaussberg (0.05 wt. %), Priestly Peak (0.11 wt. %) and the Leucite Hills (0.51 wt. %) are much lower. These low C contents were one of the reasons to hypothesise reduced conditions for lamproite sources (Foley, 1989), because only 2000 ppm C (≈ 0.7 wt. % CO_2) is soluble in silicate melts if the oxygen fugacity is too low to stabilise CO_2 (Taylor and Green, 1987). Carbonate-richer lamproites do occur, as in the Holsteinborg region of Greenland (Scott, 1981), but are not common. Use of the same name lamproite implies a similar primary melt

type, whereas the sources and melting conditions may differ substantially. This role of carbonate could raise the question as to whether the orangeites are more related to ultramafic lamprophyres than to lamproites: the main compositional differences are in the higher CaO and CO_2 and lower SiO_2 of the ultramafic lamprophyres (Figs. 3.5 and 3.8).

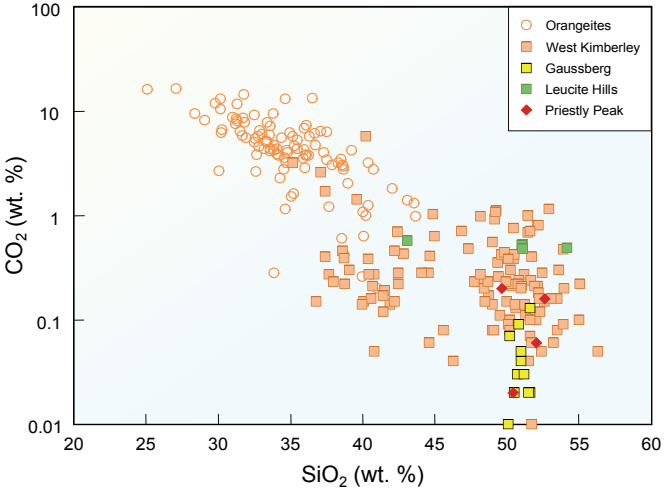


Figure 3.12 Lamproites and orangeites contrasted using CO_2 contents (note log scale). Orangeites almost exclusively contain >1 wt. % CO_2, whereas many lamproites from different localities contain considerably less CO_2. These low CO_2 contents are thought to be due to melting in a reduced environment where graphite or diamond is stable rather than CO_2.

The most insecure issue in the composition of the primary mantle derived melts is probably the abundances of volatile components in the source. Many of the expectations around volatile contents in the sources of sodic compositions, such as high H_2O in basanites and high CO_2 in melilitites, are constrained more by experiments than by analyses of the rocks. Lamprophyres are generally accepted to be volatile-rich expressions of more common rocks and yet available direct analyses yield confusing and conflicting results. For example, the average analysed H_2O content of alkaline lamprophyres is higher than the average loss on ignition, prompting the suspicion that investigations have concentrated on rocks thought to be especially rich in volatiles. Indirect indications may be of little help: Currie (1976) notes that carbonate melts need to be relatively dry to crystallise olivine, and yet aillikites and alnöites both have extremely variable olivine abundances (Rock, 1986) and often non-trivial amounts of phlogopite.

4. MANTLE ROCKS AND MANTLE METASOMATISM

Do not chiefly direct your attention to those intellectual positions which its exponents feel it necessary explicitly to defend. There will be some fundamental assumptions which adherents … unconsciously presuppose. Such assumptions appear so obvious that people do not know what they are assuming because no other way of putting things has ever occurred to them.

Whitehead (1926)

There have been many reviews summarising the mineralogy and composition of mantle peridotites from xenoliths and from peridotite massifs – blocks emplaced tectonically into the crust. Following the development of thermobarometry in the 1970s based on the compositions of coexisting minerals, it was shown that xenoliths in kimberlites have deeper origins (>150 km) than those elsewhere in the continents. These pressure-temperature estimates have been refined over the years based on a greater range of experimental conditions and an ever increasing data set of mantle rock compositions (Sudholz *et al.*, 2022). Two excellent reviews of cratonic (Harte, 1983) and non-cratonic (Menzies, 1983) xenoliths documented both the distinct original compositions and the later modification of the mantle, but the access to these reviews unfortunately suffered due to the closure of the publisher shortly afterwards. Later reviews have emphasised the evolution of the lithosphere through time, recognising the importance of strongly depleted peridotites formed as residues of extensive melting in the Archean that now make up the sub-cratonic lithosphere, but also changes between Proterozoic and Phanerozoic mantle processes (Menzies, 1990).

I do not wish to duplicate these reviews; I will concentrate here on samples that document the evidence for the mineralogical and compositional constitution of the neosomes of mantle migmatites. These range from distinct, mostly hydrous, mantle rock assemblages, through composite xenoliths that manifest the infiltration and solidification of melts in mantle veins and dykes, to new mineral assemblages disseminated within peridotites that are the products of reactions between infiltrating melts and pre-existing minerals. A persistent irritation hounding hypotheses for the origin of kimberlites, melilitites, ultramafic lamprophyres and other alkaline melts from CO_2-rich mantle has been the scarcity of direct evidence for carbonates in natural mantle samples. I follow the evolution of this discussion because of the requirement for carbonates to be involved in many melt sources.

4.1 Timeline for Mantle Petrology and Mantle Metasomatism

Ancient and rooted prejudices do often pass into principles; and those propositions which once attain the force and credit of a principle, are not only themselves, but likewise whatever is deducible from them, thought privileged from all examination.

Berkeley (1734)

When considering the mineralogical make up of the Earth's mantle it is illuminating to recap the evolution of thoughts and concepts on mantle petrology. The quote by George Berkeley has a modern expression: it takes time for old ideas to die out and for new ideas to gain acceptance. But once they do become mainstream and are entrenched in textbooks, they can be exceptionally resilient to modification, let alone dislodgement and removal. Berkeley had mathematicians in his crosshairs, not geologists (which didn't exist in name at the time), but the message is for general consumption. The principle in question for us is that mantle derived melts originate from uniform mantle peridotite, so let us remove the privilege and have a closer look at what we know about the uppermost mantle at depths where it is likely to melt.

Figure 4.1 is a timeline for mantle petrology, dating back to the first description of mica in mantle xenoliths, although I could have gone back to the coining of the term petrology by Pinkerton (1800), who spelled it petralogy. It will become apparent that newly proposed concepts may limp along for many decades before they become generally accepted. In contrast, concepts that should already be defunct can live on for decades before they slowly drift into neglect. One can find surprisingly recent statements that now look very wrong.

The book *The Evolution of the Igneous Rocks* by Norman Bowen (1928) is such a landmark publication that many modern reviews and textbooks appear to assume that the history of petrology began with it, or at least they imply this by ignoring everything that preceded it. This is reminiscent of Kuhn's (1962) note that "*Textbooks… begin by truncating the scientist's sense of his discipline's history and then proceed to supply a substitute for what they have eliminated*". There are many older literature sources, but "*they are never read save by the curious student, who has leisure and inclination to dig among the cemeteries of geological literature*" (Geikie, 1905).

It has become a "well known fact" that Bowen (1928) showed that basalts originate by partial melting of mantle peridotite. However, it is less well known that his peridotite would not qualify for the name today, as it contained 21 % plagioclase and would now be termed an olivine norite! Peridotite meant little more than "olivine-rich rock" at the time; there was no requirement for it to be ultramafic as in modern definitions. Bowen's conclusion followed only shortly after suggestions that peridotite formed the uppermost mantle layer (Williamson and Adams, 1923; Washington, 1925), including a little plagioclase based on the

mineralogy of meteorites (Washington, 1925). The alternative interpretation that the upper mantle consisted of eclogite had already been suggested (Goldschmidt, 1922) and was favoured also by Holmes (1926). Williamson and Adams (1923) concluded that the upper mantle must be above its melting point whilst nevertheless maintaining high rigidity due to the high pressure. This was echoed more famously by Daly (1926) who concluded there was a glassy basalt layer in the upper mantle.

Olivine-rich xenoliths, known at the time as bombs, included in volcanic rocks are described from the Eifel region in the 1860s by Wolf (1867) who placed their origin in the crust. Daubrée (1879) placed them below the crust on account of their high density, marking the kick-off for the concept of mantle xenoliths. Considering the divergent opinions circulating at the time, the publication by Wagner (1928) is all the more remarkable. Based on xenoliths in kimberlites from South Africa, he described the mantle as we know it today, consisting of "a bewildering variety of peridotites, dominated by lherzolites and harzburgites, which greatly outnumber pyroxenites." He also described two types of eclogites as minor lenses within peridotite, anticipating both the subdivision of eclogites into two groups as well as the small proportion of eclogite in the cratonic lithosphere.

The peridotite-eclogite question, although resolved, is a story that seems to have become mandatory for historical narratives in textbooks (I am doing it here too, but my reason is to draw attention to this). Opposing models have a tendency to live on in siloed specialisations which do not adequately consider all the available evidence. Bowen favoured peridotite over eclogite for the mantle, noting that *"geophysical data help very little"* (Bowen, 1928, p. 319). Much later, Carmichael *et al.* (1974) agreed that there was still no decision based on geophysical data alone; *"to go further we must turn to the evidence of petrology and experimental geochemistry"* (Carmichael *et al.*, 1974, p. 341). For many geophysicists, either peridotite or eclogite could explain the seismic velocities in the upper 400 km (Anderson and Bass, 1984), which perhaps explains the curious persistence of the discussion when petrologists had discounted eclogite and opted for peridotite much earlier.

In the early 1960s, the concept of pyrolite emerged at about the same time that high pressure experiments on melt sources became possible (Section 5). Ringwood's (1962) pyrolite paper formalised the peridotite mantle and he was successful in drawing the attention of geophysicists, petrologists and geochemists to it because he himself worked across all these fields. This was followed closely by details of the mineralogical make up of the peridotite mantle (Green and Ringwood, 1963), and by the first experiments on melting of peridotite (O'Hara, 1963).

Although this period firmly established peridotite as the main upper mantle rock after more than 30 years of uncertainty, concepts such as a glassy basalt layer and especially eclogite were not completely dislodged. Barth (1962) still thought peridotite to be most likely formed by pneumatolytic replacement

of ocean-floor rocks and noted "*we may tentatively revert to the hypothesis of a 'glassy basaltic sublayer'* for the mantle"! Turner and Verhoogen (1960) preferred peridotite as the main mantle rock but stated that "*the answer appears to be no*" regarding whether water depresses its melting point. As late as 1976, Yoder's textbook on the origin of basalts carried a subheading "*Garnet peridotite accepted as source – tentatively*" (Yoder, 1976, p. 42).

This reminds me of a quote from Max Planck, which I have often used in courses for graduate students on philosophy and ethics in Earth Science: "*A new scientific truth does not triumph by convincing its opponents … but rather because [they] eventually die, and a new generation grows up that is familiar with it*" (Planck, 1949). This has passed into the vernacular as "*science advances one funeral at a time*".[13] A stark geological example is the death of Abraham Gottlob Werner in 1817, which "*was felt by many… to bring a relief from the despotism which he had long exercised*" and "*greatly contributed to the progress of geology in Germany*" (Geickie, 1905).

In the 1960s, although the question of whether the upper mantle consisted of peridotite or eclogite was still debated, its homogeneity was not doubted. We learn from this brief overview that, whenever a major advance appears to be made, observations pointing in that direction were often there at least 40 years previously. This gives me cautious optimism for progress concerning the now aged concepts of mantle metasomatism and mantle heterogeneity, which have been with us in a vague state for 40–50 years now.

4.2 A Brief History of Mantle Metasomatism

The words need not be the right words: but the time, which no-one could predict beforehand, had to be the right time.

Snow (1963)

The paper by Oxburgh (1964) on amphibole in the mantle was extremely influential (especially for a relatively poorly cited paper!) as it opened the era of the hydrous mantle, mantle metasomatism and mantle heterogeneity which was very productive for about 20 years (Fig. 4.1). In it, he argued that amphibole (or alternatively phlogopite) was necessary in the mantle sources of basalts in order to explain the K_2O contents of the melts, since the recently defined mineralogy of four phase peridotite (Green and Ringwood, 1963) could not provide enough K_2O. Green and Ringwood (1963) had considered that pyrolite may consist of an olivine-amphibole assemblage, which they named ampholite (a name that never caught on), under some conditions, but thought it to be limited to the crust.

13. When I included this quote in a presentation to biologists and chemists in 2019, a discussion ensued as to who they thought ought to die next.

Oxburgh's arguments were mostly theoretical, also foreseeing the essential role of incongruent melting of amphibole, but he was able to back this up by citing known occurrences of xenoliths with amphiboles. In fact, dark mica was described almost 100 years earlier from the Eifel in Germany (Wolf, 1867), one of the two regions featured in the first paper on mantle metasomatism (Lloyd and Bailey, 1975), but its significance was not recognised. Oxburgh's paper came just at the right time and appears to have triggered the following: [1] both amphibole and mica were soon found in xenoliths, [2] high pressure experiments, which would not have been possible 10 years earlier, investigated the stability of amphiboles and mica at mantle pressures, and [3] the concept of mantle metasomatism was born. All of these occurred within 10 years of Oxburgh's paper. Amphiboles were described in lherzolites (Varne, 1970) and as megacrysts in basalts (Best, 1970), as was phlogopite (Dawson and Powell, 1969). Even K richterite was described from xenoliths consisting mostly of phlogopite that would now be referred to as glimmerites (Erlank and Finger, 1970). First experiments on the stability of calcic amphibole (Lambert and Wyllie, 1968), phlogopite (Kushiro *et al.*, 1967; Yoder and Kushiro, 1969) and K richterite (Kushiro and Erlank, 1970) at mantle pressures were conducted at about the same time, closely followed by experiments on hydrous peridotites (Green and Ringwood, 1969; Green, 1973; Section 5).

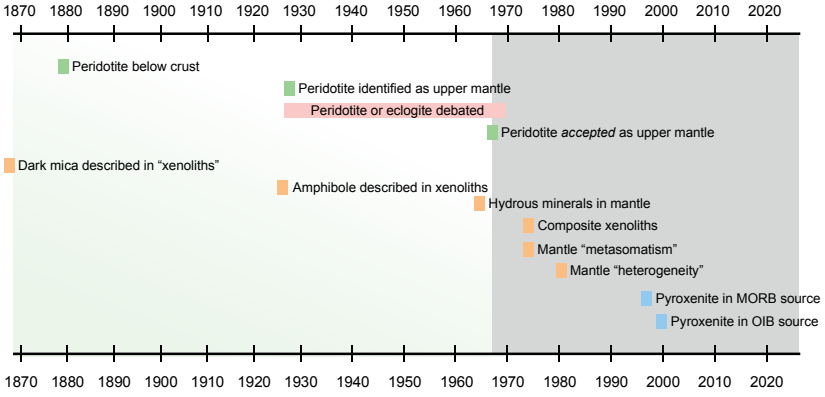


Figure 4.1 Timeline for mantle and mantle metasomatism. Grey marks the period of general acceptance of peridotite as the main upper mantle rock.

Improvements in trace element analysis and modelling brought additional advances consistent with the mantle metasomatism concept. Alkali basalts were shown to be explainable by 3–7 % partial melting of peridotite, and two components were recognised in lherzolite xenoliths: the first a residue after melt loss and the second corresponding to enrichment by a partial melt of peridotite (Frey and Green, 1974). The conclusion, shown in Figure 4.2, was that an unenriched peridotite could not give rise to the trace element patterns

typical of basanites or melilitites. Degrees of partial melting of 0.5–1 % would be needed to explain the rare earth element (REE) patterns, and this amount of melt was thought too low to be extractable from the source (Fig. 4.2; Frey *et al.*, 1978). Degrees of melting of 4–6 %, consistent with results of experiments on peridotites (Section 5), would require sources with REE patterns that were already enriched relative to chondrites or primitive mantle estimates.

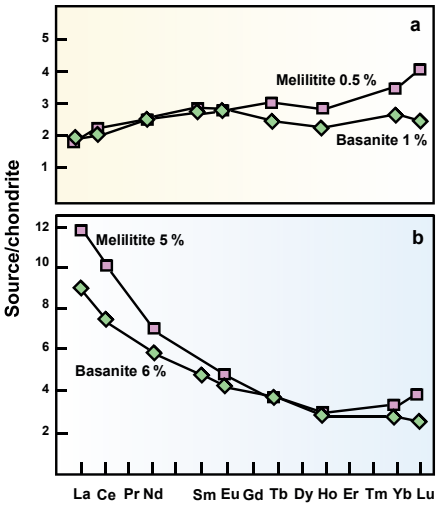
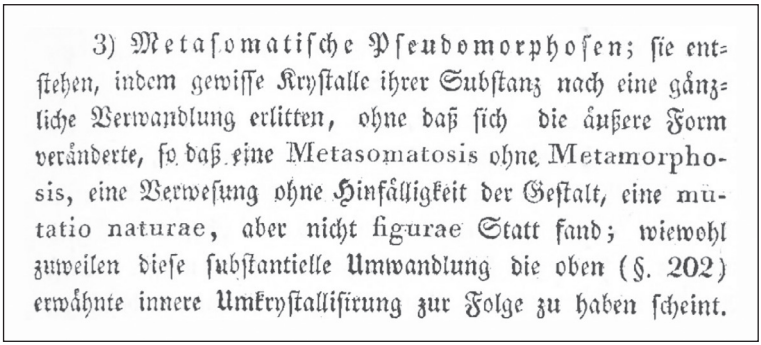


Figure 4.2 Projected rare earth element patterns of sources for basanites and olivine melilitites. Top: for an unenriched source, 0.5–1 % melting would be required, whereas 4 to 6 % melting suggests a source that is enriched in the light REE (bottom). Simplified from Frey *et al.* (1978).

The concept of mantle metasomatism was introduced at the *First International Kimberlite Conference* to explain the appearance of phlogopite, amphibole and other minerals that were interpreted to replace pre-existing peridotite minerals (Harte *et al.*, 1975; Lloyd and Bailey, 1975). Other minerals included titanomagnetite, titanite and apatite in Eifel and Ugandan samples, and rutile, ilmenite and sulfides in Kaapvaal Craton xenoliths. Soon afterwards, the interest that this sparked resulted in reports of two now well known suites, the MARID (mica-amphibole-rutile-ilmenite-diopside) suite (Dawson and Smith, 1977) and a suite termed GPP-PP-PKP that was interpreted to indicate progressive metasomatism of garnet peridotites. The latter was fully described by Erlank *et al.* (1987) but parts had appeared piecemeal in the 1970s (Erlank and Rickard, 1977). The amphibole in MARID and the K in PKP (Phlogopite-K richterite Peridotite) both refer to the alkali-amphibole K richterite.

Another point of contention (and confusion) surrounding mantle metasomatism is whether the agent of mantle metasomatism is a fluid or a melt. In the early years, it was usually assumed to be a fluid consisting largely of H_2O and perhaps CO_2 with dissolved components, particularly alkalis, and not a melt. This is perhaps partly due to the co-opting of the word *metasomatism* from crustal metamorphic processes, where it is frequently the effect of fluid infiltration. The term metasomatism crops up for the first time, to my knowledge, in the book by Carl Friedrich Naumann (1828, p. 209; Fig. 4.3), in which he notes that metasomatism would be a better word than metamorphism for a change in composition rather than form.



3) Metaſomatiſche Pſeudomorphoſen; ſie entſtehen, indem gewiſſe Kryſtalle ihrer Subſtanz nach eine gänzliche Verwandlung erlitten, ohne daß ſich die äußere Form veränderte, ſo daß eine Metasomatosis ohne Metamorphosis, eine Verweſung ohne Hinfälligkeit der Geſtalt, eine mutatio naturae, aber nicht figurae Statt fand; wiewohl zuweilen dieſe ſubſtantielle Umwandlung die oben (§. 202) erwähnte innere Umkryſtalliſirung zur Folge zu haben ſcheint.

Figure 4.3 The first appearance of the term metasomatism, used for a change in composition as opposed to metamorphism, which indicates a change in form. From Naumann (1828).

As time went by, large, banded MARID xenoliths were described (Dawson, 1987) as was a composite xenolith containing a MARID vein cutting a PKP assemblage (Waters *et al.*, 1989) and a MARID cutting a peridotite (Boyd, 1991). These made the magmatic origin of mantle metasomatism more likely.

A vein pictured by Harte *et al.* (1975) containing phlogopite, rutile and sulfides is probably the result of melt infiltration although it was interpreted as due to fluid action at the time (Fig. 4.4). If we now apply more recent geobarometers to the mineral compositions of the host peridotite in that sample, a pressure of 51 kbar results, which shows that cracks, and thus veins and dykes, can form in the deep lithosphere to a depth of at least 165 km. Similar veins consisting of phlogopite, rutile, ilmenite and sulfide were described by Harte and Gurney (1975).

Wyllie (1980) discussed whether melt infiltration should be referred to as metasomatism at all in deference to the metamorphic origins of the word, and others combined the two or explained most hydrous minerals as due to the

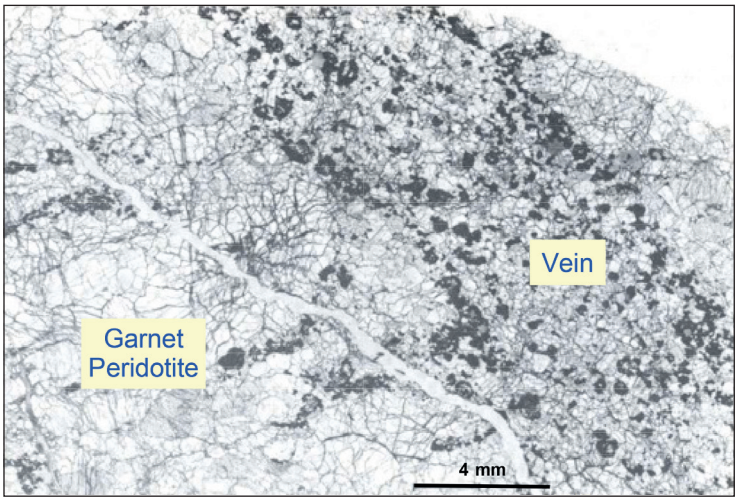


Figure 4.4 Vein rich in phlogopite, rutile and sulfides. Thermobarometric estimates in the peridotite host place this vein at 51 kbar (≈170 km). Modified from Harte *et al.* (1975).

injection of melt, but with fluids released in the distal extremities due to the build up of volatiles in the residual melt after it was largely crystallised (Wilshire, 1987; Fig. 4.5).

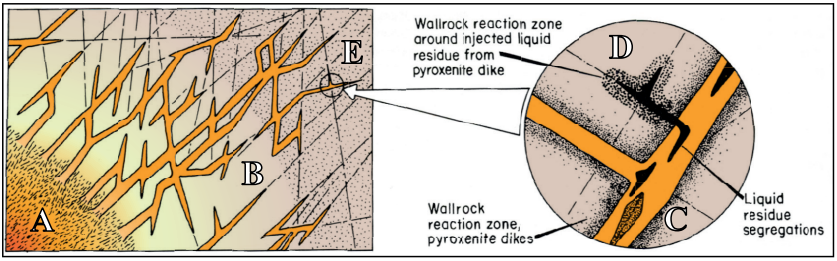


Figure 4.5 Model of mantle metasomatism assigning fluids a relatively restricted role following fractionation and extensive crystallisation of melts. Modified from Wilshire (1987).

By the 1990s, the possibility of the percolation of very small amounts of melt was generally accepted (McKenzie, 1989) and the review by Harte *et al.* (1993) considered most mantle metasomatism to be due to the infiltration of melt. We now recognise that many pockets of enriched assemblages in peridotites are probably closed, annealed segments of once continuous veins (*e.g.*, Fig. 4.6;

Rehfeldt *et al.*, 2008). This may be an expression of the explanation by Sleep (1988) of discontinuous migrating veins at the depth of transition between channelled and porous flow.

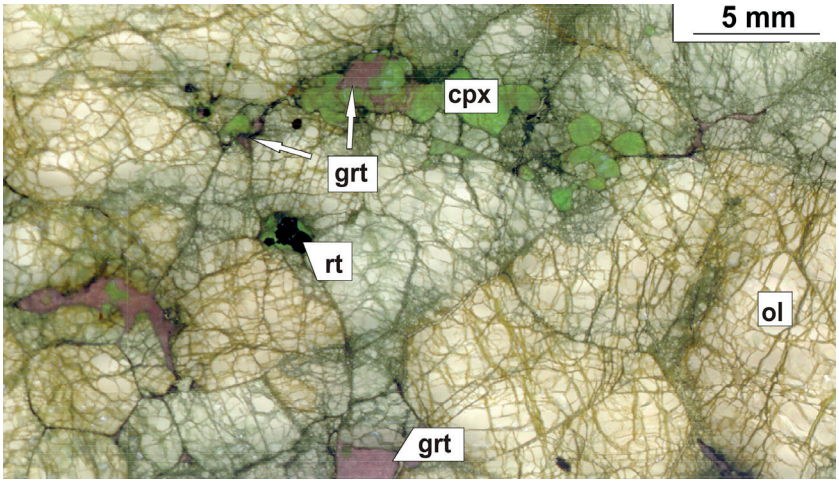


Figure 4.6 Closed vein consisting of clinopyroxene, garnet and rutile. Many elongate patches rich in Cpx and garnet are interpreted as pinched off veins resulting from melt infiltration. This may be widespread at the transition depth from channelled to porous flow. From Rehfeldt *et al.* (2008) with permission from Elsevier.

This century, we have moved on to discussions about the compositions of the infiltrating melts, which is central to the generation of the mantle migmatite neosomes, and there is more discussion of reaction zones between melts and wall rock peridotites. Potassium-rich hydrous melts such as lamproite, basanite-nephelinite or carbonate-rich melts are most commonly invoked as the metasomatic agents.

4.3 The Rock Record – Metasomatic Minerals and Rocks

Before reviewing the scope of metasomatic processes seen in xenoliths, it is instructive to consider larger samples of mantle in oceanic and continental areas, many of which are emplaced on land as ophiolites or massif peridotites. These have the major advantage of their large size, giving a spatial context much more favourable than that in xenoliths, which seldom exceed 80 cm across and are mostly much smaller. The major advantage of xenoliths is that they mostly originate from much greater depths in the mantle. Peridotites in both oceanic and continental environments have commonly experienced upward flow of melt from

below which transforms their composition and mineralogical makeup. Wehrlites and amphibole-bearing rocks occur in uplifted lithosphere at Zabargad Island in the Red Sea (Bonatti *et al.*, 1986) and dykes and veins of amphibole pyroxenites occur in the lower oceanic lithosphere either resulting from partial melts that did not escape their sources at mid-ocean ridges or from the re-solidification of melts before they escape the mantle. The latter are seen as alkaline melts and as amphiboles in ocean island xenoliths and in Petit-Spot volcanoes. I take a brief look at massif peridotites because they provide spatial context for xenoliths; mineralogical changes close to hydrous veins can be tracked away from the veins into the surrounding peridotite – information that is usually lost in xenoliths.

There are many examples of hydrous minerals in peridotites, whereas evidence for carbonates in mantle rocks is much sparser and more controversial. I focus in this section on ultramafic rocks other than peridotite, whether formed by fluid metasomatism or melt infiltration. Of most interest are discrete rock types other than peridotite, since these are potentially the neosomes of mantle migmatites whose melting leads to the alkaline rocks outlined in Section 3. These include megacrysts – generally thought to represent pieces of mantle pegmatite – which were in earlier times labelled discrete nodules.

The purpose of this overview is to stake out the known range of volatile-rich ultramafic rocks about which we need to improve our knowledge of their melting points and the melt compositions they may produce. This is essential if we are to be more specific about the currently vague concepts of mantle heterogeneity and mantle metasomatism.

4.3.1 Massif Peridotites

Dykes and veins of pyroxenite are commonly found in orogenic peridotite massifs and are commonly classified as Cr-diopside series and Al-Augite series, as at Balmuccia, Italy (Shervais, 1979). Many are anhydrous (*e.g.*, Fig. 4.7). Hydrous minerals associated with such dykes have been studied particularly at Lherz (French Pyrenees) and Finero (Italian Alps) where they occur either in peridotites adjacent to pyroxenite veins or as discrete amphibole pyroxenite and hornblendite veins (Bodinier *et al.*, 1990). Metasomatised peridotites may contain amphibole, apatite, spinel, carbonate (dolomite) and sulfide (Bodinier *et al.*, 1990; Morishita *et al.*, 2003), whereas amphibole-rich veins may have minor ilmenite, rutile and sulfides (Lorand, 1989). Bodinier *et al.* (2004) studied samples drilled in an amphibole vein and at regular intervals from the vein, finding an amphibole zone in the peridotite close to the vein and a more distal zone of apatite crystallisation. They attributed this to coexisting hydrous silicate and carbonate melts, whereby the carbonate melts originated as differentiates of the original hydrous silicate melt. Other studies have interpreted the metasomatising melt to be an alkali basaltic, basanitic, or carbonatitic melt, with the last of these invoked to explain amphibole- and apatite-rich rocks. These studies show that apatite may occur in metasomatically enriched peridotites, although the saturation

levels of P_2O_5 in basaltic melts are thought to make it unlikely that apatite could remain residual after later partial melting. Apatite can, in principle, remain stable in peridotite throughout the mantle lithosphere as long as sufficient P_2O_5 is introduced (Konzett *et al.*, 2012). The oxidation state of massif peridotites varies over several log units fO_2 (approximately FMQ -4 to +1), with more oxidised rocks associated with the introduction of amphibole by infiltrating melts. We will meet the association of oxidation with metasomatism and of carbonatite melts with amphibole and apatite in xenoliths again below.



Figure 4.7 Example of a dry vein rich in clinopyroxene cutting peridotite in the Ronda massif, southern Spain. The hammer is 55 mm long.

4.3.2 Megacrysts

Megacrysts are basically single crystal xenoliths, commonly 5–10 cm and occasionally up to 20–30 cm (Fig. 4.8). Those occurring in kimberlites have been most intensively studied but they occur also in non-cratonic continental mantle in alkali basalts and picrites. Their origin has been debated for a long time owing to the lack of textural context because of their size: they are most frequently attributed to crystallisation at high pressures from melts allied to the host magma, or from failed magmas of similar composition that never made it

out of the mantle. In the case of kimberlitic megacrysts, explanations frequently feature one of the vaguest and most misused terms in igneous petrology: protokimberlite, which appears to mean different things to different people.



Figure 4.8 Single crystal clinopyroxene megacryst from Kandidushka, Vitim plateau, Siberia. Thin rind of host basalt is visible at left. Hammer is 55 mm long.

For the purposes of relevance to mantle migmatite neosomes, megacrysts can be thought of as mechanically disrupted mantle pegmatites. Since the grain size of pegmatites is favoured by a hydrous environment of crystallisation, it is not surprising that hydrous minerals are commonly present.

In kimberlites, garnet, clinopyroxene, olivine, ilmenite and orthopyroxene are common xenocrysts. The abundance of ilmenites indicates a melt rich in Ti and Fe, which is also commonly invoked for metasomatism of peridotites and attributed to a kimberlite-like melt (Pasteris *et al.*, 1979). Minerals with similar compositions to megacryst features in polymict xenoliths, where the assemblage of phlogopite, ilmenite, rutile and sulfides is the same as in IRPS suite xenoliths (Ilmenite-Rutile-Phlogopite-Sulfide; Harte *et al.*, 1987, see below). Composite megacrysts are a rare but important source of context and document the association of garnet and ilmenite, in which phlogopite, amphibole, ilmenite, and rarely calcite occur as inclusions (Pivin *et al.*, 2009). Whereas most attribute silicate and oxide megacrysts to magmatic crystallisation (Pasteris *et al.*, 1979), others invoke reaction of similar melts with peridotitic wall rocks (Pivin *et al.*, 2009), possibly involving carbonate-rich silicate melts.

In alkaline basaltic rocks, garnet and ilmenite are less common, whereas calcic amphibole, clinopyroxene, apatite and titanomagnetite are more abundant. Amphibole is most commonly associated with clinopyroxene in alkali basalts and basanites (*e.g.*, Becker, 1977), whereas phlogopite occurs in more potassic or melilititic host magmas (*e.g.*, Johnson *et al.*, 1997). The association of apatite with amphibole is most famous in eastern Australia (Wass, 1979) but occurs also in other places such as Scotland, where it is also attributed to the interaction of carbonate melts with peridotite (Long *et al.*, 1994). Less common are sodic richterites in olivine nephelinites, which have been described from Morocco (Wagner *et al.*, 1993).

4.3.3 Xenoliths

From the enormous literature on mantle xenoliths, I will concentrate on sifting out material relevant to characterising the identity of the migmatite neosomes. It should be remembered here that the deepest neosomes are likely to be missing in the xenolith record because they are the ones that melt first and give rise to the melts that bring xenoliths from higher levels to the surface. This implies that veins (cracks) form at deeper levels than the deepest composite xenoliths known (Fig. 4.4). When I first presented my veined mantle source model at the 5th Kimberlite Conference, I ran into objections that K richterite could not be important in source regions because it is rare in xenoliths. We now know from experiments that K richterite will melt first and rapidly if it is present and so it is logical that it is rare in xenoliths, since it contributes greatly to the melts that bring xenoliths to the surface (Foley *et al.*, 1999, 2022).

Hydrous minerals (phlogopite, amphibole, apatite) commonly occur, whereas the evidence for carbonate is much harder to find. I begin with xenoliths in kimberlites from thick, cratonic lithosphere; the evidence has historically relied heavily on the Kaapvaal Craton of southern Africa and so there is a danger that our impressions of neosomes at 150–200 km depth are skewed in favour of Kaapvaal assemblages. I then consider other areas with thinner continental and oceanic lithosphere. This is a brief and selective summary that includes some of my favourite early papers: some of these may seem relatively obscure, but this serves to illustrate how common these xenolith types are and that they deserve to be taken seriously as mantle migmatite neosome assemblages.

4.3.3.1 Fe-Ti-rich assemblages: the PIC and IRPS suites

The detailed study of Kaapvaal mantle xenoliths began in earnest at approximately the same time as the International Kimberlite Conferences in the 1970s. It soon became apparent that the deep Kaapvaal mantle is represented in xenoliths by two sets of peridotites: one was garnet peridotites (lherzolites and harzburgites) that lack extensive mineralogical or geochemical signs of mantle metasomatism and the second was enriched in Fe and Ti and contains conspicuous amounts of oxide minerals (Harte, 1983). Later studies delineated subdivisions of the

Kaapvaal Fe-Ti-enriched metasomatised xenoliths and their separation from the MARID suite (Dawson and Smith, 1977). Grégoire *et al.* (2002) introduced the PIC suite, which stands for phlogopite-ilmenite-clinopyroxene and contrasted it with MARID in that it does not contain K richterite. This dualism has been used in later detailed investigations (Fitzpayne *et al.*, 2018) and is thought to indicate the type of melt that may have formed them – Group 1 kimberlite for PIC and orangeite for MARID.

Erlank *et al.* (1987) described a series of xenoliths from GP (garnet peridotite), GPP (garnet-phlogopite-peridotite through PP (phlogopite peridotite) to PKP (phlogopite-K richterite-peridotite), which they interpreted as progressive metasomatism in this order. I refer to this series here as PKP (phlogopite-K richterite-peridotite) for brevity. Harte *et al.* (1987) described the IRPS suite (Ilmenite-Rutile-Phlogopite-Sulfide), which is also enriched in Fe and Ti. The question of the origin and inter-relationships of these various groups remains to be fully resolved. I retain here the division of K richterite-free from K richterite-bearing (next section) for convenience, which does not necessarily imply lack of a common origin.

Examples of the Fe-Ti-rich suite are illustrated in samples from Matsoku and Kimberley in Figure 4.9. They are rich in phlogopite and ilmenite, the latter of which is commonly too abundant to represent full crystallisation of a melt *in situ* and rather indicates crystallisation from a passing melt. In contrast with many non-cratonic mica clinopyroxenites, these rocks are not rich in clinopyroxene.

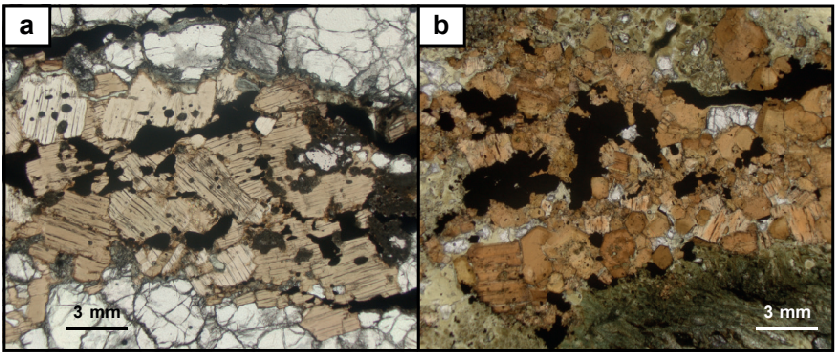


Figure 4.9 Veins cutting peridotite rich in phlogopite and ilmenite from Matsoku **(a)** and Kimberley **(b)**.

The IRPS suite appears to be common but is not often given this name, for example, the vein in Figure 4.4 has the appropriate assemblage. The same assemblage occurs in polymict xenoliths. Although described in detail from Kaapvaal, Fe-Ti-rich xenoliths occur associated with kimberlites on other

continents, such as the Arkhangelsk and Siberia and so are related to a globally relevant melt type. Fe-Ti enrichment is also widely observed in minerals from disaggregated mantle xenoliths.

4.3.3.2 The MARID and PKP suites

The PKP and MARID suites of the Kaapvaal Craton both contain the alkali amphibole K richterite, but this mineral is usually subordinate in abundance to phlogopite in both suites. Accessory zircon was later found in many MARID xenoliths and became the most efficient way of determining their age of crystallisation (Konzett *et al.*, 2000). The PKP suite was described fully by Erlank *et al.* (1987), after partial descriptions that concentrated on the occurrence of K richterite, which prompted the first experiments on the stability of K richterite (Kushiro and Erlank, 1970). Orthopyroxene occasionally occurs (Dawson and Smith, 1977). Apatite is conspicuously absent.

Regarding MARIDs, even relatively recent publications note that it is "hotly debated" whether they form by pervasive metasomatism of peridotite or are veins of magmatic material. I cannot understand this: the textural evidence from the rare, but important, composite xenoliths clearly favours a cumulate magmatic origin. There are cases of banded xenoliths up to 17 cm across (Dawson and Smith, 1977; Dawson 1987), MARID veins cutting peridotites with sharp contacts (Boyd, 1991), and MARID cutting PKP (Waters *et al.*, 1989). On the basis of Re-Os, U-Pb and Lu-Hf isotope data, Hinde *et al.* (2025) argue for an origin by reaction of a pure MARID magma and lithospheric peridotite, whereby "*their parental magmas are derived from a subduction related source*".

In contrast, PKP series rocks lack sharp contacts and are diffusely intergrown with peridotite. The most likely explanation of the relationship between them is that PKP are reaction products between MARID melts and peridotites, with the spatial context as shown in Figure 4.5 (Wilshire, 1987). This is supported by mineral chemistry of phlogopite and clinopyroxene which relates MARID and phlogopite peridotites to the same agents (Fitzpayne *et al.*, 2020).

The MARID suite is thought to be related to orangeite magmas in contra-distinction to the PIC suite which is related to kimberlite melts (Grégoire *et al.*, 2002). However, whether orangeite melts crystallise to form MARID or whether MARIDs melt to form orangeites or lamproites is debated (Waters, 1987; Giuliani *et al.*, 2015). Zircon dating has shown that MARID and PKP range in age from similar ages to the host kimberlite to ≈40 Myr older (Konzett *et al.*, 2000; Hinde *et al.*, 2025). MARID is widespread in the Kaapvaal mantle, as illustrated by its sampling in kimberlites as well as orangeites.

Orangeites are mostly a peculiarly regional magma type of the Kaapvaal Craton, as illustrated by their label "lamproite var. Kaapvaal" (Mitchell, 2020; see Section 3). However, they do occur on other continents and MARID-type rocks have been invoked for magma sources around the world, including cratonic areas in East Africa, North Atlantic Craton, India, and Finland. This makes the relationship between MARID and orangeite melts a question of global relevance.

MARID- and PKP-like assemblages in other continents are indicated by PKP-like xenoliths in Amazonia (Gervasoni *et al.*, 2022), and MARIDs in a non-cratonic setting in Morocco (Wagner *et al.*, 1996). LIMA minerals (LIndsleyite-MAthiasite; alkali-chrome titanate minerals) have been described from Brazil, Venezuela, China and Russia, which may indicate widespread PKP-like metasomatism. However, the label MARID is overused and applied to a variety of assemblages rich in mica that are not related to orangeite magmas, perhaps grading into mica clinopyroxenites and glimmerites (see below). Examples are rocks with calcic amphiboles in instead of K richterite, nevertheless referred to as MARID-type (*e.g.*, Cooper *et al.*, 2024). Regardless of their classification, these represent potential mantle neosome assemblages.

4.3.3.3 Mica clinopyroxenites

Hydrous ultramafic rocks rich in dark mica and/or calcic amphibole, usually together with clinopyroxene, form the most common hydrous dykes and veins in the mantle. They are also the most widespread hydrous minerals in peridotites and thus are the main mineralogical expression of modal mantle metasomatism. High pressure experiments on volcanic rock compositions at near-solidus conditions, although not usually planned for this purpose, show consistently that rocks rich in mica or amphibole and clinopyroxene are the expected results of crystallisation of mafic melts at upper mantle pressures (see Section 5). Mica- and amphibole-rich assemblages are here treated separately, but the two minerals commonly occur together. The stability limits of calcic amphiboles limit them to around 100 km in peridotites, but Ti- and F-rich minerals may be stable to in excess of 140 km alone or in other ultramafic assemblages. At deeper levels, phlogopite is stable with clinopyroxene to at least 260 km, and with garnet and K richterite to in excess of 300 km (Konzett *et al.*, 1997). This means it may be present throughout the lithospheric mantle, as also documented by a diamond-bearing phlogopite xenolith from Mir (Ponomarenko and Tsepin, 1987). There is a huge literature on hydrous ultramafic mantle rocks: the ones cited here are just a few examples chosen to illustrate the mineralogical and geographical scope.

Mica clinopyroxenites range in composition from glimmerites (>90 % mica) to clinopyroxenites (>90 % clinopyroxene) and are known from the mantle from all continents except Antarctica. In North America, they occur both on and off cratons: in Montana, they include glimmerites and orthopyroxene in mica pyroxenites (O'Brien *et al.*, 1991). Further east in Ontario and in the Northwest Territories potassic igneous rocks contain glimmerites with minor carbonates, amphibole, titanite, rutile, zircon and sulfides (Peterson and Le Cheminant, 1993). Mica clinopyroxenites also occur in minettes in the Mexican volcanic belt (Righter and Rosas-Eiguera, 2001).

In Asia, many pyroxenites containing both mica and calcic amphibole are well documented from alkali basalts and picrites in the Vitim Plateau east of Lake Baikal in southern Siberia and from neighbouring Mongolia. At Bartoy in southern Baikal, more than 50 % of peridotites contain disseminated amphibole

and phlogopite, and veins of amphibole and phlogopite or hydrous pyroxenites occur together with megacrysts of garnet, ilmenite and titanomagnetite (Litasov *et al.*, 2005). Further northeast on the Vitim plateau, veins include phlogopite-ilmenite and garnet-phlogopite varieties (Ashchepkov *et al.*, 2011). Phlogopite clinopyroxenites occur in basalts in northeastern China (Dai *et al.*, 2020) and further west in Shaanxi province, alkaline lamprophyres carry xenoliths of phlogopite-amphibole pyroxenites that contain minor apatite, perovskite, ilmenite and titanite (Xu *et al.*, 1999), reminiscent of the well described xenoliths in western Uganda (see below). Discrete glimmerites have been described from central Asia (Tajikstan; Lutkov *et al.*, 1989) and phlogopite and amphibole pyroxenites occur in northern Iran, India and the Arabian Peninsula.

Moving to Europe, mantle derived mica pyroxenites are widespread, occurring in ultramafic lamprophyres in the Kola Peninsula (Beard *et al.*, 1996) and in alkali basalts and picritic lamprophyres in the western Pannonian Basin in Hungary (Bali *et al.*, 2018), and in Romania (Vaselli *et al.*, 1995). The most extensively described occurrences are in the Eifel region of Germany and in Italy. In the Eifel, one of the birthplaces of mantle metasomatism, phlogopite and calcic amphibole are common in peridotites but also occur as discrete veins of phlogopite pyroxenite with and without amphibole, as mica hornblendite (Witt-Eickschen *et al.*, 1998). In Italy, phlogopite clinopyroxenites occur as veins in peridotite and as discrete xenoliths as far south as Sicily (Gianetti and Luhr, 1990).

In Africa, mica clinopyroxenites are well documented in southwestern Uganda, but also occur in Algeria (Bossière and Megartsi, 1982) and in Kenya (Kaeser *et al.*, 2006). The xenoliths of western Uganda occur in three volcanic fields and have been famous since the classic publications of Holmes and Harwood (1932, 1937). Although the label "olivine-biotite-pyroxene series" (OBP; Holmes and Harwood, 1937) has become well known, olivine is only common in the southernmost Ugandan field (Bufumbira), which is the northeastern corner of the Virunga volcanic field, most of which lies in Rwanda and Congo. The two northern fields, in which the host volcanic rocks consist of extremely silica-poor compositions from kamafugites to carbonatites, lack peridotites and carry xenoliths dominated by biotite and clinopyroxene (Muravyeva and Senin, 2018). Examples from the central Katwe-Kikorongo field are shown in Figure 4.10; these are micro-XRF composite element maps that highlight the mineralogical range. The origin of the Ugandan xenoliths has been controversial, with some favouring *in situ* metasomatic replacement of peridotites (Lloyd and Bailey, 1975). However, in our collection of 602 xenoliths from the central Katwe-Kikorongo field, olivine was found in only two. Lloyd (1981) noted that neither orthopyroxene nor garnet has been found in the xenoliths of southwest Uganda, which she interpreted as indicating complete obliteration of previously existing four phase lherzolite. The xenoliths are undoubtedly complex and some show three phases of infiltration and veining. My preferred interpretation is that most are magmatic cumulates formed in large dykes or sills in the mantle and that only a vanishingly small proportion may be modified peridotites, presumably close to the margins of the

mica pyroxenite dykes in a similar way to the relationship between MARID and PKP series. I concur with Nixon and Hornung (1973) that the OBP series concept has been over emphasised; xenoliths are either dominated by magmatic mica and clinopyroxene in the Toro Ankole fields *or* by modified peridotites in Virunga.

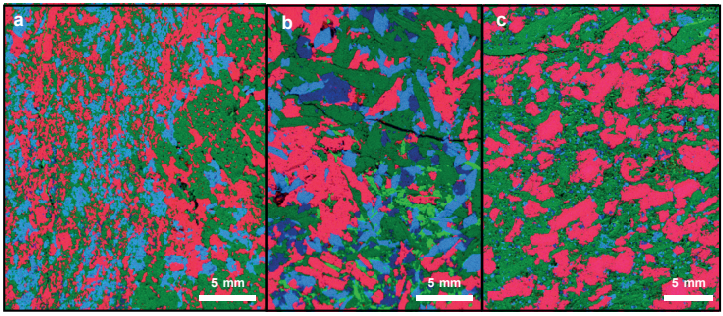


Figure 4.10 False colour micro-XRF maps of thin sections of biotite pyroxenite suite xenoliths from western Uganda. **(a)** Sample 212B with approximately equal amounts of phlogopite, clinopyroxene and titanite with very minor Ti magnetite. **(b)** Sample KX-143 contains a higher proportion of Ti magnetite and several percent apatite. **(c)** Sample 122D with perovskite, also contains apatite, but Ti magnetite and titanite are lacking. Red = phlogopite, green = clinopyroxene, light blue = titanite, dark blue = Ti magnetite, green = apatite, light blue in **(c)** = perovskite.

Mica pyroxenites also occur in intraplate regions on smaller continental blocks or in oceanic environments. These include Hawaii, where they have been implicated as source materials in the production of alkaline lavas and in seamounts to the northeast of Papua New Guinea. Lamprophyres in New Zealand contain xenoliths with mica and amphibole veins (Tulloch and Nathan, 1990).

4.3.3.4 Amphiboles and the amphibole-apatite suite

Calcic amphiboles, mostly pargasite-kaersutite, are common in xenoliths in alkali basaltic to nephelinitic and lamprophyric rocks that are erupted on thinner lithosphere (<100 km) all around the world.

Amphiboles are commonly associated with apatite in a series that was brought to our attention by many examples in eastern and southern Australia (Wass, 1979; Wass *et al.*, 1980). Associated minerals are clinopyroxene, phlogopite, Ti-rich phases including ilmenite, Ti magnetite and rutile, sulfides, carbonates and minor olivine (Wass *et al.*, 1980; Bonadiman *et al.*, 2021). In this area, amphibole and apatite are also common as disseminated phases in peridotites (O'Reilly and Griffin, 2000).

Broadly similar xenoliths with amphibole and apatite, with or without clinopyroxene and phlogopite, occur on almost all continents and in different tectonic settings. In the Americas amphibole has been shown to vary from a more Ti-rich kaersutitic composition in veins to pargasite in neighbouring peridotite wall rocks. Amphibole occurs in xenoliths in Arizonan ultramafic rocks together with titanoclinohumite and magnesite. Patagonian peridotite xenoliths with amphibole, apatite and phlogopite have been attributed to metasomatism by carbonatite melts (Gorring and Kay, 2000), whereas others discount this origin due to high TiO_2 contents (Rivalenti *et al.*, 2004). In China and Mongolia, xenoliths with amphibole, apatite and phlogopite have been described from basalts, diorites and ultramafic lamprophyres (Liu *et al.*, 2021) and in alkali basalts in Vietnam (Konzett *et al.*, 2019).

Moving to Europe, a similar assemblage is found in the eastern part of Spitzbergen and hornblendites and amphibole clinopyroxenites occur in the Eifel, in southeastern Spain, Israel, and close to the Bohemian Massif in Hungary and the Czech Republic.

The East African Rift's eastern branch has xenoliths with amphibole and apatite occurring together with graphite and orthopyroxene in Kenya (Suzuki, 1979; Kaeser *et al.*, 2006), but further north in Ethiopia, xenoliths were found to have experienced minimal metasomatism except for the introduction of apatite (Bedini and Bodinier, 1999). Recall from the section on massif peridotites that this may be a distal expression of metasomatism around amphibole dykes. Amphibole pyroxenites, some with phlogopite, and kaersutites are found in alkaline lamprophyres and rushayites (a sodic equivalent of katungite; Denaeyer, 1966) in Algeria and Morocco (Bossière and Megartsi, 1982; Wagner *et al.*, 2003) and in Madagascar (Cuccinello *et al.*, 2023). In New Zealand's South Island, ultramafic lamprophyres also carry amphibole and apatite-bearing xenoliths, which are said to be MARID-like except that the amphibole is pargasite and not K richterite (Cooper *et al.*, 2024).

The occurrence of amphibole and apatite in peridotites and veins is commonly attributed to metasomatism by carbonatite or carbonated silicate melts (*e.g.*, Gorring and Kay, 2000; Morishita *et al.*, 2003). This is supported by the occurrence of carbonate and of carbonatite melt inclusions in xenoliths with apatite. The widespread occurrence of apatite in mantle rocks indicates that the metasomatic agents must be relatively oxidised, as P is taken up in silicate minerals (olivine, clinopyroxene; Goodrich, 1984) in reducing conditions in which metals are stable.

4.3.3.5 Where is the carbonate?

It became clear during the 1970s that CO_2 is needed in the source of many alkaline rocks such as kimberlites and melilitites and high pressure experiments showed that carbonates should be stable in the mantle (Section 5). However, little evidence could be found for carbonates in the xenolith record: reviews of xenoliths through the 1980s documented many kinds of metasomatism but

not carbonates. At this time two cases of carbonate occurring as inclusions in other minerals had been reported; dolomite in garnet (McGetchin and Besancon, 1973)[14] and magnesite in diamond (Bulanova, 1986), as well as what were interpreted to be the decomposition products of dolomite in a wehrlite (Berg, 1986), to which I will return later. The paucity of carbonates was interpreted to be due to the extremely rapid decomposition of carbonate during ascent, which was supported by experiments. Indeed, Canil (1990) concluded from his experiments that "*it is difficult to reconcile why carbonates are found in* any *mantle assemblage*".

This century, many more instances of interstitial carbonates or carbonatite melt inclusions in minerals in peridotites have been reported, but few of convincing interstitial melts. Interstitial carbonate minerals have been attributed to accumulation from carbonatite melts or fluids in a variety of geological settings. These are not thought to represent melt compositions, but minerals deposited from melts. Potential melt compositions can be reconstructed from inclusions trapped in minerals (Golovin and Kamenetsky, 2023): these are usually found to have alkali carbonatite compositions, particularly sodic.

However, the most convincing examples are pools of carbonate melt that are preserved exclusively in wehrlites. This makes perfect sense, since a carbonatite melt infiltrating a peridotite will react with orthopyroxene to produce olivine and clinopyroxene (Yaxley *et al.*, 1991). Therefore, carbonatite melt can only be preserved once the orthopyroxene has been completely eliminated. Pools were described in a garnet wehrlite from Greenland by Kiseeva *et al.* (2024) which were interpreted to indicate an alkali carbonate melt composition. Although compromised by low temperature alteration, the pools had analogues preserved as inclusions in spinels in the same rock. The melt pools contained dolomite, serpentine, calcite, apatite, phlogopite, sulfides and spinel, whereas the melt inclusions additionally preserved brucite, alkali carbonates and periclase.

Figure 4.11 illustrates similar melt pools in wehrlites from Kimberley in the Kaapvaal Craton, presented at the 2022 Goldschmidt conference (Ramokgaba *et al.*, 2022). We interpret these as equivalent to the features described by Berg (1986), but which he did not describe as melt pools. These have perhaps been overlooked in the past because they are quite ugly in thin section, appearing as irregular brown masses. Micro-XRF images detail the mineralogy, which is dominated by magnesite (dark blue) in Figure 4.11a or dolomite (now a mixture of calcite and brucite; light blue/white; Fig. 4.11b). Minor constituents are fluorapatite and immiscible sulfide-rich melts.

14. This is a 3 page note in *Earth and Planetary Science Letters* at a time when the journal lived up to its name. I recall talking to Friedrich Begemann, an editor at the time, during my years at the Max Planck Institut für Chemie when I was considering a submission to EPSL. He told me the ideal letter was five pages long and had no more than five references *per* printed page. Several other journals have since tried opening letters sections, which often closed again: we geoscientists seem to make it a very difficult thing to do.

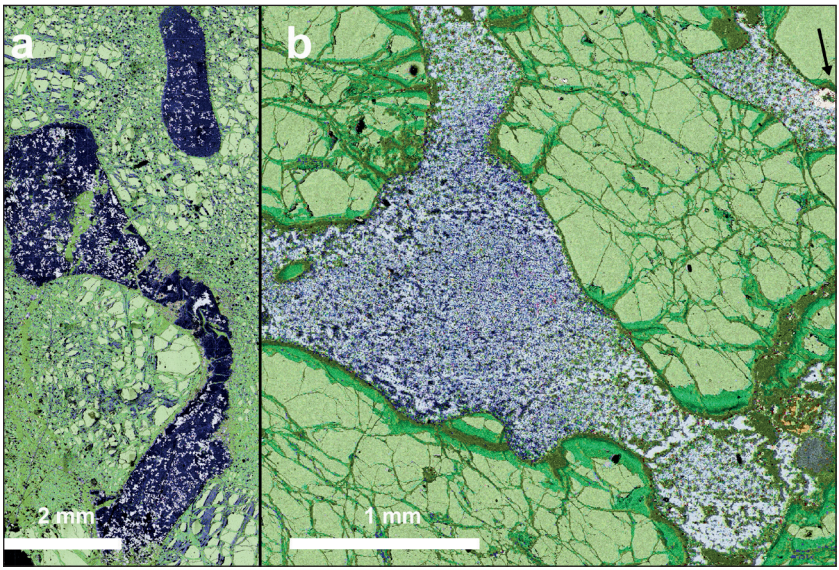


Figure 4.11 Element maps of carbonate melt pools in wehrlites from the Kaapvaal Craton, South Africa. **(a)** Sample with magnesite-rich melt composition. **(b)** Sample with dolomitic melt composition (now mixture of magnesite, calcite and brucite). Arrow at top right highlights immiscible sulfide-rich melt. Rock dominated by olivine rimmed by serpentine. Dark blue = magnesite, light blue = calcite, light grey = brucite, light green = olivine, dark green = serpentine, buff = sulfide.

Together these pools document a range of carbonatite melts in wehrlites of the cratonic mantle, including alkali, dolomitic and magnesitic carbonatites from different depths of the lithosphere. Other examples of carbonates in wehrlites have been described from the Canary Islands (Kogarko *et al.*, 1995).

The mineralogical effects of reaction of peridotite with carbonatite melts have been described from a variety of different geological settings, including the oceanic lithosphere (Hauri *et al.*, 1993), a subduction setting in Italy (Ammannati *et al.*, 2016), rifting close to the Cameroon Line (Nkouandou *et al.*, 2022) and the Tan-Lu fault, a major shear zone in China (Lin *et al.*, 2020).

4.3.3.6 Reaction zones

The recognition of xenolith assemblages as stemming from reaction zones used to depend greatly on similarity of assemblages and/or mineral compositions to composite xenoliths or outcrops of massif peridotites. In recent years, it has additionally benefitted from the results of high pressure reaction experiments in which two rock types or a rock and a melt are included in the same experimental capsule. These have been summarised recently by Prelević *et al.* (2024).

I noted above that the PKP suite probably represents a reaction zone between MARID melt and peridotite and that the IRPS suite of xenoliths is related to the general enrichment in Fe and Ti close to kimberlitic melts. Two other assemblages that probably result from reaction of peridotite with infiltrating melts are rich in [1] mica and clinopyroxene, and [2] orthopyroxene. Many mica clinopyroxenites have sharp boundaries with peridotite and originate as wall crystallisation from melts in cracks, but mica and clinopyroxene have also been produced in reaction zones between peridotite and melts of clastic sediments, basanite or phonolite. Orthopyroxene-rich assemblages containing amphibole and phlogopite (Ertan and Leeman, 1996) and/or ilmenite (Litasov *et al.*, 2000) have been found in natural composite xenoliths. In experiments, orthopyroxene is a common product in reaction zones, triggered by reaction with melts as varied as those derived from clastic sediments (Mallik *et al.*, 2016), rhyolite (Mallik *et al.*, 2015), phyllite (Wang *et al.*, 2017), carbonated ocean crust (Mallik and Dasgupta, 2013) and limestone (Wang *et al.*, 2024). Phlogopite pyroxenites are predicted products of mantle wedge metasomatism in subduction zones (Sekine and Wyllie, 1982a,b; see Section 5) and are seen in natural examples from appropriate areas, such as Papua New Guinea (McInnes and Cameron, 1994).

5. EXPERIMENTS ON MELTING PERIDOTITE

There are also superficial reasoning men, who think themselves qualified to write... [about] those regions of the earth which can never be seen… from having… looked into the bottom of a little crucible.

Hutton (1795)

This was James Hutton's opinion about the very first experiments on the melting of basalt by James Hall, performed in 1790 but published after Hutton's book (Hall, 1798). In this section I summarise the salient findings from experiments on mantle peridotites and alkaline melts, and on reactions between them. Conclusions based on experiments have, methinks, become a little less superficial with time.

The previous history of the peridotite is also important here because peridotites with different degrees of geochemical depletion will show different reactivities with melts. The experimental characterisation of melts produced from a variety of potential mantle migmatite neosome components and their reactions with peridotite is needed to understand which melts will arise from the complete melting event. These are considered in Section 6.

Our knowledge of the melt compositions produced from peridotites are relatively mature. The results are now used in software packages that calculate melt compositions from a given bulk peridotite mantle composition: programs such as pMELTS and Thermocalc are dependent on the results of high pressure experiments. However, the least known components in these models are, unfortunately, those of greatest interest for alkaline rocks (Na, K, Ti, P and volatiles), so one must be cautious when using these programmes for alkaline melt compositions.

5.1 A Short History of Experimental Studies Relevant to Mantle Derived Alkaline Melts

Although "*Hall can deservedly be called the father of experimental petrology*" (Loewinson-Lessing, 1954), high temperature melting experiments with accurate temperature measurements had to await the development of calibrated thermocouples at the newly established Geophysical Laboratory of the Carnegie Institution of Washington in the earliest 20th century. The philosophy of this laboratory opened the way for 50 years of experiments on minerals and simple systems before high pressure experiments at upper mantle conditions became possible with the development of the piston cylinder apparatus around 1960 (Boyd and England, 1958). The dominance of the Washington laboratory was not as complete as described by Young (2003), however, as Niggli was also conducting experiments in Europe (Niggli, 1916, 1919). The availability of high pressure

experiments approximately coincided with the almost universal acceptance of peridotite as the principal mantle rock (Fig. 5.1; see Section 4) and was assisted by the advent of the electron microprobe, which was used for experimental charges from 1967 onwards (Green and Ringwood, 1967)[15]. After this, progress was rapid on melting of peridotite, initially dry and then with added H_2O and then CO_2, as well as on liquidus experiments on candidate primary magmas (Fig. 5.1). Experiments on other mantle rocks and on reaction between peridotites with partial melts of these other rocks came later, initially to model processes involving subducted materials.

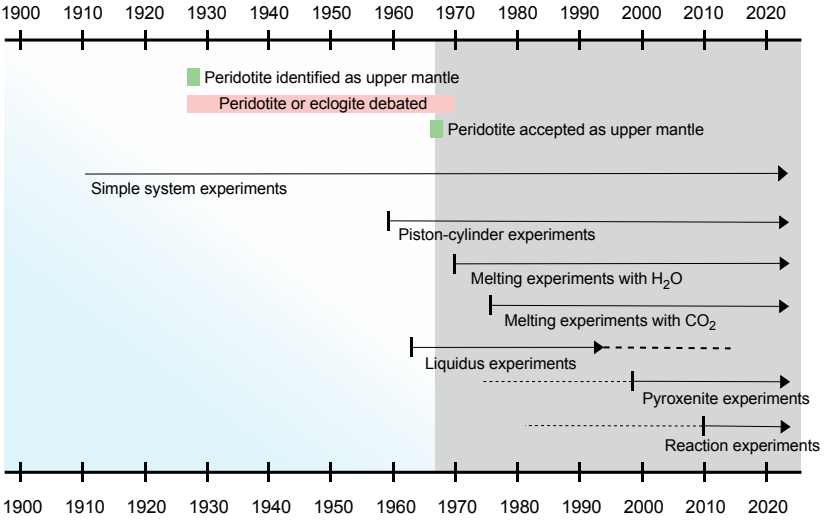


Figure 5.1 Timeline for experiments. Experiments on simple systems and mineral joins followed accurate measurements of high temperatures. Experiments at mantle pressures were possible from the 1960s and first experiments on peridotite melting, initially dry and then with various volatiles, followed from 1967. Melting experiments on pyroxenites and reaction experiments relevant to melt source regions started in earnest around 2000. Liquidus surveys of supposed primary magmas were common in the 1970s and 1980s, but interest waned after they failed to match expectations.

Two broad groups of experiments, termed forward and inverse by Wyllie (1987, 1988) were conducted in different laboratories. Forward experiments refer to partial melting of source rocks, whereas inverse experiments begin with a

15. This classical study of experimental petrology was published in *Contributions to Mineralogy and Petrology* shortly after it was re-named from the "*Beiträge zur Mineralogie und Petrographie*". It is a little known fact that it was rejected by the *Journal of Petrology*, whose editors at the time were Yoder and Tilley, the authors of the 1962 190 page tome, which it threatened to outdate. Tilley was Green's PhD supervisor.

supposed primary melt composition and survey the liquidus curve at different pressures, looking for minerals that correspond to those in the source (Fig. 5.2). Both approaches have their advantages and disadvantages. In forward experiments, different starting materials can be chosen, such as natural ground mineral mixes, mixes of synthetic analytical reagents, glasses and gels. All of these have their supporters and vehement critics, but the details of this are a bit off topic here. Of greater concern for our purposes is that it is very difficult to isolate and analyse melt compositions at very low degrees of melting, but it is these melts that are enriched in alkalis. This problem is worse in volatile-bearing experiments, where mats of dendritic crystals can form very rapidly from the melt during the couple of seconds between the experiment being turned off and the charge dropping below the melting point. This makes analyses of melt compositions at best approximate. Methods to circumvent this problem include iterative sandwich experiments and the use of melt traps made of diamond or amorphous carbon to separate the melt from the matrix into pools of analysable size (see 5.3.2 and 5.3.4). These improve knowledge of partial melt compositions but can never be entirely satisfactory.

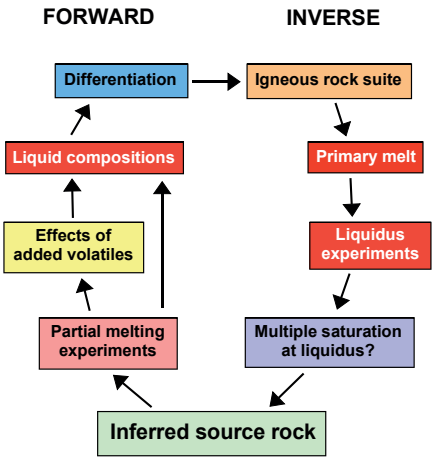


Figure 5.2 Forward and inverse experimental approaches to melting processes in the mantle, simplified after Wyllie (1987).

In inverse experiments (Fig. 5.2), one of the main problems is the choice of a starting composition, which can be either the most primitive of a volcanic rock series or an estimated primary melt composition but may not correspond to a true primary melt. Added to this, the correct contents of volatile components can only be estimated because they are partly lost during eruption. The concept is that forward and inverse experiments are mutually complementary, as illustrated in Figure 5.3. If we assume that partial melt compositions of peridotite as a function

of pressure and temperature can be mapped out, despite the issues noted above, then the phases present at a given temperature and pressure (minerals and melt) should all appear together at the liquidus of another experiment on the volcanic rock composition at the same temperature and pressure. This assumes that the primary melt has been correctly identified for the inverse experiment. Clearly, the uncertainties in both forward and inverse experiments are large enough to allow ample wiggle room, so that in many cases where exact matches were not found, they were explained away as close enough and accepted. These are classic examples of *ad hoc* sticking plasters for the cuts and bruises suffered by the concept of complementarity in forward and inverse experiments (see Section 5.6). Liquidus experiments were much less commonly carried out from 1990 onwards, probably largely because they usually failed to identify conditions of multiple saturation in peridotite minerals.

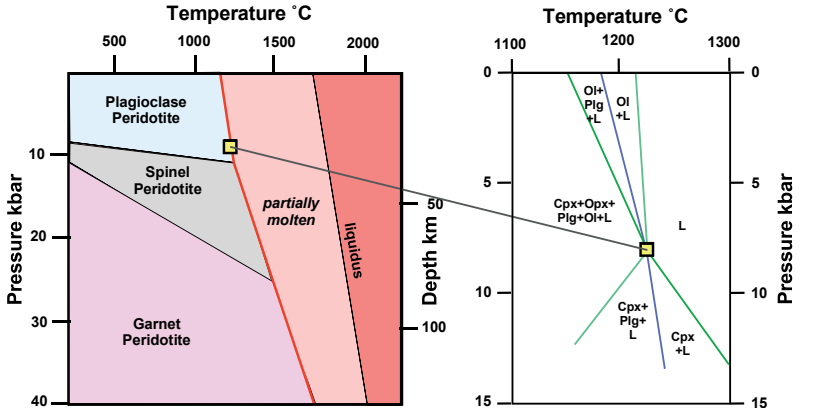


Figure 5.3 Illustration of mutual complementarity of forward (left) and inverse (right) approaches to experimental igneous petrology. An initial melt of peridotite (yellow square, left) is in equilibrium with olivine, orthopyroxene, clinopyroxene and plagioclase at 8 kbar. The same melt must have all four minerals coexisting with melt when run at the same temperature and pressure (right). After Wyllie *et al.* (1981) and Fujii and Kushiro (1977).

Text Box 5.1 – The Debt to Peter Wyllie

I doubt that any scientists with whom I have *not* worked have had a greater effect on how I see and do my science than Peter Wyllie. Apart from developing a deep interest in phase petrology from his papers, I learned more about his methods osmotically by talking at length to my first post-doc, Sieger van der Laan, who came to Göttingen after working with Peter at Caltech.

I only met Peter a few times at conferences, the first time at the 4^{th} *International Kimberlite Conference* in Perth when I was still a student. He came across as quite formal (as did many people at conferences in those days), and usually wore

smart blazers. But this was clearly his speciality, as the chairman Steve Haggerty introduced him as Captain Wyllie and asked us to fasten our seatbelts for the ride. I was intrigued by his method for projecting coloured diagrams, which were just coming into vogue after generations of black and white slides. Most people, me included, would photograph colour pictures or simply draft figures on a coloured background, but Peter is the only one I have seen who would take a felt pen to his black and white slides and colour in lines or areas on the 35 x 23 mm diapositives! It was very effective.

After this first meeting, he responded to my requests for reprints not by just throwing them in the mail. They would always be commented (*e.g., "I only received a few of these and yours is about the only one from Mainz that I have been able to satisfy"*) often with remarks about my own work (*e.g., "a great paper you sent. I wanted to write a letter, but no time – off to India for a month"*) that were very encouraging for an early career scientist who held his work in awe.

Apart from well known reviews linking experiments and geodynamics in the 1970s and 1980s, many of Peter's papers were relevant to various aspects of petrology important to the development of the mantle migmatites concept, particularly the role of volatile components. He produced the first experimental work adding CO_2 to magmatic systems in experiments (Wyllie and Tuttle, 1959) and, in the first ever paper in the *Journal of Petrology*, described the first experiments with igneous carbonatite melts (Wyllie and Tuttle, 1960). With Gus Koster van Groos, he described silicate-carbonate liquid immiscibility a decade before more well known papers on this subject (Koster van Groos and Wyllie, 1963) and studied the effects of HF and NH_3 on melts (Wyllie and Tuttle, 1961). The 1970s work on CO_2 in melts and on pure carbonate systems (Irving and Wyllie, 1973) is still extremely topical for melting in the mantle of all sorts of rocks, including subducted carbonates and the deep carbon cycle (see Sections 5 and 7). It was Wyllie (1980) who first suggested that the initial source regions of kimberlites are deep in the asthenosphere and contain reduced carbon.

The final experimental innovation that is relevant to mantle migmatites that I wish to mention was the introduction of reaction experiments containing either mixed rock powders or two separate blocks of rock such that one rock melts and reacts with the second rock to produce a new hybrid rock (Sekine and Wyllie, 1982a). This type of experiment has become common since 2010 and is essential to understanding the melting process in mixed source regions (Sections 6 and 7).

5.2 Early Experiments in Simple Systems

Experiments at room pressure in the period 1910–1960 were dominated by the Carnegie Institution's Geophysical Laboratory, which followed the philosophy of experimenting either on minerals, joins between minerals or their end members, or on chemically simple systems that became progressively complex as the years progressed. Experiments on plagioclase and olivine (Bowen, 1913; Bowen and Anderson, 1914) were quickly followed by a ternary system in a paper that also contained a phase diagram tutorial for petrologists (Bowen, 1914).

Perhaps the most famous simple system for basaltic rocks is CMAS (CaO-MgO-Al_2O_3-SiO_2), which was widely used for peridotite and basaltic compositions because it accounts for approximately 85 % of the rock compositions (O'Hara, 1968; Yoder, 1976). The elimination of FeO (8–10 %) was understandable, because it is notoriously difficult to control due to reacting with noble metal capsules, but, unfortunately for the alkaline rocks, the other missing 5–7 % contains Na_2O, K_2O, TiO_2 and P_2O_5, all of which are important constituents in mantle derived alkaline rocks. Work on alkali-bearing systems also started with simple systems, such as Na_2O-Al_2O_3-SiO_2 and K_2O- Al_2O_3-SiO_2 (Schairer and Bowen, 1947), eventually incorporating Fe (Na_2O-Al_2O_3-Fe_2O_3-SiO_2; Bailey and Schairer, 1966). However, because of the systems chosen, the focus of most discussion based on these results was on granitic and syenitic rocks and the evolved products of fractionation and not on primitive mantle derived alkaline rocks. The summary of available experiments on alkaline rocks by Edgar (1974) discusses simple systems almost entirely.

The exception is the basalt tetrahedron introduced by Yoder and Tilley (1962) which has as its four bounding corners (1) Clinopyroxene [Cpx] (or simplified to Diopside [Di]), (2) Quartz [Qz], (3) Olivine [Ol] (or forsterite [Fo], and (4) Nepheline [Ne]. This was used to represent the compositions of basalts using the concept of silica undersaturation, meaning that alkaline rocks fall in the volume Ne-Ol-Cpx-Plagioclase [Plg]. I note with some amusement that this was Yoder's first foray into experiments outside simple systems into natural basalts, which he undertook at the insistence of the Australian Cecil Tilley (Young, 2003; p. 500).

Another kind of experiment in simple systems that I found very influential and used myself was to investigate the effects of minor components on melt compositions by logging the deflection of phase boundaries (Kushiro, 1975). Figure 5.4 shows that the boundary lines between olivine and protoenstatite, and between protoenstatite and silica minerals, move away from MgO and towards SiO_2 as the alkalis are added. This can be interpreted in terms of the melt structure, which is "depolymerised", thus enhancing the stability of the less silica-rich phase. We will study the effects of volatiles using this concept later.

There has been much discussion of the value of experiments in simple systems *versus* those on natural rocks, with a main argument for simple systems being the wish to fully understand phase relations and/or thermodynamic behaviour without the uncertainties introduced by systems that are too complex. Although simple system experiments *"excel at predicting what happens in the sanitized conditions of the laboratory"* (Lewens, 2016), we should not forget that they are sophisticated studies of idealised conditions. In contrast, supporters of the use of natural rocks argue that it is the real Earth that is of interest, which simple chemical systems cannot satisfactorily reproduce. I generally favour experiments on natural rocks and minerals, since important components are usually shortened out of simple systems. The experiments we now have on natural systems have brought us much further than we would have come

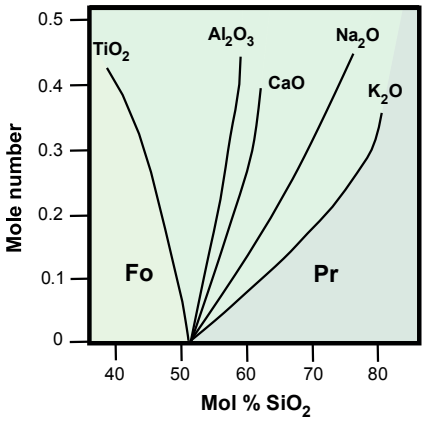


Figure 5.4 Shift of the cotectic lines between forsterite (Fo)and protoenstatite (Pr) with addition of oxides to the mixture. Note K_2O and Na_2O have the strongest effects, moving the cotectic lines away from MgO and towards SiO_2. Modified after Kushiro (1975).

if experimentalists were still slowly adding more components to their simple systems. The preference for complex over simple systems goes beyond geology and is nicely summarised by the physicist Hans-Peter Dürr (1988): "*When I look at precision, I have to try to separate certain parts from the whole, because I can only capture exactly what is isolated... I have what is isolated, so to speak, under control and can then describe it very precisely. But relevance has to do with the embedding of the part in the whole... If I want to be relevant, I have to forego local exactness*".

5.3 Melting Experiments on Peridotites

My aim in this section is to summarise the experimental evidence for the range of melt compositions that could result from melting of mantle peridotite and so define the approximate limits for peridotite melts. This will reveal which melts *cannot* be explained if we cling to the model of a uniform peridotite mantle.

It must be remembered here that experiments can be performed at a range of pressures and temperatures, regardless of whether these are realistic for the Earth. The variation in realistic geothermal gradients is therefore important, but these, in turn, have considerable uncertainties since direct measurements in the mantle are not possible. A crude comparison of average mantle geotherms with the melting curve for peridotite is shown in Figure 5.5 (modified from Wyllie, 1988), from which we can conclude that under normal (*i.e.* average) circumstances, the mantle should never melt and so we should have no volcanoes! Figure 5.5 also summarises the subsolidus mineralogy of the mantle, showing the depth

limit for garnet peridotite at around 400 km, where it transforms to majorite and high pressure polymorphs of olivine. The first experiments that outlined this on the pyroxene-garnet transformation (Ringwood, 1967) and the olivine-spinel transition (Ringwood and Major, 1966) also took place in the late 1960s, so it is no wonder that Ted Ringwood would write in 1967 *"it is difficult not to be an optimist in this field today"* (Fifield, 1985). The partially molten field in Figure 5.5 thins at higher pressures due to solidus-liquidus convergence, explained in one of my favourite theoretical petrology papers (Walker, 1986).

Figure 5.5 does not give an impression of the number of experiments on which the diagram is based: this is well constrained at depths down to about 120 km, which are accessible by piston cylinder apparatuses, but sketchy in comparison at higher pressures, which depend on experiments in multi-anvil apparatuses. These suffer from larger temperature gradients and are often done at larger pressure intervals (*e.g.*, 5 GPa = 50 kbar between experiments) in order to cover the large pressure range.

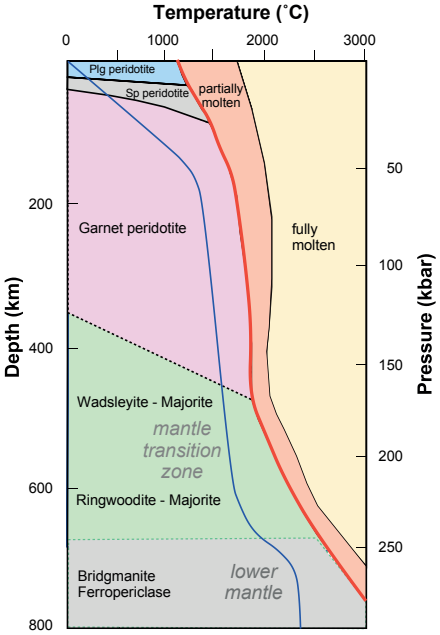


Figure 5.5 Comparison of average geothermal gradient (blue line) with the experimentally delineated solidus for dry peridotite (red line; modified after Wyllie, 1988). The geothermal gradient assumes a conductive lid, which is thicker beneath continents. The increase in temperature in the lower mantle assumed a separate convection layer, which is now thought unlikely.

Since we clearly do have volcanoes, the average conditions depicted in Figure 5.5 often do not apply. Instead, there are special conditions that allow the mantle solidus to meet the geothermal gradient, resulting in partial melting. Five special conditions are summarised in Figure 5.6, namely (1) an increase in temperature, (2) a decrease in pressure, (3) addition of water and/or other volatile components that depress the solidus, (4) a change in redox conditions (redox melting), and (5) the presence of other rocks with lower melting points than peridotite. Somewhat counter-intuitively, option 1 – an increase in temperature – may be the least important of the five. Decreasing pressure is common in areas of mantle upwelling, especially under mid-ocean ridges where the solidus can be crossed along an adiabatic path. The effect of the addition of water or other volatiles is not to change the position of the rock composition that is melting (Fig. 5.6), but to move the melting curve to lower temperatures, as has been discussed particularly for subduction zones. The same is true for redox melting (option 4), since here the oxidation of CH_4 increases the activity of H_2O, thus depressing the melting point (Taylor and Green, 1987, 1988). Option 5, the role of other rocks than peridotite in the mantle is central to mantle migmatite melting; the experiments on these rocks are presented and discussed in Section 6. Of these scenarios, all except decompression melting are likely to vary in intensity on a relatively small spatial scale, as discussed by Foley *et al.* (2024).

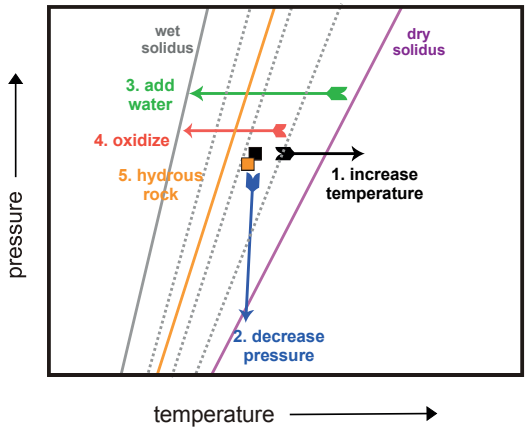


Figure 5.6 Five special conditions that facilitate melting in the mantle, of which (1-4) apply for peridotites: (1) increase in temperature is the least likely for mantle conditions, (2) decompression melting is common for parcels of upwelling mantle, particularly beneath mid-ocean ridges, (3) addition of H_2O (or other volatiles) depresses the melting temperature, placing the rock above the melting point, (4) redox melting results from oxidation of CH_4, which increases the activity of H_2O with a similar effect to (3), and (5) the rock is not a peridotite but has a lower melting temperature and thus lies above its own solidus.

5.3.1 Volatile-Free Peridotites

The earliest high pressure experiments I know of on melting of peridotite were conducted by O'Hara (1963), published as a preliminary report in the *Yearbook of the Carnegie Institution of Washington*, but which never appeared in full in journals. These yearbooks are little known outside experimental petrology but were a great joy to research students (and, I suppose, more mature researchers too) up until the mid-1980s, since you could have them sent to you free every year! They contain many scientific gems that never found their way into mainstream journals. General folklore assigns the first experiments on peridotite to Ito and Kennedy (1967), which were done in an apparatus that I later met in person when Reinhard Boehler brought it to Mainz in 1986.

However, the composition of partial melts of peridotite was controversial at this time, because nobody had yet analysed them and estimates were based on the proportion of phases present and the interpretation of phase diagrams. It was not until the early 1970s that the first electron probe analyses of experimental glasses became available. Nevertheless, a petrogenetic grid mapping out the variation of melts of peridotite with pressures to 35 kbar (≈100 km) was presented by Green (1970) and is reproduced here as Figure 5.7. This was based largely on liquidus experiments on various basalts, combined with a smaller number of experiments on peridotite, but which were not aimed at determining melt compositions (Green and Ringwood, 1967, 1969). These included the first experiments on the stability of amphibole in peridotite, which allowed surprisingly accurate prediction of low degree melt compositions (Fig. 5.7).

Figure 5.7 gives broad pressure-temperature ranges for specific basaltic rock compositions as well as the maximum stability of residual phases, resulting in the residual rock types noted. This is a remarkable compilation for its time and represented a major step forward in the understanding of which melt types can be produced by melting of mantle peridotite in the uppermost 100 km of the mantle. All studies since this are minor improvements and iterations in comparison. This petrogenetic grid could be said to define the best exemplar for the mantle peridotite paradigm (Lewens, 2016, p. 61) notwithstanding the complaint that "*it is time to bring to a close the paradigm of revolutionary paradigm shifts*" (Lewens, 2016, p. 83).

Silica oversaturated (quartz-normative) primary melts are restricted to very low pressures, whereas silica undersaturated (foid-normative = alkaline) rocks occur at higher pressures. They are especially favoured at very low degrees of melting (<10 %) but may occur at higher degrees of melting at higher pressures of around 30 kbar or more. This figure set the scene for many further experiments which have refined the melt compositions and studied peridotites with different degrees of depletion.

The broad conclusions and effects of progressive melting and pressure summarised in Figure 5.8 hold today and many of the generalisations have been shown to be extrapolatable to higher pressures and for other ultramafic rocks.

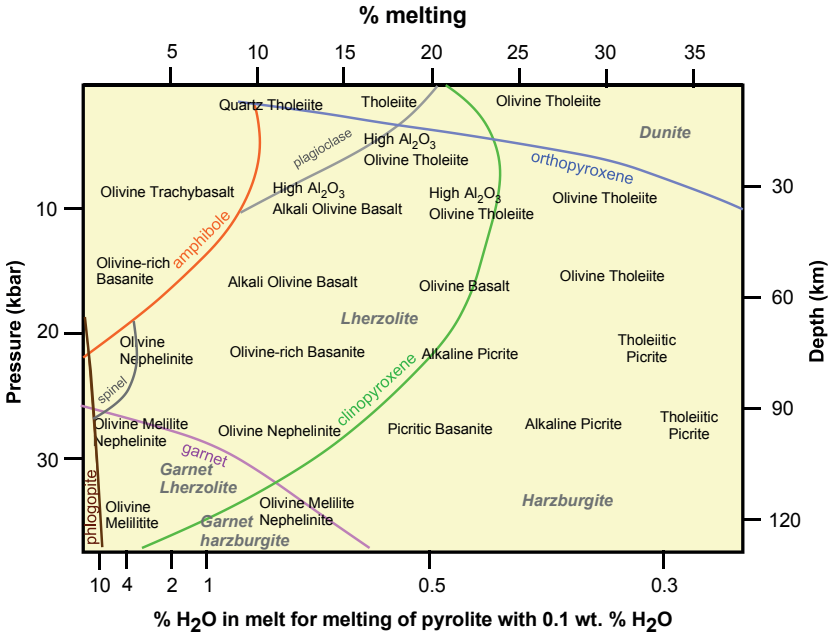


Figure 5.7 Petrogenetic grid of Green (1970) showing melt types as a function of pressure and degree of melting. Labelled minerals are stable in the residue to the left of the coloured lines. Stability of phlogopite and origin of nephelinites and melilitites were estimated at this time. Modified from Green (1970).

Initial melts are either close to the line of silica undersaturation or are alkaline but become less alkaline with progressive melting (Fig. 5.8a). All melts at 10 kbar remain in the silica saturated field (Fig. 5.8a; Falloon and Green, 1987) and further experiments have confirmed that quartz-normative melts are only produced at pressures less than 10 kbar (Falloon *et al.*, 1997). At pressures at and above 15 kbar, initial melts are alkaline and become more strongly so with increasing pressure (Fig. 5.8b,c; Jaques and Green, 1980; Falloon and Green, 1988).

We come away from this summary with the generalities that the SiO_2 content of melts decreases with increasing pressure, and the MgO contents of melts increase. The pressure effects were confirmed in multi-anvil experiments that increased the pressure range to 14 GPa (≈ 450 km) where komatiitic liquids with >30 wt. % MgO were found (Takahashi and Scarfe, 1985). These first experiments were very short and suffered from severe temperature gradients. Later experiments with stepped furnaces and longer durations (Walter, 1998) indicate 18–20 % MgO for 10 % melting at 40 kbar, increasing to 23–25 % MgO at 15–20 % melting and 60–70 kbar. K_2O contents were less than1 % and Na_2O

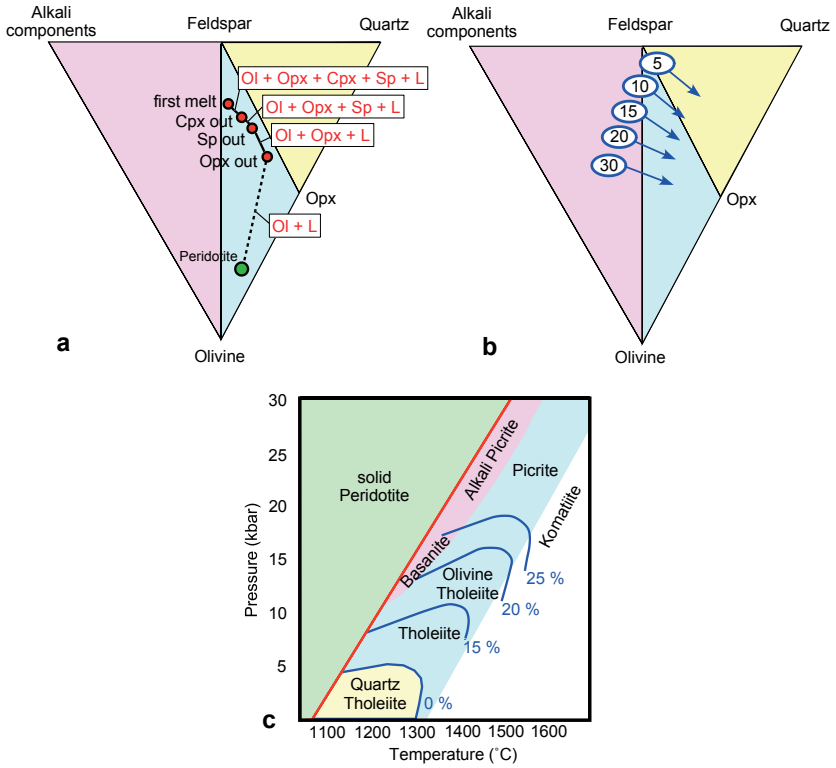


Figure 5.8 Compositions of melts of dry peridotite on floor of the basalt tetrahedron (left and centre) and on a pressure-temperature plot (right). **(a)** Melts are entirely in the Ol + Opx + Plg triangle at 10 kbar. **(b)** Melts are mildly alkaline at low degrees of melting at high pressures (pink background). Blue arrows indicate evolution of melt compositions as degree of melting increases. **(c)** Melt compositions on a pressure-temperature plot using same colour scheme as in **(a)**-**(b)**. Blue lines indicate percentage of normative olivine in the melt composition. After Falloon and Green (1987), Jaques and Green (1980).

less than 1.2 % regardless of pressure between 3 and 7 GPa. The reason for the main pressure effects is that olivine is a peritectic phase at low pressures, meaning that the modal abundance of this low SiO_2, high MgO phase increases in the residue, and melts show the opposite trend. These experimental results are the state of the art currently, as there is little reason to expect that melting in the mantle at pressures much higher than 70 kbar (225 km) occurs, except in subducted materials.

These experiments on anhydrous melting of peridotite are used very effectively in the generalised model for decompression melting of mantle peridotite (McKenzie and Bickle, 1988). Common basalts appear to be explained well, with mid-ocean ridge basalts originating at depths of 40 km and less, whereas ocean island basalts emanate from a melting zone that starts deeper and ends below the lid of the oceanic lithosphere. Alkaline rocks, however, are not so easily explained without recourse to multi-stage processes and/or non-peridotitic source rocks (Pilet *et al.*, 2008).

Whilst the differences in degree of depletion of the peridotite due to previous melt loss do not appear to move cotectic boundaries very significantly (Falloon *et al.*, 1988), the effects of minor components such as alkalis and P were investigated more closely after claims that initial melts are enriched in Na_2O and SiO_2. Later experiments showed that at 3 % melting, Na_2O and SiO_2 contents were below 4 % and ≈ 49 % (Robinson *et al.*, 1998) and that K_2O also has only a small effect on the SiO_2 content of melts (Davis and Hirschmann, 2013).

The determination of melt compositions directly by the electron microprobe analysis of glass between crystals is fraught with difficulties due to growth of individual quench crystals and rims around residual grains. This can be a problem even in anhydrous experiments but is especially problematic in hydrous melts. Several methods have been employed to attempt to circumvent this problem, which I discuss briefly here because the problem is exacerbated at low degrees of melting where melts are alkaline and volatiles are relevant to the production of alkaline melt compositions. Here, melts are surrounded more closely by crystals for quench rims to nucleate on.

The oldest method was developed for other purposes: in the days before investigation by electron microprobe and scanning electron microscope was possible, researchers would grind a part of the quenched experimental charge under oil on a glass side and identify the phases present with a petrographic microscope. However, minor phases can easily be missed because the charge consists mostly of olivine. Green (1973) introduced the method of conducting experiments on a peridotite that had already had 40 % olivine components subtracted from the bulk peridotite composition, which helped him to identify minor phases more easily. The logic is that olivine is still present in the residue and so equilibrium of the melt with all peridotite minerals is maintained. This was continued in many later papers on "pyrolite minus 40 % olivine". The relevance for alkaline rocks is that this subtraction of olivine artificially enriches the bulk composition in alkalis, and since the alkalis are incompatible components, the melts may also have artificially high Na_2O and K_2O contents.

Different types of reversal experiments have been used to demonstrate equilibrium and thus the accuracy of the estimated melt composition despite the quench modification. The simplest of these reversals is similar to the inverse method noted above (Fig. 5.3): the melt composition analysed in the forward experiment is used in a second experiment to look for the peridotite minerals at the liquidus. Later, more sophisticated methods included sandwich

experiments, iterative sandwich experiments, and synthetic fluid inclusions. In sandwich experiments, the melt from the first (forward) experiments is packed as a layer between two blocks of peridotite and run at the original pressure and temperature. If the melt changes composition, then a further sandwich experiment is run with the new melt composition, and so on until equilibrium is achieved. This slow method requires patience and application and has mostly been used for low degree melts of peridotite (*e.g.,* Robinson *et al.*, 1998) but has also been used successfully for carbonatite melts (Wallace and Green, 1988).

Another set of techniques seeks to separate the melt from matrix in pools large enough for analysis, thus eliminating the problem of overlap with neighbouring crystals by using melt traps made of either diamonds Hirose and Kushiro (1993) or vitreous carbon (Foley *et al.*, 1999). Other methods, used mostly for volatile-bearing experiments and whose effectiveness was partly serendipitous, trap melt in healed cracks within olivine or graphite capsules, or between olivine and platinum capsules. The result is the same: larger areas of melt are separated from the matrix, facilitating analysis.

5.3.2 Peridotite with Water

Many experiments have now been performed on peridotite with added H_2O, beginning with Green and Ringwood (1969) who showed that amphibole is stable in peridotite below the solidus to at least 28.8 kbar (≈ 93 km). Figure 5.9 summarises the melting curves determined for peridotite under water-undersaturated conditions. These show a maximum pressure stability for amphibole of 26–30 kbar and a temperature maximum of 1030 to 1150 °C, depending on the fertility of the bulk composition. The lower temperature applies to depleted lherzolite, whereas the highest is for Hawaiian pyrolite, a fertile composition (Wallace and Green, 1991). It should be noted that almost all the curves in Figure 5.9 are on compositions with 40 % olivine subtracted, so that these are enriched compositions. The addition of only 0.5 % H_2O depresses the solidus temperature by hundreds of degrees at mantle pressures, which anticipates the incipient melting concept developed later. At pressures above the amphibole stability limit, the solidus temperature falls further; phlogopite is stable at the solidus (Fig. 5.9) and persists to around 70 kbar (≈ 225 km: Konzett and Ulmer, 1999).

Quench problems render the direct analysis of unmodified glass almost impossible in hydrous experiments (Fig. 5.10). However, the general indication is that the depolymerising effect of water on aluminosilicate structures in melts results in an increase in the silica content of melts, which is supported by phase boundary shifts that expand the stability field of olivine relative to enstatite (Kushiro, 1975).

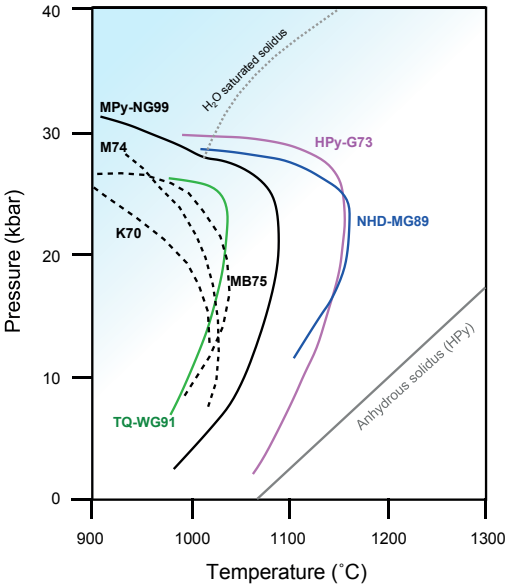


Figure 5.9 Stability of amphibole and melting curves of water-undersaturated peridotite. Dehydration melting of amphibole defines the solidus below ≈ 30 kbar. HPy-G73 = Hawaiian Pyrolite (Green, 1973); M74 = Millhollen *et al.*, 1974; K70 = Kushiro, 1970; MG75 = Mysen and Boettcher, 1975; NHD = Depleted peridotite (Mengel and Green, 1989); TQ-WG91 = Tinaquillo (depleted peridotite; Wallace and Green, 1991); MPy-NG99 = MORB Pyrolite (Niida and Green, 1999). Note: the four amphiboles with the highest P-T stability (solid lines) are all from bulk compositions with 40 % olivine subtracted.

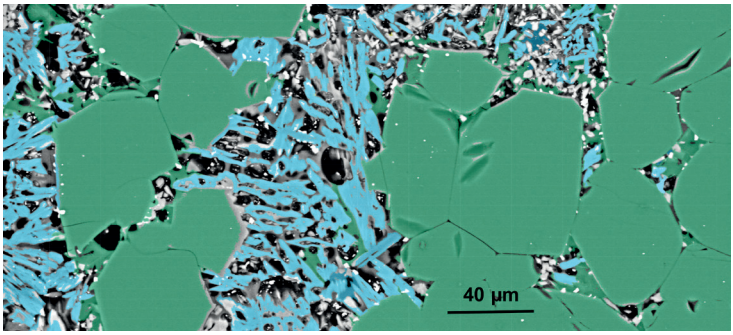


Figure 5.10 False colour image of quenched volatile-rich melt showing interlocking meshwork of quench crystals (blue, white and black) grown in place of high pressure melt. Green = stable high pressure silicate crystals. Image by Zairong Liu.

Under pressures of 15–30 kbar (50–100 km), initial melts are dominated by components derived from the melting of amphiboles and so are silica undersaturated and basanite to nephelinite in composition, concurring with the pressure effect promoting alkali-rich compositions (Fig. 5.8) and the composition of amphibole itself ($SiO_2 \leq 40$ %). At higher pressures, initial melts have potassic compositions buffered by phlogopite and low SiO_2 contents due to the same pressure effects.

5.3.3 Peridotite with Carbon Dioxide

I briefly summarise work on CO_2 alone before progressing to H_2O + CO_2 mixtures, which are more relevant to natural mantle conditions. Despite these being less realistic for the real world, experiments on mantle melting with CO_2 alone are still commonly done 40 years after the first experiments on the effect of CO_2 on melting processes. Carbon dioxide-rich melts are unquenchable, which led to considerable debate about interpretation of experimental textures after experiments on CO_2 started in the 1970s. Early experiments, therefore, investigated the effect of CO_2 by determining its solubility in simple systems and in melts (Eggler *et al.*, 1974). The effect on peridotite melting was first predicted from theoretical analysis (Eggler and Holloway, 1977; Wyllie, 1978). Together with subsolidus experiments, these showed that carbonates are unstable at the solidus below 20 kbar, above which dolomite and then magnesite (above about 40 kbar) can be stable in peridotite. Calcite (or aragonite) is not stable in peridotites at all under mantle conditions. The solubility of CO_2 in mafic melts increases greatly at pressures greater than 20–22 kbar, leading to a back bend in the solidus, with depressed melting temperatures relative to dry conditions at higher pressures. However, the solidus depression is not as severe as with H_2O.

Phase boundary responses to CO_2 addition show that the SiO_2 content of melts decreases strongly with addition of CO_2 with the phase volume for orthopyroxene expanding relative to olivine (Fig. 5.11). This can cause melts to move to strongly silica undersaturated compositions (Eggler, 1974), an effect which is augmented in alkali-rich melts (Wendlandt and Eggler, 1980a). The reason for this is interpreted as the bonding of divalent cations with carbonate ions in the melt, resulting in polymerisation of the aluminosilicate network (Ni and Keppler, 2013). This effect of CO_2 brings strongly undersaturated melts such as melilitites, kamafugites and ultramafic lamprophyres into focus, which could not be explained in dry or hydrous conditions.

Later studies on peridotite or peridotite-like simple systems with CO_2 at high pressures indicate that initial melts would be carbonatitic, passing to compositions intermediate between carbonate and silicate as melting progresses. However, whether this change is sudden or gradual has been the subject of some debate (Dalton and Presnall, 1998; Dasgupta and Hirschmann, 2006), which is relevant to low degree melting of the mantle in oxidised conditions.

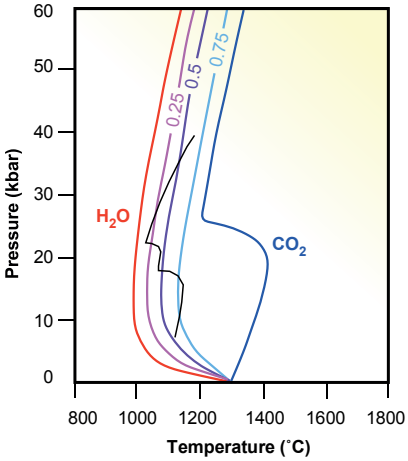


Figure 5.11 Theoretical estimations of the melting curve for peridotite with various mixtures of H_2O and CO_2 (after Wyllie, 1978). Black line is the later experimental curve from Olafsson and Eggler (1983). All of these curves pre-date the discovery of sodic carbonatites in 1988.

5.3.4 Peridotite with Mixed H_2O and CO_2

For mixed $H_2O + CO_2$ systems, elaborate theoretical phase diagrams were rapidly produced in the late 1970s with solidus curves for intermediate volatile mixtures between those for pure H_2O and CO_2 (Eggler and Holloway, 1977; Wyllie, 1978). These now serve as a cautionary tale, since later experiments using the iterative sandwich method with small amounts of both H_2O and CO_2 found the solidus to be even lower than for pure H_2O (Wallace and Green, 1988; Fig. 5.12). The difference between this result and earlier experiments (Fig. 5.12; Olafsson and Eggler, 1983) was the recognition that polishing of the experimental charges needs to be water-free because of the presence of water soluble sodium-rich carbonate melt. Over a pressure range between 20 and 31 kbar this melt coexists with amphibole (Fig. 5.12). The sodic melt composition led to comparisons with the Oldoinyo Lengai sodium carbonatite lavas.

At pressures higher than the stability limit of amphibole, the solidus lies at lower temperatures, but initial melts are still carbonatite (Foley *et al.*, 2009: Fig. 5.13a). This solidus has been tracked to approximately 70 kbar (225 km; Fig. 5.13b; Pintér *et al.*, 2021) and continues approximately linearly to higher pressures assuming it continues to run parallel to the better studied solidus with CO_2 alone (Dasgupta and Hirschmann, 2006; Brey *et al.*, 2008). Despite the difficulty with quenching the melts progress has been made with characterising partial melt compositions, which change gradually from dolomitic carbonatite

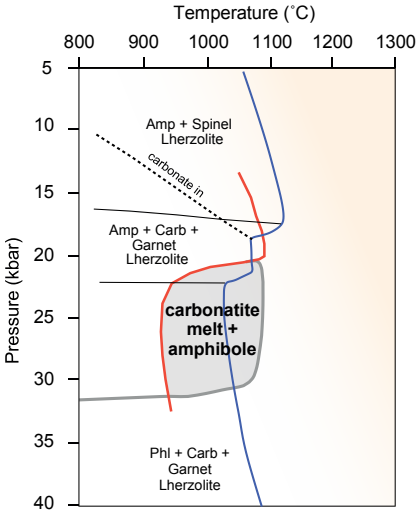


Figure 5.12 Solidus of peridotite with small amounts of CO_2 and H_2O (red line from Wallace and Green, 1988) compared to estimated solidus before the dissolution of alkali carbonates was recognised (blue line from Olafsson and Eggler, 1983). Grey field is alkali carbonatite melt.

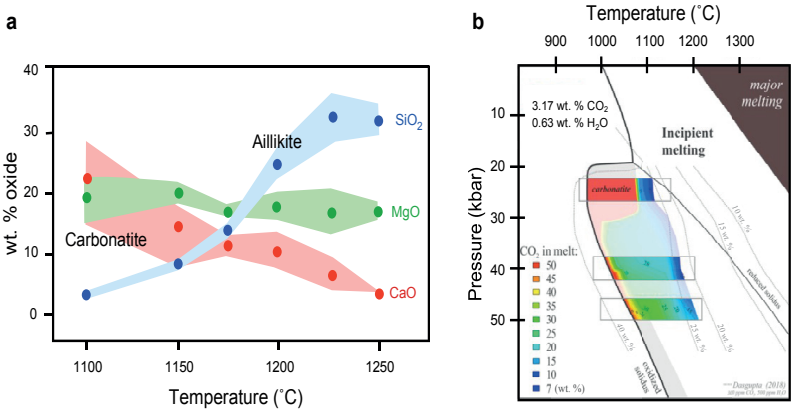


Figure 5.13 Melt compositions and solidi of K-enriched peridotite with CO_2 and H_2O. **(a)** Progressive melting at 50 kbar changes melts continuously from carbonatite (35 % SiO_2) to carbonated silicate melts similar to aillikites (Foley *et al.*, 2009). **(b)** Solidus of peridotite with melts coloured for CO_2 contents of melt (from Pintér *et al.*, 2021) which are similar to estimates of CO_2 contents in melts from Dasgupta (2018, grey lines).

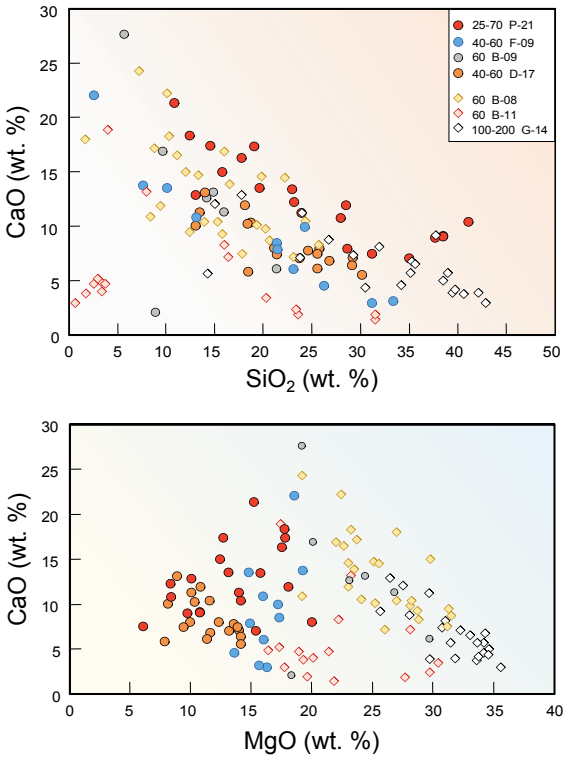


Figure 5.14 Melt compositions from peridotite (Pd) with CO_2 and with CO_2 + H_2O. **(a)** Continuous range from carbonatite (<10 % SiO_2) to silicate melts. **(b)** High MgO melts are mostly at the highest pressures. Sources: P-21 = Pintér *et al.* (2021, 25–70 kbar); F-09 = Foley *et al.* (2009, 40–60 kbar); B-09 = Brey *et al.*(2009, 60–100 kbar); D-17 = Dvir and Kessel (2017, 40–60 kbar); B-08 = Brey *et al.* (2008, 60–100 kbar); B-11 = Brey *et al.* (2011, 60–100 kbar); G-14 = Ghosh *et al.* (2014, 100–200 kbar).

to more silicate-rich compositions akin to aillikites at 50 kbar (Fig. 5.13a). Sodic carbonatite melts are not found at pressures higher than the amphibole stability limit (≈ 30 kbar). These confirm earlier predictions that a transition through carbonated silicate melts to carbonatites should occur at high pressures (Wendlandt and Eggler, 1980b) based on phase stabilities in experiments in which melt compositions could not be analysed.

Using a time consuming method of volume integration of melt pools, Pintér *et al.* (2021) mapped out the CO_2 contents of melts in the pressure range 25–50 kbar (Fig. 5.13b). These melts are compared to those from other studies in Figure 5.14, either peridotite with both CO_2 and H_2O or with just CO_2. These

melts show a continuum from carbonatite to carbonated silicate melts, filling the space for foidite in Figure 3.5, and are broadly similar to some ultramafic lamprophyres and kimberlites, but all have K_2O/Na_2O much lower than natural kimberlites (Fig. 5.14b; Foley *et al.*, 2019).

These results allow us to define a much clearer picture of the *incipient melting regime,* as first coined by Green and Falloon (1998). This label expresses the wide temperature range below the dry solidus of peridotite in which only a small proportion of partial melt exists. Figure 5.15a shows that for conditions with both H_2O and CO_2, the amount of melt in the incipient melting regime is seen to vary as a function of pressure. At 20–30 kbar where amphibole is stable, a wide temperature range exists were the degree of melting remains very small, and melts remain carbonatitic until amphibole melts. At this point the melts change suddenly from sodic carbonatite to undersaturated silicate melts akin to nephelinite. At higher pressures the change from carbonatite to carbonated silicate melts is more gradual and the degree of melting shows a correspondingly continuous increase (Figs. 5.14a and 5.15a). This incipient melting regime is the ultimate source of much mantle metasomatism, since these low degree melts can exist over a wide range of temperature and pressure and are free to move long before major melting of the silicate minerals will occur. Their mobility is facilitated in the case of carbonate-rich melts by the low viscosity and high interconnectivity of carbonate-rich melts, whether sodic or dolomitic (Minarik and Watson, 1995). In later sections, I will consider to what extent melts like those in the experiments can reach the surface in modified form, or if they mostly react within the mantle to form other rocks, meaning they form the neosomes for mantle migmatites.

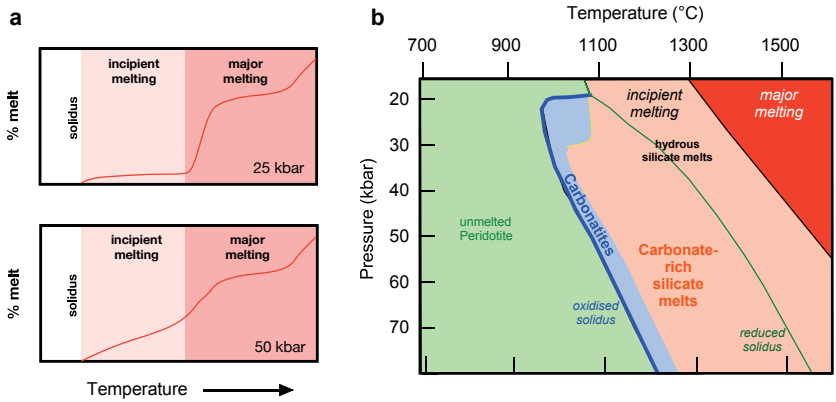


Figure 5.15 Degree of melting in the incipient melting regime at 25 and 50 kbar in oxidised mantle peridotite. At 25 kbar very low degree melting persists in the carbonatite field, at 50 kbar, melt progress more quickly to intermediate between carbonate and silicate melt (aillikite).

5.3.5 Other Volatiles

Other volatile components are unlikely to be as abundant in melt sources as H_2O and CO_2, but I consider them briefly here as they could have effects on melt structure and so the melt compositions produced. The elements F, Cl, and N are often classed as volatile components in igneous petrology. This status for F is debatable, because where hydrous minerals are present, mineral/melt partition coefficients are typically >1 and decrease from apatite to phlogopite to amphibole (Flemetakis *et al.*, 2021; Ezad and Foley, 2022; Foley and Ezad, 2024). It could thus be considered a compatible element in hydrous assemblages. Reports differ as to whether addition of F results in polymerisation or depolymerisation of silicate melts (Foley *et al.*, 1986; Dalou *et al.*, 2015), but this may be due to differing usage of the concept of melt polymerisation. Addition of F moves the Opx + Ol cotectic line to lower SiO_2 (Fig. 5.16; Foley *et al.*, 1986) implying a polymerising effect on the melt, but the melt viscosity is decreased, which can be explained by bonding of F in smaller structural units accompanied by polymerisation only in the larger units of the aluminosilicate network.

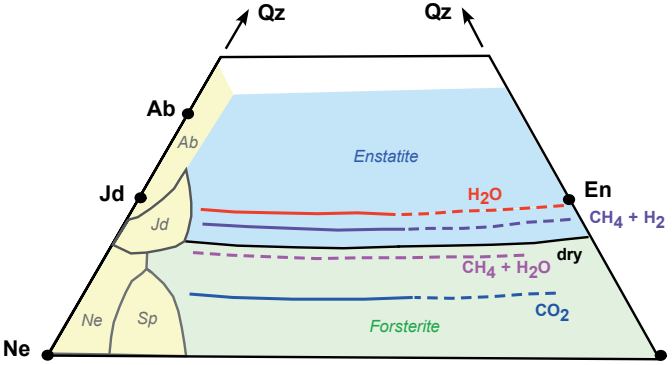


Figure 5.16 Phase boundary responses to added volatiles. Water and methane cause a slight increase in the stability of olivine relative to orthopyroxene, resulting in melts richer in SiO_2. CO_2 has a strong opposite effect and results in SiO_2-undersaturated melt compositions. Modified after Foley (2022).

Analyses of hydrous minerals in xenoliths indicate that F/Cl ratios are high (50–160; Hecker *et al.*, 2020), implying a minor role for Cl unless aqueous fluids are involved (Andersen *et al.*, 1984). This may restrict the importance of Cl to subduction related processes, although this may be different in Proterozoic and earlier times when more Cl was returned to the deep mantle, as possibly witnessed by inclusions in diamonds and in the sources of some kimberlites (Kamenetsky *et al.*, 2009; Weiss *et al.*, 2015; Förster *et al.*, 2019; Chen *et al.*, 2023). The effect of Cl on silicate melts is to polymerise the melt and expand the field of pyroxene relative to olivine (Dalou *et al.*, 2015).

Nitrogen was long considered essentially inert as well as rare in the mantle, but it is unusually enriched in potassic rocks such as lamproites and lamprophyres (Johnson and Goldblatt, 2015). A significant role for N may be limited to melting in reducing conditions, where its solubility increases as NH_3 and NH_2^+ complexes as long as H_2O is also present (Mysen *et al.*, 2008). Here it will have a depolymerising effect on melts.

5.3.6 Melting in Reduced Mantle

Carbon speciation in the upper mantle is controlled by the oxygen fugacity. Under oxidised conditions, $CO_2 + H_2O$ mixtures are expected in melts and carbonate minerals are stable, whereas under reduced conditions CH_4 coexists with H_2O in melts and solid carbon as graphite or diamond are stable. The solubility of CH_4 in silicate melts is much lower than that of CO_2 (≈ 0.2 wt. % *vs.* > 20 wt. %; Taylor and Green, 1987) and CH_4 has a mildly depolymerising effect (Fig. 5.16). The $H_2O/(CH_4 + H_2O)$ ratio varies greatly over a small range in oxygen fugacity (fO_2) from ≈ 0.35 at the iron-wüstite buffer to >0.99 just 2 log units higher (Taylor and Green, 1987), This corresponds to a large variation in the position of the solidus, because CH_4 acts essentially as a dilutant for the effect of H_2O on the solidus depression (Fig. 5.17; Taylor and Green, 1988); a higher proportion of CH_4 in the mixture corresponds to a higher solidus temperature and thus less likelihood that melting will occur.

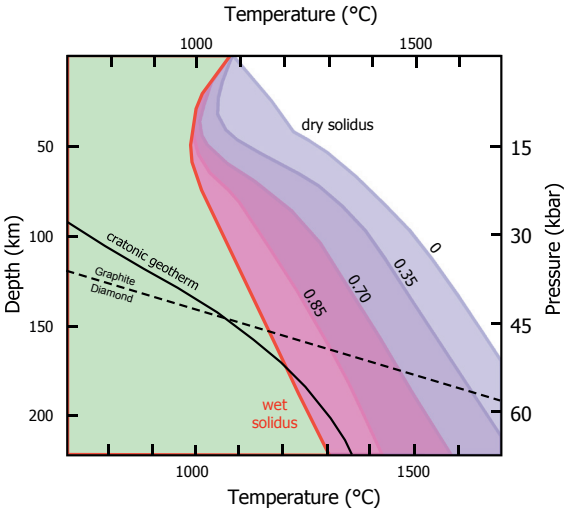


Figure 5.17 Solidi of peridotite with various H_2O-CH_4 mixtures in reducing conditions (after Taylor and Green, 1988). Numbers on melting curves refer to values of $H_2O/(H_2O + CH_4)$.

Melting with these reduced C-O-H mixtures is nevertheless relevant for high pressures, since the oxygen fugacity in the mantle is expected to decrease with increasing depth such that the fO_2 of the iron-wüstite buffer is probably reached at a depth of ≈ 250 km (Rohrbach *et al.,* 2007). This depth is exceeded by the deepest peridotite xenoliths in cratonic areas: oxybarometers confirm fO_2 conditions close to iron- wüstite for peridotites at the base of the cratonic lithosphere (Woodland and Peltonen, 1999). Reducing conditions may also have been more common at shallower depths early in Earth history (Aulbach and Stagno, 2016; Foley, 2022).

There are disappointingly few experimental determinations of melt compositions under reducing conditions, partly due to bad quenching problems (Fig. 5.18). The study by Liu *et al.* (2023) confirms expectations from the depolymerising effect of both CH_4 and H_2O that melts are silica-richer (53–55 wt. % SiO_2) than in anhydrous conditions. The high pressure melts measured by Jakobsson and Holloway (2008) were also SiO_2 saturated and exceptionally MgO-rich (25–33 wt. %) at 50 to 125 kbar. However, melting at the high end of this pressure range may not occur in the Earth's mantle. These melts are richer in MgO than melts of dry peridotite at equivalent pressures (Fig. 5.18; Walter, 1998). To circumvent severe quenching problems, Litasov *et al.* (2014) calculated melt compositions by mass balance, finding high MgO contents (25–31 wt. %) and SiO_2 saturation with only one nepheline-normative melt. They also interpreted texturally different but compositionally similar areas as fluid. We desperately need more information on melt compositions under reducing conditions, but it has proven extremely difficult to find funding to conduct such experiments.

5.4 Melting of Dry Pyroxenites

A few melting experiments on dry pyroxenites took place from the 1970s, focussing on xenoliths or lenses of pyroxenite in peridotite bodies such as Beni Bousera in Morocco (Kornprobst, 1970). However, although melts were produced, no compositions of melts were reported (hence the dotted line in Fig. 5.1), as the aim of the experiments was to investigate subsolidus mineralogy and whether the rocks could originate as high pressure cumulates from melts.

There was a rejuvenation of interest in the 1990s to address the question of the fate of subducted ocean crust in the peridotite mantle. Partial melts of eclogite were found to be silicic and would react with peridotite to produce pyroxenites (Yaxley and Green, 1998). Geochemical studies showed that the garnet signature in mid-ocean ridge basalts could be explained by a mixed peridotite + pyroxenite source (Hirschmann and Stolper, 1996). This progressed to the suggestion that the Hawaiian basalt source might be completely free of olivine (Sobolev *et al.*, 2005) before opinion settled to explain a range of oceanic and continental basalts as being formed by melting of different proportions of pyroxenite and peridotite (Lambart *et al.*, 2009).

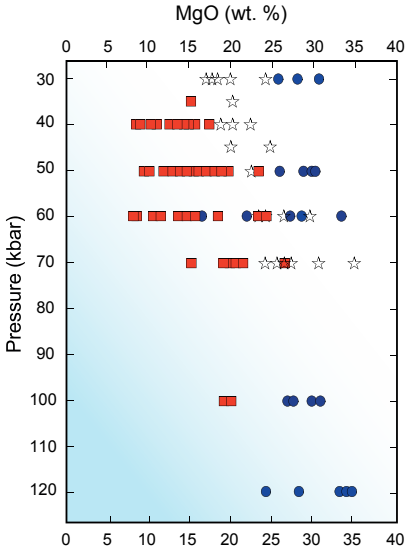


Figure 5.18 Compilation of MgO in experimental melts of peridotite. Blue circles = reduced conditions (H_2O + CH_4), red squares = oxidised conditions (CO_2 + H_2O), open stars = no volatiles added. Reduced melts have consistently higher MgO than oxidised melts and often higher than anhydrous melts.

A rash of experimental studies around 2000 addressed this question and showed that pyroxenites consistently melt at temperatures several tens of degrees lower than peridotite, thus in principle being the equivalent of dry mantle migmatites. A summary by Kogiso *et al.* (2004) concluded that a critical factor affecting melt compositions was whether the pyroxenite was silica saturated or not: silica-excess pyroxenites produced melts with >52 wt. % SiO_2, whereas silica undersaturated ones produced melts with <48 wt. % SiO_2. However, at this time the number of experiments on pyroxenites was limited, so this summary used many experiments on basalts in addition to pyroxenites. Of interest for the generation of alkaline basalts, further experiments on garnet pyroxenites showed that low degree melts could be silica undersaturated (nepheline-normative), possibly explaining the alkaline members of ocean island basalt suites. With continued melting, melts became silica saturated in a similar way to partial melts of peridotite in Figure 5.8.

5.5 Liquidus Experiments on Alkaline Rocks

Experiments investigating the liquidus temperatures and minerals in volcanic rocks or estimated primary melt compositions have had two principal purposes,

namely delineating fractionation at various pressures and defining the pressure, temperature and source rock that melted to produce them.

This type of experiment was common from the early 1960s through to the late 1980s (Fig. 5.1; Yoder and Tilley, 1962; Green and Ringwood, 1967; Esperança and Holloway, 1987), but became much rarer after that, with later studies focusing on specialist applications such as unusual compositions or tracing the fractionation of melts at mantle pressures with a series of experiments (Pilet *et al.*, 2009). My attention here will focus on just two aspects: (1) the survey of liquidus phases in a search for multiple saturation in source minerals, corresponding to the inverse approach of Figure 5.3, and (2) near-solidus experiments that were mostly a by product of the same experiments.

The philosophy of the inverse approach was to search for a point along the liquidus at which multiple saturation in several minerals occurred, which could then be interpreted to represent the mineral assemblage of the source rock. However, the common result was that several points occurred where two minerals coexist together at the liquidus, but three, let alone four minerals as in a peridotite source, were extremely rare. Examples are reproduced in Figure 5.19 for anhydrous tholeiites, olivine tholeiites and basanites (from Cohen *et al.*, 1967; Thompson, 1972; Arculus, 1975; Yoder, 1976).

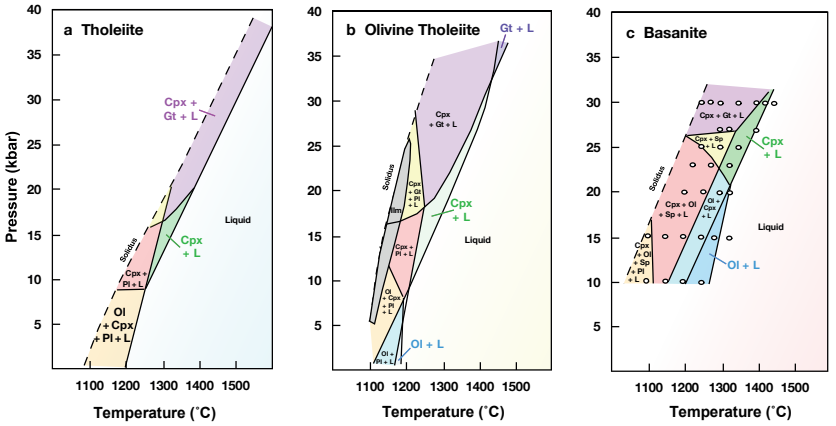


Figure 5.19 Examples of liquidus experiments on dry rocks. Diagrams are often published with no points representing individual experiments; these are left in in **(c)**. Note that lines can be moved several tens of degrees and still be compatible with experiments, illustrating the uncertainty in interpretations of these diagrams. Solidus dashed because no subsolidus experiments were run. Grey area in **(b)** has ilmenite in addition to minerals in fields at higher temperatures. Liquidus assemblages shown in coloured text. Diagrams after Cohen *et al.* (1967), Thompson (1972) and Arculus (1975).

Figure 5.3b shows a "successful" study in which all four peridotite minerals occur at a single point at the liquidus (from Fujii and Kushiro, 1977). Even here, there is leeway for interpretation, as most lines drawn on these diagrams can be moved 50 °C or 3–5 kbar and still be consistent with the experiments (Fig. 5.19c), depending on the number of experiments conducted and their closeness in pressure-temperature space. This means that a multiple saturation point is optimistic even where it is interpreted to exist.

The reason for failure to find multiple saturation in peridotite minerals was usually assumed to be attributable to poor choice of starting materials, namely that the volcanic composition chosen was not the primary melt it was thought to be. The obvious alternative, namely that the source was not a four phase peridotite, was not usually contemplated. I concur with Wyllie (1987) who noted that a negative result may be more useful than a positive one in the case of liquidus experiments.

When liquidus experiments on alkaline rocks became more fashionable, the uncertainties were compounded by the need for volatile components. Phlogopite featured strongly under near-liquidus conditions (Fig. 5.20), but it became clear that its appearance at the liquidus was promoted when large amounts of water were added, and it remained sub-liquidus in (more realistic) water undersaturated conditions (Barton and Hamilton, 1979; Foley, 1989). Furthermore, the choice of primary magma compositions on which to conduct the experiments often implicitly followed the peridotite melting model. The "identification" of a primary melt often uses the criteria of Mg# of 70 or higher and high Ni contents (Section 3). These are based on equilibrium between olivine and melt and intended for peridotite sources; they might not apply for a mica pyroxenite or other olivine-free source. A liquidus study of kimberlite was optimistically interpreted to show that the source was a phlogopite carbonate garnet lherzolite, but the phlogopite and carbonate were found only hundreds of degrees below the liquidus (Fig. 5.21; Eggler and Wendlandt, 1979).

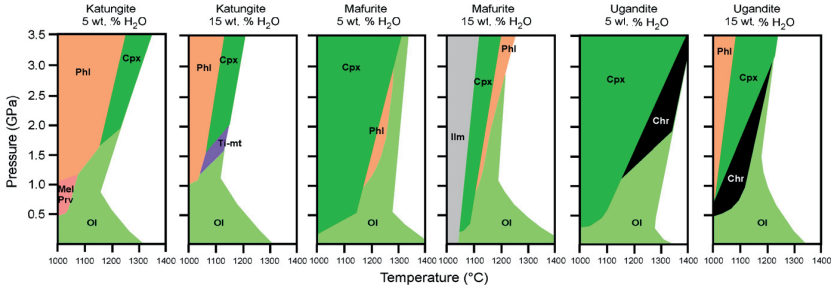


Figure 5.20 Liquidus experiments on kamafugites, showing large stability fields for clinopyroxene and phlogopite with minor oxide minerals at high pressures. Olivine is restricted to low pressures. After Foley *et al.* (2012).

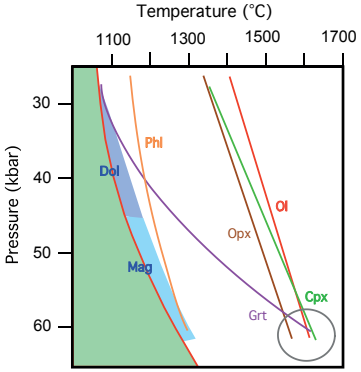


Figure 5.21 Liquidus experiments for kimberlite. The four peridotite minerals olivine, clinopyroxene, garnet and orthopyroxene are at or close to the liquidus at 60 kbar, but phlogopite and carbonate occurred only well below the liquidus. After Eggler and Wendlandt, 1979.

We are left with multiple uncertainties for liquidus experiments on alkaline rocks: maybe the rock chosen was not primary, we may not have enough experiments to define multiple saturation at the liquidus, or the choice of volatiles and/or their amounts may be wrong. A conservatively minded petrologist could form the opinion that these collected uncertainties were large enough to retain the *status quo* and allow the homogeneous peridotite mantle to survive. Strongly alkaline melts appear to require phlogopite or amphibole in their sources, but these could still, for many, be housed as additional minor phases in peridotite. These are classic *ad hoc* fixes, which are *"additional assumptions whose function is to save and protect the main hypothesis or theory in the case of contradictory facts"* (Sattler, 1986).

For the purposes of this article, there are largely overlooked results in many of these liquidus experiments which I wish to emphasise, namely the few near-solidus experiments. These show us which rocks would result if the melts were to solidify before they leave the mantle, forming the dykes, veins and patches that are seen in mantle rocks (Section 4). Since H_2O is an incompatible component, it will be concentrated in any melt and so anhydrous experiments are of limited usefulness here. Nevertheless, Figure 5.19 shows a tendency for oxide minerals (ilmenite, spinel) to appear close to the solidus, which is also the case for strongly alkaline rocks (Fig. 5.20). Other than this, we see clinopyroxene and phlogopite as the main silicate phases, with olivine retreating into the background as pressure rises. This is compatible with conclusions that many source rocks for lamproites and other potassic rocks are rich in mica and clinopyroxene (Barton and Hamilton, 1979; Arima and Edgar, 1983; Foley, 1989; Shu *et al.*, 2024), and with the occurrence of mica pyroxenites as mantle xenoliths from all continents (Section 4).

5.6 Reaction Experiments

Reaction experiments are a relatively new addition to the toolbox of experimental petrology, commonly used only since around 2010 (Fig. 5.1). They have been used to investigate two stages in the origin of alkaline melts: firstly, the reaction between melts and peridotite to produce non-peridotitic ultramafic rocks in the mantle (the neosomes of mantle migmatites) and, secondly, the melting of these migmatites to produce the alkaline rocks seen at the Earth's surface. We will look at the first type here, whereas the second type is considered in Section 6.

There are three main types of experimental set ups used, each with its own capsule configuration (Fig. 5.22). In the first (and oldest) two distinct rock types are placed one above the other as drilled cylinders, mineral mixes or rock powders (Fig. 5.22a). In the second, the two rock powders are mixed into a uniform starting material (Fig. 5.22b), and in the third, a melt trap is placed at one end of the capsule to capture melt that has flowed through the peridotite in order to assess changes in melt composition due reaction with the peridotite (Fig. 5.22c). The first two of these are relevant for the formation of migmatite neosomes. A problem with the first, layered, method is that the thickness of the reaction zone varies with melt viscosity: the zone can be very thin in the case of high viscosity silicic melts (*e.g.,* Wang *et al.,* 2017), whereas low viscosity carbonate melts may flow through the entire capsule, even in short experiments (*e.g.,* Gervasoni *et al.,* 2017). In several studies, both layered and mixed experiments were conducted, documenting that the results are similar, producing the same assemblages as reaction products (Wang and Foley, 2018; Zhang *et al.,* 2020).

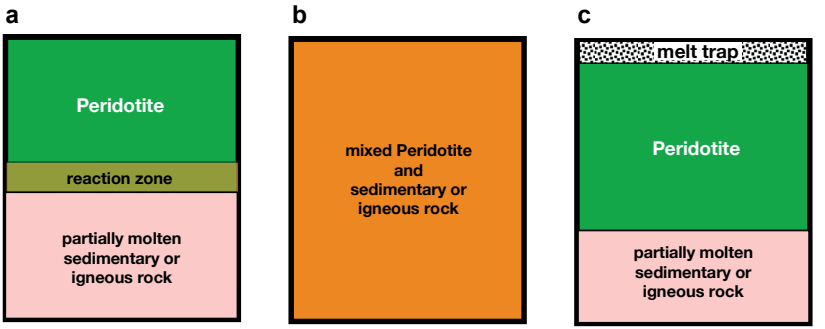


Figure 5.22 Capsule configurations for reaction experiments. **(a)** Two blocks of rock or rock powder are juxtaposed; the sedimentary or igneous rock partially melts, resulting in a reaction zone in contact with the peridotite. **(b)** The peridotite and second rock are intimately mixed before the experiment. **(c)** Flow through experiment: partial melt from the sedimentary or igneous rock (bottom) is drawn through the peridotite into a melt trap at the other end of the capsule. Modified after Prelević *et al.* (2024).

The first layered reaction experiments followed theoretical predictions that mica pyroxenites should result from the reaction of silicic melts released from the melting of sedimentary rocks in subducting slabs (Sekine and Wyllie, 1982b). Experiments juxtaposing peridotite with granite, chosen to represent the melt of sedimentary rocks, confirmed that metasomes always contained Opx and Cpx, plus phlogopite and garnet. Olivine is reacted out of the former peridotite, and in extreme cases of high melt/rock ratios, quartz may appear. The principle is reproduced in Figure 5.23, showing that a phlogopite pyroxenite plane lies between the melt and peridotite. This scheme was worked out by Sekine and Wyllie (1982b) on the basis of experiments in a simple system conducted at just 3 kbar. Later experiments at 28 and 55 kbar (Fig. 5.23b,c) showed that the stability field of phlogopite expands enormously towards high pressures, underlining the importance of these reaction processes and making the formation of mica pyroxenite metasomes unavoidable at high pressures. Later experiments in the dry system (*i.e.* without phlogopite) showed that garnet and the two pyroxenes form a thermal divide at high pressures, separating basaltic from dacite-like melts (Yaxley and Green, 1998) and that pyroxenites containing alkali-amphiboles could form at low melt/rock ratios (Rapp *et al.*, 1999). The latter study also presented a first characterisation of the trace element contents of melts in these mixed systems.

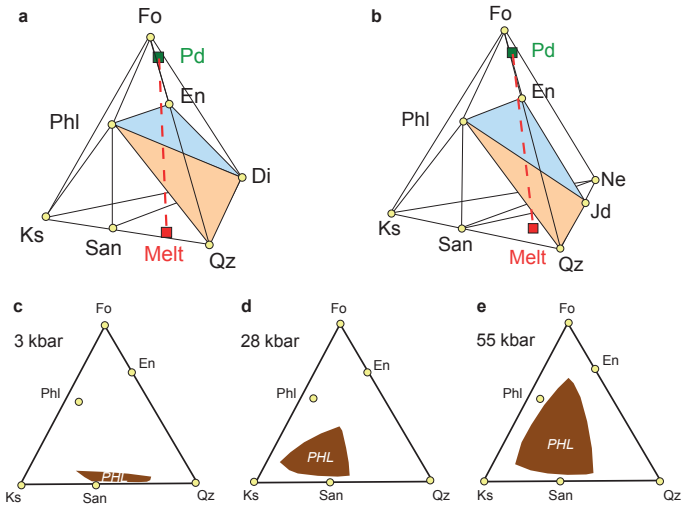


Figure 5.23 Sketches illustrating the reaction relationship between silicic melt with peridotite (Pd) to form phlogopite pyroxenites. Addition of melt to peridotite will move the system along the dotted line towards the melt composition, meeting the blue triangle Phl + Opx + Cpx. Quartz may result for high melt/rock ratios. Concept developed by Sekine and Wyllie (1982b) based on experiments at 3 kbar (lower left). Later experiments at higher pressure (Foley *et al.*, 1986 and unpublished data) showed a greatly enlarged phlogopite field, confirming this reaction at mantle pressures.

There was a resurgence in reaction experiments from 2010 onwards to study the effects of silicic melts in subduction zones and of carbonatite melts. These have consistently shown the dominance of phlogopite and clinopyroxene in the metasome assemblages formed by reaction with melts of clastic sediments and even basanite with peridotite (Förster *et al.*, 2019, 2020), including some spectacular images of the reaction zones (Fig. 5.24). Calcic amphiboles can be formed in these reaction zones at 15–25 kbar (Zhang *et al.*, 2020; Grützner *et al.*, 2023) and K richterite at 50 kbar (Gao *et al.*, 2023).

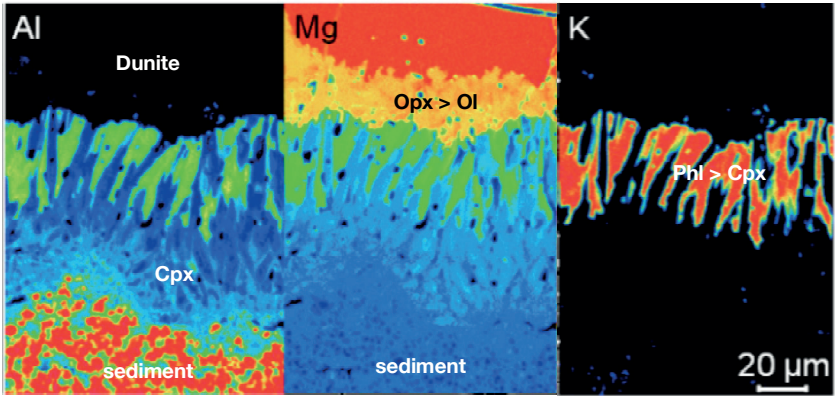


Figure 5.24 Laser-ICP-TOFMS maps of reaction zones produced by reaction of melts of marine sediment with dunite in experiments. Original sediment (bottom) and dunite (top) are separated by 3 zones: (1) reaction of olivine to orthopyroxene at the leading edge of melt migration (orange). (2) Phlogopite + clinopyroxene, (3) Clinopyroxene. Zones rich in Opx and Phl + Cpx are known from natural xenoliths and near-solidus experiments on alkaline volcanic rocks. From Förster *et al.*, 2021.

Following the experimental recognition that Na carbonatite may be produced by melting of peridotite in the presence of H_2O and CO_2 (Section 5.3.4), Green and Wallace (1988) predicted that the end product of reaction of this melt with lherzolite would be an apatite-bearing spinel wehrlite, and natural wehrlite xenoliths were soon identified and explained as the result of this process (Yaxley *et al.*, 1991). In the past few years, experiments have investigated the solid products of reaction of peridotite with carbonatite and carbonated silicate melts. Gülmez *et al.* (2023) found wehrlite products but also identified amphibole and carbonates, whereas Gervasoni *et al.* (2017) noted abundant oxide minerals in addition to the expected wehrlite. Websterite may also result in experiments between melted limestone and peridotite (Wang *et al.*, 2024). The publications by Gervasoni *et al.* (2017) and Pintér *et al.* (2022) describe products from reaction with carbonatite and carbonated silicate melts similar to the aillikitic melts found in high pressure experiments on peridotite with H_2O and CO_2 (Fig. 5.15). These agree that the main product from carbonatite + peridotite reaction is

clinopyroxene, whereby hydrous carbonatite also leads to the crystallisation of phlogopite and K richterite at 5 GPa (Pintér *et al.*, 2022). For carbonated silicate melts, clinopyroxene and hydrous minerals are less abundant in the reaction products and garnet becomes volumetrically more important.

Magnesite has been found in the solid products for carbonated silicate melts at higher pressures even where carbonate is only a minor part of the mixed silicate-carbonate sediment investigated (75–125 kbar; Bulatov *et al.*, 2014). Despite the very high pressures, these experiments may be relevant to deep subduction of carbonates and their reheating in the mantle transition zone, since the carbonate components in limestones do not melt along most modern subduction geotherms and so may be deeply subducted (Chen *et al.*, 2023). Melting of foundered sedimentary carbonates or carbonated eclogites in the mantle transition zone has been invoked for the origin of some nephelinites and alkali basalts to account for stable isotope signatures (Zeng *et al.*, 2021; Shu *et al.*, 2022).

6. EXPERIMENTAL MELTING OF MANTLE NEOSOME ASSEMBLAGES

Experimental work on hydrous mantle assemblages initially treated the individual minerals, followed by investigations of their stability in peridotites (Section 5). Melting studies of dry pyroxenites followed from around 2000, catalysed particularly by the hypothesis that they are important contributors to mid-ocean ridge basalts (Hirschmann and Stolper, 1996) and ocean island basalts (Sobolev *et al.*, 2005). However, experimental studies of the melting of potential hydrous mantle migmatite neosome assemblages have been very sparse until recently: the number of experimental studies on peridotites significantly outnumbers those on all other ultramafic rocks combined, despite the large range of possible dyke and vein assemblages being much greater than those of peridotite.

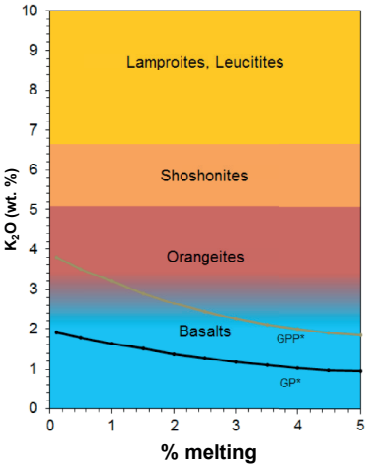


Figure 6.1 K_2O content of melts as a function of melt fraction produced by batch partial melting of garnet peridotite (GP*) and garnet phlogopite peridotite (GPP*) after Novella and Frost (2014). Vertical scale expanded to show that melting of phlogopite peridotite cannot produce shoshonites, leucitites and lamproites.

Experiments on hydrous ultramafic rocks have generally had one of three goals: (1) the characterisation of melt compositions, (2) the determination of solidus temperatures, or (3) the study of reactions between melts of them and surrounding peridotites. Although relatively uncommon amongst xenoliths compared to garnet and spinel peridotites, these hydrous rocks will be major contributors to melts as long as their melting points are lower than surrounding peridotites (Foley, 1992a; Foley *et al.*, 1999). A pioneer study by Lloyd *et al.* (1985) melted a mica pyroxenite from western Uganda expressly to examine its potential

as a source for kamafugite melts but did not report any subsolidus experiments. There has been a recent major increase in the number of studies useful for mantle migmatites (Table 6.1).

Table 6.1 Experimentally determined solidus temperatures for hydrous ultramafic rocks.

Ref	PHL	CPX	Ca-AM	Alk-AM	AP	ILM	RU	Ttn	TiMT	GRT	OPX	OL	T°C solidus	P kbar
1	42	57			1								1075 ±25	13.3
1	42	57			1								1125±25	17.7
1	42	57			1								1125±25	21.1
2			96			2		1					<1150	15
2		44	55		tr			tr					<1225	15
3	33	33		33									1037±12	15
3	32	32		32	5								1012±12	15
3	32	32				5							1012±12	15
3	32	32					5						1012±12	15
3	32	32		32		5							<1250	50
4			50		50								1048±12	15
4		10	45		45								1072±12	15
4	10		45		45								1072±12	15
4	10	10	40		40								1048±12	15
4	49	49			2								1022±12	15
3	28	28		28		15							<1125	15
3	28	28		28	15								<1125	15
3	28	28		28			15						<1150	15
5		23	63									15	1150±50	10
6		42	15			2				42			1060±20	20
7	33	33									33		1075±25	15
7	33	33									33		1125±25	30
7	33	33									33		1225±25	45
8	45	46			2	2		3	2				1225±25	27
8	20	50			3	5						22	1225±25	27
8	55	40			2				3				1225±25	27
8	45	46			2	2		3	2				≈1200	50
8	55	40			2				3				<1200	50
9	67	20		13		1	2						1260	30
9	67	20		13		1	2						<1300	35

1 = Funk and Luth (2013); 2 = Pilet *et al.* (2008); 3 = Foley *et al.* (2022); 4 = Foley *et al.* (2025); 5 = Grützner *et al.* (2023); 6 = Bowman *et al.* (2024); 7 = Shu *et al.* (2024); 8 = Innocenzi *et al.* (2024); 9 = Sweeney *et al.* (1993). The mineral mode for Ref.1 was taken from a subsolidus experiment at 13.3 kbar and for Ref. 9 from Waters (1987).

The necessity for these hydrous assemblages in addition to peridotites is illustrated in Figure 6.1. This plot is expanded from Novella and Frost (2014) to include shoshonites, lamproites and leucitites, showing clearly that these are unlikely to be produced by melting of phlogopite peridotites. Higher K_2O contents of 7–8 % were found by Condamine *et al.* (2016) in melts of a phlogopite-rich lherzolite at 30 kbar, attesting to the control of K_2O contents by the proportion of phlogopite in the source rock (Shu *et al.*, 2024). The K_2O contents of orangeites, which are barely ultrapotassic (Figs. 3.8 and 3.9), are at the limit for melts of phlogopite peridotite.

6.1 Melting Points of Hydrous Minerals

The first experiments on hydrous phases studied their stability fields in isolation. Pioneering studies of amphiboles preceded the introduction of well calibrated piston cylinder apparatuses and so were limited to crustal pressures (Boyd, 1959). These showed that amphibole melts incongruently, which has since been shown to *always* apply to amphiboles and phlogopite. Experiments on phlogopite followed the realisation that the mantle contains hydrous phases (Fig. 5.1; Kushiro *et al.*, 1967; Yoder and Kushiro, 1969) and on K richterite shortly afterwards (Kushiro and Erlank, 1970). Figure 6.2 summarises the stability fields of hydrous phases relative to the probable range of continental geotherms and shows that phlogopite, calcic amphibole and K richterite are all stable up to at least 1100 °C at lithospheric mantle depths. The green field for calcic amphibole refers to amphibole in peridotite, whereas the green dashed line shows an enlarged field is possible if the amphibole is not in peridotite, particularly if enriched in Ti or F (Forbes and Flower, 1974; Foley, 1991; Huckenholz *et al.*, 1992). In fertile peridotites the pressure limit for amphibole stability may be between these two lines (up to 38 kbar; Mandler and Grove, 2016). The phlogopite field is pinned between two experiments on MARID rocks (yellow circles; Sweeney *et al.*, 1993; Konzett *et al.*, 1997) as almost vertical at >1300 °C between 35 and 80 kbar but it reaches its pressure limit at 90–100 kbar (Trønnes, 2002). Phlogopite may be stable to higher pressures and temperatures in rocks other than peridotite, for which its maximum pressure is between 60 and 70 kbar (Konzett and Ulmer, 1999), but it may be compromised by the formation of minor phases in F-rich compositions (Harlow, 2022). The dashed brown line shows that phlogopite persists to higher temperatures in simpler bulk compositions, including phlogopite + clinopyroxene or phlogopite websterite in Fe-free systems (Modreski and Boettcher, 1973; Luth, 1997). However, these are not directly applicable to natural mantle compositions. The effect of Ti and F stabilises phlogopite by at least 100 °C in a similar way to calcic amphibole (Forbes and Flower, 1974; Foley, 1991); it was seen, for example, at 61 kbar and 1500 °C by Harlow (2022). K richterite was initially discounted as a phase at upper mantle pressures due to a reaction relationship with garnet (Kushiro and Erlank, 1970), but this need not apply in many neosome assemblages which are garnet-free

(Erlank and Finger, 1970). The K richterite maximum in Figure 6.2 is constrained by experiments from Konzett *et al.* (1997) and Trønnes (2002), and may extend to >140 kbar (Trønnes, 2002). Fluorine appears to have a huge stabilising effect (Foley, 1991) but this has not been investigated at >50 kbar.

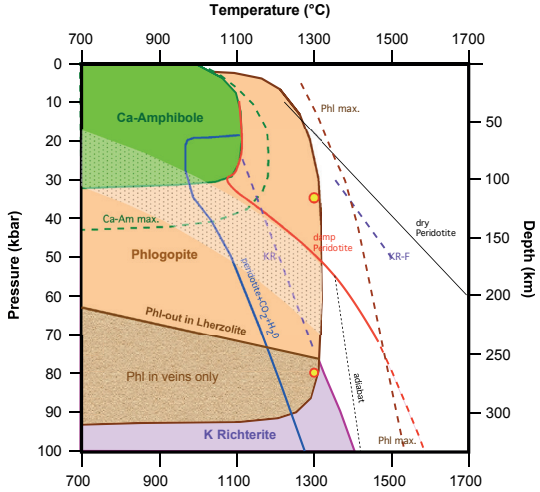


Figure 6.2 Pressure-temperature stability fields of phlogopite, Ca amphibole and K richterite from experimental constraints compared to a range of cratonic geotherms (stippled area, from Mather *et al.*, 2011) and asthenospheric adiabat for a surface temperature of 1280 °C (dotted line). Ca amphibole is stable to 100 km depth (green area) in peridotite but has an expanded maximum stability if rich in F and/or Ti (green dashed line). Phlogopite field constrained by melting of MARID (yellow circles, from Sweeney *et al.*, 1993; Konzett *et al.*, 1997). Phlogopite is replaced by K richterite in lherzolites at 60–70 kbar (brown line, Konzett and Ulmer, 1999) but may persist to higher pressures in veins (Trønnes, 2002). High pressure K richterite limits from Konzett (1997) and Trønnes (2002). Peridotite solidi from Green *et al.* (2014) for dry and damp, Pintér *et al.* (2021) for H_2O + CO_2. Dashed line for pure phlogopite from Yoder and Kushiro (1969) and Kushiro *et al.* (1967) and for Fluor-K richterite from Foley (1991).

Figure 6.2 also shows the range of cratonic geotherms from Mather *et al.* (2011). For geotherms in thinner continental areas, both calcic amphibole and phlogopite may be stable at the solidus in various ultramafic rock assemblages at 20–30 kbar and 1100–1200 °C. K-richterite may be largely restricted to cratonic mantle lithosphere and possibly deep subduction zones (Konzett and Ulmer, 1999), whereas calcic amphiboles are relevant for continental mantle away from cratons and in oceanic lithosphere.

6.2 Constraints on the Choice of Neosome Assemblages for Experimental Melting Studies

As detailed in Sections 4 and 5, the range of rock assemblages that may be relevant for the melting of mantle migmatite neosomes is immense and evidence for the mineralogical scope comes from three main sources: [1] xenolith assemblages, [2] near-solidus experiments on alkaline rocks, and [3] geochemistry and isotope signatures of igneous rocks. The most common non-peridotite xenoliths around the world are mica clinopyroxenites, which is consistent with the results of near-solidus experiments on alkaline rocks (Section 5). Such experiments are restricted in number as they are mostly by products of liquidus surveys, but they are important in showing the expected products if alkaline silicate melts were to crystallise as dykes before leaving the mantle. Many produce assemblages reminiscent of those known from xenoliths (Mitchell and Edgar, 2002; Foley *et al.*, 2012). Suggestions for source mineral assemblages from trace element and isotope geochemistry of alkaline igneous rocks includes mica and clinopyroxene for leucitites from Sierra Nevada (Elkins-Tanton and Grove, 2003), mica and orthopyroxene for Montana minettes (Carlson and Irving, 1994), and similar rocks with Ti oxides and apatite for Brazilian and Ugandan kamafugites (Carlson *et al.*, 2007; Innocenzi *et al.*, 2024) and for Australian orangeites (Choi *et al.*, 2021).

Many experiments on neosome melting have emphasised assemblages that are rare as xenoliths but may nevertheless be important in having the lowest melting temperatures and so be major contributors to igneous melt compositions. These include MARID group assemblages that occur in cratonic lithosphere and assemblages containing amphibole and apatite. The latter are known especially from eastern Australia, where these two minerals can make up 50–95 % of the rock (Wass *et al.*, 1980). Apatite in association with amphibole occurs widely, for example in Patagonia (Gorring and Kay, 2000), Laos (Konzett *et al.*, 2019) and Mongolia (Shao *et al.*, 1999), in the Red Sea (Piccardo *et al.*,1993) and in massif peridotites (Morishita *et al.*, 2003).

Reaction of infiltrating melts with peridotite may lead to enrichment in phlogopite at the margins of the dyke. A natural example is shown in Figure 6.3, showing phlogopite at the margins which is zoned from phlogopite produced by reaction at the dyke margin, to magmatic phlogopite further towards the centre before the crystallisation of clinopyroxene and richteritic amphibole follow. Many xenolith assemblages may be samples of such marginal reaction zones. An experimental analogue was produced by Foley and Pertermann (2021) who showed that phlogopite forms from the flow of lamproite melt through peridotite.

If we treat potential metasomatic agents in terms of initial melts of a primitive mantle composition at depths of 100–250 km, then two main groups of melts are to be expected for first generation metasomatic agents. The first is hydrous and rich in K based on the observation that H_2O has approximately the same incompatibility as K. Only very low amounts of both H_2O and K can be contained

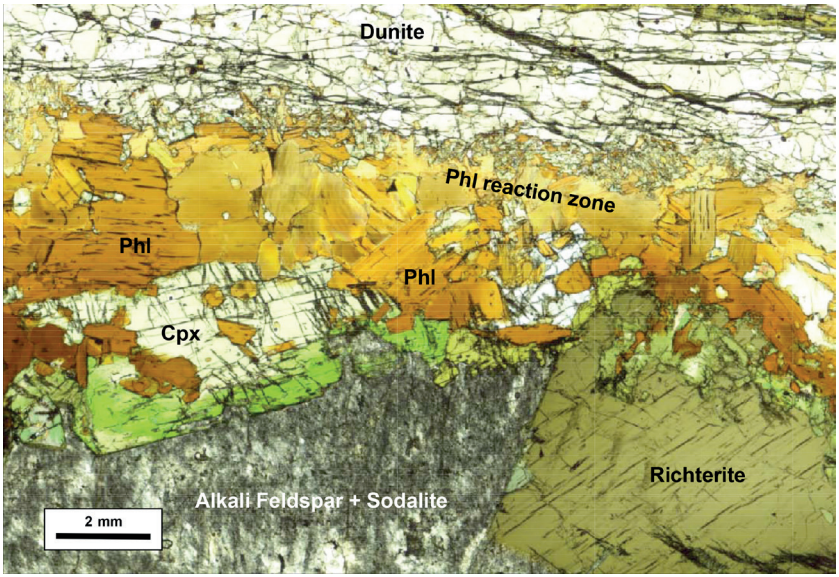


Figure 6.3 Mica + clinopyroxene-rich vein in dunite, Inagli, Siberia. Phlogopite at the edge is formed by reaction with the dunite wall rock and gives way to magmatic phlogopite (darker) towards the vein centre, followed by clinopyroxene and amphibole crystallisation.

in peridotite minerals, but both are important constituents of phlogopite, so that phlogopite-rich assemblages will inevitably result when these melts recrystallise. The second melt type is carbonated silicate melt, corresponding to partial melts of peridotite with CO_2 and H_2O, which are intermediate in composition between carbonatite and melilitite (Section 5). Hydrous K-rich assemblages have been at the centre of attention for experimental melting studies to date, whereas the crystallised products of carbonated silicate melts have been neglected up to now, even though they probably give rise to phlogopite-carbonate assemblages (Shatskiy *et al.*, 2022), which have been invoked as probable source components for alkaline magmas (Tappe *et al.*, 2008; Rosenthal *et al.*, 2009).

Phlogopite-rich ultramafic rocks have been summarised under the group name hydrous pyroxenites (Foley *et al.*, 2022; Ezad *et al.*, 2024a) but this is inappropriate for some that are not rich in pyroxenes, such as glimmerites, hornblendites, or amphibole-apatite rocks. The term hydrous ultramafic rocks appears unnecessarily cumbersome and does not exclude peridotites. Turner (1901) proposed the name perknite as a group name including mixtures of phlogopite, amphibole, clinopyroxene, orthopyroxene and mica. This was revived by Currie (1976), who suggested it be used for amphibole-pyroxene rocks. Perhaps we should resurrect this name for many mantle migmatite neosome

assemblages; if we allow apatite or garnet as a major constituent and limit olivine to less than 25 % then this covers all the rocks considered thus far in melting experiments in the next section.

The main question for melting experiments is to ascertain whether the melting points of the veins/dykes are lower than that of peridotite wall rocks. Figure 6.2 shows that hydrous rocks should be stable at the water-undersaturated ("damp") solidus of peridotite. Melting of peridotite in the presence of both CO_2 and H_2O may occur at lower temperatures, but such enrichment in CO_2 is likely to be very localised and not widespread (Foley *et al.*, 2024).

6.3 Melting Points of Hydrous Ultramafic Assemblages

Merely by dint of solid proof and reasoning I can never hope to overcome the inveterate prejudices of mankind.

Hume (1739)

Table 6.1 summarises experimental studies that have determined or approximated the melting points of hydrous ultramafic rocks at mantle pressures. Studies that did not bracket the solidus temperature (*i.e.* include a subsolidus experiment) but had a trace of melt are listed, whereas studies in which the lowest temperature experiment had a significant proportion of melt and so lay appreciably above the solidus are not. An example of the latter is the pioneering study of Lloyd *et al.* (1985) in which the lowest melt proportion was 23 %; the melting point must have been well below 1200 °C but was not determined.

Although the most common xenolith types are phlogopite clinopyroxenites, there have been relatively few experimental surveys of their melting points. On average, the hydrous assemblages in Table 6.1 have melting points 65 °C below the damp peridotite solidus and 285 °C below the dry peridotite solidus. There are even fewer experiments on melting of amphibole-clinopyroxene assemblages. The uncertainties in the melting temperatures listed are purely from the spacing of experiments; in some cases an interpolation of experiments at different pressures has allowed a more exact estimation. An example is the 30 kbar result for phlogopite websterite from Shu *et al.* (2024; ref. 7 in the table), which must be close to 1140 °C.

Several experimental surveys were not designed to bracket the solidus, concentrating instead on melt compositions. The temperatures given by Pilet *et al.* (2008; Ref.2) were listed as having a trace of melt present: just 15 °C higher there was 25 % melt for the hornblendite and 37 % melt at 1250 °C for the hornblende clinopyroxenite. Grützner *et al.* (2023) noted melt at 1100 °C in reaction experiments with peridotite, whereas the solidus of the pure clinopyroxene hornblendite at 1150 ± 50 °C is estimated from just two very widely spaced experiments.

In other experiments, Konzett (1997) noted a melting point for MARID of approximately 1100 °C between 30 and 60 kbar; this is not listed because of the lack of mineral modes. Förster *et al.* (2018) found 90 % melt in a MARID at 10 kbar and 1300 °C and Förster *et al.* (2017) found 77 % melt of a glimmerite at 10 kbar and 1225 °C.

Figure 6.4 compares the melting points of various hydrous neosome assemblages at 15 kbar – the pressure at which the widest variety of rocks have been studied. The solidi given for Funk and Luth (2013), Grützner *et al.* (2023) and Bowman *et al.* (2024) are interpolated from results at slightly different pressures (13.3–20 kbar).

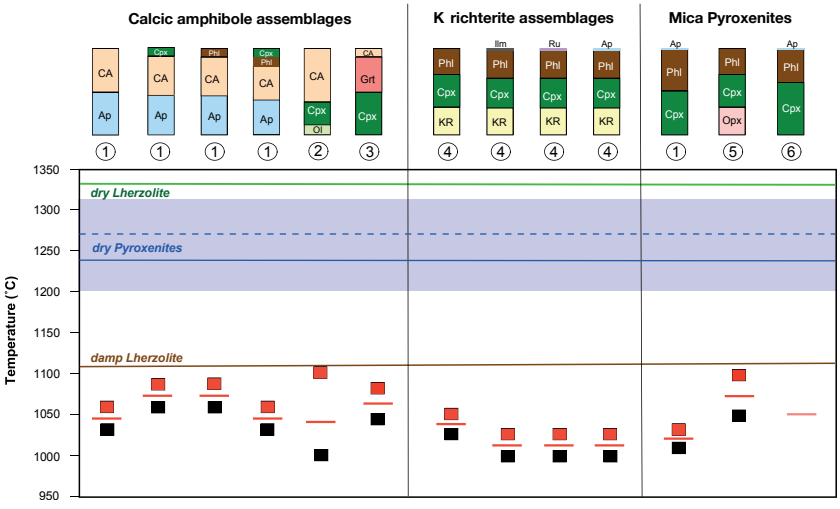


Figure 6.4 Experimentally determined solidi of mantle migmatite neosome proxy assemblages at 15 kbar. Mineral modes for hydrous rocks are shown as at the top. Red (suprasolidus) and black (subsolidus) squares indicate experiments that bracket the melting temperature (red dashes). Melting points for peridotite (dry = green; brown = damp) from Falloon and Green (1988). Blue lines are melting points of dry pyroxenites from Borghini *et al.* (2017) and Borghini and Fumagalli (2018, dashed) and Lambart *et al.* (2016; solid). Blue band = range of dry pyroxenite melting points (Lambart *et al.*, 2016). All hydrous ultramafic rocks melt before damp peridotite. Source references (in circles): 1 = Foley *et al.* (2025), 2 = Grützner *et al.* (2023), 3 = Bowman *et al.* (2024), 4 = Foley *et al.* (2022), 5 = Shu *et al.* (2024), 6 = Funk and Luth (2013).

The damp (water-undersaturated) peridotite solidus lies at a temperature (1105–1110 °C; Green *et al.*, 2014) which is 30–90 °C higher than all the assemblages with abundant amphibole and/or mica. The hydrous dykes would melt 225–285 °C below dry peridotite and at least 120 °C lower than dry pyroxenites (Fig. 6.4). There is thus a systematic difference between the melting points of hydrous ultramafic rocks and peridotite (dry or hydrous) and dry pyroxenites.

6.4 Melt Compositions

The following series of figures are chosen to show the variety of melts and to distinguish between melts of the various source mineral assemblages.

The Al_2O_3 *vs.* CaO plot (Fig. 6.5) clearly separates melts of assemblages containing alkali amphibole from those containing calcic amphiboles. Phlogopite pyroxenites occupy an intermediate position. Melting reactions consistently show that a high proportion of amphibole dominates the initial melt compositions: melting of K richterite therefore leads to low CaO and Al_2O_3 (Konzett, 1997; Förster *et al.*, 2017, 2018; Foley *et al.*, 2022); this applies also to MARIDs with low proportions of K richterite because K richterite melts before phlogopite. Pargasite/karersutite, on the other hand, has higher CaO and Al_2O_3, resulting in melts with >12 % Al_2O_3 (Pilet *et al.*, 2008; Bowman *et al.*, 2024; Foley *et al.*, 2025). The central position of phlogopite pyroxenite on this plot is more marked than first appears, because the few points with very low Al_2O_3 or CaO are from experiments in a simplified system with no FeO or Na_2O (Luth, 1997). Phlogopite pyroxenite melts form a zone that continues towards melts of phlogopite peridotite with higher Al_2O_3 and lower CaO (crosses; Conceiçao and Green, 2004; Condamine and Médard, 2014, Condamine *et al.*, 2016). Melts of phlogopite peridotite may have higher Al_2O_3 than any other hydrous ultramafic rocks and current data suggest they are also restricted to ≤10 % CaO (Fig. 6.5). The melts with highest Al_2O_3 are from phlogopite spinel peridotite, whereas phlogopite garnet peridotites give rise to lower Al_2O_3. Phlogopite melts more slowly than amphibole wherever they coexist (Foley *et al.*, 2022) but still dominates melting of hydrous rocks without amphiboles.

In the peridotite mantle paradigm, low CaO and Al_2O_3 in igneous melts is usually attributed to a harzburgite source that had been previously depleted by the loss of a basaltic melt from an initially fertile lherzolite. The experiments on hydrous ultramafic rocks show that low CaO and Al_2O_3 in igneous rocks (Fig. 6.5) is not necessarily a sign of a depleted peridotite source.

The K_2O *vs.* Na_2O plot (Fig. 6.6) narrates a relatively simple and logical story. Melts of rocks containing calcic amphibole have low K_2O/Na_2O, mostly <1, following the composition of the amphibole because these dominate the melting reaction. The green squares with lower Na_2O and K_2O/Na_2O up to 2 signify assemblages that are mostly calcic amphibole and apatite but also include a small proportion of phlogopite. Phlogopite-rich assemblages produce melts with high K_2O and K_2O/Na_2O commonly 5–10, that are, therefore, clearly separated from melts of calcic amphibole rocks in this plot. The circles signify assemblages with high contents of both K richterite and phlogopite. Natural K richterites have K_2O/Na_2O ≈1: the melts have higher ratios partly because of the presence of phlogopite and partly because both melt incongruently. This increases the K_2O content of melts, since no other minerals take up K_2O whereas Na_2O is sponged up by clinopyroxene. The Na_2O contents and K_2O/Na_2O thus depend on the

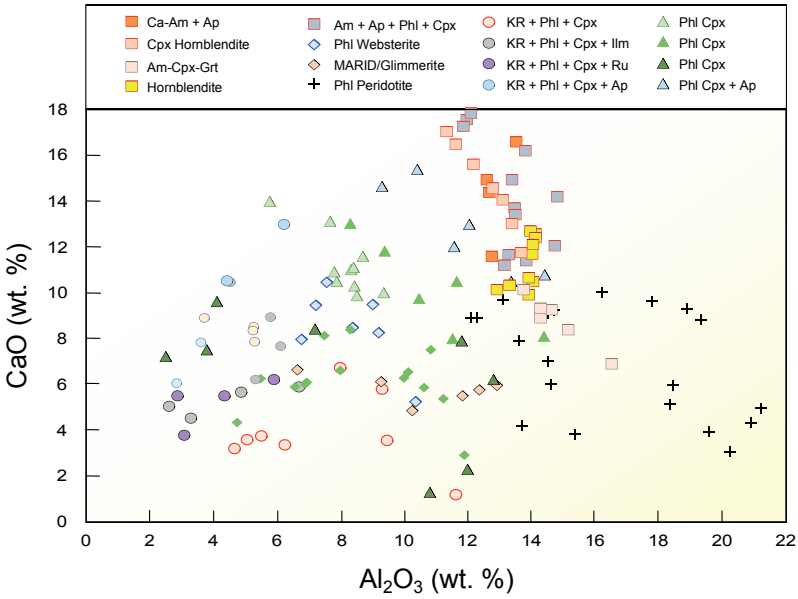


Figure 6.5 Compositions of experimental melts in Al_2O_3 and CaO. Melts from assemblages with Ca amphiboles are closely grouped in Al_2O_3 with high CaO, showing restricted overlap with melts of phlogopite peridotite. Melts of phlogopite clinopyroxenites lie between these and melts of K richterite assemblages. Square symbols designate mineral assemblages with calcic amphibole, triangles are mostly phlogopite clinopyroxenites, and orange diamonds are rocks dominated by phlogopite, namely MARIDs and glimmerites. Circles are rocks that contain abundant K richterite and blue diamonds are phlogopite websterites. Black crosses are melts of phlogopite peridotite for comparison.

abundance of clinopyroxene and pressure. The melts of phlogopite peridotite with Na_2O contents of 4–5 % are all low degree melts for which clinopyroxene contributes to the melts (Condamine and Médard, 2014; Condamine *et al.*, 2016).

The SiO_2 contents of melts also express the composition of the amphiboles in the melting assemblage: melts of calcic amphibole-rich assemblages are closely grouped at 35–45 % SiO_2, whereas K richterite assemblages have 48–56 % SiO_2 except where their compositions are influenced by additional apatite or Ti oxides. Apatite has been found to influence the incongruent melting reaction of K richterite, eliminating peritectic olivine in favour of clinopyroxene and orthopyroxene (Foley *et al.*, 2022). Lower SiO_2 in the melt results from the higher SiO_2 of pyroxenes (>50 %) relative to olivine (≈40 %). Pyroxenes are stabilised by increasing pressure, which leads to lower SiO_2 and higher MgO in melts

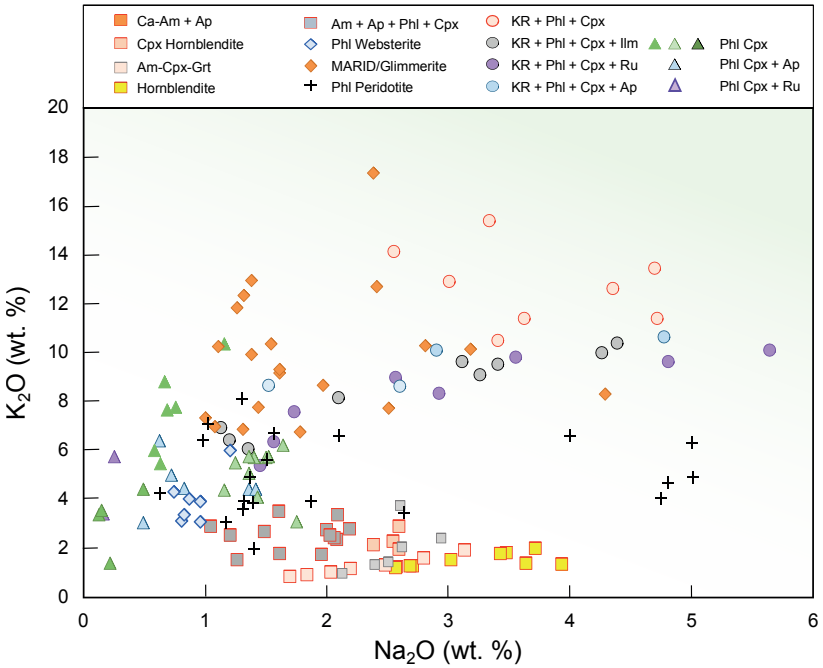


Figure 6.6 Na_2O and K_2O compositions of experimental melts, distinguishing high Na_2O/K_2O melts from Ca-amphibole assemblages from low Na_2O/K_2O melts from phlogopite and K richterite assemblages.

produced at high pressures. This explains the occurrence of olivine lamproite melts at pressures of the lower lithosphere (Foley, 1993). Melts of mica pyroxenites without K richterite are richer in CaO and Al_2O_3 and do not resemble lamproites.

Note the log scale for the K_2O/Na_2O axis in Figure 6.7, which was chosen to emphasise differences in the melts of calcic amphibole assemblages: the dark grey squares contain phlogopite in addition to abundant calcic amphibole (Foley *et al.*, 2025). Phlogopite-rich rocks, including those with abundant clinopyroxene (triangles), MARID and glimmerites mostly have intermediate SiO_2 (45–50 %), which is also true of phlogopite websterites. An exception is the biotite pyroxenites from Uganda (dark green triangles): these contain 8–9 % of titanite, magnetite and apatite, which can depress the SiO_2 of melts as noted above for K richterite-phlogopite assemblages. Initial melts of phlogopite peridotite melts may be extremely rich in SiO_2 (58–63.5 wt. %; Condamine and Médard, 2014). These are low degree melts (1–6 %) which have no analogues in the experiments on K richterite assemblages, which melt rapidly to 20–50 % melting (Foley *et al.*, 2022).

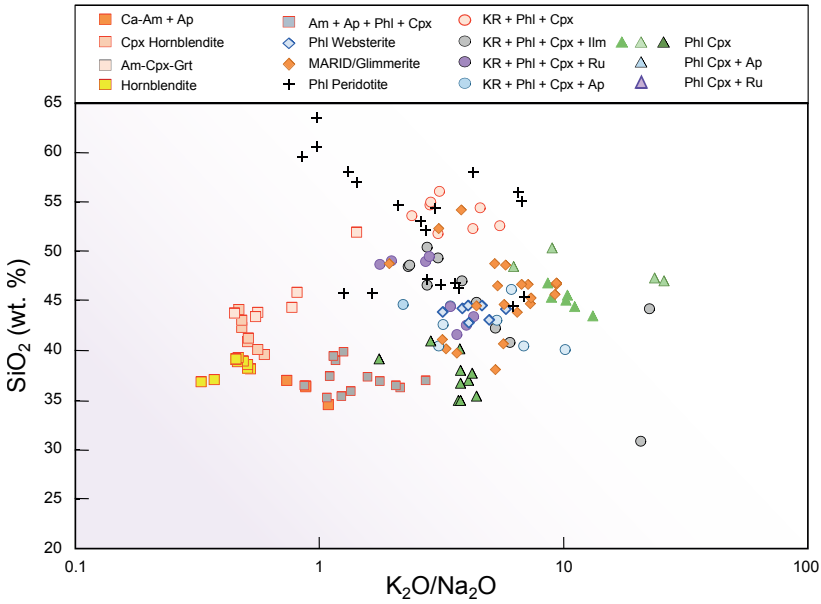


Figure 6.7 K_2O/Na_2O *vs.* SiO_2 compositions of experimental melts. Log scale chosen for K_2O/Na_2O to distinguish between melts of different Ca amphibole assemblages.

Figure 6.8 investigates the TiO_2 contents of melts, which are known to be variable and geographically characteristic for potassic rocks (Fig. 3.11). For lamproites, this has been attributed to differences and abundances of accessory phases (Foley and Ezad, 2024). In Figure 6.8, experimental melts of the different starting assemblages occupy distinct regions with little overlap. Of the calcic amphibole-rich assemblages, only the hornblendite with the highest TiO_2 contain minor Ti oxides (Pilet *et al.*, 2008); the moderate TiO_2 of the other melts is gained from the silicate minerals, particularly amphibole.

The phlogopite pyroxenite melts with the highest TiO_2 (6–8 wt. %) are those from Uganda, which contain oxide minerals and also Ti-rich micas. MARIDs and glimmerites contain ilmenite and rutile but only as minor constituents. The TiO_2 richest of these melts (5.6–9.2 wt. %) are from a synthetic mix with 5 % added TiO_2 (Konzett, 1997). The effect of accessory oxides on melt compositions is best illustrated by the K richterite assemblages (circles; Foley *et al.*, 2022): the red circles have no accessory phases, whereas there is 5–15 % ilmenite (grey) or rutile (purple) for the circles with very high TiO_2. Mixes with rutile contribute more TiO_2 than those with ilmenite, with 14–17 wt. % TiO_2 for 15 wt. % rutile, 11–13.4 wt. % TiO_2 for 15 % ilmenite, 11.0–11.6 wt. % TiO_2 for 5 % rutile, and 6.6–10.7 wt. % TiO_2 for 5 % ilmenite.

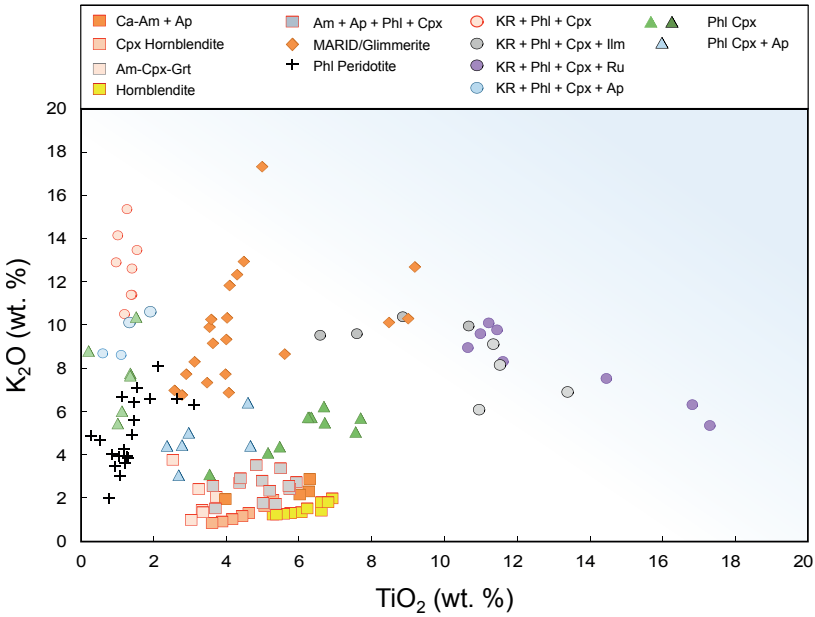


Figure 6.8 TiO_2 *vs.* K_2O compositions of experimental melts. Extremely high TiO_2 contents are melts of assemblages with high amounts of ilmenite or rutile.

The P_2O_5 contents in igneous melts, particularly alkaline melts, are often too high to be explained by partial melting of apatite-free peridotite. Therefore, experiments on hydrous assemblages with apatite may be especially relevant for the melting of mantle migmatite neosomes. Several assemblages containing apatite have been studied, some as minor, unintended accessories, others in higher proportions (5, 15 or 50 %) and with differing F/OH. Results are plotted in Figure 6.9, in which melts with less than 2 % P_2O_5 are all from assemblages with trace amounts of apatite (experiments with no apatite are excluded from this plot). The blue circles are melts from experiments with 5 or 15 % apatite (5.6–9.0 % P_2O_5) but are NOT the assemblages that contained the most apatite. The calcic amphibole-apatite series of experiments (Foley *et al.*, 2025) have 42.5–50 % apatite, yet P_2O_5 contents are only 2 to 5 % in the melt. This difference is due to the higher F contents of the calcic amphibole-apatite series of experiments, showing a huge effect on the melting reaction and thus the P_2O_5 contents of the melts. Where the apatite is more OH-rich, it contributes more to the melt, resulting in higher P_2O_5 contents in melts. The F/OH ratio will also affect the melting temperature, with higher temperatures for fluorine richer compositions. This means that the melting temperature of the calcic amphibole-apatite series

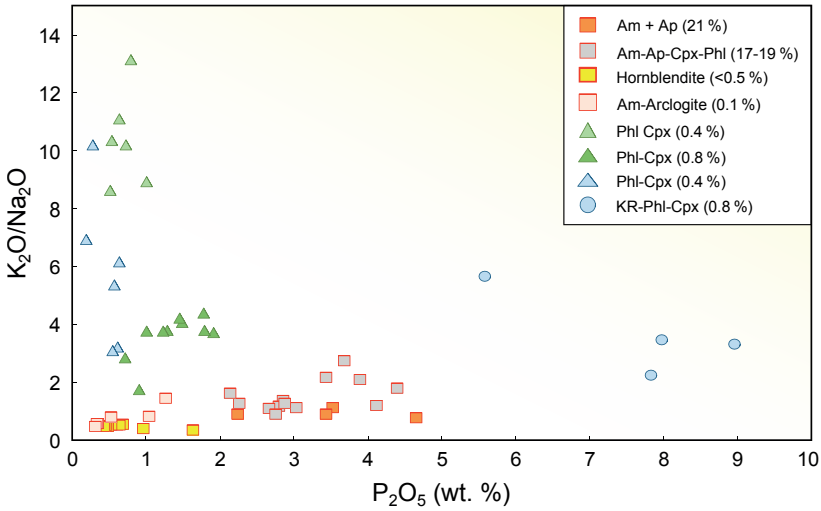


Figure 6.9 Phosphorus compositions of experimental melts as a function of K_2O/Na_2O. Figures in brackets are P_2O_5 (wt. %) contents of the unmelted bulk assemblages. Highest P_2O_5 contents are in melts with just 5 % apatite in the starting mix. Arclogite is a garnet clinopyroxenite with minor calcic amphibole.

would be further depressed if the apatite were richer in the hydroxyapatite component, enhancing the lower melting temperature relative to peridotite and dry pyroxenite shown in Figure 6.4.

It is clear from the numerous experiments now available for the melting of hydrous ultramafic rocks that assemblages with accessory phases produce melts exceptionally rich in P_2O_5 or TiO_2, depending on the phases present (Figs. 6.8 and 6.9). These melts are unlike any melts seen at the Earth's surface, but they may be important agents of mantle metasomatism seen in xenoliths. This may explain veins and dykes of the IRPS series (ilmenite-rutile-phlogopite-sulphide; (Harte *et al.*, 1987) of the Kaapvaal craton and phlogopite pyroxenites rich in Ti oxides and/or apatite in Uganda (Fig. 4.11).

7. WHERE ARE WE AND WHERE ARE WE GOING?

In this section, I consider how much progress has been made in the 3–4 decades of development of the mantle migmatites concept to assess where we are and what future research should concentrate on. I integrate the information and conclusions from earlier sections and summarise what has changed.

7.1 Comparison of Experimental Melts with Primitive Mantle Derived Melts

We are now in a position to compare the primitive mantle derived melts outlined in Section 3 with melt compositions that could be produced by melting of hydrous ultramafic rocks (Section 6). The data basis for this comparison has improved vastly over the last few years: at the time of my vein melting hypothesis in 1992 only 10 melt compositions were available from a single experimental study of biotite pyroxenite (Lloyd *et al.*, 1985) that was not designed to locate the solidus. The plots in Sections 6 and 7 show that we now have an additional 151 melt compositions from a dozen further studies. This is still a pittance compared to experiments on peridotite.

Several studies on neosome assemblages by my own research group have been slow in producing published results: they were started in the 1990s (van der Laan and Foley, 1994; Foley *et al.*, 1999) but published in full only recently (Foley *et al.*, 2022, 2025; Foley and Ezad, 2024). These delays were mostly due to the success of post-docs in moving on to jobs outside academia coupled with my own moves between universities and countries and other responsibilities such as Centre Directorship. I have often thought I could have been more productive if I had been able to stay in one place throughout my career as many people do.

The identity of likely neosome assemblages has been constrained by natural xenoliths and by forward and reverse experimental approaches. Nevertheless, the mineralogical scope of possible neosome assemblages is huge and only a restricted number have been studied experimentally to date. More experimental studies are needed, particularly to investigate assemblages that contain carbonate and reaction products formed during the infiltration of melts. Evidence from xenoliths and reaction experiments has outlined mineral assemblages of potential interest, but melting experiments still lag behind.

Figures 7.1 to 7.3 compare the compositions of experimental melts of hydrous ultramafic assemblages with approximate compositions of primitive mantle derived alkaline rocks. The data are the same as used in Figures 6.5 to 6.7 and Figures 3.9 to 3.11, whereby coloured areas for the natural rocks encompass around 90 % of data points instead of presenting average compositions as in Section 3. The variation in CaO and Al_2O_3 (Fig. 7.1) shows melts of K richterite

+ phlogopite assemblages overlapping mostly with anorogenic lamproites and orangeites, whereas melts of calcic amphibole assemblages broadly mimic the basanite-nephelinite-melilitite series. This is the same plot used to distinguish ultrapotassic (Fig. 3.3) and other mantle derived rocks (Fig. 3.10). Melts of phlogopite pyroxenites fall together with the basanite-nephelinite-melilitite series but also with kamafugites, alnöites and the Ca-richest orogenic lamproites. Melts of MARIDs and glimmerites plot together with lamproites. Also plotted in Figure 7.1 are Roman Province leucitites, the third original end member of ultrapotassic rocks (Fig. 3.3): their high Al_2O_3 and CaO is not reproduced by any of the hydrous ultramafic rocks yet investigated experimentally. This may support the interpretation that they are derived from veins rich in both calcic amphibole and phlogopite, which have not yet been the subject of experimental melting studies (Bracco-Gartner *et al.*, 2023).

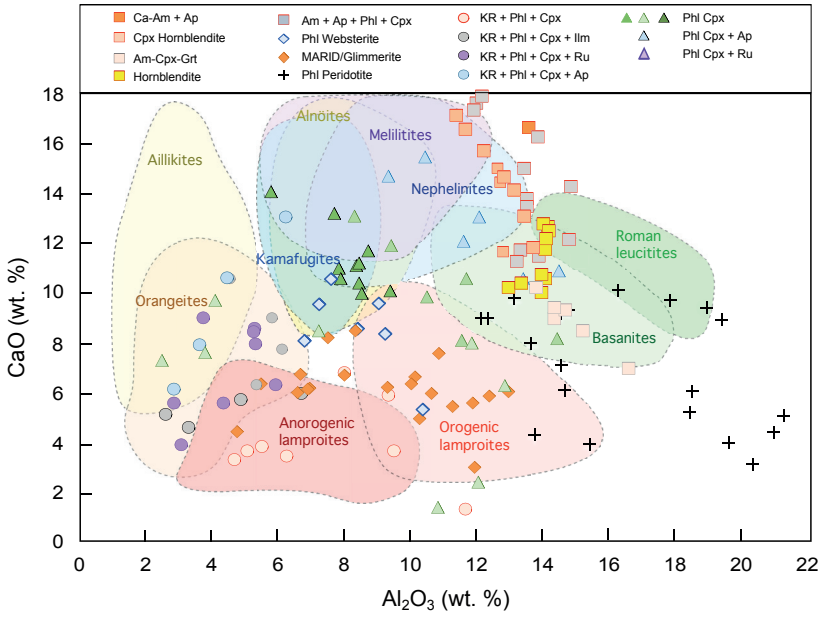


Figure 7.1 Comparison of Al_2O_3 and CaO contents of experimental melts of hydrous ultramafic rocks with primitive mantle derived alkaline rocks. Squares = mineral assemblages with calcic amphibole; triangles = phlogopite clinopyroxenites; orange diamonds = rocks dominated by phlogopite (MARIDs and glimmerites); circles = rocks with abundant K richterite; blue diamonds = phlogopite websterites. Black crosses = melts of phlogopite peridotite. Coloured areas for natural rocks encompass ≈90 % of data points, eliminating outliers.

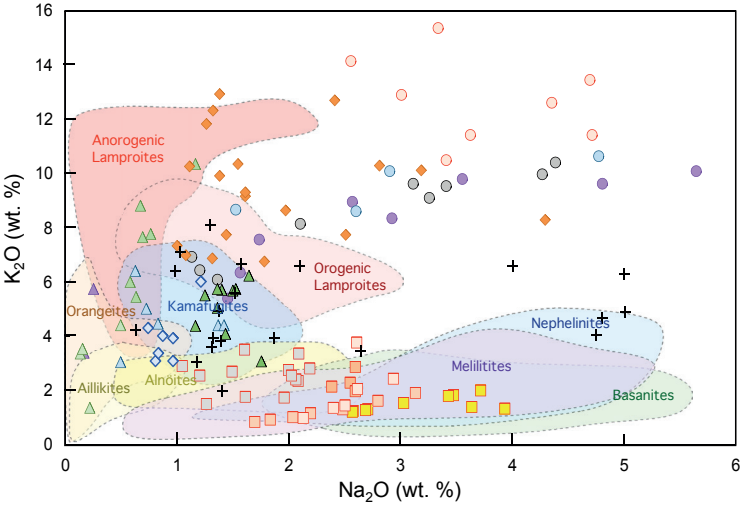


Figure 7.2 Comparison of Na_2O and K_2O contents of experimental melts of hydrous ultramafic rocks with primitive mantle derived alkaline rocks. Coloured areas for natural rocks encompass ≈90 % of data points, eliminating outliers. Symbols for experimental melts as in Figure 7.1.

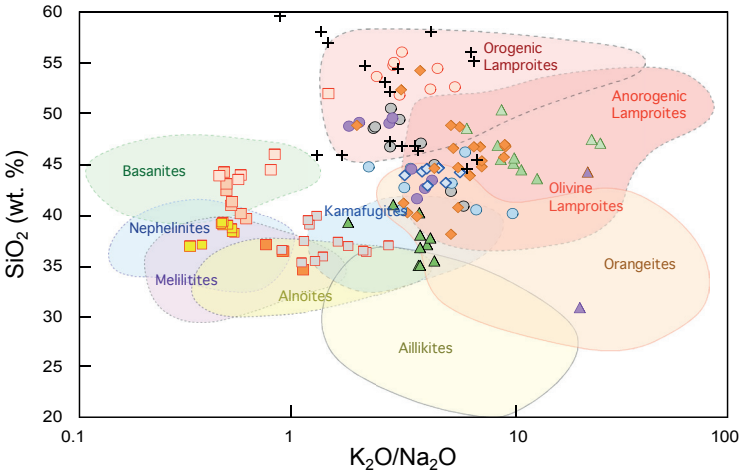


Figure 7.3 Comparison of K_2O/Na_2O and SiO_2 contents of experimental melts of hydrous ultramafic rocks with primitive mantle derived alkaline rocks. Coloured areas for natural rocks encompass ≈90 % of data points, eliminating outliers. Symbols for experimental melts as in Figure 7.1.

The Na_2O *vs*. K_2O plot (Fig. 7.2) shows calcic amphibole assemblages overlapping almost exclusively with the basanite-nephelinite-melilitite series. Addition of 10 % phlogopite moves the melts away from the basanite-nephelinite-melilitite series to overlap with alnöites. Many melts of phlogopite-rich ultramafic rocks may give rise to melts with high K_2O/Na_2O. Many melts from the K richterite series experiments have high K_2O similar to anorogenic lamproites, but higher Na_2O, which probably indicates that the K richterite/phlogopite ratio in the experiments is too high. The Na_2O content of the melts will also be lower at high pressures (30–60 kbar) where residual clinopyroxene sequesters more Na_2O. I reiterate that melts of many of the assemblages with apparently excessive amounts of ilmenite, rutile or apatite may not be seen as melts at the surface but may nevertheless be realistic for metasomatic agents in the mantle. This plot appears to eliminate the possibility that calcic amphibole can be present in the source of lamproites, kamafugites, orangeites and aillikites.

The K_2O/Na_2O *vs*. SiO_2 plot was used by Foley *et al.* (2025) to distinguish melts from phlogopite and calcic amphibole assemblages. Figure 7.3 is a modification in which the x axis is shown as a log scale to better illustrate the relationship of calcic amphibole assemblages to the basanite-nephelinite-melilitite series. This confirms the overlap of the basanite-nephelinite-melilitite series with melts of calcic amphibole assemblages, as well as the similarity of phlogopite-bearing examples to alnöites. Phlogopite and K richterite-rich assemblages produce melts similar to lamproites and there is once again a good match between kamafugites and melts of the Ugandan biotite pyroxenites (Fig. 7.3). The low SiO_2 at high K_2O/Na_2O of aillikites and many orangeites cannot be reproduced by melting of the assemblages that have been experimentally studied to date.

The experimental melts outlined in Figures 7.1 to 7.3 and in Section 6 consistently document an association of calcic amphibole-rich assemblages with the basanite-nephelinite-melilitite series and of phlogopite + K richterite with lamproites. However, they also illustrate the need for melting experiments on additional assemblages, most likely carbonate-bearing ones for orangeites and aillikites and rocks rich in both phlogopite and calcic amphibole for some leucitites. The results also consistently support an alternative origin to melting of peridotites for some members of the basanite-nephelinite-melilitite series by melting of calcic amphibole-rich source rocks (Foley *et al.*, 1999; Pilet *et al.*, 2008).

7.2 An Update on the Concept of Vein Melting

New ideas are like delicate plants; they deserve to be nurtured for a time and allowed to unfold their full potential, rather than being uprooted and dismissed out of hand.

Peat (1987)

The vein melting model (Foley, 1992a) followed reviews of liquidus experiments on potassic rocks (Foley, 1992b) that showed that derivation from a peridotite source, even if hydrous, was unlikely. Until this time, I had favoured a mica harzburgite source based on experiments in the simple system kalsilite (Ks) - forsterite (Fo) - silica (Qz) (Foley *et al.*, 1986).

Experiments on ultrapotassic rocks aimed at finding peridotite minerals at the liquidus found olivine restricted to low pressures: only strained explanations claiming artificially expanded olivine stability could approach peridotite mineral saturation at 25–30 kbar for SiO_2-rich lamproites (Barton and Hamilton, 1982). In contrast, phlogopite and clinopyroxene were common in experiments on all three ultrapotassic rock groups (Fig. 5.23; Barton and Hamilton, 1979, 1982; Arima and Edgar, 1983; Nicholls and Whitford, 1983). A separate rock rich in phlogopite and clinopyroxene appears to be the best explanation (Section 5) and hydrous veins in composite xenoliths support this conclusion.

The implied paradigm shift from melting of peridotite to the melting of mantle migmatite neosomes revealed immense gaps in our understanding of the melting of olivine-free ultramafic rocks. The non-peridotitic parts of the upper mantle are largely olivine-free ultramafic rocks, very few of which had been studied experimentally. Due to the prevailing peridotite model a massive number of experiments had been done on peridotite, but few on other ultramafic rocks, despite their mineralogical scope being much larger. In this experimental vacuum, my 1992 paper concentrated on the effects of variations in wall rock composition and melting conditions rather than on the melt compositions derived from the veins (Fig. 10 in Foley, 1992a); the data was simply not available. It later transpired that volcanic rock compositions are primarily controlled by the nature of the veins and not of peridotite wall rocks.

Several lines of evidence favoured the vein melting model. Composite xenoliths from around the world indicated that hydrous minerals commonly form dykes in peridotite (Section 4). Thermobarometric results showed that veins can form to depths of at least 170 km (Harte *et al.*, 1975) and probably much deeper in cold conditions, indicating the operation of channelled flow in the deepest lithosphere. The now mostly abandoned liquidus experiments failed to yield the sought for multiple saturation in peridotite minerals at the liquidus but instead showed that clinopyroxene and phlogopite are widespread as near-solidus minerals (Section 5). Radio-isotope studies often required multiple components that are easy to accommodate in sources composed of mixed rock types; these studies also indicated varied periods of storage in the lithosphere prior to re-melting. Trace elements provided geochemical indicators for minerals involved during melting that are not stable in peridotite, including several Ti oxides and apatite.

The melting mechanism envisaged in 1992 is illustrated in Figure 7.4. The concept was originally applied to deep continental mantle conditions and did not consider the full scope of melts considered in this review. Under these conditions the amphibole is K richterite, whereas calcic amphiboles will generally

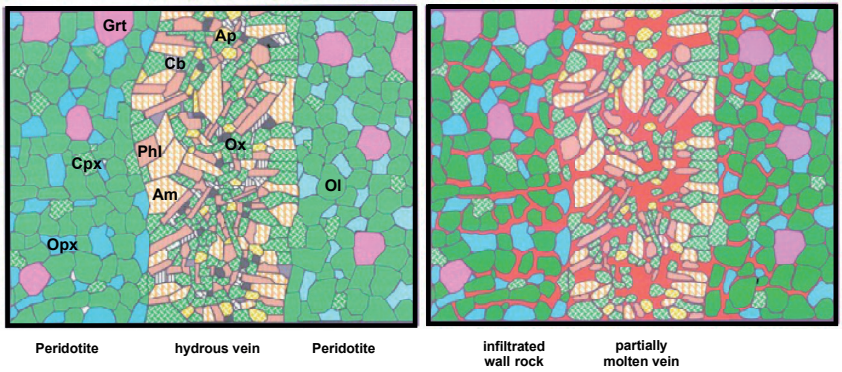


Figure 7.4 Sketches illustrating the 1992 vein melting model. **(a)** The unmelted hydrous vein was hypothesised to have abundant phlogopite, clinopyroxene and K richterite plus accessory oxides, apatite and carbonate. **(b)** Melting begins in the vein due to the lower melting point of the hydrous assemblage and melt is drawn into the peridotite wall rock along chemical potential gradients stemming from disequilibrium between the alkaline vein melt and peridotite. Modified from Foley (1992a).

be present in shallower mantle regions. There are exceptions to this geodynamic simplification, such as K richterite xenocrysts in olivine nephelinites in Morocco and edenite-pargasite in kimberlitic xenoliths at Jagersfontein in South Africa.

That melting occurs first in hydrous veins can now be taken as accepted: Figure 7.5 is a modification of a figure in the 1992 paper, replacing the concept diagram with melting curves now defined by experiments. This shows that hydrous veins may be as much as 60 % molten when the lowest melting temperature for peridotite is reached. This leaves a temperature interval in which vein melts exist alone, which corresponds to the activity of mantle metasomatic agents. Melting may proceed to involve wall rocks, particularly in the cases of decompression due to lithospheric thinning or of an external heat source. The latter applies to flood basalt regions, where alkali picrites may be the eventual products that bear the incompatible element signatures of hydrous veins but major element compositions typical for substantial melting of peridotite (Ellam and Cox, 1991).

Figure 7.4b foresaw the infiltration of vein melts into the peridotite wall rock. This was recognised as depending on the incongruent melting reactions of the hydrous phases and on factors such as the Mg-number of the minerals in vein and wall rock. As the veins form, minerals formed in a proximal position will have higher Mg/Fe than those formed in a distal position after the melt has fractionated during flow through the crack (Fig. 7.6). The question in the 1990s was whether melt would infiltrate the wall rock or not, whereas later work

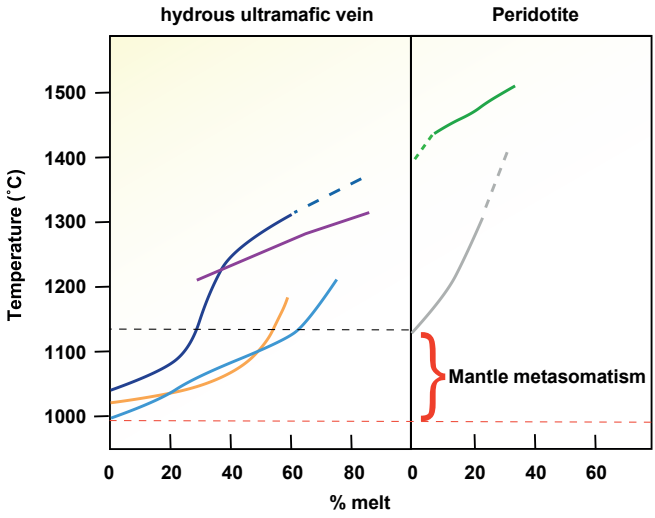


Figure 7.5 Different melting points of hydrous vein assemblage (left and peridotite). The conceptual diagram from Foley (1992a) is replaced here with experimentally determined curves: Light Blue = Clinopyroxene + Phlogopite + Apatite (Foley *et al.*, 2025); Yellow = Amphibole + Apatite + 10 % Clinopyroxene (Foley *et al.*, 2025); Dark blue = Phlogopite Clinopyroxenite (Funk and Luth, 2013); Purple = Biotite Pyroxenite (Lloyd *et al.*, 1985); Grey = damp Peridotite (Green *et al.*, 2010; Condamine and Médard, 2014; Condamine *et al.*, 2016); Green = dry Peridotite.

based on reaction experiments and natural xenoliths showed that the importance of reaction assemblages at the junction between vein and wall rock had been overlooked.

Particularly the experimental work published since around 2010 has permitted much more specific models for the mineral assemblages in the source regions of alkaline rocks, both in hydrous veins and in peridotites. Examples are clinopyroxene/phlogopite ratios ranging from 2/3 for Italian kamafugites to 3 for Brazilian kamafugites, both with minor amounts of apatite and Ti oxides (Innocenzi *et al.*, 2024) and varying proportions of calcic amphibole, phlogopite and clinopyroxene for Italian potassic volcanic rocks (Bracco-Gartner *et al.*, 2023). Italian orogenic lamproites are modelled as involving mixtures of melts, some of which have K richterite in the source despite being in a post-subduction environment (Bracco-Gartner *et al.*, 2025).

Many studies continue to appeal to very low degree melts of peridotites containing hydrous minerals or carbonates, but these are questionable because of the uncertainty surrounding the separability of such small amounts of melt.

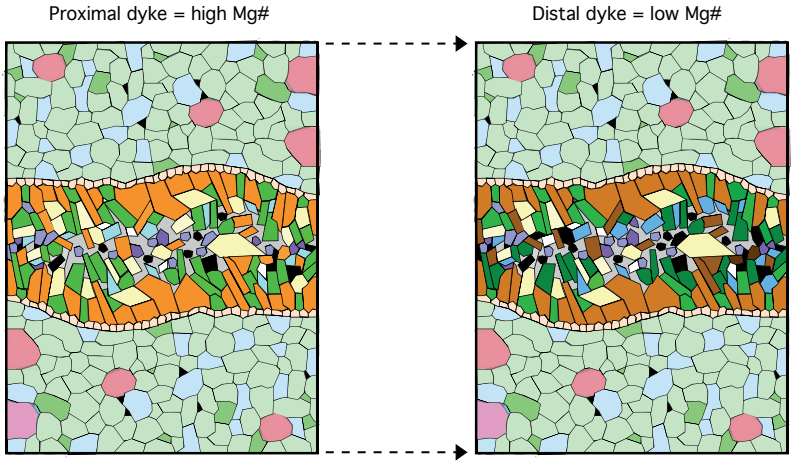


Figure 7.6 Variation of mineralogy of dyke assemblages at proximal and distal positions along a dyke due to fractionation of melt as it flows. Darker shading (right) illustrates higher Fe and lower Mg# for phlogopite (orange/brown) and clinopyroxene (green) and higher proportion of oxide minerals (black). Phlogopite rim formed by reaction with peridotite is unchanged (yellow) owing to buffering by the Mg# of peridotite minerals.

Examples are the alleged production of Brazilian nephelinites from as low as 0.1 % melting of enriched garnet lherzolite (Bongiolo *et al.*, 2015) or East African nephelinites from 0.2 % melting of carbonate-phlogopite-garnet peridotite (Baudouin and Parat, 2020). For melilitites, 1 % melting of amphibole and phlogopite-bearing mantle should suffice for Tanzanian rocks (Mattson *et al.*, 2013) and the same low volume of melting of a source with carbonate and possibly phlogopite for Madagascan examples (Mazzeo *et al.*, 2021). More realistic are vein derived melts for those ascribed the lowest melting degrees, with involvement of surrounding peridotite for the less alkaline rocks. An example is Sicilian rocks, for which the most alkaline rocks were ascribed to a phlogopite + amphibole +/- apatite +/- carbonate source, whereas less alkaline rocks were attributed to melting of variously enriched spinel lherzolites (Bianchini *et al.*, 1998).

In the past few years, it has become clear that mineral assemblages in reaction zones between the veins and peridotite wall rocks are important components of sources and increase the mineralogical scope of later source assemblages greatly. Examples of the type of vein assemblages and reaction products are shown in Figure 7.7 for a selection of probable percolating melt types in the mantle. Phlogopite and clinopyroxene are vein minerals produced by all infiltrating melts, supplemented by K richterite in the case of lamproitic melts. The reaction zone here is dominated by phlogopite as in Figure 6.3: coated

channels of phlogopite appear necessary to successfully extract lamproitic melts from the mantle, as they would lose their lamproitic composition if they react extensively with peridotite (Foley and Pertermann, 2021).

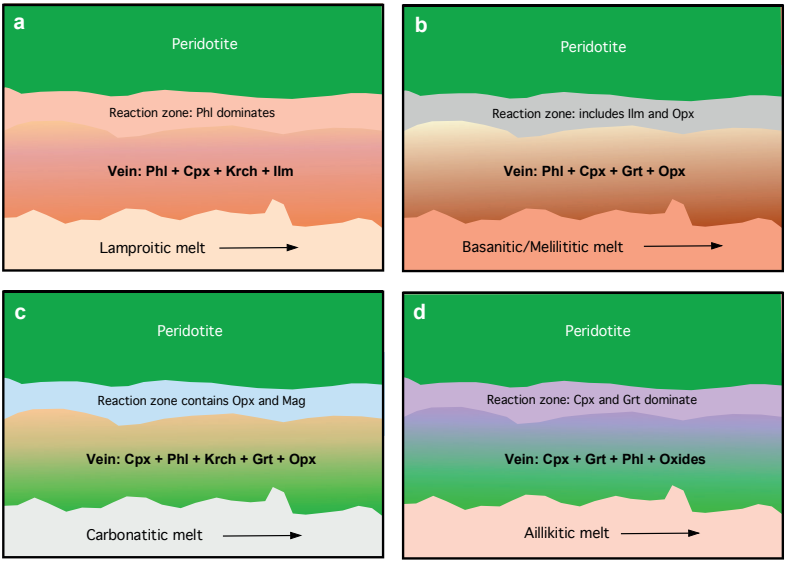


Figure 7.7 Illustration of variation of mineral assemblages in dykes and reaction zones formed from different melts in the dyke. **(a)** Lamproitic melt forms dyke with phlogopite, clinopyroxene, K richterite and minor ilmenite. Reaction zone is dominated by phlogopite (Foley *et al.*, 2025). **(b)** Sodic basanite to melilitite melts commonly have orthopyroxene and garnet, but no K richterite (Shu *et al.*, 2024). Ilmenite and orthopyroxene occur in the reaction zone. **(c)** Carbonatitic melt results in silicate phases as carbonate melt flows on. Phlogopite, clinopyroxene, K richterite, garnet and orthopyroxene have been found in reaction experiments (Pintér *et al.*, 2022). Phlogopite + carbonate results when the fractionated carbonate melt stops. Reaction zone may contain orthopyroxene and magnesite. **(d)** For aillikite melt, precipitation includes carbonates and abundant oxides. Reaction zone dominated by clinopyroxene and garnet (Pintér *et al.*, 2022).

Basanite or melilitite melts result in orthopyroxene and garnet in veins and ilmenite and orthopyroxene in reaction zones (Fig. 7.7b), explaining the occurrence of the latter as xenoliths. Ilmenite and/or rutile and apatite are essential to explain the TiO_2, P_2O_5 and trace element compositions of many alkaline igneous rocks produced by melting of veins. Although usually thought of as accessory minerals, they may be much more abundant in veins. In the case of carbonatite and aillikite melts, evidence comes mostly from experiments, in which K richterite and garnet have been produced as reaction products between carbonatite and peridotite (Pintér *et al.*, 2022). Crystallisation and reaction of

aillikite melt with wall rock peridotite produces silicate minerals (Pintér *et al.*, 2022), causing the melt to evolve towards carbonatite (Yaxley *et al.*, 1991; Chen *et al.*, 2025). Carbonate-phlogopite veins are expected to be the end product in distal vein positions and have been noted as a required source component for carbonated silicate igneous rock types (Tappe *et al.*, 2008; Rooney *et al.*, 2020). Experiments indicate that carbonate-phlogopite vein networks may result from infiltration of hydrous carbonate-rich melts (Shatskiy *et al.*, 2022). Natural examples are not common, probably because they easily re-melt, but magnesite is known from late stage veins in phlogopite-garnet peridotites in a Sulu borehole (Zhang *et al.*, 2007) and calcite and carbonate + phlogopite pockets are the latest stage in Obnazhennaya xenoliths (Ionov *et al.*, 2018). Late stage carbonates occur in reacted peridotites once they have been transformed into wehrlite (Gülmez *et al.*, 2023) and are preserved in natural wehrlites (Ramokgaba *et al.*, 2022; Kiseeva *et al.*, 2024; Fig. 4.13).

7.3 Where Are We Going? New and Missing Information

Despite the progress that has been made in recent years, there are many open questions that illuminate avenues for research. I review these first before moving on to consider new sources of information and other factors that will influence the formation conditions of neosome assemblages and the conditions of their later re-melting.

Figure 7.8 illustrates five questions; some are partly answered but only for the specific assemblages that have already been studied experimentally. Major advances have been made recently in addressing the first two questions – which assemblages are relevant for veins and at what temperature they melt? More experimental work is needed in determining the melting temperatures and melt compositions of other neosome assemblages, particularly those with orthopyroxene or carbonate. Working as a forward model from an assumed initially uniform peridotite, I view two principal types of melt as the most likely melts of peridotite at depths of 120–200 km, namely potassic silicate melts and carbonated silicate melts. The first follows logically from the similar incompatibility of K and H_2O: both of these components are expelled by peridotite minerals and enter the melt, and both are major and essential components of phlogopite, which therefore forms whenever these melts crystallise. Carbonated silicate melts form if CO_2 is also present (Section 5), which probably became more frequent as the Earth's mantle became older and more fractionated due to multiple melting and recrystallisation events. The oxidation state of the mantle at depths of 120–200 km probably also increased with time (Aulbach and Stagno, 2016; Foley, 2022) – particularly due to metasomatism, which often causes oxidative reactions – making carbonate-rich melts more likely in more recent geological times. Most experiments on neosome melting to date have

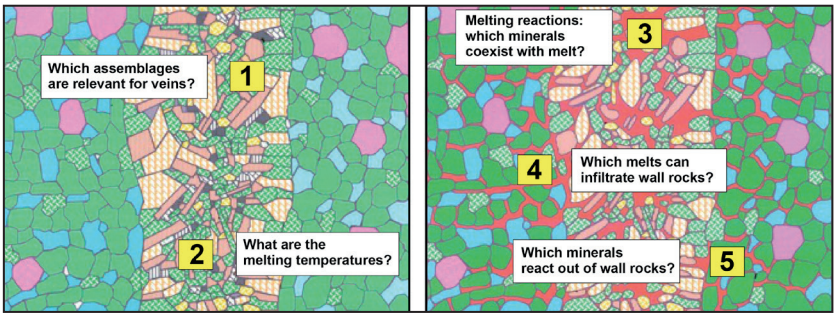


Figure 7.8 Major research questions for melting processes in hydrous veins and reaction of melts with wall rocks. [1] Vein assemblages differ according to melt type and tectonic environment, [2] Melting temperatures are needed to confirm they are below that of peridotite wall rocks, [3] Melting reactions include incongruent melting and govern which minerals can stay in the residue and coexist with melt, [4] Infiltration of melts into wall rocks may be blocked by impenetrable reaction zone or favoured by low viscosity and melt-wall rock disequilibrium, and [5] Melting reactions control which minerals will be dissolved from the wall rocks.

concentrated on potassic silicate assemblages (Section 6), whereas experiments on carbonate-rich rocks have been neglected. This needs to be rectified in the coming years.

It is now well known that the melting reactions of hydrous neosomes are governed by incongruent melting of hydrous minerals and that this affects the compositions of melts produced: if olivine is a peritectic phase (low pressures) then melts are relatively SiO_2-rich, whereas at higher pressures peritectic pyroxene formation leads to lower SiO_2 and higher MgO in melts.

The final two questions concern which melts can infiltrate wall rocks and which mineral components will be added to the melt have been transformed by the work on reaction zones: early work assumed that infiltration is governed by a chemical potential gradient but neglected the formation of reaction zones, concentrating instead on dissolution of peridotite minerals (Foley *et al.*, 1999). The advent of reaction experiments has outlined a complication relative to the simple picture in Figure 7.8. The extent of reaction zones may depend on the viscosity of the melt (Section 5.6): in the case of relatively viscous melts, the reaction zone may rapidly form an impenetrable barrier (Wang *et al.*, 2017) leading to coated channels, whereas low viscosity melts such as carbonatites may progress much further into the wall rock (Gervasoni *et al.*, 2017). Fluids may infiltrate further than melts (Förster *et al.*, 2021; Fig. 5.27) but are limited in volume due to being released from the melts.

Future work will constrain the types of neosome assemblages, the composition of melts produced from them, and decipher which magmas in the transitional (or "overlap") region of Figure 3.3 are derived from mantle migmatite sources and which from volatile-rich peridotites. We can look now at the old criteria used to propose the vein model and also at new lines of evidence that have appeared since. Near-solidus experiments could be explored further to define vein assemblages. Reaction zones in experiments and in natural rocks have opened a new range of possible source rock types. Bearing in mind this variety of likely mantle migmatite assemblages, it is important that our research programme includes experimental melting of a corresponding array of candidate ultramafic rocks. Radio-isotope systems were instrumental in recognising that many metasomatic components are old and therefore stored in the lithosphere, and these have been joined by many new systems, particularly the non-traditional stable isotopes, which have helped to track components deeper than the lithosphere and into the mantle transition zone. Melts have been preferred to fluids as metasomatic agents, but experiments on fluids are under-represented, mainly due to experimental difficulties.

A greater range of geodynamic settings should be investigated: most melting experiments to date have looked at stable continents, relevant to either cratonic or thinner lithosphere, whereas other environments may house different source rock types that need to be investigated. Of particular interest here are regions of post-collisional extension, where potassic rocks are common (Prelevic *et al.*, 2013). Trace elements are increasingly used as geochemical indicators for the role of minerals that are not stable in peridotite. The study of trace and minor elements in olivines for defining the involvement of pyroxenites at source is now relatively mature (Foley *et al.*, 2013), whereas other indicators for hydrous and accessory minerals would benefit from further development. The next few sections outline recent advances for some of these factors and to some degree present a manifesto for future work.

7.3.1 Melt/Rock Reaction Products

The importance of reactions between melts and peridotite in producing new rock assemblages in the mantle was not adequately addressed early in the development of the mantle migmatite concept. Reaction experiments are becoming widespread (Prelević *et al.*, 2024) and are helping to clarify the origin of previously described natural xenolith assemblages for which interpretations were compromised by their limited size. We now have a greatly expanded palette of reaction assemblages that need to be included in the fold of mixed rock sources.

Several suites of hybridisation experiments have emphasised the importance of the *formation* of hydrous pyroxenites, particularly during subduction (Sekine and Wyllie, 1982a; Förster *et al.*, 2019), and later from other experiments that employed a variety of melts from rhyolite to carbonatite (Mallik *et al.*, 2015; Gervasoni *et al.*, 2017). However, these experiments have not ascertained the

compositions of melts that may be derived from these rocks at a later stage. Currently, melting experiments have concentrated on potential vein assemblages but not yet on the reaction zones around them.

Reaction experiments show that orthopyroxene and garnet may be common in reaction zone assemblages. Orthopyroxene-bearing reaction products are seen in experiments with silicic melts of clastic sedimentary material and rhyolitic melts (Mallik *et al.*, 2015; Wang *et al.*, 2017) and similar features are seen in natural rocks. The formation of orthopyroxene has been attributed also to basanitic melts for natural rocks (Litasov *et al.*, 2000). Wehrlitic rocks resulting from the removal of orthopyroxene commonly occur by reaction with carbonate-rich melts (Yaxley *et al.*, 1991), which may apply to any low SiO_2 melt such as nephelinite. Other experiments with carbonated silicate melts have produced websterites (Mallik and Dasgupta, 2013) and diamonds in reaction zones (Palyanov *et al.*, 2005; Pintér *et al.*, 2022).

7.3.2 New Isotope Systems

The radio-isotope systems Rb-Sr, Sm-Nd and U-Th-Pb were instrumental in the recognition that many components of alkaline magmas must have been stored in the lithosphere for considerable time and that recycled materials including clastic sediments must be involved in their origins (Nelson *et al.*, 1986). More recently, other isotope systems, particularly the heavy stable isotopes, have introduced new evidence for interpretations alternative to traditional peridotite sources. This includes especially the involvement of subducted carbonates as limestones, marls or components in clastic sediments, and has extended the depth of mixed sources into the mantle transition zone. The references that follow are just a small selection from a booming field, but they show the direction and scope of research relevant to hydrous and carbonate minerals in mixed sources.

The involvement of recycled carbonates in the sources of carbonatites, carbonated silicate rocks and ultrapotassic rocks is indicated by light isotope signatures of both Ca (Chen *et al.*, 2024a) and Mg (Ansellem *et al.*, 2023). Using the example of Ca isotopes, partial melting of peridotite results in limited fractionation: $\delta^{44/40}$Ca of mantle peridotite is 0.94 ± 0.11 ‰ and of basalts ≈ 0.86 ‰ (Chen *et al.*, 2019) whereas much lower values are found for many carbonatites (0.26 ± 0.25 ‰; Ansellem *et al.*, 2020), ultramafic lamprophyres (0.67–0.75; Chen *et al.*, 2024a), leucitites (0.64–0.79 ‰; Ren *et al.*, 2024), and lamproites (0.58–0.66 ‰; Chen *et al.*, 2024a). The low $\delta^{44/40}$Ca of the lamproites was attributed to K richterite in the source, as Cu isotopes revealed metal-bearing conditions too reducing for carbonate stability (Chen *et al.*, 2024b). Low $\delta^{44/40}$Ca of 0.51–0.70 for basaltic rocks at Dalihu, China shows that rocks normally assigned to a peridotite source can also be from mixed source rocks (He *et al.*, 2023).

Zinc isotope ratios in recycled carbonates are characteristically high and commonly coupled with Mg or Ca isotopes, reviewed by Chen and Chen (2026). These have given evidence for recycled carbonates in the source of kamafugites and also of rocks traditionally assigned to purely peridotite sources, including basalts and basanites (Zhu *et al.*, 2021).

Hydrous minerals in sources may be tracked by K isotopes, since slab-derived fluids have heavy K isotopes. These have been combined with Fe isotopes to indicate phlogopite pyroxenite sources for Turkish and Tibetan ultrapotassic rocks (Du *et al.*, 2024). Indications from other isotope systems relevant to hydrous source minerals include B, Ba and Mo isotopes expressing sediments in the source of continental basalts (Cheng *et al.*, 2025; Chen and Chen, 2026).

7.3.3 Which Metasomatic Agents?

A brief summary of the probable identities of metasomatic agents is opportune now that we have staked out the natural rock evidence for the mineralogical scope of migmatite neosomes. As noted in Section 4.2, the first descriptions of mantle metasomatism appealed to poorly defined fluids as the responsible agents. Attention mostly switched to melts from the late 1980s, but we still see many authors sit on the fence and attribute metasomatism to "fluids/melts".

Lloyd and Bailey (1975) and Lloyd (1981) consistently attributed the Ugandan mica pyroxenite suite to pervasive alteration of peridotite by fluids, whereas I argued above that they are magmatic in origin. Other examples of early descriptions had an Fe-rich fluid for amphibole-rich veins, "kimberlitic/carbonatitic fluid" for the amphibole-apatite suite and specifically fluid and *not melt* for amphiboles and phlogopite in Eifel xenoliths (Stosch and Lugmair, 1986). In the early 1980s, rare earth element enrichment was attributed to CO_2-rich fluids, but experiments a few years later showed that CO_2 fluids do not carry much REE (Meen *et al.*, 1989).

Experiments aimed at identifying the major and trace element compositions of solutes in fluids are notoriously difficult and consequently not numerous. Analysis of experimental solutes quenched out of high pressure fluids led Schneider and Eggler (1986) to conclude that fluid solutes are silica-rich and close to K feldspar in composition, which would appear inappropriate for most types of mantle metasomatism. However, the use of pure H_2O and CO_2 in experiments may lead to misleading results for natural conditions: the addition of halogens to fluids may considerably increase trace element solubility (Keppler, 1996). Apatite has been shown to effectively scavenge Cl from fluids and appreciable contents of Cl in metasomatic apatite may favour the action of fluids (O'Reilly and Griffin, 2000). As the F/Cl ratio for most examples of mantle hydrous minerals is high (Flemetakis *et al.*, 2021), the role of fluids may be implied to be minor.

The mobility of H_2O-CO_2 fluids and their spatial reach have been debated. Kramers (1991) concluded that pervasive metasomatism by fluids was unlikely because they would move preferentially in fractures. Bonadiman *et al.* (2021)

also considered fluids to be a poor long distance transport medium because they are highly reactive. This is consistent with many models that restrict the effects of fluids to reaction zones marginal to melt infiltration pathways following fluid saturation (Fig. 4.5; Wilshire, 1987). CO_2-rich fluids are even less likely than aqueous fluids as dihedral angle measurements show they are less mobile (Watson and Brenan, 1987).

Experiments on fluids at mantle pressures suffer from problems of quenching and retention of all the material dissolved in the fluids. Experimental methods have evolved from calculating the fluid solute by mass balance, quenching solutes in the outer of two capsules (Schneider and Eggler, 1986) to separation of fluid from residues in a diamond trap, which was then analysed by Laser Ablation-ICP-MS, capturing the fluid solute trapped between the diamonds, but not the fluid itself. The best method currently available freezes the unopened capsule and analyses the fluid together with solutes directly by laser Ablation-ICP-MS in a diamond trap whilst frozen in a cryocell.

This method appears to indicate that substantially higher concentrations of trace elements are transported in melts or supercritical fluids (Kessel *et al.*, 2005) than in aqueous or CO_2 fluids. At the present time, I see the statements by Eggler (1987) that "*the presently available data place rather severe limits on the ability of fluid to accomplish all that trace-element and isotope geochemists might want them to accomplish*" and "*many more experiments are obviously needed*" as still reflecting the state of affairs nearly 40 years later.

Metasomatism by silicate melts is usually attributed to broadly basanitic melts in the continental mantle away from cratons, whereas a subdivision in cratonic mantle between PIC and MARID-type assemblages are generally ascribed to kimberlitic and lamproitic melts (Waters, 1987; Grégoire *et al.*, 2002; Fitzpayne *et al.*, 2018). A persistent irritation here is the enigmatic but frequently invoked "protokimberlite" melt, since its origin and composition are seldom defined.

Carbonatite and carbonated silicate melts intermediate between melilitite and carbonatite have a somewhat chequered history as metasomatic agents. A role for CO_2 was originally based in theoretical and experimental studies (Section 4), but evidence from natural rocks has slowly accumulated. Alkali carbonatite melts have received the most attention: they were first found in experiments and then as melt inclusions in olivine and clinopyroxene before pools were described in a wehrlite by Kiseeva *et al.* (2024). Dolomitic and magnesitic melt pools have now been described (Ramokgaba *et al.*, 2002; Section 4). My personal suspicion is that there has been an over emphasis on alkali carbonatites due to the unique composition of the Oldoinyo Lengai lavas and that more examples of less alkaline carbonatites will be found in wehrlites.

Carbonated silicate melts approximating aillikite in composition have been shown in experiments to be the products of melting of peridotite at 40–70 kbar if both CO_2 and H_2O are present (Foley *et al.*, 2009; Pintér *et al.*, 2021; Section 5). These experiments show that carbonatite melts would occur at the solidus, but

continental geotherms will lie above the solidus so that melt proportions of <1 %, which are frequently invoked, are unreasonable. The carbonatite melts may occur in subduction zones where cold material is heated. Aillikitic melts are expected to evolve towards carbonatite in composition as silicate minerals and apatite crystallise during reaction with peridotite, enriching carbonate in the melt (Yaxley *et al.*, 1991; Chen *et al.*, 2025). These carbonate melts are likely to accumulate beneath craton margins, providing an explanation for the spatial association of carbonate-rich magmatism with craton edges (Chen *et al.*, 2025). An association of sulfides with carbonates in metasomatism has long been known, and recent experiments have also shown that carbonatite melts can transport considerable amounts of S and will leach many metals of economic interest from silicates as they pass (Ezad *et al.*, 2024b). This is consistent with the mobilisation of PGE by recycled carbonate (He *et al.*, 2020).

7.3.4 Accessory Phases and Neglected Tectonic Environments

Most mantle metasomatic features have been described in continental regions, which are relevant for the sources of a variety of alkaline magmas including kimberlites, carbonatites, lamproites and ultramafic lamprophyres, as well as for members of the basanite-nephelinite-melilitite series. However, many potassic rocks are associated with subduction zones, more specifically post-collisional extension settings at convergent margins. Many of these are also the sites of economically important metal deposits. Hydrous pyroxenites have long been considered the probable reaction products between slab-derived melts and peridotite, but a widespread implicit assumption has been that of an archetypal Andean-style subduction process in which melting occurs at >100 km depth. Many of these are expected to contain orthopyroxene, which has not been the subject of many melting experiments yet. Also, few experiments have investigated garnet-rich assemblages other than dry eclogites and garnet pyroxenites.

In contrast, post-collisional potassic melts are produced at relatively low mantle pressures during lithospheric thinning, commonly in areas of complex, new lithosphere that was formed or substantially modified during the collision process under low temperature conditions much closer to the surface than in Andean-type subduction (Prelević *et al.*, 2013). These lithosphere blocks may result from low pressure imbrication of subducted crust and mantle wedge blocks in accretionary orogens that have never taken the components deeper than 70–80 km (Prelević *et al.*, 2013; Wang *et al.*, 2021). This widens the scope of potential neosome assemblages to include minerals such as lawsonite, phengite, muscovite, Na richterite and clinohumite, some of which have been investigated for the stability of single minerals but not for melting temperatures and melt compositions in neosome assemblages.

Accessory phases may often be a misnomer in the context of mantle migmatite neosomes. Apatite, Ti oxide minerals and sulfides are usually regarded as accessory phases but may be abundant in mantle neosome veins.

Apatite must be common in source regions in order to explain P_2O_5 and REE contents of melts both in experiments and natural rocks and are commonly found as megacrysts crystallised in the mantle (Long *et al.*, 1994). Apatite is rare in peridotites, crystallising only in areas that have been affected by carbonatite metasomatism (Morishita *et al.*, 2003). Peridotite minerals themselves cannot sequester high levels of P_2O_5, as the garnet/melt partition coefficient – the highest by far of any peridotite mineral – ranges from only 0.03–0.08 at relevant mantle pressures (Konzett *et al.*, 2012). The contribution of apatite to melts will vary enormously as a function of F, OH and Cl contents. Melts in experiments with 5 % F-poor apatite contained 6–8 wt. % P_2O_5 (Foley *et al.*, 2022), whereas melts in experiments with 40–50 % F-rich apatite contained only 2–4 wt. % P_2O_5 (Foley *et al.*, 2025). More systematic experiments are needed on apatite-bearing assemblages to understand these effects. The F-OH ratio will have strong effects on the contribution of phlogopite to melts (Seenstra *et al.*, 2024).

The images in Figure 4.11 show that several Ti-rich oxide and silicate phases can be abundant. The Ugandan xenoliths contain ilmenite, rutile, magnetite, titanite, perovskite and chromite. Ilmenite and rutile are widespread in other xenoliths, but assemblages containing them have not yet been subjected to many melting experiments. Feldspars occasionally occur as megacrysts (Litasov *et al.*, 2005) but feldspar-bearing neosome assemblages have not yet been experimented on at all. Rutile and zircon have been found as reaction products in experiments (Odling, 1995) as well as in natural MARID xenoliths. Further investigations into trace element partitioning between minerals and melts should help to identify which minerals are involved in migmatite neosomes, for example, to choose between the effects of various Ti oxides. This work should enlarge the scope of minerals that can be used to assess source mineral identities, modelled on the success of trace elements in olivine.

Sulfide minerals are also likely to be common in mantle veins. They are abundant enough to define part of the name of the IRPS suite (ilmenite-rutile-phlogopite-sulfide; Harte *et al.*, 1987) and are known from amphibole and phlogopite veins, in polymict breccias, amphibole megacrysts, as well as metasomatic products in peridotites (Lorand, 1989). The sulfides may be Ni pyrrhotite or mono-sulfide solid solutions with minor pentlandite and chalcopyrite (Lorand, 1989). Sulfides also occur in the apatite-rich layers in the Finero massif (Morishita *et al.*, 2003) that are interpreted to result from reaction with carbonate melts, and in wehrlite zones associated with carbonate in the Canary Islands (Kogarko *et al.*, 1995). This is in keeping with experiments that show that carbonate melts transport S (Ezad *et al.*, 2024b).

7.3.5 Redox Effects on Melting of Mantle Migmatites

Many of the minerals needed in neosomes to explain alkaline melt compositions are sensitive to redox conditions. The oxygen fugacity (fO_2) of mantle rocks decreases with increasing depth by around 0.7 log units *per* Gigapascal, which

results in reducing conditions and metal saturation at depths somewhere between 250 and 300 km (Rohrbach *et al.*, 2007, 2011). The redox state of cratonic mantle lithosphere appears to be even more reduced, with much of its lower reaches indicating fO_2 of around the iron-wüstite buffer to 2 log units higher. This lower fO_2 may be partly due to generally more reducing conditions in the early Earth (Foley, 2011; Aulbach and Stagno, 2016) and partly due to melting residues being more reduced due to redox fractionation during melting (Foley, 2022).

There is now abundant evidence from oxybarometry that reactions between peridotites and infiltrating melts produce more oxidising products in most cases. There is probably a locally complex interplay between oxidised conditions in which carbonates exist and reduced conditions in which diamonds, but no carbonates are stable (Stagno, 2019; Foley *et al.*, 2024). Redox freezing occurs where a more oxidised melt meets a reduced environment and commonly deposits diamonds (Foley, 2008; Rohrbach and Schmidt, 2011). Redox melting will occur where diamonds are remobilised as carbonates during lithosphere extension or erosion (Foley, 2008; Stagno, 2019). Diamonds may be formed by either reduction of carbonate or CO_2 or the oxidation of CH_4 (Stachel and Harris, 2008; Jacob *et al.*, 2016, 2025), indicating locally complex environments. Redox effects such as these have only rarely been investigated in experiments: more experiments juxtaposing rocks and melts with different oxidation states like those of Pintér *et al.* (2022) and Gao *et al.* (2025) are needed.

A distinct environment that may result in strong redox contrasts between melts and surrounding rocks is deep subduction of surface rocks. The foundering of sedimentary carbonate rocks in the mantle transition zone has been finding increasing favour to explain trace element and isotope compositions of a range of rocks from carbonatite to nephelinite (Zeng *et al.*, 2021). Melting here will initially produce carbonatite melts because it corresponds to a rare case of melting due to temperature increase (Fig. 5.6). The fate of these melts depends on the temperature of the rocks that the melt meets: if cold, the carbon is trapped by redox freezing, whereas in hotter environments it migrates further as a carbonate-rich melt (Gao *et al.*, 2025). These melts may produce abundant diamond when they encounter the more reduced craton base, or carbonate-rich assemblages if they crystallise without encountering a sufficiently reducing environment.

7.3.6 Kimberlite Models and the Origin of Carbon in Carbonated Silicate Melts

There are several claims that phlogopite-carbonate veins must be significant source components for some alkaline melts (Foley *et al.*, 2002b; Tappe *et al.*, 2008; Rooney *et al.*, 2020) and that these can form by fractionation of carbonated silicate melts at mantle pressures (Shatskiy *et al.*, 2022). Carbonate-rich neosomes formed close to the lithosphere-asthenosphere boundary most likely ultimately result from the solidification of partial melts of peridotite with CO_2 and H_2O. Natural rocks that have been assigned to peridotite + CO_2 + H_2O

sources include melilitites, kimberlites and ultramafic lamprophyres (specifically aillikites). Melilitites most likely form at lower pressures (100–140 km), whereas both kimberlites and aillikites carry peridotite xenoliths that attest to depths of origin greater than 195 km (Röper *et al.*, 2025).

The question then arises as to the difference between kimberlites and aillikites and the origin of the carbonate. A recent summary by Giuliani *et al.* (2025) sees the CO_2 contents of ultramafic lamprophyres and kimberlites as very similar at 5.3–6.2 wt. %, but the former includes both aillikites and alnöites, which have distinct compositions (Figs. 7.1 to 7.3). Lumping them together gives ultramafic lamprophyres as a whole a higher average SiO_2 content than aillikites alone. Also, the assimilation of lithospheric cargo by kimberlites during ascent increases SiO_2 relative to the original mantle melt, which was originally richer in carbonate. If kimberlites and aillikites are similar melts at the time they exist at the base of the lithosphere, then why do kimberlites almost universally collect considerable lithospheric cargo, whereas aillikites do not? Xenocrystic olivine cores do exist in aillikites (Veter *et al.*, 2017; Nosova *et al.*, 2018), but there is no evidence for wholesale assimilation of lithospheric peridotite material as is usual for kimberlites; peridotite xenoliths are rare in aillikites!

Two current models for kimberlite genesis both see the initial melts for kimberlites originating deeper than 250 km in regions that are not enriched in carbon and are too reducing to contain carbonate (Foley *et al.*, 2019; Giuliani *et al.*, 2023). In the first of these models, a small amount of reduced carbon is contained in iron alloy or diamond at 300 km and is oxidised to carbonate as the melt rises into a more oxidising environment, in keeping with the general oxygen fugacity profile of the mantle. Radiogenic isotopes support a deep asthenospheric origin (Woodhead *et al.*, 2019) and Ca isotopes see no evidence for an origin of the carbonate from subducted carbon (Antonelli *et al.*, 2023). However, this model requires extremely small degrees of melting (0.2–0.5 %; Giuliani *et al.*, 2023) to attain the high CO_2 contents of kimberlites after oxidation, and it is not clear that these would be extractable. Melts of reduced mantle at 250 km and deeper are MgO-rich silicate melts and silica saturated (Fig. 5.18). Extractability at very low degrees of melting is hampered by the growth of flat faces on minerals, which increases the volume of pore spaces between minerals (Faul, 1997), and by SiO_2-rich compositions (Marquardt and Faul, 2018). Melt extraction and thus mobility is much easier at shallower mantle depths where carbonate melts can exist, as shown by the rapid movement of carbonate melts in experiments (Gervasoni *et al.*, 2017). The initial melts in this model must therefore be less mobile than later on in their evolution.

In the second model the initial reduced melt has limited C-carrying capacity (Foley *et al.*, 2019). Although this reduced C will oxidise on ascent as in the Giuliani model, the amount of carbonate so formed will not approach the levels seen in kimberlites. Additional carbonate is supplied from veins and dykes (migmatite neosomes) in the transitional region between the lithosphere and asthenosphere. This region is not a sharp line but a thermal boundary layer

(TBL) in which veins may form and will increase in abundance upwards (Fig. 7.9; Foley *et al.*, 2024). Carbonates may be common here but not in the deeper, reduced asthenosphere nor in the more reduced cratonic lithosphere above. The redox conditions in this zone are sufficiently oxidising for carbonates to occur, but these will be locally restricted: melting may occur where carbonate is present from earlier melt movements and re-melting here will produce aillikitic melts. In most places, however, there will not be enough CO_2 or H_2O to supress the melting point of peridotite and large tracts of the TBL will contain no melt at all (Fig. 7.9). Kimberlites may thus be a mixture of three, not two, components: an initial reduced silicate melt, carbonate-rich components acquired in the TBL, and lithospheric peridotite components picked up at shallower levels. The second of these may be related to melts as in the Giuliani model, but which solidify and are picked up by later kimberlites.

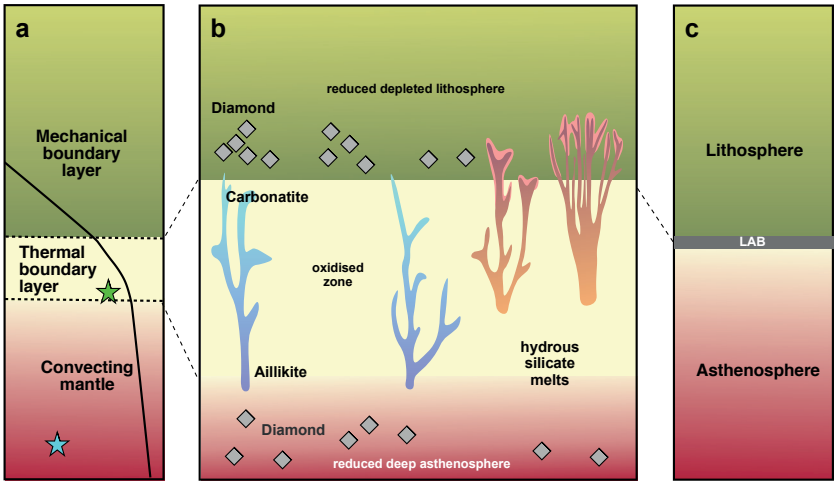


Figure 7.9 Redox conditions and melting zones in the sub-cratonic continental mantle. **(a)** The thermal boundary layer approximates a zone in which redox conditions are sufficiently oxidised for carbonate-rich melts (green star). Below this, carbon is in reduced form in diamond, metals or carbides (grey diamonds); this is the ultimate source of kimberlites (blue star). **(b)** Aillikite melts (dark blue) originate in the oxidised zone (yellow) and fractionate to carbonate melts (light blue) as they migrate (also laterally). Carbonate melts are trapped and reduced to form diamond in the lower cratonic lithosphere. Hydrous silicate melts fractionate to less magnesian compositions (light brown) and may release fluids (pink) only at a very late stage. The influx of melts from the oxidised zone reflects the common oxidising effect of mantle metasomatism. Aillikitic melt will form wherever CO_2 and H_2O is present in the oxidised zone, but large expanses of this zone will contain no melt. **(c)** The oxidised zone lies immediately below the lithosphere- asthenosphere boundary (LAB; right). Modified from Foley *et al.* (2024).

Although melts in this TBL zone are aillikitic, natural aillikites seen at the surface are more complex than simple melts of peridotite with CO_2 and H_2O. Trace element and isotope characteristics indicate that they contain additional vein components (Tappe *et al.*, 2006, 2008). Calcium isotope measurements are still sparse, but indicate possible recycled carbonate (Chen *et al.*, 2024a) in aillikites. This would align the aillikites with nephelinites, melilitites and basanites that contain evidence for recycled carbonates (Section 7.3.2). The similarity of aillikites to experimental melts of peridotite may thus be fortuitous and overstated.

8. OUTLOOK

Where nobody knows anything,
there is no point in changing your mind.

Russell (1952)

Russell's maxim may have been appropriate for keeping to the peridotite model 20–30 years ago, but there is now a large foundation of information for assessing the mantle migmatite model, including descriptions of natural mantle xenoliths, high pressure experiments on both peridotite and hydrous ultramafic mantle dyke assemblages, and characterisation of mantle derived melt compositions by petrological and geochemical means. Several lines of future investigation can focus on improving our understanding of the sources and the mechanisms of melting.

Melts of peridotite generally dominate voluminous basaltic melt types but melting most likely *usually* begins by involving the melting of other components that are usually ignored in the common cases where melting has proceeded to include peridotite. Neosome contributions to melts are logically widespread but diluted in many cases. We recognise clear cases such as alkali picrites (Ellam and Cox, 1991; Litasov and Taniguchi, 2002), but we are likely to miss many others if the neosome contribution is more subdued. It needs to be more widely recognised that if melt sources are mixtures of two or three components, as has frequently been hypothesised, then the ensuing melts will often be transitional between accepted rock definitions. Classification methods that run on the philosophy "*there can be no transitional magma types*" (Mitchell, 2020, p. 126) are not helpful. Strict adherence to rock classifications may confuse elucidation of melt origins because they may not carry over into magmas. We need to see beyond rock classifications and characterise the range of mantle derived melts and the components contained in them. This is to read the books and not just erect libraries in the analogy of Lovelock (1988; Section 3.3).

There is a range of primitive mantle derived melts including nephelinites, melilitites and basanites that have alternative and competing explanations for their origin; in some cases, both may be possible. One is melting of volatile-bearing peridotite alone and the other is melting of dykes mixed with components from neighbouring peridotite, either by further melting or dissolution (Fig. 3.3). The dichotomy lies to a large extent in the abundances of volatiles. It is well known that estimates of the original volatile contents of primitive magmas are problematic: this was noted for kimberlites by Giuliani *et al.* (2023), and fluids are often expelled by aillikites during ascent (Tappe *et al.*, 2006; Ezad *et al.*, 2025). In this near vacuum of knowledge, there is a tendency to give too much credence to experiments, the results of which are taken too seriously in the absence of real data when their applicability is often questionable. Volatile contents in

experiments are often unrealistically high as an experimental convenience. We need more analyses of abundances of individual volatile species in rocks and estimates of contents in primitive magmas, whether by direct analysis or using trace element proxies.

The scope of melting experiments on mantle dyke-forming assemblages needs to be extended, including reaction assemblages that have recently been characterised. Forward modelling from an initially uniform mantle dictates that two types of melt should be the genetic roots of most mantle migmatite neosomes, namely hydrous potassic silicate melts and ultramafic lamprophyre-like melts with compositions between carbonatite and melilitite containing around 20 wt. % SiO_2. The latter are produced in the oxidised zone of the uppermost asthenosphere (Section 7.3.6). Amongst hydrous assemblages, calcic amphibole commonly occurs together with phlogopite in xenoliths and in models for source rocks (*e.g.,* Bonadiman *et al.,* 2021), but this pair has not yet been subjected to many melting experiments.

In melting experiments to date, there has been a bias towards hydrous assemblages, with carbonate-bearing rocks neglected almost completely despite them being commonly invoked as metasomatic agents. Only simple system experiments are available at present and are restricted to magnesite (Mag), whereas other carbonates may be relevant in non-peridotitic assemblages. Phlogopite-carbonate rocks have been invoked as sources and have been shown to form by reaction (Enggist *et al.,* 2012; Shatskiy *et al.,* 2022). The few experiments available indicate a negative slope for the melting curve of Mag + Phl + Cpx (Enggist *et al.,* 2012; Enggist and Luth, 2016), unlike for hydrous ultramafic rocks, which may promote the importance of these melts below cratonic lithosphere. Carbonates may be complex at high pressures of 30–60 kbar corresponding to source regions for melts; these have not yet been considered at all in melting experiments. The relationship of oxides and sulfides to carbonate melts also needs to be researched in detail, as economically important metals have been shown to be mobilised in carbonate melts (Ezad *et al.,* 2024b).

Orogenic lamproites seem to plot close to melts of mica-rich pyroxenites according to currently available experiments (Figs. 7.1 to 7.3), but the appropriate source assemblages may not have been studied yet. Post-collisional extension may reactivate recently formed lithosphere that may contain low pressure subducted material such as blueschists (Tommasini *et al.,* 2011; Wang *et al.,* 2017) that contain several minerals that have not been investigated in melting experiments.

Reaction zones produced by reaction of infiltrating melts with peridotite have been produced in experiments, but there are no experiments yet to determine the composition of melts they later produce on re-melting. This is especially relevant to post-collisional extension, where re-melting may follow shortly after formation. Experiments so far have helped to interpret natural rocks, such as those rich in orthopyroxene and ilmenite, but there is scope for many more. Given the common redox contrast between melts and peridotites, experiments with deliberate and controlled redox contrasts should be very helpful.

The old question of whether melts or fluids are responsible for mantle metasomatism is not completely resolved, mostly because of the dearth of accurate experiments on fluids. The effects of fluids are now thought to be restricted to small areas around the extremities of dykes (Figs. 4.5 and 7.9), but they are also relevant for processes during the ascent of volatile-rich magmas. The limited number has been due to experimental difficulties, but with cryocell ablation analysis the techniques now exist to make major progress here. Flow through experiments comparing melts before and after reaction with peridotite (Foley and Pertermannn, 2021) represent another under-used experimental technique that could be more widely used. Near-solidus experiments have been very useful but seldom intended to investigate the constitution of mantle neosomes, being instead just a by product of the now largely discontinued liquidus studies.

An important quest that will be helped by the experiments outlined above in the next few years is to demystify the currently vague concepts of mantle metasomatism and mantle heterogeneity by being more specific about the petrological identity of mixed source components. Things are looking brighter with the advent of heavy stable isotope analyses, which help to identify melt and volatile origins but are also not specific about most of the minerals in the source. These have been of limited value for neosome identification as yet but in combination with further experiments may help to clear the fog and shed light on the petrological identity of source components. We need to avoid the mantle migmatite concept becoming locked into fictive and vague statistical end members as with radiogenic isotope components of chemical geodynamics which have been stuck in the mud in a similar manner to metasomatism for 40 years. These are now treated as canonical end members arising from recycling of oceanic and continental crust but still lack specific petrological identities. Mantle metasomatism now needs *disruptive science* that pushes science in new directions as opposed to *consolidating science* that merely adds to the *status quo* (Park *et al.*, 2023). Personally, I prefer to use confirmative to consolidating.

The widespread loose usage of terms is joined by *low degree melting* in the context of melting of mixed sources (Fig. 8.1). The proportions of vein and peridotite will vary widely, as will the solidus temperature difference between vein and peridotite (Fig. 7.5). For example, the meaning of 10 % melting is unambiguous for pure peridotite but loses its meaning for either of the mixtures in areas 2 and 3 of Figure 8.1.

Specification of source minerals will also benefit from the use of minor and trace element contents in potential source minerals, which has been very successful for olivine in identifying pyroxenite in mantle sources. First row transition elements are particularly promising here due to their different compatibility (Lang and Lambart, 2022): these elements are regularly at least as compatible in hydrous neosome minerals as in olivine (Cannaò *et al.*, 2022; Ezad *et al.*, 2024a, 2025) and have been used to identify the role of phlogopite

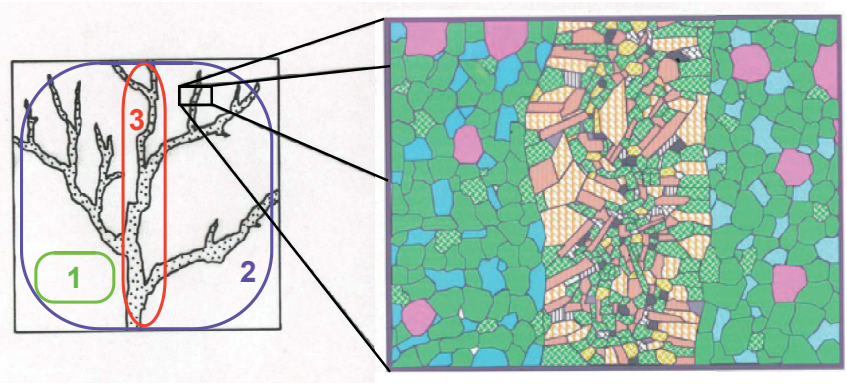


Figure 8.1 Illustration of the problematic use of the concept of low degree melting. The meaning in the peridotite model is clear (area 1), whereas mixed source rocks (areas 2 and 3) will contain widely differing proportions of vein and peridotite components. Area 3 may contain 60 % melt in the vein before melting of the peridotite begins (Fig. 7.5). Modified from Foley (1992a).

in mantle sources (Veter *et al.*, 2017). The next few years should bring many more characterisations of trace elements in other minerals relevant to mantle neosomes, such as apatites and prising apart the signatures of Ti oxides.

My final comments concern the geodynamic setting of melting of the mantle migmatites. There is often debate as to the relative contributions to melts from the lithosphere and asthenosphere, whereas many melts may come from a transition zone in which one slowly becomes the other in a thermal boundary layer (Fig. 7.9). The lithosphere-asthenosphere boundary (LAB) may be an artificial boundary that misleads interpretations in a similar way to those in rock classification (Section 3), leading people to imagine a single line that is really a transition zone. The longevity of veins and dykes is not a yes or no question: it increases with decreasing depth within the thermal boundary zone. However, small-scale lateral variations may be even more important; where volatiles are enriched, the melting point may be severely depressed, whereas where volatiles are lacking (most likely the more widespread case) no melting will occur (Fig. 7.9). These lateral variations may correspond to strong redox variations, thus solving *"the dichotomy of the mantle redox state"* (Woerman and Rosenhauer, 1985). The melt infusions that lead to the volatile-rich parts are episodic, leading to petrological complexity on a relatively small scale, well within the resolution of geophysical sensing methods. The unthinking application of Occam's razor can cause a tendency to appeal to just one or two developmental stages or mixing components when the expanse of geological time dictates that several are more likely (Section 3). These small scale variations in mineral assemblages means that specific rocks need to be subjected to future experiments. We need to move away from global averages such as GLOSS (global subducted sediment), which has been included in experiments even though it doesn't exist anywhere; the petrological behaviour of an average has no meaning.

Another interpretation with a long history that needs reassessment concerns the heat source that is often assumed necessary to cause re-melting. There have been many claims that plumes are the cause of melting to produce strongly alkaline intraplate melts. The low melting points of hydrous ultramafic dykes and carbonate-rich hydrous assemblages now confirmed by experiments (Section 6) do not necessarily require any thermal input to induce re-melting of migmatite neosomes. In my opinion, tomographic images are commonly interpreted optimistically as plumes derived from very deep in the mantle, corresponding to what I call a Campbell-Griffiths plume – a vertically continuous column with a large plume head (Campbell and Griffiths, 1990), which leads to a huge temperature anomaly when it impinges on the base of the lithosphere. Real images seldom present this picture but may be interpreted in this way despite the real image showing no continuous shaft or plume head. Past studies have provided the alternative interpretations that melting due to apparent plume activity may be related to either volatile input (wet spots, as opposed to hot spots; Green and Falloon, 1998) or to compositional differences. Coupled with extension and associated decompression melting, both would lead to re-melting of migmatite neosomes without the need for additional thermal input.

Apart from extension related to either rifting or post-collisional relaxation, melting may occur in response to reactivation of stable lithosphere through erosion at the base, which remobilises especially material in the thermal boundary layer (Gernon *et al.,* 2023). This may occur due to small scale edge-driven convection, shear driven upwelling, or lateral advection of solid material, which releases stored carbon during oxidation after dislodgement (Muirhead *et al.*, 2020). Lateral migration of melts beneath cratons is likely during which they evolve compositionally (Chen *et al.*, 2025) and may become trapped in topographic ridges where they solidify to produce mantle migmatites. As geophysical methods advance, it is becoming increasingly clear that the location of surface volcanism is frequently associated with *changes,* or steps, in the thickness of lithosphere (*e.g.,* Manassero *et al.*, 2024), as are economically important metal deposits (Hoggard *et al.*, 2020; Huston *et al.*, 2023). Substantial advances in our understanding of the relationship between alkaline and carbonate melts and the margins of cratons and other geological provinces can be expected in the next few years as the search for critical metals develops.

Finally, I leave the reader with a more philosophical note that, as both mantle migmatites and the peridotite model are replete with inductive assumptions and extrapolations, we should not be over confident with our conclusions. It is safer to follow Bertrand Russell (1931): *"When a man tells you he knows the exact truth about anything, you are safe in inferring that he is an inexact man",* than Ludwig Wittgenstein (1918): *"The truth of the thoughts presented here seems to me to be untouchable and definitive. In my opinion, I have finally solved these problems. If I am not wrong, then the value of this work is that it shows how little needed to be done to solve the problems."*

REFERENCES

ALLAN, J.F., CARMICHAEL, I.S.E. (1984) Lamprophyre lavas in the Colima graben, SW Mexico. *Contributions to Mineralogy and Petrology* 88, 203-216.

ALLEGRE, C.J., TURCOTTE, D.L. (1986) Implications of a two-component marble-cake mantle. *Nature* 323, 123-127.

AMMANNATI, E., JACOB, D.E., AVANZINELLI, R., FOLEY, S.F., CONTICELLI, S. (2016) Trace elements in olivine distinguish silicic and carbonate metasomatism in the mantle. *Earth and Planetary Science Letters* 444, 64-74.

ANDERSEN, T., O'REILLY, S.Y., GRIFFIN, W.L. (1984) The trapped fluid phase in upper mantle xenoliths from Victoria, Australia: implications for mantle metasomatism. *Contributions to Mineralogy and Petrology* 88, 72-85.

ANDERSON, D.L., BASS, J.D. (1984) Mineralogy and composition of the upper mantle. *Geophysical Research Letters* 11, 637-640.

ANDREWS, J.R. (1969) A kimberlite dyke in the Nigerdlikasik area, Frederikshåb district. *Grønlands Geologiske Undersoegelse Rapport* 19, 35-37.

ANSELLEM, E., MOYNIER, F., BERTRAND, H., BOUYON, A., MATA, J., TAPPE, S., DAY, J.M.D. (2020) Calcium isotopic evidence for the mantle sources of carbonatites. *Science Advances* 6, eaba3269.

ANSELLEM, E., SCHILLAR, M., KLAUSEN, M., BOUYON, A., ROJAS, V., BIZZARRO, M. (2023) Origin of carbonatites and associated silicate rocks revealed by Mg triple-isotope approach. *Chemical Geology* 636, 121663.

ANTONELLI, M.A., GIULIANI, A., WANG, Z.C., WANG, M.L., ZHOU, L., FENG, L.P., LI, M., ZHANG, Z.F., LIU, F., DRYSDALE, R.N. (2023) Subducted carbonates not required: deep mantle melting explains stable Ca isotopes in kimberlite magmas. *Geochimica et Cosmochimica Acta* 348, 410-427.

ARCULUS, R.J. (1975) Melting behaviour of two basanites in the range 10 to 35 kbar and the effect of TiO_2 on the olivine-diopside reactions at high pressures. *Carnegie Institution of Washington, Yearbook* 74, 512-515.

ARIMA, M., EDGAR, A.D. (1983) A high pressure experimental study on a magnesian-rich leucite lamproite from the West Kimberley area, Australia: petrogenetic implications. *Contributions to Mineralogy and Petrology* 84, 228-234.

ASHCHEPKOV, I.V., ANDRÉ, L., DOWNES, H., BELYATSKY, B.A. (2011) Pyroxenites and megacrysts from Vitim picrite-basalts (Russia): polybaric fractionation of rising melts in the mantle? *Journal of Asian Earth Sciences* 42, 14-37.

AULBACH, S., STAGNO, V. (2016) Evidence for a reducing Archean ambient mantle and its effects on the carbon cycle. *Geology* 44, 751-754.

AYER, A.J. (1973) *The central questions of philosophy*. Pelican Books, Harmondsworth, U.K., 243 pp.

BAILEY, D.K., SCHAIRER, F. (1966) The system Na_2O-Al_2O_3-Fe_2O_3-SiO_2 at 1 atmosphere, and the petrogenesis of alkaline rocks. *Journal of Petrology* 7, 114-170.

BALI, E., HIDAS, K., GUDFINSSON, G.H., KOVACS, Z., TÖRÖK, K., ROMAN-ALPISTE, M.J. (2018) Zircon and apatite-bearing pyroxene hornblendite mantle xenolith from Hungary, Carpathian-Pannonian region. *Lithos* 316, 19-32.

BARTH, T.W. (1962) *Theoretical Petrology*, John Wiley & Sons, London, 2nd edition, 416 pp.

BARTH, T.W., RAMBERG, I.B. (1966) The Fen circular complex. In: Tuttle, O.F., Gittins, J. (Eds.) *Carbonatites*. John Wiley & Sons, London, 225-257.

BARTON, M., HAMILTON, D.L. (1979) The melting relationships of a madupite from the Leucite Hills, Wyoming, to 30 kb. *Contributions to Mineralogy and Petrology* 69, 133-142.

BARTON, M., HAMILTON, D.L. (1982) Water-undersaturated melting experiments bearing upon the origin of potassium-rich magmas. *Mineralogical Magazine* 45, 267-278.

BAUDOUIN, C., PARAT, F. (2020) Phlogopite-olivine nephelinites erupted during early-stage rifting, North Tanzanian Divergence. *Frontiers in Earth Science* 8, 277.

BEARD, A.D., DOWNES, H., VETRIN, V., KEMPTON, P.D., MALUSKI, H. (1996) Petrogenesis of Devonian lamprophyre and carbonatite minor intrusions, Kandalasha Gulf (Kola Peninsula, Russia). *Lithos* 39, 93-119.

BEATTIE, P., DRAKE, M., JONES, J., LEEMAN, W., LONGHI, J., MCKAY, G., NIELSEN, R.L., PALME, H., SHAW, D., TAKAHASHI, E., WATSON, E.B. (1993) Terminology for trace element partitioning. *Geochimica et Cosmochimica Acta* 57, 1605-1606.

BECKER, H.J. (1977) Pyroxenites and hornblendites from the maar-type volcanoes of the Westeifel, Federal Republic of Germany. *Contributions to Mineralogy and Petrology* 65, 45-52.

BEDINI, R.M., BODINIER, J.-L. (1999) Distribution of incompatible trace elements between the constituents of spinel peridotite xenoliths: ICP-MS data from the East African Rift. *Geochimica et Cosmochimica Acta* 63, 3883-3900.

BERG, W. (1986) Evidence for carbonate in the mantle. *Nature* 324, 50-51.

BERKELEY, G. (1734) *The Analyst; or, a discourse addressed to an infidel mathematician*. J. Tonson, London, U.K.

BEST, M.G. (1970) Kaersutite-peridotite inclusions and kindred megacrysts in basanitic lavas, Grand Canyon, Arizona. *Contributions to Mineralogy and Petrology* 27, 25-44.

BIANCHINI, G., CLOCCHIATTI, R., COLTORTI, M., JORON, J.L., VACCARO, C. (1998) Petrogenesis of mafic lavas from the northernmost sector of the Iblean district (Sicily). *European Journal of Mineralogy* 10, 301-315.

BODINIER, J.-L., VASSEUR, G., VERNIERES, J., DUPUY, C., FABRIÉS, J. (1990) Mechanisms of mantle metasomatism: geochemical evidence from the Lherz orogenic peridotite. *Journal of Petrology* 31, 597-628.

Bodinier, J.-L., Menzies, M.A., Shimizu, N., Frey, F.A., McPherson, E. (2004) Silicate, hydrous and carbonate metasomatism at Lherz, France: contemporaneous derivatives of silicate melt-harzburgite reaction. *Journal of Petrology* 45, 299-330.

Bonadiman, C., Brombin, V., Andreozzi, G.B., Benna, P., Coltorti, M., Curetti, N., Faccini, B., Merli, M., Pelorosso, B., Stagno, V., Tesauro, M., Pavese, A. (2021) Phlogopite-pargasite coexistence in an oxygen reduced spinel-peridotite ambient. *Scientific Reports* 11, 11829.

Bonatti, E., Ottonello, G., Hamlyn, P.R. (1986) Peridotites from the island of Zabargad (St. John), Red Sea: petrology and geochemistry. *Journal of Geophysical Research* 91, 599-631.

Bongiolo, E.M., Pires, G.L.C., Geraldes, M.C., Santos, A.C., Neumann, R. (2015) Geochemical modeling and Nd-Sr data links nephelinite-phonolite successions and xenoliths of Trinidade Island (South Atlantic Ocean, Brazil). *Journal of Volcanology and Geothermal Research* 306, 58-73.

Borghini, G., Fumagalli, P. (2018) Subsolidus phase relations in a mantle pyroxenite: an experimental study from 0.7 to 1.5 GPa. *European Journal of Mineralogy* 30, 333-348.

Borghini, G., Fumagalli, P., Rampone, E. (2017) Partial melting of secondary pyroxenite at 1 and 1.5 GPa, and its role in upwelling heterogeneous mantle. *Contributions to Mineralogy and Petrology* 172, 70.

Bossière, G., Megartsi, M. (1982) Pétrologie des nodules pyroxénolites associés à la rushayite d`In Teria (N.E. d`Illisi, ex Fort Polignac), Algérie. *Bulletin de Minéralogie* 105, 89-98.

Bowen, N.L. (1913) The melting phenomena of the plagioclase feldspars. *American Journal of Science* 35, 577-599.

Bowen, N.L. (1914) The ternary system diopside-forsterite-silica. *American Journal of Science* 38, 207-264.

Bowen, N.L. (1928) *The evolution of the igneous rocks*. Dover Publications, New York, 332 pp.

Bowen, N.L., Anderson, O. (1914) The binary system MgO-SiO_2. *American Journal of Science* 37, 487-500.

Bowman, E.E., Mallik, A., Ducea, M.N. (2024) Partial melting of arclogite and petrogenesis of alkaline-silicate complexes. *Earth and Planetary Science Letters* 645, 118952.

Boyd, F.R. (1959) Hydrothermal investigations of the amphiboles. In: Abelson, P.H. (Ed.) Researches in Geochemistry, John Wiley and Sons, New York, 377-396.

Boyd, F.R. (1991) Mantle metasomatism: evidence from a MARID-harzburgite compound xenolith. *Carnegie Institution of Washington Yearbook* 1990-1991, 18-23.

Boyd, F.R., England, J.L. (1958) Development of high-pressure apparatus. *Carnegie Institution of Washington Yearbook* 57, 170-173.

Bracco-Gartner, A.J.J., Nikogosian, I.K., Davies, G.R., Koornneef, J.M. (2023) The multi-component mantle source of Roman province ultrapotassic magmas revealed by melt inclusions. *Geochimica et Cosmochimica Acta* 355, 266-281.

Bracco-Gartner, A.J.J., Nikogosian, I.K., Aartsen, J.M., Karlas, E.L., Davies, G.R., Koornneef, J.M. (2025) Orogenic lamproites from Italy formed by mixing of highly potassic and shoshonitic melts. *Earth and Planetary Science Letters* 654, 119228.

Brey, G. (1978) Origin of olivine melilitites – chemical and experimental constraints. *Journal of Volcanology and Geothermal Research* 3, 61-88.

Brey, G.P., Weber, R., Nickel, K.G. (1990) Calibration of a belt apparatus to 1800˚C and 6 GPa. *Journal of Geophysical Research* 95, 15603-15610.

Brey, G.P., Bulatov, V.K., Girnis, A.V., Layahe, Y. (2008) Experimental melting of carbonated peridotite at 6-10 GPa. *Journal of Petrology* 49, 797-821.

Brey, G.P., Bulatov, V.K., Girnis, A.V. (2009) Influence of water and fluorine on melting of carbonated peridotite at 6 and 10 GPa. *Lithos* 112S, 249-259.

Brey, G.P., Bulatov, V.K., Girnis, A.V. (2011) Melting of K-rich carbonated peridotite at 6-10 GPa and the stability of K-phases in the upper mantle. *Chemical Geology* 281, 333-342.

Brøgger, W.C. (1921) *Die Eruptivgesteine des Kristianiagebietes IV. Das Fengebiet in Telemark, Norwegen*. Videnskapsselskapets Skrifter 1. Mat.+ Naturv. Klasse 1920, 9, 408 pp.

Bulanova, G. (1986) Compositional evolution of syngenetic inclusions of ultrabasic association in Yakutian diamond. *Extended Abstracts of the 4th International Kimberlite Conference (Geological Society of Australia Abstract Series* 16), 371-373.

Bulatov, V.K., Brey, G.P., Girnis, A.V., Gerdes, A., Höfer, H. (2014) Carbonated sediment-peridotite interaction and melting at 7.5-12 GPa. *Lithos* 200-201, 368-385.

Burnham, C.W. (1979) The importance of volatile constituents. In: Yoder, H.S. (Ed.) The evolution of the igneous rocks: fiftieth anniversary perspectives. Princeton University Press, Princeton, New Jersey, 439-482.

Campbell, I.H., Griffiths, R.W. (1990) Implications of mantle plume structure for the evolution of flood basalts. *Earth and Planetary Science Letters* 99, 79-93.

Canil, D. (1990) Experimental study bearing on the absence of carbonate in mantle-derived xenoliths. *Geology* 18, 1011-1013.

Cannao, E., Tiepolo, M., Borghini, G., Langone, A., Fumagalli, P. (2022) The influence of oxygen fugacity and chlorine on amphibole-liquid trace element partitioning at upper-mantle conditions. *European Journal of Mineralogy* 34, 35-57.

Carlson, R.J., Irving, A.J. (1994) Depletion and enrichment history of subcontinental lithospheric mantle: an Os, Sr, Nd and Pb isotopic study of ultramafic xenoliths from the northwestern Wyoming Craton. *Earth and Planetary Science Letters* 126, 457-472.

Carlson, R.J., Araujo, A.L.N., Junqueira-Brod, T.C., Gaspar, J.C., Brod, J.A., Petrinovic, I.A., Hollanda, M.H.B.M., Pimentel, M.M., Sichel, S. (2007) Chemical and isotopic relationships between peridotite xenoliths and mafic-ultrapotassic rocks from southern Brazil. *Chemical Geology* 242, 415-434.

Carmichael, I.S.E., Turner, F.J., Verhoogen, J. (1974) *Igneous Petrology*. McGraw-Hill, New York, 739 pp.

Chen, C.F., Dai, W., Wang, Z.C., Liu, Y.S., Li, M., Becker, H., Foley, S.F. (2019) Calcium isotope fractionation during magmatic processes in the upper mantle. *Geochimica et Cosmochimica Acta* 249, 121-137.

Chen, C.F., Förster, M.W., Foley, S.F., Shcheka, S.S. (2023) Subduction of carbonate-rich crustal rocks drives the deep carbon and chlorine cycles. *Nature* 620, 576-581.

Chen, C.F., Foley, S.F., Tappe, S., Ren, H.E., Feng, L.P., Liu, Y.S. (2024a) Calcium isotopes track volatile components in the mantle source of alkaline rocks and associated carbonatites. *Earth and Planetary Science Letters* 625, 118489.

Chen, C.F., Foley, S.F., Shcheka, S.S., Liu, Y.S. (2024b) Copper isotopes track the Neoproterozoic oxidation of cratonic mantle roots. *Nature Communications* 15, 4311.

Chen, C.F., Förster, M.W., Shcheka, S.S., Ezad, I.S., Shea, J.J., Liu, Y.S., Jacob, D.E., Foley, S.F. (2025) Sulfide-rich continental roots at cratonic margins form by migration of carbonated melts. *Nature* 637, 615-621.

Chen, Y., Chen, Y.X. (2026) Origin of mantle-derived ultrapotassic rocks constrained by non-traditional stable isotopes. *Earth-Science Reviews* 279, 105544.

Cheng, Y.S., Liu, J.Q., Chen, L.H., Chen, K.Y., Zhao, J., Wang, X.J., Zeng, G., Zhang, H.F. (2025) Barium isotope fingerprint for recycled ancient sediment in the source of EM1-type continental basalts. *Geophysical Research Letters* 52, e2024GL111960.

Choi, E., Fiorentini, M.L., Giuliani, A., Foley, S.F., Maas, R., Graham, S. (2021) Petrogenesis of Proterozoic alkaline ultramafic rocks in the Yilgarn Craton, Western Australia. *Gondwana Research* 93, 197-217.

COHEN, L.H., ITO, K., KENNEDY, G.C. (1967) Melting and phase relations in an anhydrous basalt to 40 kbar. *American Journal of Science* 265, 475-518.

CONCEIÇAO, R.V., GREEN, D.H. (2004) Derivation of potassic (shoshonitic) magmas by decompression melting of phlogopite + pargasite lherzolite. *Lithos* 72, 209-229.

CONDAMINE, P., MÉDARD, E. (2014) Experimental melting of phlogopite-bearing mantle at 1 GPa: implications for potassic magmatism. *Earth and Planetary Science Letters* 397, 80-92.

CONDAMINE, P., MÉDARD, E., DEVIDAL, J.L. (2016) Experimental melting of phlogopite-peridotite in the garnet stability field. *Contributions to Mineralogy and Petrology* 171, 95.

COOPER, N.P., SCOTT, J.M., BRENNA, M., PALMER, M.C., ROUX, P.J.L., COOPER, A.F., REID, M.R., STIRLING, C.H. (2024) Hydrous veined mantle lithosphere and implications for the source of Zealandia intraplate magmas. *Lithos* 478, 107608.

COPI, I.M. (1978) *Introduction to Logic*. Collier-MacMillan, London, U.K., Fifth edition, 590 pp.

COSGROVE, M.E. (1972) The geochemistry of the potassium-rich Permian volcanic rocks of Devonshire, England. *Contributions to Mineralogy and Petrology* 36, 155-170.

CROSS, W. (1902) The development of systematic petrography in the nineteenth century. Part II. *Journal of Geology* 10, 451-499.

CUCINELLO, C., GRIFA, C., DE'GENNARO, R., FRANCIOSI, L., ROCCO, I., MORRA, V., MELLUSO, L. (2023) Alkaline rocks of the Bobaomby volcanic field point to a petrogenetic link between Comoros and northern Madagascar lithosphere. *International Journal of Earth Sciences* 112, 707-723.

CURRIE, K.L. (1976) The alkaline rocks of Canada. *Geological Survey of Canada Bulletin* 239, 228 pp.

DAI, H.K., WANG, Z.C., LIU, Y.S., CHEN, C.F., ZONG, K.Q., ZHOU, L., ZHANG, G.L., LI, M., MOYNIER, F., HU, Z.C. (2020) Calcium isotope compositions of mantle pyroxenites. *Geochimica et Cosmochimica Acta* 270, 144-159.

DALOU, C., LE LOSQ, C., MYSEN, B.O., CODY, G.D. (2015) Solubility and solution mechanisms of chlorine and fluorine in aluminosilicate melts at high pressure and high temperature. *American Mineralogist* 100, 2272-2283.

DALTON, J.A., PRESNALL, D.C. (1998) The continuum of primary carbonatitic–kimberlitic melt compositions in equilibrium with lherzolite: data from the system $CaO–MgO–Al_2O_3–SiO_2–CO_2$ at 6 GPa. *Journal of Petrology* 39, 1953-1964.

DALY, R.A. (1910) Origin of alkaline rocks. *Geological Society of America Bulletin* 21, 87-118.

DALY, R.A. (1926) *Our mobile Earth*. Charles Scribner's Sons, New York, 342 pp.

DASGUPTA, R. (2018) Volatile-bearing partial melts beneath oceans and continents – where, how much, and of what compositions? *American Journal of Science* 318, 141-165.

DASGUPTA, R., HIRSCHMANN, M.M. (2006) Melting in the Earth's deep upper mantle caused by carbon dioxide. *Nature* 440, 659-662.

DAUBRÉE, A. (1879) *Études synthétiques de Géologie expérimentale*. Libraire des Corps des Ponts et Chaussées, des Mines et des Télégraphes, Paris, 828 pp.

DAVIS, F.A., HIRSCHMANN, M.M. (2013) The effects of K_2O on the compositions of near-solidus melts of garnet peridotite at 3 GPa and the origin of basalts from enriched mantle. *Contributions to Mineralogy and Petrology* 166, 1029-1046.

DAWSON, J.B. (1962) Sodium carbonate lavas from Oldoinyo Lengai, Tanganyika. *Nature* 195, 1075-1076.

DAWSON, J.B. (1987) The MARID suite of xenoliths in kimberlite: relationship to veined and metasomatised peridotite xenoliths. In: Nixon, P.H. (Ed.) *Mantle Xenoliths*. John Wiley & Sons, London, U.K., 465-473.

DAWSON, J.B., POWELL, D.G. (1969) Mica in the upper mantle. *Contributions to Mineralogy and Petrology* 22, 233-237.

Dawson, J.B., Smith, J.V. (1977) The MARID (mica-amphibole-rutile-ilmenite-diopside) suite of xenoliths in kimberlite. *Geochimica et Cosmochimica Acta* 41, 309-323.

de la Roche, H. (1986) Classification et nomenclature des roches ignées; un essai de restauration de la convergence entre systématique quantitative, typologie d'usage et modélisation génétique. *Bulletin de la Société Géologique de France* 2, 337-353.

Denaeyer, M.E. (1966) La rushayite, homologue sodique de la katungite, lave nouvelle du volcan Nyiragongo (Kivu, Afrique centrale). *Bulletin de la Société française de Minéralogie et de Cristallographie* 89, 107-119.

Dilek, Y., Furnes, H. (2014) Ophiolites and their origins. *Elements* 10, 93-100.

Dimroth, E. (1970) Meimechites and carbonatites of the Castignon Lake Complex, New Quebec. *Neues Jahrbuch für Mineralogie, Abhandlungen* 112, 239-278.

Du, D.H., Luo, X.L., Wang, X.L., Palmer, M.R., Ersoy, E.Y., Li, W.Q. (2024) A recipe for making potassium-rich magmas in collisional orogens: new insights from K and Fe isotopes. *Earth and Planetary Science Letters* 632, 118642.

Dürr, H.-P. (1988) *Das Netz des Physikers. Naturwissenschaftliche Erkenntnis in der Verantwortung.* Carl Hanser Verlag, München, 490 pp.

Dvir, O., Kessel, R. (2017) The effect of CO_2 on the water-saturated solidus of K-poor peridotite between 4 and 6 GPa. *Geochimica et Cosmochimica Acta* 206, 184-200.

Edgar, A.D. (1974) Experimental studies. In: Sørensen, H. (Ed.) *The alkaline rocks*. John Wiley and Sons, Chichester, U.K., 355-389.

Eggler, D.H. (1974) Effect of CO_2 on the melting of peridotite. *Carnegie Institution of Washington Yearbook* 73, 215-224.

Eggler, D.H. (1987) Solubility of major and trace elements in mantle metasomatic fluids: experimental constraints. In: Menzies, M.A., Hawkesworth, C.J. (Eds.) *Mantle Metasomatism*. Academic Press, London, U.K., 21-41.

Eggler, D.H., Holloway, J.R. (1977) Partial melting of peridotite in the presence of H_2O and CO_2: principles and review. Magma Genesis. *Bulletin, State of Oregon, Department of Geology and Mineral Resources* 96, 15-36.

Eggler, D.H., Wendlandt, R.F. (1979) Experimental studies on the relationship between kimberlite magmas and partial melting of peridotite. In: Boyd, F.R., Meyer, H.O.A. (Eds.) *Kimberlites, diatremes and diamonds: their geology, petrology, and geochemistry*. American Geophysical Union, Washington, USA, 330-338.

Eggler, D.H., Mysen, B.O., Seitz, M.G. (1974) The solubility of CO_2 in silicate liquids and crystals. *Carnegie Institution of Washington Yearbook* 73, 226-228.

Ehrenburg, S.N. (1979) Garnetiferous ultramafic inclusions in minettes from the Navajo volcanic field. In: Boyd, F.R., Meyer, H.O.A. (Eds.) *The Mantle Sample: Inclusions in Kimberlites and Other Volcanics*. Proceedings of the Second International Kimberlite Conference (American Geophysical Union), 330-344.

Elkins-Tanton, L.T, Grove, T.L. (2003) Evidence for deep melting of hydrous metasomatized mantle: Pliocene high-potassium magmas from the Sierra Nevadas, *Journal of Geophysical Research* 108, 2350.

Ellam, R.M., Cox, K.G. (1991) An interpretation of Karoo picrite basalts in terms of interaction between asthenospheric magmas and the mantle lithosphere. *Earth and Planetary Science Letters* 105, 330-342.

Enggist, A., Luth, R.W. (2016) Phase relations of phlogopite and pyroxene with magnesite from 4 to 8 GPa: KCMAS-H_2O and KCMAS-H_2O-CO_2. *Contributions to Mineralogy and Petrology* 171, 88.

Enggist, A., Chu, L., Luth, R.W. (2012) Phase relations of phlogopite with magnesite from 4 to 8 GPa. *Contributions to Mineralogy and Petrology* 163, 467-481.

ERLANK, A.J., FINGER, L.W. (1970) The occurrence of potassic richterite in a mica nodule from the Wesselton kimberlite, South Africa. *Carnegie Institution of Washington Yearbook* 68, 442-443.

ERLANK, A.J., RICKARD, R.S. (1977) Potassic richterite bearing peridotites from kimberlite and the evidence they provide for upper mantle metasomatism. *Extended Abstracts of the Second International Kimberlite Conference*, 93-95.

ERLANK, A.J., WATERS, F.G., HAWKESWORTH, C.J., HAGGERTY, S.E., ALLSOPP, H.L., RICKARD, R.S., MENZIES, M. (1987) Evidence for mantle metasomatism in peridotite nodules from the Kimberley pipes, South Africa. In: Menzies, M.A., Hawkesworth, C.J. (Eds.) *Mantle Metasomatism*. Academic Press, London, U.K., 221-311.

ERNST, T. (1965) Do peridotitic inclusions in basalts represent mantle material? In: Smith, C.H., Sorgenfrei, T. (Eds) *The Upper Mantle Symposium, New Delhi (1964)*. IUGS, Copenhagen, 180-185.

ERTAN, I.E., LEEMAN, W.P. (1996) Metasomatism of Cascades subarc mantle: evidence from a rare phlogopite orthopyroxenite xenolith. *Geology* 24, 451-454.

ESPERANÇA, S., HOLLOWAY, J.R. (1987) On the origin of some mica lamprophyres: experimental evidence from a mafic minette. *Contributions to Mineralogy and Petrology* 95, 207-216.

EZAD, I.S., FOLEY, S.F. (2022) Experimental partitioning of fluorine and barium in lamproites. *American Mineralogist* 107, 2008-2019.

EZAD, I.S., SHCHEKA, S.S., BUHRE, S., BUHRE, A., GOROJOVSKY, L.R., SHEA, J.J., FÖRSTER, M.W., FOLEY, S.F. (2023) Rapid quench piston-cylinder apparatus: an improved design for the recovery of volatile-rich geological glasses from experiments at 0.5-2.5 GPa. *Reviews of Scientific Instruments* 94, 055107.

EZAD, I.S., BLANKS, D.E., FOLEY, S.F., HOLWELL, D.A., BENNETT, J., FIORENTINI, M.L. (2024a) Hydrous minerals in metasomatized mantle lithologies control Ni metal availability in magmatic sulfide deposits. *Mineralium Deposita* 59, 227-236.

EZAD, I.S., SAUNDERS, M., SHCHEKA, S.S., FIORENTINI, M.L., GOROJOVSKY, L.R., FÖRSTER, M.W., FOLEY, S.F. (2024b) Incipient carbonate melting drives metal and sulfur mobilization in the mantle. *Science Advances* 10, eadk5979.

EZAD, I.S., SHEA, J.J., FOLEY, S.F. (2025) Hydrous minerals are sinks for first row transition elements in the mantle: an experimental partitioning study. *Chemical Geology* 690, 122883.

FALLOON, T.J., GREEN, D.H. (1987) Anhydrous partial melting of MORB pyrolite and other peridotite compositions at 10 kbar and implications for the origin of primitive MORB glasses. *Mineralogy and Petrology* 37, 181-219.

FALLOON, T.J., GREEN, D.H. (1988) Anhydrous partial melting of peridotite from 8 to 35 kb and the petrogenesis of MORB. *Journal of Petrology Special Lithosphere Issue*, 379-414.

FALLOON, T.J., GREEN, D.H., HATTON, C.J., HARRIS, K.L. (1988) Anhydrous partial melting of a fertile and depleted peridotite from 2 to 30 kb and application to basalt petrogenesis. *Journal of Petrology* 29, 1257-1282.

FALLOON, T.J., GREEN, D.H., O'NEILL, H. ST.C., HIBBERSON, W.O. (1997) Experimental tests of low degree peridotite partial melt compositions: implications for the nature of anhydrous near-solidus peridotite melts at 1 GPa. *Earth and Planetary Science Letters* 152, 149-162.

FAUL, U.H. (1997) Permeability of partially molten upper mantle rocks from experiments and percolation theory. *Journal of Geophysical Research* 102, 10299-10311.

FENNER, C.N. (1948) Immiscibility of igneous magmas. *American Journal of Science* 246, 465-502.

FIFIELD, R. (Ed.) (1985) *The making of the Earth*. Basil Blackwell Ltd., Oxford, U.K., 336 pp.

FITZPAYNE, A., GIULIANI, A., PHILLIPS, D., HERGT, J., WOODHEAD, J.D., FARQUHAR, J., FIORENTINI, M.L., DRYSDALE, R.N., WU, N. (2018) Kimberlite-related metasomatism recorded in MARID and PIC mantle xenoliths. *Mineralogy and Petrology* 112, 571-584.

Fitzpayne, A., Giuliani, A., Hergt, J., Woodhead, J.D., Maas, R. (2020) Isotopic analyses of clinopyroxenes demonstrate the effects of kimberlite melt metasomatism upon the lithospheric mantle. *Lithos* 370-371, 105595.

Flemetakis, S., Klemme, S., Stracke, A., Genske, F., Berndt, J., Rohrbach, A. (2021) Constraining the presence of amphibole and mica in metasomatized mantle sources through halogen partitioning experiments. *Lithos* 180-181, 105859.

Foley, S.F. (1984) Liquid immiscibility and melt segregation in alkaline lamprophyres from Labrador. *Lithos* 17, 127-137.

Foley, S.F. (1985) The oxidation state of lamproitic magmas. *Tschermaks Mineralogische und Petrographische Mitteilungen* 34, 217-238.

Foley, S.F. (1989) The genesis of lamproitic magmas in a reduced, fluorine-rich mantle. In: Ross, J. *et al*. (Eds.) *Kimberlites and related rocks, volume 1: their composition, occurrence, origin and emplacement*. Geology Society of Australia Special Publication 14, Blackwell Scientific Publications, Melbourne, 616-632.

Foley, S.F. (1991) High-pressure stability of the fluor- and hydroxy end-members of pargasite and K-richterite. *Geochimica et Cosmochimica Acta* 55, 2689-2694.

Foley, S.F. (1992a) Vein-plus-wall-rock melting mechanisms in the lithosphere and the origin of potassic alkaline magmas. *Lithos* 28, 435-453.

Foley, S.F. (1992b) Petrological characterization of the source components of potassic magmas: geochemical and experimental constraints. *Lithos* 28, 187-204.

Foley, S.F. (1993) An experimental study of olivine lamproite: first results from the diamond stability field. *Geochimica et Cosmochimica Acta* 57, 483-489.

Foley, S.F. (2008) Rejuvenation and erosion of the cratonic lithosphere. *Nature Geoscience* 1, 503-510.

Foley, S.F. (2011) A reappraisal of redox melting in the Earth's mantle as a function of tectonic setting and time. *Journal of Petrology* 52, 1363-1391.

Foley, S.F. (2022) Mantle redox melting. In: Moretti, R., Vielzeuf, D. (Eds.) Mantle redox geochemistry. *Geophysical Monograph* 266, 93-113.

Foley, S.F., Wheller, G.E. (1990) Parallels in the origin of the geochemical signatures of island arc volcanics and continental potassic igneous rocks: the role of residual titanates. *Chemical Geology* 85, 1-18.

Foley, S.F., Pintér, Z. (2018) Primary melt compositions in the Earth's mantle. In: Kono, Y., Sanloup, C. (Eds.) *Magmas under pressure: advances in high-pressure experiments on structure and properties of melts*. Elsevier, Amsterdam, 3-42.

Foley, S.F., Pertermann, M. (2021) Dynamic metasomatism experiments investigating the interaction between migrating potassic melt and garnet peridotite. *Geosciences* 11, 432.

Foley, S.F., Ezad, I.S. (2024) Melting of hydrous pyroxenites with alkali amphiboles in the continental mantle: 2. Trace element compositions of melts and minerals. *Geoscience Frontiers* 15, 101692.

Foley, S.F., Taylor, W.R., Green, D.H. (1986) The effect of fluorine on phase relationships in the system $KAlSiO_4$-Mg_2SiO_4-SiO_2 at 28 kbar and the solution mechanism of fluorine in silicate melts. *Contributions to Mineralogy and Petrology* 93, 46-55.

Foley, S.F., Venturelli, G., Green, D.H., Toscani, L. (1987) The ultrapotassic rocks: characteristics, classification and constraints for petrogenetic models. *Earth Science Reviews* 24, 81-134.

Foley, S.F., Musselwhite, D.S., van der Laan, S.R. (1999) Melt compositions from ultramafic vein assemblages in the lithospheric mantle: a comparison of cratonic and non-cratonic settings. In: Gurney, J.J., Gurney, J.L., Pascoe, M.D., Richardson, S.H. (Eds.) *Proceedings of the 7th International Kimberlite Conference*. Red Roof Publishers, Cape Town, 238-246.

Foley, S., Tiepolo, M., Vannucci, R. (2002a) Growth of early continental crust by melting of amphibolite in subduction zones. *Nature* 417, 837-840.

FOLEY, S., ANDRONIKOV, A.V., MELZER, S. (2002b) Petrology of ultramafic lamprophyres from the Beaver Lake area of eastern Antarctica and their relation to the breakup of Gondwanaland. *Mineralogy and Petrology* 74, 361-384.

FOLEY, S., ANDRONIKOV, A.V., JACOB, D.E., MELZER, S. (2006) Spinel + garnet lherzolites from the Jetty Peninsula, East Antarctica: mineralogical features and geotherms in the mantle beneath a developing rift. *Geochimica et Cosmochimica Acta* 70, 3096-3120.

FOLEY, S.F., YAXLEY, G.M., ROSENTHAL, A., BUHRE, S., KISEEVA, E.S., RAPP, R.P., JACOB, D.E. (2009) The composition of near-solidus melts of peridotite in the presence of CO_2 and H_2O at 40 – 60 kbar. *Lithos* 112S, 274-283.

FOLEY, S.F., LINK, K., TIBERINDWA, J.V., BARIFAIJO, E. (2012) Patterns and origin of igneous activity around the Tanzanian craton. *Journal of African Earth Sciences* 62, 1-18.

FOLEY, S.F., PRELEVIC, D., REHFELDT, T., JACOB, D.E. (2013) Minor and trace elements in olivines as probes into early igneous and mantle melting processes. *Earth and Planetary Science Letters* 363, 181-191.

FOLEY, S.F., YAXLEY, G.M., KJARSGAARD, B.A. (2019) Kimberlites from source to surface: experimental petrology perspectives. *Elements* 15, 393-398.

FOLEY, S.F., EZAD, I.S., VAN DER LAAN, S.R., PERTERMANN, M. (2022) Melting of hydrous pyroxenites with alkali amphiboles in the continental mantle: melting relations and major element composition of melts. *Geoscience Frontiers* 13, 101380.

FOLEY, S.F., CHEN, C.F., JACOB, D.E. (2024) The effects of local variations in conditions on carbon storage and release in the continental mantle. *National Science Review* 11, nwae098.

FOLEY, S.F., EZAD, I.S., SHU, C.T., FÖRSTER, M.W. (2025) Melting of amphibole-apatite-rich metasomes in the continental mantle and comparison of melt compositions with natural igneous rocks. *Lithos* 500-501, 107976.

FORBES, W.C., FLOWER, M.F.J. (1974) Phase relations of titan-phlogopite, $K_2Mg_4TiAl_2Si_6O_{22}(OH)_4$: a refractory phase in the upper mantle. *Earth and Planetary Science Letters* 22, 60-66.

FORD, C.E., RUSSELL, D.G., CRAVEN, J.A., FISK, M.R. (1983) Olivine-liquid equilibria: temperature, pressure and composition dependence of the crystal/liquid cation partition coefficients for Mg, Fe2+, Ca and Mn. *Journal of Petrology* 24, 256-265.

FÖRSTER, M.W., PRELEVIC, D., SCHMÜCK, H.R., BUHRE, S., VETER, M., MERTZ-KRAUS, R., FOLEY, S.F., JACOB, D.E. (2017) Melting and dynamic metasomatism of mixed harzburgite + glimmerite mantle source: implications for the origin of orogenic potassic magmas. *Chemical Geology* 455, 182-191.

FÖRSTER, M.W., PRELEVIC, D., SCHMÜCK, H.R., BUHRE, S., MARSCHALL, H.R., MERTZ-KRAUS, R., JACOB, D.E. (2018) Melting phlogopite-rich MARID: lamproites and the role of alkalis in olivine-liquid partitioning. *Chemical Geology* 476, 429-440.

FÖRSTER, M.W., FOLEY, S.F., MARSCHALL, H.R., ALARD, O., BUHRE, S. (2019) Melting of sediments in the deep mantle produces saline fluid inclusions in diamonds. *Science Advances* 5, eaau2620.

FÖRSTER, M.W., BUHRE, S., XU, B., PRELEVIC, D., MERTZ-KRAUS, R., FOLEY, S.F. (2020) Two-stage formation of K-enrichment in ultrapotassic magmatism confirmed by a novel experimental approach. *Minerals* 10, 41.

FÖRSTER, M.W., BUSSWEILER, Y., PRELEVIC, D., DACZKO, N.R., BUHRE, S., MERTZ-KRAUS, R., FOLEY, S.F. (2021) Sediment-peridotite reaction controls fore-arc metasomatism and arc-magma geochemical signatures. *Geosciences* 11, 372.

FREY, F.A., GREEN, D.H. (1974) The mineralogy, geochemistry and origin of lherzolite inclusions in Victorian basanites. *Geochimica et Cosmochimica Acta* 38, 1023-1069.

FREY, F.A., GREEN, D.H., ROY, S.D. (1978) Integrated models of basalt petrogenesis: a study of quartz tholeiites to olivine melilitites from South Eastern Australia utilizing geochemical and experimental petrological data. *Journal of Petrology* 19, 463-513.

Fujii, T., Kushiro, I. (1977) Melting relations and viscosity of an abyssal tholeiite. *Carnegie Institution of Washington Yearbook* 76, 461-467.

Funk, S.P., Luth, R.W. (2013) Melting phase relations of a mica-clinopyroxenite from the Milk River area, southern Alberta, Canada. *Contributions to Mineralogy and Petrology* 166, 393-409.

Gao, M.D., Foley, S.F., Xu, H.J., Wang, Y. (2023) Mantle metasomatism induced by water-fluxed melting of the continental crust at ultra-high pressures. *Geology* 51, 1127-1131.

Gao, M.D., Wang, Y., Foley, S.F., Xu, Y.G. (2025) Variable mantle redox states driven by deeply subducted carbon. *Science Advances* 11, eadu4985.

Geikie, A. (1905) The Founders of Geology. Macmillan and Co., London & New York, 486 pp.

Gernon, T.M., Jones, S.M., Brune, S., Hincks, T.K., Palmer, M.R., Schumacher, J.C., Primiceri, R.M., Field, M., Griffin, W.L., O'Reilly, S.Y., Keir, D., Spencer, C.J., Meredith, A.S., Glerum, A. (2023) Rift-induced disruption of cratonic keels drives kimberlite magmatism. *Nature* 620, 344-350.

Gervasoni, F., Klemme, S., Rohrbach, A., Grützner, T., Berndt, J. (2017) Experimental constraints on mantle metasomatism caused by silicate and carbonate melts. *Lithos* 282-283, 173-186.

Gervasoni, F., Jalowitsji, T., Rocha, M.P., Weska, R.K., Novais-Rodrigues, E., Rodrigues, R.A.D., Busweiler, Y., Borbosa, E.S.R., Berndt, J., Dantas, E.L., Souza, V.D., Klemme, S. (2022) Recycling process and proto-kimberlite melt metasomatism in the lithosphere-asthenophere boudnary beneath the Amazonian Craton recorded by garnet xenocrysts and mantle xenoliths from the Carolina kimberlite. *Geoscience Frontiers* 13, 101429.

Ghosh, S., Litasov, K.D., Ohtani, E. (2014) Phase relations and melting of carbonated peridotite between 10 and 20 GPa: a proxy for alkali- and CO_2-rich silicate melts in the deep mantle. *Contributions to Mineralogy and Petrology* 167, 964.

Gianetti, B., Luhr, J.F. (1990) Phlogopite-clinopyroxenite modules from high-K magmas, Roccamonfina volcano, Italy: evidence for a low-pressure metasomatic origin. *Earth and Planetary Science Letters* 101,404-424.

Giuliani, A., Phillips, D., Woodhead, J.D., Kamenetsky, V.S., Fiorentini, M.L., Maas, R., Soltys, A., Armstrong, R.A. (2015) Did diamond-bearing orangeites originate from MARID-veined peridotites in the lithospheric mantle? *Nature Communications* 6, 6837.

Giuliani, A., Schmidt, M.W., Torsvik, T.H., Fedortchouk, Y. (2023) Genesis and evolution of kimberlites. *Nature Reviews Earth and Environment* 4, 738-753.

Giuliani, A., Dalton, H., Pearson, D.G. (2025) Kimberlites: the deepest chemical probes of Earth. *Treatise on Geochemistry* 3e, 159-230.

Goldschmidt, V.M. (1922) Der Stoffwechsel der Erde. *Zeitschrift für Elektrochemie und angewandte physikalische Chemie* 28, 411-421.

Golovin, A.V., Kamenetsky, V.S. (2023) Compositions of kimberlite melts: a review of melt inclusions in kimberlite minerals. *Petrology* 31, 143-178.

Goodrich, C.A. (1984) Phosphoran pyroxene and olivine in silicate inclusions in natural iron-carbon alloy, Disko Island, Greenland. *Geochimica et Cosmochimica Acta* 48, 1115-1126.

Gorring, M.L., Kay, S.M. (2000) Carbonatite metasomatized peridotite xenoliths from southern Patagonia: implications for lithospheric processes and Neogene plateau magmatism. *Contributions to Mineralogy and Petrology* 140, 55-72.

Green, D.H. (1964) The petrogenesis of the high-temperature peridotite intrusion in the Lizard area, Cornwall. *Journal of Petrology* 5, 134-188.

Green, D.H. (1970) The origin of basaltic and nephelinitic magmas. *Transactions of the Leicester Philosophical and Literary Society* 64, 28-54.

GREEN, D.H. (1973) Experimental melting studies on a model upper mantle composition at high pressure under water-saturated and water-undersaturated conditions. *Earth and Planetary Science Letters* 19, 37-53.

GREEN, D.H., RINGWOOD, A.E. (1963) Mineral assemblages in a model mantle composition. *Journal of Geophysical Research* 68, 937-945.

GREEN, D.H., RINGWOOD, A.E. (1967) The genesis of basaltic magmas. *Contributions to Mineralogy and Petrology* 15, 103-190.

GREEN, D.H., RINGWOOD, A.E. (1969) Mineralogy of peridotitic compositions under upper mantle conditions. *Physics of the Earth and Planetary Interiors* 3, 359-371.

GREEN, D.H., WALLACE, M.E. (1988) Mantle metasomatism by ephemeral carbonatite melts. *Nature* 336, 459-462.

GREEN, D.H., FALLOON, T.J. (1998) Pyrolite: a Ringwood concept and its current expression. In: Jackson, I.N.S. (Ed.) *The Earth's Mantle: Composition, Structure and Evolution*, Cambridge University Press, Cambridge, U.K. 311-380.

GREEN, D.H., HIBBERSON, W.O., KOVÁCS, I., ROSENTHAL, A. (2010) Water and its influence on the lithosphere-asthenosphere boundary. *Nature* 467, 448-451.

GREEN, D.H., HIBBERSON, W.O., ROSENTHAL, A., KOVÁCS, I., YAXLEY, G.M., FALLOON, T.J., BRINK, F. (2014) Experimental study of the influence of water on melting and phase assemblages in the upper mantle. *Journal of Petrology* 55, 2067-2096.

GRÉGOIRE, M., BELL, D.R., LE ROUX, A.P. (2002) Trace element geochemistry of phlogopite-rich mafic mantle xenoliths: their classification and their relationship to phlogopite-bearing peridotites and kimberlites revisited. *Contributions to Mineralogy and Petrology* 142, 603-625.

GREIG, J.W. (1927) Immiscibility in silicate melts, Part 1. *American Journal of Science Fifth Series* 13, 1-44.

GRÜTZNER, T., PRELEVIC, D., BERNDT, J., KLEMME, S. (2023) The origin of Na-alkaline lavas revisited: new constraints from experimental melting of amphibole-rich metasomes + lherzolite at uppermost mantle pressure. *Contributions to Mineralogy and Petrology* 178, 73.

GÜLMEZ, F. PRELEVIC, D., FÖRSTER, M.W., BUHRE, S., GÜNTHER, J. (2023) Experimental production of K-rich metasomes through sediment recycling at the slab-mantle interface in the fore-arc. *Scientific Reports* 13, 19608.

GÜMBEL, C.W. (1874) *Die Paläolithischen Eruptivgesteine des Fichtelgebirges*. Weis, Munich, Germany, 50 pp.

GÜMBEL, C.W. (1879) *Der Palaeolith. Eruptivgesteine des Fichtelgebirges*. Munich, Germany.

HALL, J. (1798) Experiments on whinstone and lava. *Transactions of the Royal Society of Edinburgh* 5, 43-75.

HAMILTON, D.L., FREESTONE, I.C., DAWSON, J.B., DONALDSON, C.H. (1979) Origin of carbonatites by liquid immiscibility. *Nature* 279, 52-54.

HARLOW, G.E. (2002) Diopside + F-rich phlogopite at high P and T: systematics, crystal chemistry and the stability of $KMgF_3$, clinohumite and chondrodite. *Geological Materials Research* 4, 28 pp.

HARTE, B. (1983) Mantle peridotite and processes – the kimberlite sample. In: Hawkesworth, C.J., Norry, M.J. (Eds.) *Continental basalts and mantle xenoliths*. Shiva, Nantwich, U.K., 46-91.

HARTE, B., GURNEY, J.J. (1975) Ore mineral and phlogopite mineralisation within ultramafic nodules from the Matsoku kimberlite pipe, Lesotho. *Carnegie Institution of Washington Yearbook* 75, 528-536.

HARTE, B., COX, K.G., GURNEY, J.J. (1975) Petrography and geological history of upper mantle xenoliths from the Matsoku kimberlite pipe. *Physics and Chemistry of the Earth* 9, 477-506.

HARTE, B., WINTERBURN, P.A., GURNEY, J.J. (1987) Metasomatic and enrichment phenomena in garnet peridotite facies mantle xenoliths from the Matsoku kimberlite pipe, Lesotho. In: Menzies, M.A., Hawkesworth, C.J. (Eds.) *Mantle Metasomatism*. Academic Press, London, U.K., 145-220.

HARTE, B., HUNTER, R.H., KINNY, P.D. (1993) Melt geometry, movement and crystallization, in relation to mantle dykes, veins and metasomatism. *Philosophical Transactions of the Royal Society of London, Series A*, 342, 1-21.

HAURI, E., SHIMIZU, N., DIEU, J.L., HART, S.R. (1993) Evidence for hotspot-related carbonatite metasomatism in the oceanic upper mantle. *Nature* 365, 221-227.

HE, D.T., LIU, Y.S., MOYNIER, F., FOLEY, S.F., CHEN, C.F. (2020) Platinum group element mobilization in the mantle enhanced by recycled sedimentary carbonate. *Earth and Planetary Science Letters* 541, 116262.

HE, D.T., LIU, Y.S., MOYNIER, F., FOLEY, S.F., CHEN, C.F., ZHU, Y.T., LU, X.T., ZHANG, G.L., ZONG, K.Q. (2023) Tightly coupled Ca-Zn-Sr isotope co-variations in basalts caused by recycled calcium carbonate in the mantle source. *Chemical Geology* 637, 121678.

HEARN, B.C. (1994) Composite megacrysts and megacryst aggregates from the Williams kimberlite, Montana, USA: multiple products of mantle melts. In: Meyer, H.O.A., Leonardos, O.H. (Eds.) *Kimberlites, Related Rocks and Mantle Xenoliths*. Proceedings of the Fifth Kimberlite Conference, Araxa, Brazil 1, 388-404.

HECKER, J.G., MARKS, M.A.W., WENZEL, T., MARKL, G. (2020) Halogens in amphibole and mica from mantle xenoliths: implications for the halogen distribution and halogen budget of the metasomatized continental lithosphere. *American Mineralogist* 105, 781-794.

HEISENBERG, W. (1973) *Der Teil und das Ganze: Gespräche im Umkreis der Atomphysik*. Deutsche Taschenbuch Verlag, München, Germany, 288 pp.

HINDE, J.C., JANNEY, P.E., LEGROS, H., WOODLAND, S., PEARSON, D.G. (2025) A subduction-modified source for the parental melts of MARID assemblages from the Kaapvaal craton lithosphere and their formation via hybridisation with wall-rock peridotite. *Mineralogy and Petrology* 119, 863-882.

HIROSE, K., KUSHIRO, I. (1993) Partial melting of dry peridotites at high pressures: determination of compositions of melts segregated from peridotite using aggregates of diamond. *Earth and Planetary Science Letters* 114, 477-489.

HIRSCHMANN, M.M., STOLPER, E.M. (1996) A possible role for garnet pyroxenite in the origin of the "garnet signature" in MORB. *Contributions to Mineralogy and Petrology* 124, 185-208.

HÖGBOM, A.G. (1895) Über das Nephelinsyenitegebiet auf der Insel Älno, Parts 1 and 2. *Geologiska Föreningen i Stockholm Förhandlingar* 17, 100-158, 214-254.

HOGGARD, M.J., CZARNOTA, K., RICHARDS, F.D., HUSTON, D.L., JAQUES, A.L., GHELICHKHAN, S. (2020) Global distribution of sediment-hosted metals controlled by craton edge stability. *Nature Geoscience* 13, 504-510.

HOLMES, A. (1926) The structure of continents. *Nature* 118, 586.

HOLMES, A., HARWOOD, H.F. (1932) Petrology of the volcanic fields east and south-east of Ruwenzori, Uganda. *Quarterly Journal of the Geological Society* 88, 370-439.

HOLMES, A., HARWOOD, H.F. (1937) The petrology of the volcanic area of Bufumbira. *Geological Survey of Uganda Memoir* 3, 299 pp.

HUCKENHOLZ, H.G., GILBERT, M.C., KUNZMANN, T. (1992) Stability and phase relations of calcic amphiboles crystallized from magnesio-hastingsite compositions in the 1 to 45 kbar pressure rage. *Neues Jahrbuch für Mineralogie, Abhandlungen* 164, 229-268.

HUME, D. (1739) *A treatise of human nature*. 1969 edition. Penguin Books, Harmondsworth, UK, 678 pp.

Huston, D.L., Champion, D.C., Czarnota, K., Duan, J.M., Hutchens, M., Paradis, S., Hoggard, M., Ware, B., Gibson, G.M., Doublier, M.P., Kelley, K., McCafferty, A., Hayward, N., Richards, F., Tessalina, S., Carr, G. (2023) Zinc on the edge – isotopic and geophysical evidence that cratonic edges control world-class shale-hosted zinc-lead deposits. *Mineralium Deposita* 58, 707-729.

Hutchinson, M.T., Frei, D. (2009) Kimberlite and related rocks from Garnet Lake, West Greenland, including their mantle constituents, diamond occurrence, age and provenance. *Lithos* 112S, 318-333.

Hutton, J. (1795) *Theory of the Earth, with proofs and illustrations*. William Creech, Edinburgh, 620 pp.

Innocenzi, F., Ezad, I.S., Ronca, S., Agostini, S., Lustrino, M., Foley, S. (2024) Experimental petrology constraints on kamafugitic magmas. *European Journal of Mineralogy* 36, 899-916.

Innocenzi, F., Ronca, S., Agostini, S., Gaurino, V., Foley, S., Lustrino, M. (2025) Kamafugites. *Earth-Science Reviews* 270, 105238.

Ionov, D.A., Doucet, L.S., Xu, Y.G., Golovin, A.V., Oleinikov, O.B. (2018) Reworking of Archean mantle in the NE Siberian craton by carbonatite and silicate melt metasomatism: evidence from a carbonate-bearing, dunite to websterite xenolith suite from the Obnazhennaya kimberlite. *Geochimica et Cosmochimica Acta* 224, 132-153.

Irving, A.J. (1980) Petrology and geochemistry of composite ultramafic xenoliths in alkalic basalts and implications for magmatic processes within the mantle. *American Journal of Science* 280A, 389-426.

Irving, A.J., Wyllie, P.J. (1973) Melting relationships in CaO-CO_2 and MgO-CO_2 to 36 kilobars with comments on CO_2 in the mantle. *Earth and Planetary Science Letters* 20, 220-225.

Irving, A.J., Price, R.C. (1981) Geochemistry and evolution of lherzolite-bearing phonolitic lavas from Nigeria, Australia, East Germany and New Zealand. *Geochimica et Cosmochimica Acta* 45, 1309-1320.

Ito, K., Kennedy, G.C. (1967) Melting and phase relations in a natural peridotite to 40 kilobars. *American Journal of Science* 265, 519-538.

Jacob, D.E. (2006) High sensitivity analysis of trace element-poor geological reference glasses by Laser-Ablation-Inductively-Coupled-Mass-Spectrometry (LA-ICP-MS). *Geostandards and Geoanalytical Research* 30, 221-235.

Jacob, D.E., Viljoen, K.S., Grassineau, N., Jagoutz, E. (2000) Remobilization in the cratonic lithosphere recorded in polycrystalline diamond. *Science* 289, 1182-1185.

Jacob, D.E., Piazolo, S., Schreiber, A., Trimby, P. (2016) Redox-freezing and nucleation of diamond via magnetite formation in the Earth's mantle. *Nature Communications* 7, 11891.

Jacob, D.E., Stern, R.A., Czas, J., Reutter, M., Piazolo, S., Stachel, T. (2025) Polycrystalline diamond aggregates and their role in Earth's deep carbon cycle. *Geochmica et Cosmochimica Acta* 389, 136-156.

Jakobsson, S., Holloway, J.R. (2008) Mantle melting in equilibrium with an iron-wüstite-graphite buffered COH-fluid. *Contributions to Mineralogy and Petrology* 155, 247-256.

Jaques, A.L., Green, D.H. (1980) Anhydrous melting of peridotite at 0–15 kb pressure and the genesis of tholeiitic basalts. *Contributions to Mineralogy and Petrology* 73, 287-310.

Jaques, A.L., Lewis, J.D., Smith, C.B., Gregory, G.P., Ferguson, J., Chappell, B.W., McCulloch, M.T. (1984) The diamond-bearing ultrapotassic (lamproitic) rocks of the West Kimberley region, Western Australia. In: Kornprobst, J. (Ed.) *Kimberlites 1. Kimberlites and related rocks*. Elsevier, Amsterdam, 225-254.

Jaques, A.L., Yaxley, G.M., Harley, S.L. (2025) David Headley Green 1936-2024. *Historical Records of Australian Science* 36, HR250006. doi:10.1071/HR25006.

Jenner, G.A., Foley, S.F., Jackson, S.E., Green, T.H., Fryer, B.J., Longerich, H.P. (1993) Determination of partition coefficients for trace elements in high-pressure-temperature experimental run products by Laser Ablation Microprobe - inductively coupled mass spectrometry (LAM-ICP-MS). *Geochimica et Cosmochimica Acta* 57, 5099-5103.

Johannsen, A. (1938) *A descriptive petrography of the igneous rocks. Volume IV.* University of Chicago Press, Chicago, USA.

Johnson, B., Goldblatt, C. (2015) The nitrogen budget of the Earth. *Earth-Science Reviews* 148, 150-173.

Johnson, L.H., Jones, A.P., Church, A.A., Taylor, W.R. (1997) Ultramafic xenoliths and megacrysts from a melilitite tuff cone, Deeti, northern Tanzania. *Journal of African Earth Sciences* 25, 29-42.

Kaeser, B., Kalt, A., Pettke, T. (2006) Evolution of the lithospheric mantle beneath the Marsabit volcanic field (northern Kenya): constraints from textural, P-T and geochemical studies on xenoliths. *Journal of Petrology* 47, 2149-2184.

Kamenetsky, V.S., Maas, R., Kamenetsky, M.B., Paton, C., Phillips, D., Golovin, A.V., Gornova, M.A. (2009) Chlorine from the mantle: magmatic halides in the Udachnaya-East kimberlite, Siberia. *Earth and Planetary Science Letters* 285, 96-104.

Keppler, H. (1996) Constraints from partitioning experiments on the composition of subduction zone fluids. *Nature* 380, 237-240.

Kessel, R., Schmidt, M.W., Ulmer, P., Pettke, T. (2005) Trace element signature of subduction zone fluids, melts and supercritical fluids at 120-180km depth. *Nature* 437, 724-727.

Kiseeva, E.S., Kamenetsky, V.S., Chayka, I.F., Maas, R., Nielsen, T.F.D. (2024) Carbonate melts in the cratonic mantle evidenced by a xenolith from the Majuagaa kimberlite, West Greenland. *Geology* 53, 89-93.

Kline, M. (1980) *Mathematics: the loss of certainty.* Oxford University Press, Oxford, UK, 366 pp.

Kogarko, L., Henderson, C.M.B., Pacheco, H. (1995) Primary Ca-rich carbonatite magma and carbonate-silicate-sulfide liquid immiscibility in the upper mantle. *Contributions to Mineralogy and Petrology* 121, 267-274.

Kogiso, T., Hirschmann, M.M., Pertermann, M. (2004) High-pressure partial melting of mafic lithologies in the mantle. *Journal of Petrology* 45, 2407-2422.

Konzett, J. (1997) Phase relations and chemistry of Ti-rich K-richterite-bearing mantle assemblages: an experimental study to 8.0 GPa in a Ti-KNCMASH system. *Contributions to Mineralogy and Petrology* 128, 385-404.

Konzett, J., Ulmer, P. (1999) The stability of hydrous potassic phases in lherzolitic mantle – an experimental study to 9.5 GPa in simplified and natural bulk compositions. *Journal of Petrology* 40, 629-652.

Konzett, J., Sweeney, R.J., Thompson, A.B., Ulmer, P. (1997) Potassium amphibole stability in the upper mantle: an experimental study in a peralkaline KNCMASH system to 8.5 GPa. *Journal of Petrology* 38, 537-568.

Konzett, J., Armstrong, R.A., Günther, D. (2000) Modal metasomatism in the Kaapvaal craton lithosphere: constraints on timing and genesis from U-Pb zircon dating of metasomatized peridotites and MARID-type xenoliths. *Contributions to Mineralogy and Petrology* 139, 704-719.

Konzett, J., Rhede, D., Frost, D.J. (2012) The high PT stability of apatite and Cl partitioning between apatite and hydrous potassic phases in peridotite: an experimental study to 19 GPa with implications for the transport of P, Cl and K in the upper mantle. *Contributions to Mineralogy and Petrology* 163, 277-296.

KONZETT, J., HAUZENBERGER, C., KRENN, K., JOACHIM-MROSKO, B., STALDER R., GRÖBNER, K., SIEBERER, A.K., HOANG, N., KHOI, N.N. (2019) Neogene metasomatism in the subcontinental lithosphere beneath SE Asia – evidence from modal and cryptic phosphorus enrichment in peridotites and pyroxenites from southern Laos. *Journal of Petrology* 60, 2413-2447.

KORNPROBST, J. (1970) Les péridotites et les pyroxénolites du massif ultrabasique des Beni Bouchera: une étude expérimentale entre 1100 et 1550°C, sous 15 à 30 kilobars de pression sèche. *Contributions to Mineralogy and Petrology* 29, 290-309.

KORNPROBST, J. (Ed.) (1984) *Kimberlites I: Kimberlites and related rocks*. Elsevier, Amsterdam, 466 pp.

KOSTER VAN GROOS, A.F., WYLLIE, P.J. (1963) Experimental data bearing on the role of liquid immiscibility in the genesis of carbonatites. *Nature* 199, 801-802.

KRAMERS, J.D. (1991) Paradoxes of the mantle lithosphere underneath Archaean continents and models for its origin. *Schweizerische Mineralogische und Petrographische Mitteilungen* 71, 175-186.

KRANCK, E.H. (1953) Bedrock geology of the seaboard of Labrador between Domino Run and Hopedale. *Geological Survey of Canada Bulletin* 26, 16-32.

KRESTEN, P., ÅHMAN, E., BRUNFELT, A.O. (1981) Alkaline ultramafic lamprophyres and associated carbonatite dykes from the Kalix area, northern Sweden. *Geologische Rundschau* 70, 1215-1231.

KUHN, T.S. (1962) *The structure of scientific revolutions*. 2nd edition. University of Chicago Press, Chicago, USA, 210 pp.

KUSHIRO, I. (1970) Stability of amphibole and phlogopite in the upper mantle. *Carnegie Institution of Washington Yearbook* 68, 245-247.

KUSHIRO, I. (1975) On the nature of silicate melt and its significance in magma genesis: regularities in the shift of phase boundaries involving olivine, pyroxene, and silica minerals. *American Journal of Science* 275, 411-431.

KUSHIRO, I., ERLANK, A.J. (1970) Stability of potassic richterite. *Carnegie Institution of Washington Yearbook* 68, 231-233.

KUSHIRO, I., SYONO, Y., AKIMOTO, S. (1967) Stability of phlogopite at high pressures and possible presence of phlogopite in the Earth's upper mantle. *Earth and Planetary Science Letters* 3, 197-203.

LAMBART, S., LAPORTE, D., SCHIANO, P. (2009) An experimental study of pyroxenite partial melts at 1 and 1.5 GPa: implications for the major-element composition of mid-ocean ridge basalts. *Earth and Planetary Science Letters* 288, 335-347.

LAMBART, S., BAKER, M.B., STOLPER, E.M. (2016) The role of pyroxenite in basalt genesis: Melt-PX, a melting parameterization for mantle pyroxenites between 0.9 and 5 GPa. *Journal of Geophysical Research* 121, 5708-5735.

LAMBERT, I.B., WYLLIE, P.J. (1968) Stability of hornblende and a model for the low velocity zone. *Nature* 219, 1240-1241.

LANG, O.I., LAMBERT, S. (2022) First-row transition elements in pyroxenites and peridotites: a promising tool for constraining source mineralogy. *Chemical Geology* 612, 1211378.

LARSEN, L.M., PEDERSEN, A.K., SUNDVOLL, B., FREI, R. (2023) Alkali picrites formed by melting of old metasomatized lithospheric mantle: Manîtdlat Member, Vaigat Formation, Palaeocene of West Greenland. *Journal of Petrology* 44, 3-38.

LE BAS, M.J. (1978) Nephelinite volcanism at plate interiors. *Bulletin of Volcanology* 41, 459-462.

LE MAITRE, R.W. (2002) *Igneous rocks: a classification and glossary of terms: recommendations of the International Union of Geological Sciences, Subcommission on the systematics of igneous rocks*. Cambridge University Press, Cambridge, U.K., 236 pp.

LEWENS, T. (2016) *The meaning of science*. Basic Books, Perseus, New York, 254 pp.

LEWIS, H.C. (1888) The matrix of the diamond. *Geological Magazine* 5, 129-131.

Lin, A.B., Zheng, J.P., Aulbach, S., Xiong, Q., Pan, S.K., Gerdes, A. (2020) Causes and consequences of wehrlitization beneath a trans-lithospheric fault: evidence from Mesozoic basalt-borne wehrlite xenoliths from the Tan-Lu Fault Belt, North China Craton. *Journal of Geophysical Research* 125, e2019JB019084.

Litasov, K.D., Taniguchi, H. (2002) Mantle evolution beneath the Baikal Rift. *Centre for Northeast Asian Studies, Tohoku University Monograph* 5, 221 pp.

Litasov, K.D., Foley, S.F., Litasov, Y.D. (2000) Magmatic modification and metasomatism of the subcontinental mantle beneath the Vitim volcanic field (East Siberia): evidence from trace element data on pyroxenite and peridotite xenoliths from Miocene picrobasalt. *Lithos* 54, 83-113.

Litasov, K.D., Litasov, Y.D., Malkovets, V.G. (2005) Metasomatism and transformations of the upper mantle beneath the southern Baikal territory: evidence from xenoliths of the Bartoy volcanic area. *Geochemistry International* 43, 242-267.

Litasov, K.D., Shatskiy, A., Ohtani, E. (2014) Melting and subsolidus phase relations in peridotite and eclogite systems with reduced C O H fluid at 3–16 GPa. *Earth and Planetary Science Letters* 391, 87-99.

Liu, L.H., Zhang, Z.C., Chang, Z.G., Santosh, M., Liu, B.X., Li, H.X. (2021) Ultramafic xenoliths from aillikites in the Tarim large igneous province: implications for Alaskan-type affinity and role of subduction. *Lithos* 380, 105902.

Liu, Y.S., He, D.T., Gao, C.G., Foley, S., Gao, S., Hu, Z.C., Zong, K., Chen, H.H. (2015) First direct evidence of sedimentary carbonate recycling in subduction-related xenoliths. *Scientific Reports* 5, 11547.

Liu, Z., Rohrbach, A., Tiraboschi, C., Foley, S.F., Berndt, J., Klemme, S. (2023) The effect of COH fluids on partial melting of eclogite and lherzolite under moderately oxidizing and reducing conditions. *Chemical Geology* 616, 121219.

Lloyd, F.E. (1981) Upper-mantle metasomatism beneath a continental rift: clinopyroxenes in alkali mafic lavas and nodules from South West Uganda. *Mineralogical Magazine* 44, 315-323.

Lloyd, F.E., Bailey, D.K. (1975) Light element metasomatism of the continental mantle: the evidence and the consequences. *Physics and Chemistry of the Earth* 9, 389-416.

Lloyd, F.E., Arima, M., Edgar, A.D. (1985) Partial melting of a phlogopite-clinopyroxenite nodule from south-west Uganda: an experimental study bearing on the origin of highly potassic continental rift volcanics. *Contributions to Mineralogy and Petrology* 91, 321-329.

Loewinson-Lessing, F.Y. (1954) *A historical survey of petrology*. Oliver & Boyd, Edinburgh, 105 pp.

Long, A.M., Menzies, M.A., Thirlwall, M.F., Upton, B.G.J., Aspen, P. (1994) Carbonate-mantle interaction: a possible origin for megacryst/xenolith suites in Scotland. In: Meyer, H.O.A., Leonardos, O.H. (Eds.) *Kimberlites, related rocks and mantle xenoliths*. Proceedings of the Fifth Kimberlite Conference, Araxa, Brazil, 467-477.

Lorand, J.L. (1989) The Cu-Fe-Ni sulfide component of the amphibole-rich veins from the Lherz and Freychinède spinel peridotite massifs (northeastern Pyrenees, France): a comparison with mantle-derived megacrysts from alkali basalts. *Lithos* 23, 281-298.

Lovelock, J.E. (1988) *The ages of Gaia. A biography of our living Earth*. Oxford University Press, Oxford, U.K., 252 pp.

Luth, R.W. (1997) Experimental study of the system phlogopite-diopside from 3.5 to 17 GPa. *American Mineralogist* 82, 1198-1209.

Lutkov, V.S., Shakhalyeva, Z.A., Lutkova, V.Ya., Sharapov, N.V., Vladykin, N.V., Kholdoyanidi, V.K. (1989) Glimmerite xenolith in alkali basaltoids of southern Tien Shan. New evidence of mantle metasomatism. *Doklady Akademii Nauk Earth Science Section* 305, 1219-1224.

MALLIK, A., DASGUPTA, R. (2013) Reactive infiltration of MORB-eclogite-derived carbonated silicate melt into fertile peridotite at 3 GPa and genesis of alkalic magmas. *Journal of Petrology* 54, 2267-2300.

MALLIK, A., NELSON, J., DASGUPTA, R. (2015) Partial melting of fertile peridotite fluxed by hydrous rhyolitic melt at 2-3 GPa: implications for mantle wedge hybridization by sediment melt and generation of ultrapotassic magmas in convergent margins. *Contributions to Mineralogy and Petrology* 169, 48.

MALLIK, A., DASGUPTA, R., TSUNO, K., NELSON, J. (2016) Effects of water, depth and temperature on partial melting of mantle-wedge fluxed by hydrous sediment-melt in subduction zones. *Geochimica et Cosmochimica Acta* 195, 226-243.

MALPAS, J., FOLEY, S.F., KING, A.F. (1986) Alkaline mafic and ultramafic lamprophyres from the Aillik Bay area, Labrador. *Canadian Journal of Earth Sciences* 23, 1902-1918.

MANASSERO, M.C., ÖZAYDIN, S., AFONSO, J.C., SHEA, J.J., EZAD, I.S., KIRKBY, A., THIEL, S., FOMIN, I., CZARNOTA, K. (2024) Lithospheric structure and melting processes in southeast Australia: new constraints from joint probabilistic inversions of 3D magnetotelluric and seismic data. *Journal of Geophysical Research* 129, e2023JB028257.

MANDLER, B.E., GROVE, T.L. (2016) Controls on the stability and composition of amphibole in the Earth's mantle. *Contributions to Mineralogy and Petrology* 171, 68.

MARQUARDT, K., FAUL, U.H. (2018) The structure and composition of olivine grain boundaries: 40 years of studies, status and current developments. *Physics and Chemistry of Minerals* 45, 139-172.

MATHER, K.A., PEARSON, D.G., MCKENZIE, D., KJARSGAARD, B.A., PRIESTLEY, K. (2011) Constraints on the depth and thermal history of cratonic lithosphere from peridotite xenoliths, xenocrysts and seismology. *Lithos* 125, 729-742.

MATTSSON, H.B., NANDEDKAR, R.H., ULMER, P. (2013) Petrogenesis of the melilititic and nephelinitic rock suites in the Lake Natron-Engaruka monogenetic volcanic field, northern Tanzania. *Lithos* 179, 175-192.

MAXWELL, N. (1984) From knowledge to wisdom. A revolution in the aims and methods of science. Basil Blackwell, Oxford, UK, 298 pp.

MAZZEO, F.C., ROCCO, I., TUCKER, R.D., MORRA, V., D'ANTONIO, M., MELLUSO, L. (2021) Olivine melilitites, mantle xenoliths, and xenocrysts of the Takarindiona district: petrogenesis, magmatic evolution, and the sub-continental lithospheric mantle of east-central Madagascar. *Journal of African Earth Sciences* 174, 104059.

MCGETCHIN, T.R., BESANCON, J.R. (1973) Carbonate inclusions in mantle-derived pyropes. *Earth and Planetary Science Letters* 18, 408-410.

MCINNES, B.I.A., CAMERON, E.M. (1994) Carbonated, alkaline hybridizing melts from a sub-arc environment – mantle wedge samples from the Tabar-Lihir Tang-Feni arc, Papua New Guinea. *Earth and Planetary Science Letters* 122, 125-141.

MCKENZIE, D.P. (1989) Some remarks on the movement of small melt fractions in the mantle. *Earth and Planetary Science Letters* 95, 53-72.

MCKENZIE, D.P., BICKLE, M.J. (1988) The volume and composition of melt generated by extension of the lithosphere. *Journal of Petrology* 29, 625-679.

MEEN, J.K., EGGLER, D.H., AYERS, J.C. (1989) Experimental evidence for very low solubility of rare-earth elements in CO_2-rich fluids at mantle conditions. *Nature* 340, 301-303.

MENGEL, K., GREEN, D.H. (1989) Stability of amphibole and phlogopite in metasomatized peridotite under water-saturated and water-undersaturated conditions. In: Ross, J., Green, D.H. (Eds.) *Kimberlites and Related Rocks Volume 1: Their composition, occurrence, origin, and emplacement*. Blackwell, Melbourne, 571-581.

MENZIES, M.A. (1983) Mantle ultramafic xenoliths in alkaline magmas: evidence for mantle heterogeneity modified by magmatic activity. In: Hawkesworth, C.J., Norry, M.J. (Eds.) *Continental basalts and mantle xenoliths*. Shiva, Nantwich U.K., 92-110.

MENZIES, M.A. (1990) Archean, Proterozoic, and Phanerozoic lithospheres. In: Menzies, M.A. (Ed.) *Continental mantle*. Oxford University Press, Oxford, U.K., 67-86.

MILLHOLLEN, G.L., IRVING, A.J., WYLLIE, P.J. (1974) Melting interval of peridotite with 5.7 percent water to 30 kilobars. *Journal of Geology* 82, 575-587.

MINARIK, W.G., WATSON, E.B. (1995) Interconnectivity of carbonate melt at low melt fractions. *Earth and Planetary Science Letters* 133, 423-437.

MITCHELL, R.H. (1970) Kimberlite and related rocks – a critical reappraisal. *Journal of Geology* 78, 686-704.

MITCHELL, R.H. (1979) The alleged kimberlite-carbonatite relationship: additional contrary mineralogical evidence. *American Journal of Science* 279, 570-589.

MITCHELL, R.H. (1986) *Kimberlites*. Plenum Press, New York, 442 pp.

MITCHELL, R.H. (1995) *Kimberlites, orangeites, and related rocks*. Plenum Press, New York, 410 pp.

MITCHELL, R.H. (2020) Igneous Rock Associations 26. Lamproites, exotic potassic alkaline rocks: a review of their nomenclature, characterization and origins. *Geoscience Canada* 47, 118-142.

MITCHELL, R.H., BERGMAN, S.C. (1991) *Petrology of lamproites*. Plenum Press, New York, 447 pp.

MITCHELL, R.H., EDGAR, A.D. (2002) Melting experiments on SiO_2-rich lamproites to 6.4 GPa and their bearing on the sources of lamproite magmas. *Mineralogy and Petrology* 74, 115-128.

MODRESKI, P., BOETTCHER, A.L. (1973) Phase relationships of phlogopite in the system K_2O-MgO-CaO-Al_2O_3-SiO_2-H_2O to 35 kilobars: a better model for micas in the interior of the Earth. *American Journal of Science* 273, 385-414.

MORISHITA, T., ARAI, S., TAMURA, A. (2003) Petrology of an apatite-rich layer in the Finero phlogopite-peridotite, Italian Western Alps; implications for evolution of a metasomatising agent. *Lithos* 69, 37-49.

MUIRHEAD, J.D., FISCHER, T.P., OLIVA, S.J., LAIZER, A., VAN WIJK, J., CURRIE, C.A., LEE, H., JUDD, E.J., KAZIMOTO, E., SANO, Y., TAKAHATA, N., TIBERI, C., FOLEY, S.F., DUFEK, J., REISS, M.C., EBINGER, C.J. (2020) Displaced cratonic mantle concentrates deep carbon during continental rifting. *Nature* 582, 67-72.

MURAVYEVA, N.S., SENIN, V.G. (2018) Xenoliths from Bunyaruguru volcanic field: some insights into lithology of East African Rift upper mantle. *Lithos* 296, 17-36.

MYSEN, B.O., BOETTCHER, A.L. (1975) Melting of a hydrous mantle: I. Phase relations of a natural peridotite at high pressures and temperatures with controlled activities of water, carbon dioxide, and hydrogen. *Journal of Petrology* 16, 520-548.

MYSEN, B.O., YAMASHITA, S., CHERTKOVA, N. (2008) Solubility and solution mechanisms of NOH volatiles in silicate melts at high pressure and temperature-amine groups and hydrogen fugacity. *American Mineralogist* 93, 1760-1770.

NAUMANN, C.F. (1828) *Lehrbuch der Mineralogie*. August Rüder, Berlin, 643 pp.

NELSON, D.R., MCCULLOCH, M.T., SUN, S.S. (1986) The origins of ultrapotassic rocks as inferred from Sr, Nd and Pb isotopes. *Geochimica et Cosmochimica Acta* 50, 231-245.

NEWTON-SMITH, W.H. (1981) *The rationality of science*. Routledge and Keegan-Paul, London, 294 pp.

NGWENYA, N.S., TAPPE, S. (2021) Diamondiferous lamproites of the Luangwa Rift in central Africa and links to remobilized cratonic lithosphere. *Chemical Geology* 568, 120019.

NI, H., KEPPLER, H. (2013) Carbon in silicate melts. *Reviews in Mineralogy and Geochemistry* 75, 251-287.

NICHOLLS, I.A., WHITFORD, D.J. (1983) Potassium-rich volcanic rocks of the Muriah Complex, Java, Indonesia: products of multiple magma sources? *Journal of Volcanology and Geothermal Research* 18, 337-359.

NIGGLI, P. (1916) Gleichgewichte zwischen TiO_2 und CO_2, sowie SiO_2 und CO_2 in Alkali-, Kalk-Alkali und Alkali-Aluminatschmelzen. *Zeitschrift für Anorganische und Allgemeine Chemie* 98, 241-326.

NIGGLI, P. (1919) Untersuchungen an Karbonat- und Chloridschmelzen. *Zeitschrift für Anorganische und Allgemeine Chemie* 106, 241-326.

NIGGLI, P. (1923) *Gesteins und Mineralprovinzen*. Gebrüder Borntraeger, Berlin, 586 pp.

NIGGLI, P., BEGER, P.J. (1923) *Gesteins und Mineralprovinzen*. Gebrüder Borntraeger, Berlin, 586 pp.

NIIDA, K., GREEN, D.H. (1999) Stability and chemical composition of pargasitic amphibole in MORB pyrolite under upper mantle conditions. *Contributions to Mineralogy and Petrology* 135, 18-40.

NIXON, P.H., HORNUNG, G. (1973) The carbonatite lavas and tuffs near Fort Portal, western Uganda. *Overseas Geology and Mineral Resources, Institute of Geological Sciences* 41, 168-179.

NKOUANDOU, O.F., BARDINTZEFF, J.M., NDASSA, Z.N.N., MEFIRE, A.F., HAMAN, A. (2022) Wehrlite xenoliths and petrogenetic implications, Hossere Do Guessa volcano, Adamawa Massif, Cameroon. *Open Geosciences* 14, 1075-1091.

NOSOVA, A.A., SAZONOVA, L.V., KARGIN, A.V., SMIRNOVA, M.D., LAPIN, A.V., SHCHERBAKOV, V.D. (2018) Olivine in ultramafic lamprophyres: chemistry, crystallisation, and melt sources of Siberian pre- and post-trap aillikites. *Contributions to Mineralogy and Petrology* 173, 55.

NOVELLA, D., FROST, D.J. (2014) The composition of hydrous partial melts of garnet peridotite at 6 GPa: implications for the origin of Group II kimberlites. *Journal of Petrology* 55, 2097-2124.

O'BRIEN, H.E., IRVING, A.J., MCCALLUM, I.S. (1991) Eocene potassic magmatism in the Highwood Mountains, Montana – petrology, geochemistry, and tectonic implications. *Journal of Geophysical Research* 96, 13237-13260.

ODLING, N.W.A. (1995) An experimental replication of upper-mantle metasomatism. *Nature* 373, 58-60.

O'HARA, M.J. (1963) Melting of garnet peridotite at 30 kilobars. *Carnegie Institution of Washington, Yearbook* 62, 71-76.

O'HARA, M.J. (1968) The bearing of phase equilibria studies in synthetic and natural systems on the origin and evolution of basic and ultra-basic rocks. *Earth-Science Reviews* 4, 69-113.

O'HARA, M.J., MERCY, E.L.P. (1963) Petrology and petrogenesis of some garnetiferous peridotites. *Transactions of the Royal Society of Edinburgh* 65, 251-314.

OLAFSSON, M., EGGLER, D.H. (1983) Phase relations of amphibole, amphibole-carbonate, and phlogopite-carbonate peridotite: petrologic constraints on the asthenosphere. *Earth and Planetary Science Letters* 64, 305-315.

O'REILLY, S.Y., GRIFFIN, W.L. (2000) Apatite in the mantle: implications for metasomatic processes and high heat production in Phanerozoic mantle. *Lithos* 53, 217-232.

OXBURGH, E.R. (1964) Petrological evidence for the presence of amphibole in the upper mantle and its petrogenetic and geophysical implications. *Geological Magazine* 101, 1-19.

PALME, H. (2018) Cosmochemistry along the Rhine. *Geochemical Perspectives* 7, 116 pp.

PALYANOV, Y.N., SOKOL, A.G., TOMILENKO, A.A., SOBOLEV, N.V. (2005) Conditions of diamond formation through carbonate-silicate interaction. *European Journal of Mineralogy* 17, 207-214.

PARK, M., LEAHEY, E., FUNK, R.J. (2023) Papers and patents are becoming less disruptive over time. *Nature* 613, 138-144.

PASTERIS, J.D., BOYD, F.R., NIXON, P.H. (1979) The ilmenite association at the Frank Smith Mine, R.S.A. In: Boyd, F.R., Meyer, H.O.A. (Eds.) *The mantle sample: inclusions in kimberlites and other volcanics*. Proceedings of the Second International Kimberlite Conference (American Geophysical Union), 265-278.

Pearson, D.G., Woodhead, J., Janney, P.E. (2019) Kimberlites as geochemical probes of Earth's mantle. *Elements* 15, 387-392.

Peat, F.D. (1987) *Synchronicity: the bridge between matter and mind*. Bantam Books, Toronto, 245 pp.

Petersen, T.D., Le Cheminant, A.N. (1993) Glimmerite xenoliths in early Proterozoic ultrapotassic rocks from the Churchill Province. *Canadian Mineralogist* 31, 801-819.

Peterson, N., Barrett, N., Russell, J.K. (2025) Carbonated mantle lithosphere from the western Canadian Cordillera. *Canadian Journal of Earth Sciences* 62, 892-908.

Piccardo, G., Rampone, E., Vannucci, R., Shimizu, N., Ottolini, L., Bottazzi, P. (1993) Mantle processes in the sub-continental lithosphere: the case study of the rifted sp-lherzolites from Zabargad (Red Sea). *European Journal of Mineralogy* 5, 1039-1056.

Pilet, S., Baker, M.B., Stolper, E.M. (2008) Metasomatized lithosphere and the origin of alkaline lavas. *Science* 320, 916-919.

Pilet, S., Ulmer, P., Villiger, S. (2009) Liquid line of descent of a basanitic liquid at 1.5 GPa: constraints on the formation of metasomatic veins. *Contributions to Mineralogy and Petrology* 159, 621-643.

Pinkerton, J. (1800) *Petralogy. A treatise on rocks*. White, Cochrane, and Co., London, 599 pp.

Pintér, Z., Foley, S.F., Yaxley, G.M., Rosenthal, A., Rapp, R.P., Lanati, A.W., Rushmer, T. (2021) Experimental investigation of the composition of incipient melts in upper mantle peridotites in the presence of CO_2 and H_2O. *Lithos* 396-397, 106224.

Pintér, Z., Foley, S.F., Yaxley, G.M. (2022) Diamonds, dunites and metasomatic rocks formed by melt/rock reaction in craton roots. *Communications Earth and Environment* 3, 296.

Pivin, M., Féménias, O., Demaiffe, D. (2009) Metasomatic mantle origin for the Mbuji-Mayi and Kundelungu garnet and clinopyroxene megacrysts (Democratic Republic of Congo). *Lithos* 112S, 951-960.

Planck, M. (1949) *Scientific autobiography and other papers*. New York. 102 pp.

Ponomarenko, A.I., Tsepin, A.I. (1987) First occurrence of diamond-bearing phlogopite xenolith from the pipe Mir, Yakutia. *Doklady Akademii Nauk, Earth Science Section* 297,1210-1213.

Prelević, D., Jacob, D.E., Foley, S.F. (2013) Recycling plus: a new recipe for the formation of Alpine-Himalayan orogenic mantle lithosphere. *Earth and Planetary Science Letters* 362, 187-197.

Prelević, D., Förster, M.W., Buhre, S., Gülmez, F., Grützner, T., Wang, Y., Foley, S.F. (2024) Recent advances made by reaction experiments on melting of heavily metasomatized hydrous mantle. *Earth-Science Reviews* 256, 104881.

Ramokgaba, L., Foley, S.F., Howarth, G.H. (2022) Modification of harzburgite to K-rich dolomite-bearing peridotite from the Kaapvaal Craton, South Africa: evidence of metasomatism by carbonate-silicate incipient melts. *Goldschmidt 2022 Abstract* 146.

Rapp, R.P., Shimizu, N., Norman, M.D., Applegate, G.S. (1999) Reaction between slab-derived melts and peridotite in the mantle wedge: experimental constraints at 3.8 GPa. *Chemical Geology* 160, 335-356.

Rehfeldt, T., Foley, S.F., Carlson, R.W., Lowry, D., Jacob, D.E. (2008) Contrasting types of metasomatism in dunite, wehrlite and websterite xenoliths from Kimberley, South Africa. *Geochimica et Cosmochimica Acta* 72, 5722-5756.

Ren, H.E., Casalini, M., Conticelli, S., Chen, C.F., Foley, S.F., Feng, L.P., Liu, Y.S. (2024) Calcium isotope compositions of subduction-related leucite-bearing rocks: implications for the calcium isotope heterogeneity of the mantle and carbonate recycling in convergent margins. *Geochimica et Cosmochimica Acta* 364, 100-113.

Righter, K., Rosas-Eiguera, J. (2001) Alkaline lavas in the volcanic front of the western Mexican Volcanic Belt: geology and petrology of the Ayutla and Tapalpa volcanic fields. *Journal of Petrology* 42, 2333-2361.

RINGWOOD, A.E. (1962) A model for the upper mantle. *Journal of Geophysical Research* 67, 857-867.

RINGWOOD, A.E. (1967) The pyroxene-garnet transformation in the Earth's mantle. *Earth and Planetary Science Letters* 2, 255-263.

RINGWOOD, A.E., MAJOR, A. (1966) Synthesis of Mg_2SiO_4-Fe_2SiO_4 spinel solid solutions. *Earth and Planetary Science Letters* 1, 241-245.

RIVALENTI, G., ZANETTI, A., MAZZUCCHELLI, M., VANNUCCI, R., CINGOLANI, C.A. (2004) Equivocal carbonatite markers in the mantle xenoliths of the Patagonia backarc: the Gobernador Gregores case (Santa Cruz province, Argentina). *Contributions to Mineralogy and Petrology* 147, 647-670.

ROBINSON, J.A.C., WOOD, B.J., BLUNDY, J.D. (1998) The beginning of melting of fertile and depleted peridotite. *Earth and Planetary Science Letters* 155, 97-111.

ROCK, N.M.S. (1977) The nature and origin of lamprophyres: some definitions, distinctions, and derivations. *Earth-Science Reviews* 13, 123-169.

ROCK, N.M.S. (1981) How should igneous rocks be grouped? *Geological Magazine* 118, 449-461.

ROCK, N.M.S. (1983) The Permo-Carboniferous camptonite-monchiquite dyke suite of the Scottish Highlands and Islands: distribution, field and petrological aspects. *Institute of Geological Sciences Report* 82/14.

ROCK, N.M.S. (1984) The nature and origin of calc-alkaline lamprophyres: minettes, vogesites, kersantites and spessartites. *Transactions of the Royal Society of Edinburgh* 74, 193-227.

ROCK, N.M.S. (1986) The nature and origin of ultramafic lamprophyres: alnöites and allied rocks. *Journal of Petrology* 27, 155-196.

ROCK, N.M.S. (1987) The nature and origin of lamprophyres: an overview. *Geological Society of London Special Publication* 30, 191-226.

ROCK, N.M.S. (1991) *Lamprophyres*. Blackie and Son, Ltd, Glasgow.

ROHRBACH, A., SCHMIDT, M.W. (2011) Redox freezing and melting in Earth's deep mantle resulting from carbon-iron redox coupling. *Nature* 472, 209-212.

ROHRBACH, A., BALLHAUS, C., GOLLA-SCHINDLER, U., ULMER, P., KAMENETSKY, V.S., KUZMIN, D.V. (2007) Metal saturation in the upper mantle. *Nature* 449, 456-458.

ROHRBACH, A., BALLHAUS, C., ULMER, P., GOLLA-SCHINDLER, U., SCHÖNBOHM, D. (2011) Experimental evidence for a reduced metal-saturated upper mantle. *Journal of Petrology* 52, 717-731.

ROONEY, T.O., GIRARD, G., TAPPE, S. (2020) The impact on mantle olivine resulting from carbonated silicate melt interaction. *Contributions to Mineralogy and Petrology* 175, 56.

RÖPER, T., BUSWEILER, Y., HELLERS, M., MÖCKEL, F., WOMBACHER, F., MÜNKER, C. (2025) A sheared garnet peridotite mantle xenolith from the Delitzsch Carbonatite Complex, Germany – evidence for thickened Cretaceous lithosphere beneath Central Europe. *Lithos* 512, 108156.

ROSENBUSCH, H. (1877) *Mikroskopische Physiographie der massigen Gesteine*. E. Schweizerbart'sche Verlagshandlung, Stuttgart.

ROSENBUSCH, H. (1887) *Mikroskopische Physiographie der Mineralien und Gesteine. Ein Hilfsbuch bei mikroskopischen Gesteinsstudien*. E. Schweizerbart'sche Verlagshandlung, Stuttgart. 806 pp.

ROSENTHAL, A., FOLEY, S.F., PEARSON, D.G., NOWELL, G.M., TAPPE, S. (2009) Magmatic evolution at the propagating tip of a continental rift – a geochemical study of primitive alkaline volcanic rocks of the western branch of the East African Rift. *Earth and Planetary Science Letters* 284, 236-248.

RUSSELL, B. (1931) *The scientific outlook*. George Allen and Unwin, London, 285 pp.

RUSSELL, B. (1946) *A history of western philosophy*. George Allen and Unwin, London, 842 pp.

RUSSELL, B. (1952) *Dictionary of mind, matter and morals*. Philosophical Library, London, 280 pp.

SALTERS, V.J.M, DICK, H.J.B. (2002) Mineralogy of the mid-ocean ridge basalt source from neodymium isotopic composition in abyssal peridotites. *Nature* 418, 68-72.

SATTLER, R. (1986) *Biophilosophy: analytic and holistic perspectives*. Springer-Verlag, Berlin, 284pp.

SCHAIRER, F., BOWEN, N.L. (1947) Melting relations in the system Na_2O-Al_2O_3-SiO_2 and K_2O-Al_2O_3-SiO_2. *American Journal of Science* 245, 193-204.

SCHMIDT, M.W., GIULIANI, A., POLI, S. (2024) The origin of carbonatites – combining the rock record with available experimental constraints. *Journal of Petrology* 65, egae105.

SCHNEIDER, M.E., EGGLER, D.H. (1986) Fluids in equilibrium with peridotite minerals: implications for mantle metasomatism. *Geochimica et Cosmochimica Acta* 50, 711-724.

SCOTT, B.H. (1981) Kimberlite and lamproite dykes from Holsteinsborg, West Greenland. *Meddelelser om Grønland* 4, 24 pp.

SEENSTRA, E.S., KLAVER, M., BERNDT, J., FLEMETAKIS, S., ROHRBACH, A., KLEMME, S. (2024) Thermal stability of F-rich phlogopite and K-richterite during partial melting of metasomatized mantle peridotite with implications for deep Earth volatile cycles. *Journal of Geophysical Research* 129, e2023JB28202.

SEKINE, T., WYLLIE, P.J. (1982a) The system granite-peridotite-H_2O at 30 kbar, with applications to hybridization in subduction zone magmatism. *Contributions to Mineralogy and Petrology* 81, 190-202.

SEKINE, T., WYLLIE, P.J. (1982b) Phase relationships in the system $KAlSiO_4$-Mg_2SiO_4-SiO_2-H_2O as a model for hybridization between hydrous siliceous melts and peridotite. *Contributions to Mineralogy and Petrology* 79, 368-374.

SHAND, S.J. (1913) On saturated and unsaturated igneous rocks. *Geological Magazine* 10, 508-514.

SHAO, J.A., HAN, Q.J., ZHANG, L.Q., MU, B.L. (1999) Cumulate complex xenoliths in the early Mesozoic in eastern Inner Mongolia. *Chinese Science Bulletin* 44, 1272-1279.

SHATSKIY, A.A., BEKHTENOVA, A., AREFIEV, A.V., PODBORODNIKOV, I.V., VINOGRADOVA, Y.G., REZVUKHIN, D.I., LITASOV, K.D. (2022) Solidus and melting of carbonated phlogopite peridotite at 3-6.5 GPa: implications for mantle metasomatism. *Gondwana Research* 101, 156-174.

SHERVAIS, J. (1979) Ultramafic and mafic layers in the Alpine-type lherzolite massif at Balmuccia, N.W. Italy. *Memoire di Scienze Geologiche, Padova* 33, 135-145.

SHU, C., FOLEY, S.F., EZAD, I.S., DACZKO, N., SHCHEKA, S.S. (2024) Phlogopite websterite in the upper mantle: a common source for ultrapotassic rocks. *Journal of Petrology* 65, agae030.

SHU, Z.T., LIU, S.A., PRELEVIC, D., CVETKOVIC, V., LI, S. (2022) Recycling of carbonates into the deep mantle beneath central Balkan Peninsula: Mg-Zn isotopic evidence. *Lithos* 432-433, 106899.

SLEEP, N.H. (1988) Tapping of melt by veins and dikes. *Journal of Geophysical Research* 93, 10255-10272.

SMITH, C.B. (1983) Pb, Sr and Nd isotopic evidence for sources of southern African kimberlites. *Nature* 304, 51-54.

SNOW, C.P. (1963) *The two cultures*. Cambridge University Press, Cambridge, U.K., 107 pp.

SOBOLEV, A.V., HOFMANN, A.W., SOBOLEV, S.V., NIKOGOSIAN, I.K. (2005) An olivine-free mantle source of Hawaiian shield basalts. *Nature* 434, 590–597.

SOBOLEV, A.V., SOBOLEV, S.V., KUZMIN, D.V., MALITCH, K.N., PETRUNIN, A.G. (2009) Siberian meimechites: origin and relation to flood basalts and kimberlites. *Russian Geology and Geophysics* 50, 999-1033.

STACHEL, T., HARRIS, J.W. (2008) The origin of cratonic diamonds – constraints from mineral inclusions. *Ore Geology Reviews* 34, 5-32.

STAGNO, V. (2019) Carbon, carbides, carbonates and carbonatitic melts in the Earth's interior. *Journal of the Geological Society* 176, 375-387.

STOSCH, H.G., LUGMAIR, G.W. (1986) Trace element and Sr and Nd isotope geochemistry of peridotite xenoliths from the Eifel (West Germany) and their bearing on the evolution of the subcontinental mantle. *Earth and Planetary Science Letters* 80, 281-298.

STRECKEISEN, A. (1976) To each plutonic rock its proper name. *Earth-Science Reviews* 12, 1-33.

STRECKEISEN, A. (1979) Classification and nomenclature of volcanic rocks, lamprophyres, carbonatites and melilitic rocks. *Neues Jahrbuch für Mineralogie, Abhandlungen* 134, 1-14.

STRONG, D.F., STEVENS, R.K., MALPAS, J., BADHAM, J.P.N. (1975) A new tale for the Lizard. *Proceedings of the Ussher Society* 3, 252.

SUDHOLZ, Z.J., GREEN, D.H., YAXLEY, G.M., JAQUES, A.L. (2022) Mantle geothermometry: experimental evaluation and recalibration of Fe-Mg geothermometers for garnet-clinopyroxene and garnet-othropyroxene in peridotite, pyroxenite and eclogite systems. *Contributions to Mineralogy and Petrology* 177, 77.

SUZUKI, K. (1979) Hornblendite inclusion from the Ngong Hills in the southern part of the Gregory Rift, Kenya. *Fourth preliminary report of African Studies, Nagoya University*, 151-162.

SWEENEY, R.J., THOMPSON, A.B., ULMER, P. (1993) Phase relations of a natural MARID composition and implications for MARID genesis, lithospheric melting and mantle metasomatism. *Contributions to Mineralogy and Petrology* 115, 225-241.

TAKAHASHI, E., IRVINE, T.N. (1981) Stoichiometric control of crystal/liquid single-component partition coefficients. *Geochimica et Cosmochimica Acta* 45, 1181-1185.

TAKAHASHI, E., SCARFE, C.M. (1985) Melting of peridotite to 14 GPa and the genesis of komatiite. *Nature* 315, 566-568.

TAPPE, S., FOLEY, S.F., JENNER, G.A., KJARSGAARD, B.A. (2005) Integrating ultramafic lamprophyres into the IUGS classification of igneous rocks: rationale and implications. *Journal of Petrology* 46, 1893-1900.

TAPPE, S., FOLEY, S.F., JENNER, G.A., HEAMAN, L.M., KJARSGAARD, B.A., ROMER, R.L., STRACKE, A., JOYCE, N., HOEFS, J. (2006) Genesis of ultramafic lamprophyres and carbonatites at Aillik Bay, Labrador: a consequence of incipient lithospheric thinning beneath the North Atlantic Craton. *Journal of Petrology* 47, 1261-1315.

TAPPE, S., FOLEY, S.F., KJARSGAARD, B.A., ROMER, R.L., HEAMAN, L.M., STRACKE, A., JENNER, G.A. (2008) Between carbonatite and lamproite – diamondiferous Torngat ultramafic lamprophyres formed by carbonate-fluxed melting of cratonic MARID-type metasomes. *Geochmica et Cosmochimica Acta* 72, 3258-3286.

TAPPE, S., STEENFELT, A., HEAMAN, L.M., SIMONETTI, A. (2009) The newly discovered Jurassic Tikuissaq carbonatite-aillikite occurrence, West Greenland, and some remarks on carbonatite-kimberlite relationships. *Lithos* 112S, 385-399.

TAPPE, S., SHAIKH, A.M., WILSON, A.H., STRACKE, A. (2022) Evolution of ultrapotassic volcanism on the Kaapvaal craton: deepening the orangeite versus lamproite debate. *Geological Society Special Publication* 513, 17-44.

TAYLOR, W.R., GREEN, D.H. (1987) The petrogenetic role of methane: effect on liquidus phase relations and the solubility mechanism of reduced C–H volatiles, In: Mysen, B.O. (Ed.) *Magmatic processes: physicochemical principles*. Geochemical Society, University Park, PA, 121-138.

TAYLOR, W.R., GREEN, D.H. (1988) Measurement of reduced peridotite-C-O-H solidus and implications for redox melting of the mantle. *Nature* 332, 239-352.

THOMPSON, R.N. (1972) Melting behaviour of two Snake River lavas at pressures up to 35 kbar. *Carnegie Institution of Washington Yearbook* 71, 406-410.

TOMMASINI, S., AVANZINELLI, R., CONTICELLI, S. (2011) The Th/La and Sm/La conundrum of the Tethyan realm lamproites. *Earth and Planetary Science Letters* 301, 469-478.

TRONNES, R.G. (2002) Stability range and decomposition of potassic richterite and phlogopite end members at 5-15 GPa. *Mineralogy and Petrology* 74, 129-148.

TULLOCH, A.J., NATHAN, S. (1990) Spinel harzburgite xenoliths in alkali basal and camptonite from North Westland and southeast Nelson, New Zealand. *New Zealand Journal of Geology and Geophysics* 33, 529-534.

TURNER, F.J., VERHOOGEN, J. (1960) *Igneous and metamorphic petrology*. McGraw-Hill, New York, 2nd Edition, 694 pp.

TURNER, H.W. (1901) Perknite (lime-magnesia rocks). *Journal of Geology* 9, 507-511.

VAN DER LAAN, S.R., FOLEY, S.F. (1994) MARIDs and mantle metasomatism. *Mineralogical Magazine* 58A, 505-506.

VARNE, R. (1970) Hornblende lherzolite and the upper mantle. *Contributions to Mineralogy and Petrology* 27, 45-51.

VARTIAINEN, H., KRESTEN, P., KAFKAS, Y. (1978) Alkaline lamprophyres from the Sokli Complex, northern Finland. *Bulletin of the Geological Society of Finland* 50, 5-68.

VASELLI, O., DOWNES, H., THIRLWALL, M.F., DOBOSI, G., CORADOSSI, N., SEGHEDI, I., SZAKACS, A., VANNUCCI, R. (1995) Ultramafic xenoliths in Plio-Pleistocene alkali basalts from the Eastern Transylvanian Basin: depleted mantle enriched by vein metasomatism. *Journal of Petrology* 36, 23-53.

VETER, M., FOLEY, S., MERTZ-KRAUS, R., GROSCHOPF, N. (2017) Trace elements in olivine of ultramafic lamprophyres controlled by phlogopite-rich mineral assemblages in the mantle source. *Lithos* 292, 81-95.

VON ECKERMANN, H. (1966) Progress of research on the Alnö carbonatite. In: Tuttle, O.F., Gittins, J. (Eds.) *Carbonatites*. John Wiley and Sons, New York, 3-31.

WAGNER, C., MOKHTARI, A., VELDE, D. (1993) Xenocrystic richterite in an olivine nephelinite: destabilisation and diffusion phenomena. *Mineralogical Magazine* 57, 515-525.

WAGNER, C., DELOULE, E., MOKHTARI, A. (1996) Richterite-bearing peridotites and MARID-type inclusions in lavas from North-Eastern Morocco: mineralogy and D/H isotopic studies. *Contributions to Mineralogy and Petrology* 124, 406-421.

WAGNER, C., MOKHTARI, A., DELOULE, E., CHABAUX, F. (2003) Carbonatite and alkaline magmatism in Taourirt (Morocco): petrological, geochemical and Sr-Nd isotope characteristics. *Journal of Petrology* 44, 937-965.

WAGNER, P.A. (1928) The evidence of the kimberlite pipes on the constitution of the outer part of the Earth. *South African Journal of Science* 25, 127-148.

WALKER, D. (1986) Melting equilibria in multicomponent systems and liquidus/solidus convergence in mantle peridotite. *Contributions to Mineralogy and Petrology* 92, 303-307.

WALLACE, M.E., GREEN, D.H. (1988) An experimental determination of primary carbonatite magma composition. *Nature* 335, 343-346.

WALLACE, M.E., GREEN, D.H. (1991) The effect of bulk composition on the stability of amphibole in the upper mantle: implications for solidus positions and mantle metasomatism. *Mineralogy and Petrology* 44, 1-19.

WALTER, M.J. (1998) Melting of garnet peridotite and the origin of komatiite and depleted lithosphere. *Journal of Petrology* 39, 29-60.

WANG, J., FOLEY, S.F., WANG, X.F., WANG, C., LIU, Y.S. (2024) Melting behaviour of impure limestone under H_2O-poor conditions: implications for the contribution of carbonate-rich sediments to arc magmatic carbon output. *Chemical Geology* 654, 122066.

WANG, Y., FOLEY, S.F. (2018) Hybridization melting between continent-derived sediment and depleted peridotite in subduction zones. *Journal of Geophysical Research* 123, 3414-3429.

WANG, Y., FOLEY, S.F., PRELEVIĆ, D. (2017) Potassium-rich magmatism from a phlogopite-free source. *Geology* 45, 467-470.

WANG, Y., FOLEY, S.F., BUHRE, S., SOLDNER, J., XU, Y.G. (2021) Origin of potassic post-collisional volcanic rocks in young, shallow, blueschist-rich lithosphere. *Science Advances* 7, eabc0291.

WASHINGTON, H.S. (1925) The chemical composition of the Earth. *American Journal of Science* 9, 351-378.

WASS, S.Y. (1979) Fractional crystallization in the mantle of late-stage kimberlitic liquids – evidence in xenoliths from the Kiama area, N.S.W., Australia. In: Boyd, F.R., Meyer, H.O.A. (Eds.) *The Mantle sample: inclusions in kimberlites and other volcanics*. Proceedings of the Second International Kimberlite Conference (American Geophysical Union), 366-373.

WASS, S.Y., HENDERSON, P., ELLIOTT, C.J. (1980) Chemical heterogeneity and metasomatism in the upper mantle: evidence from rare earth and other elements in apatite-rich xenoliths in basaltic rocks from eastern Australia. *Philosophical Transactions of the Royal Society of London Series A* 297, 333-346.

WATERS, F.G. (1987) A suggested origin of MARID xenoliths in kimberlites by high pressure crystallization of an ultrapotassic rock such as lamproite. *Contributions to Mineralogy and Petrology* 95, 523-533.

WATERS, F.G., ERLANK, A.J., DANIELS, L.R.M. (1989) Contact relationships between MARID rock and metasomatised peridotite in a kimberlite xenolith. *Geochemical Journal* 23, 11-17.

WATSON, E.B. (1982) Melt infiltration and magma evolution. *Geology* 10, 236-240.

WATSON, E.B., BRENAN, J.M. (1987) Fluids in the lithosphere. 1: Experimentally determined wetting characteristics of CO_2-H_2O fluids and their implications for fluid transport, host-rock physical properties, and fluid inclusion formation. *Earth and Planetary Science Letters* 85, 497-515.

WATTS, A. (1951) *The wisdom of insecurity*. Rider, London, 152 pp.

WEISS, Y., MCNEILL, J., PEARSON, D.G., NOWELL, G.M., OTTLEY, C.J. (2015) Highly saline fluids from subducting slab as the source for fluid-rich diamonds. *Nature* 524, 339-342.

WENDLANDT, R.F., EGGLER, D.H. (1980a) The origins of potassic magmas: 1. Melting relations in the systems $KAlSiO_4$-Mg_2SiO_4-SiO_2 and $KAlSiO_4$-MgO-SiO_2-CO_2 to 30 kilobars. *American Journal of Science* 280, 38-420.

WENDLANDT, R.F., EGGLER, D.H. (1980b) The origins of potassic magmas: 2. Stability of phlogopite in natural spinel lherzolite and in the system $KAlSiO_4$-MgO-SiO_2-H_2O-CO_2 at high pressures and high temperatures. *American Journal of Science* 280, 421-458.

WHITEHEAD, A.N. (1926) *Science and the modern world*. Cambridge University Press, Cambridge, U.K, 265 pp.

WILBER, K. (1988) *No boundary*. Shambhala Publications, Boston, USA, 160 pp.

WILLIAMSON, E.D., ADAMS, L.H. (1923) Density distribution in the earth. *Journal of the Washington Academy of Sciences* 13, 413-428.

WILSHIRE, H.G. (1987) A model of mantle metasomatism. In: Morris, E.M., Pasteris, J.D. (Eds.) *Mantle metasomatism and alkaline magmatism*. *Geological Society of America Special Paper* 215, 47-60.

WILSHIRE, H.G., SHERVAIS, J. (1975) Al augite and Cr diopside ultramafic xenoliths in basaltic rocks from the western United States. *Physics and Chemistry of the Earth* 9, 257-272.

WITT-EICKSCHEN, G., KAMINSKY, W., KRAMM, U., HARTE, B. (1998) The nature of young vein metasomatism in the lithosphere of the West Eifel (Germany): geochemical and isotopic constraints from composite mantle xenoliths from the Meerfelder Maar. *Journal of Petrology* 39, 155-185.

WITTGENSTEIN, L. (1918) *Tractatus logico-philosophicus. Logisch-philosophische Abhandlungen*. Edition Suhrkamp Verlag, Frankfurt am Main, 115 pp.

WOERMANN, E., ROSENHAUER, M. (1985) Fluid phases and the redox state of the Earth's mantle. *Fortschritte der Mineralogie* 63, 263-349.

WOLF, T. (1867) Die Auswürflinge des Laacher-Sees. *Zeitschrift der Deutschen Geologischen Gesellschaft* 19, 451-492.

Woodhead, J.D., Hergt, J., Giuliani, A., Phillips, D., Pearson, D.G., Nowell, G.M. (2019) Kimberlites reveal 2.5-billion-year evolution of a deep, isolated mantle reservoir. *Nature* 573, 578-581.

Woodland, A.B., Peltonen, P. (1999) Ferric iron contents of garnet and clinopyroxene and estimated oxygen fugacities of peridotite xenoliths from the Eastern Finland Kimberlite Province. *Proceedings of the 7th International Kimberlite Conference*, Red Roof Design, Cape Town, 904-911.

Wyllie, P.J. (1978) Mantle fluid compositions buffered in peridotite-CO_2-H_2O by carbonates, amphibole, and phlogopite. *Journal of Geology* 86, 687-713.

Wyllie, P.J. (1980) The origin of kimberlites. *Journal of Geophysical Research* 85, 6902-6910.

Wyllie, P.J. (1987) Volcanic rocks: boundaries from experimental petrology. *Fortschritte der Mineralogie* 65, 249-284.

Wyllie, P.J. (1988) Magma genesis, plate tectonics, and chemical differentiation of the Earth. *Reviews of Geophysics* 26, 370-404.

Wyllie, P.J., Tuttle, O.F. (1959) Effect of carbon dioxide on the melting of granite and feldspars. *American Journal of Science* 257, 648-655.

Wyllie, P.J., Tuttle, O.F. (1960) The system CaO-CO_2-H_2O and the origin of carbonatites. *Journal of Petrology* 1, 1-46.

Wyllie, P.J., Tuttle, O.F. (1961) Experimental investigation of silicate systems containing two volatile components, II. The effects of NH_3 and HF, in addition to H_2O, on the melting temperatures of albite and granite. *American Journal of Science* 259, 128-143.

Wyllie, P.J., Donaldson, C.H., Presnall, D.C., Irving, A.J., Stolper, E.M., Kesson, S.E., Usselman, T.M., Merrill, R.B., Walker, D. (1981) Experimental petrology of basalts and their source rocks. *Basaltic volcanism on the terrestrial planets*. Pergamon Press, Oxford, 493-630.

Xu, X.Y., Huang, H.Y., Xia, L.Q., Xia, Z.C. (1999) Features of the early Palaeozoic mantle beneath Langao County and its formation mechanism. *Acta Geologica Sinica* 73, 356-365.

Yaxley, G.M., Green, D.H. (1998) Reactions between eclogite and peridotite: mantle refertilisation by subduction of oceanic crust. *Schweizerische Mineralogische und Petrographische Mitteilungen* 78, 243-255.

Yaxley, G.M., Crawford, A.J., Green, D.H. (1991) Evidence for carbonatite metasomatism in spinel peridotite xenoliths from western Victoria, Australia. *Earth and Planetary Science Letters* 107, 305-317.

Yaxley, G.M., Kjarsgaard, B.A., Jaques, A.L. (2019) Evolution of carbonatite magmas in the upper mantle and crust. *Elements* 17, 315-320.

Yoder, H.S. (1976) *Generation of basaltic magma*. National Academy of Sciences, Washington, D.C., USA, 265 pp.

Yoder, H.S. (1993) Timetable of petrology. *Journal of Geological Education* 41, 447-489.

Yoder, H.S., Tilley, C.E. (1962) Origin of basalt magmas: an experimental study of natural and synthetic rock systems. *Journal of Petrology* 343-532.

Yoder, H.S., Kushiro, I. (1969) Melting of a hydrous phase: phlogopite. *American Journal of Science* 267-A, 558-582.

Young, D.A. (2003) *Mind over magma*. Princeton University Press, Princeton, New Jersey, USA, 686 pp.

Zeng, G., Chen, L.H., Hofmann, A.W., Wang, X.J., Liu, J.Q., Yu, X., Xie, L.W. (2021) Nephelinites in eastern China originating from the mantle transition zone. *Chemical Geology* 576, 120276.

Zhang, W.L., Shao, J.A., Xu, X.S., Wang, R.C., Chen, L.H. (2007) Mantle metasomatism by P- and F-rich melt-fluids: evidence from phosphate glass in spinel lherzolite xenolith in Keluo, Heilongjiang Province. *Chinese Science Bulletin* 52, 1827-1835.

ZHANG, Y.F., LIANG, X., WANG, C., JIN, Z.M., ZHU, L., GAN, W. (2020) Experimental constraints on the partial melting of sediment-metasomatized lithospheric mantle in subduction zones. *American Mineralogist* 105, 1191-1203.

ZHU, Y.T., LIU, Y.S., XU, R., MOYNIER, F., LI, M., CHEN, H.H. (2021) Deciphering the origin of a basanite-alkali basalt-tholeiite suite using Zn isotopes. *Chemical Geology* 585, 120585.

ZIRKEL, F. (1870) *Untersuchungen über die mikroskopische Zusammensetzung und Struktur der Basaltgesteine*. Moncur, Bonn, Germany, 208 pp.

ZOZULYA, D.R., KULLERUD, K., RIBACKI, E., ALTENBERGER, U., SUDO, M., SAVCHENKO, Y.E. (2020) The newly discovered Neoproterozoic aillikite occurrence in Vinoren (southern Norway): age, geodynamic position and mineralogical evidence of diamond-bearing mantle source. *Minerals* 10, 1029.

INDEX

A

accessory phases 194, 313, 315, 331
adiabatic 279
aillikite 227, 237–239, 247, 290, 324, 325, 330, 364, 368
alkali basalt 195
alkaline lamprophyre 227
alkaline rocks V, 193, 194, 226–230, 233, 242, 244, 258, 267, 271, 276, 283, 296–298, 306, 316–318, 322, 323, 343, 345–347, 359
alkali picrite V, 226, 321, 337
alnöite 227, 229, 247
amphibole 194, 261, 267, 322
apatite II, 194, 229, 254, 258, 259, 261, 265–268, 291, 300, 306–308, 310–315, 319–24, 329, 331, 332, 343, 350, 355, 359, 366
Archean 196, 215, 249, 343, 354, 359
asthenosphere 236, 247, 275, 333–335, 338, 340, 352, 360
Australia II, VII, 203, 206, 208, 209, 216, 221–225, 230, 234, 261, 266, 306, 342, 343, 345, 349, 350, 354, 358, 366, 367
Australian National University VI, VII, 211, 213, 223, 225

B

basalt 208, 229, 231, 237, 245, 246, 251, 260, 271, 276, 282, 293, 294, 321, 346, 348, 350, 356, 357, 363, 367–369
basalt tetrahedron 276, 282
basanite 193, 231, 242, 244, 257, 270, 286, 300, 317, 319, 324, 331, 368
belt apparatus 209–211, 214, 344

C

calc-alkaline lamprophyre 227
calcic amphibole 195, 242, 253, 261, 264, 265, 304, 305, 310–315, 317, 319, 322, 338
calcite 260, 268, 269, 325
carbon 196, 273, 275, 284, 292, 333–341, 343, 345, 346, 350, 351, 354, 359, 362, 365, 367
carbonate II, 194, 195, 227, 229, 230, 231, 238, 239, 244, 247, 248, 257, 258, 260, 261, 267–269, 275, 286, 287, 290, 292, 296–298, 301, 307, 316, 319, 321,

323–326, 328, 331–336, 338, 341, 342–346, 348, 351, 353–355, 357, 359, 360, 361, 364, 365
carbonated silicate melt 307, 358, 362
carbonatite 208, 229, 230, 245, 247, 259, 267–269, 275, 284, 287–290, 300, 301, 307, 324, 325, 327, 330–333, 338, 343, 352, 353–356, 359, 360, 362, 364, 365, 367
carbon dioxide 196, 346, 359, 367
CCFS II, 221, 222, 224
central complex kimberlites 236, 238
CH4 279, 292–294, 333
clastic sediments 270, 300, 328
clinopyroxene 193, 194, 196, 227, 233, 245, 246, 257, 259–268, 270, 274, 296, 297, 300, 301, 304, 306–308, 310–312, 319–324, 330, 361, 364, 367
clinopyroxenite 308, 315, 351, 357
composite xenolith 198, 249, 255, 261, 263, 270, 320
craton 254, 261, 263, 268, 269, 345, 351, 357, 361, 364
cratonic lithosphere 249, 251, 261, 293, 306, 335, 338, 349, 354, 358, 359

D

damtjernite 227, 238
diamond 230, 235, 248, 264, 268, 273, 292, 330, 333–335, 345, 349, 351, 353, 354, 356, 360, 361, 368
dolomite 258, 268, 286, 361
dyke V, 195, 196, 202, 229, 233, 236, 237, 239, 247, 302, 306, 323, 324, 337, 338, 342, 362

E

East African Rift 244, 267, 343, 359, 362
eclogite 251, 252, 293, 357, 358, 364, 367
experiment 208, 273–276, 283, 284, 295, 298, 303, 308

F

first row transition elements 233, 348
fluid 221, 224, 366

G

garnet 193, 196, 232, 247, 254, 257, 260–262, 264, 265, 268, 278, 293, 294, 296, 297, 299, 301, 302, 304, 308, 310, 315, 323–325, 328, 331, 332, 346, 349–351, 353, 360–362, 364, 365, 367
Gaussberg 207, 234, 247
Geocycles II, 218–224
Germany II, 203, 206, 209, 210, 212, 214, 216–220, 222–225, 252, 253, 265, 343, 352, 353, 362, 364, 366, 368
glimmerite 199, 309, 350
Göttingen I, 210, 212–214, 216, 217, 223, 274
Greifswald II, VIII, 216–218, 223, 229

H

H_2O II, IV, 194, 226, 227, 231, 239, 240, 242, 247, 248, 255, 272, 279, 284, 286–294, 297, 300, 305–308, 325, 329, 330, 333, 335, 336, 347, 350, 359, 361, 363, 365–367
habilitation II, 213–216
hornblendite 199, 258, 265, 308, 313, 343
hydrous pyroxenite 199

I

ilmenite 194, 254, 255, 258, 260–262, 265, 266, 270, 295, 297, 313–315, 319, 324, 332, 338, 347, 360
immiscibility 230, 275, 349, 352, 355, 356
Inagli 214, 307
incipient melting 284, 290

K

Kaapvaal 247, 254, 261–263, 268, 269, 315, 353, 355, 361, 364
kamafugite 227
kimberlite 203, 212, 226, 230, 235, 236, 260, 262, 263, 296, 297, 334, 342, 346–349, 351–355, 359, 364–366
Kimberlite Conference 202, 209, 254, 261, 274, 345, 347–349, 353, 357, 360, 366, 367
komatiite 227, 231, 364, 365
K-richterite 305, 349, 355, 363

L

Labrador 196, 203–206, 238, 349, 356, 358, 364
lamproite 230, 234, 236, 237, 241, 247, 257, 263, 306, 312, 343, 349, 359, 363, 364, 366
lamprophyre 202, 227, 229, 237–39, 338, 343, 369, 372
Laser-ICP-MS 204, 214, 218, 223
lherzolite 246, 247, 253, 265, 284, 296, 300, 304, 310, 323, 346, 350, 352, 354, 357, 363, 365–367
liquidus III, 194, 272, 274, 294–297
lithosphere 222, 236, 241, 249, 251, 255, 258, 259, 261, 266, 269, 283, 293, 305, 306, 312, 320, 327, 328, 331, 333–335, 338, 340, 341, 346, 347, 349, 351–356, 358, 359, 361, 362, 365, 366
low degree melt 280

M

Macquarie University VI, VII, 221
magnesite 267–269, 286, 324, 325, 338, 347
Mainz II, VIII, 206, 209–211, 216, 218–220, 222–225, 275, 280
Makkovik 204, 205
mantle heterogeneity V, VI, 194, 199, 252, 258, 339, 359
mantle metasomatism IV–VI, 194, 198, 199, 252–256, 258, 261, 264, 265, 290, 315, 329, 335, 339, 342, 343, 348, 351, 357, 360, 363–366
MARID II, 254, 255, 262–264, 266, 267, 270, 304–306, 309, 312, 330, 332, 344, 346, 347, 348, 350, 351, 353, 355, 364–366
Max Planck Institute I, 209–211, 215, 216, 218–220
megacryst 260, 353, 357
meimechite 227
melilitite 231, 242–244, 307, 317, 319, 324, 330, 331, 338, 355
metasomatic agent 247, 257, 265, 267, 306, 319, 321, 327, 329, 330–332, 338, 347, 351, 360, 361
Mg# 296, 323
Mg-value 232, 233
mica 195, 227, 236, 237, 241, 244, 250, 253, 254, 262, 264–266, 270, 296, 297, 299, 302, 306, 307, 309, 312, 320, 329, 338, 347–349, 351, 353
mica harzburgite 320
migmatite IV, V, 195, 226, 232, 236, 239, 240, 257, 260, 261, 271, 279, 298, 302, 306, 307, 309, 314, 320, 327, 329, 331, 332, 334, 337–339, 341
minette 227, 348
monchiquite 227, 362

N

neosome 195, 228, 230, 240, 241, 261, 264, 271, 302, 304, 306, 307, 309, 316, 325, 327, 331, 332, 337, 339
nephelinite 229–231, 242, 245–257, 286, 290, 317, 319, 328, 331, 333, 344, 365
Newfoundland I, VII, 197, 202–204, 206, 207, 214, 226, 238

O

olivine V, 193, 196, 207, 227, 230–234, 245, 248, 250, 252, 254, 260, 261, 265–269, 274–278, 282–286, 291, 293, 295–297, 300, 308, 311, 312, 320, 321, 326, 330, 332, 334, 339, 342–344, 349–351, 356, 358, 362, 363, 365
orangeite 227
orthopyroxene 196, 260, 264, 265, 267, 268, 270, 274, 286, 291, 297, 300, 306, 307, 311, 324, 325, 328, 331, 338
ouachitite 227
oxidation state 195, 259, 325, 349
oxygen fugacity / fO_2 232, 247, 259, 292, 293, 332–334, 345

P

partial melt 253, 273, 287, 290, 298, 348
partition coefficient 232, 332
peridotite IV–VI, 193–199, 202, 207, 226, 228, 230–234, 236, 239–242, 246, 247, 249–256, 258, 259, 261–263, 265, 267–274, 276–280, 282, 283, 284–290, 292–300, 302, 304–312, 314–317, 320–341, 343–355, 357–367
phlogopite 195, 227, 233, 241–246, 248, 252–256, 260–268, 270, 281, 284, 286, 291, 296, 297, 299–302, 304–308,

310–313, 315, 317, 319–325, 329, 332, 333, 338, 339, 346–348, 350, 352, 356–361, 363–367
picrite 231, 244, 343, 347
PKP II, 254, 255, 26–264, 266, 270
polymerise 291
primary melt 231, 239, 247, 273, 274, 294, 296
pyrolite 251, 252, 283, 284, 348, 360
pyroxenite 199, 258, 265, 266, 293, 294, 296, 299, 302, 309, 310, 313, 315, 316, 329, 339, 344, 353, 356, 357, 364

Q

quench crystals 283, 285

R

rare earth elements / REE 204, 254, 329, 332
reaction experiments 269, 272, 275, 298–300, 308, 316, 322, 324, 326, 361
reaction zone 270, 298, 323, 324, 326, 328
recycled crust V
redox 279, 332, 333, 335, 338, 340, 349, 351, 362, 364, 366
reversal experiments 283
rutile 194, 254–258, 260, 264, 266, 313–315, 319, 324, 332, 347

S

silica saturated 281, 294, 334
silica undersaturated 280, 286, 294
solidus 208, 247, 264, 278, 279, 284–290, 292, 295, 297, 300, 302, 303, 305, 306, 308, 309, 316, 320, 327, 330, 331, 339, 346–348, 350, 364, 365
solubility 286, 292, 329, 347, 358, 364
Southampton I, 197, 200–204, 207
spinel 193, 196, 232, 258, 268, 278, 297, 300, 302, 310, 323, 343, 344, 357, 362, 366, 367
stability 208, 232, 253, 263, 264, 276, 280, 284, 285, 287, 289, 291, 296, 299, 302, 304, 305, 320, 328, 331, 345, 346, 349, 352, 353, 355, 358, 363, 365
stable isotopes 327, 328, 345
subduction 212, 230, 246, 263, 269, 270, 279, 291, 300, 301, 305, 322, 327, 331, 333, 349, 353, 355, 357, 358, 361, 363, 365, 367, 368
sulfide 255, 258, 268, 269, 332, 348, 355, 357
Sunda Arc 244, 246

T

Tasmania I, VII, 204, 206, 207, 208, 209, 210, 211
thermobarometry 249
Ti magnetite 266
Ti oxides 194, 195, 306, 311, 313, 315, 320, 322, 332, 340
titanite 254, 264, 265, 266, 312, 332

U

Uganda 229, 235, 265, 266, 302, 312, 313, 315, 353, 357, 360
ultramafic lamprophyre 227
ultrapotassic 227

V

vein IV
vein melting 316, 320, 321
volatiles IV, 199, 226, 240, 248, 256, 271, 272, 276, 279, 283, 291, 294, 297, 337, 340, 359, 364

W

water 196, 213, 240, 252, 279, 284, 285, 287, 296, 308, 309, 344, 347, 351, 352, 358, 359
websterite 300
wehrlite 360

X

xenolith 255, 261, 264, 267, 269, 306, 308, 327, 343, 344, 348, 354, 355, 357, 361, 362, 366, 367



STEPHEN F. FOLEY is a petrologist and geochemist who uses high pressure experiments and trace elements to study the conditions of origin of melts in the upper mantle, integrating the petrology and composition of mantle rocks with the characteristics of the melts emanating from the mantle.

Born in Yorkshire, England, he received a B.Sc. in geology from Southampton University, England, followed by an M.Sc. at Memorial University of Newfoundland, Canada, and a Ph.D. from the University of Tasmania, Australia. Following four years of post-doctoral research at the Max Planck Institute for Chemistry in Mainz, Germany, he researched and taught at the Universities of Göttingen, Greifswald and Mainz before returning to Australia in 2013 to work as Research Coordinator at the Australian Research Council's Centre of Excellence for Core to Crust Fluid Systems based at Macquarie University in Sydney. As professor at the University of Mainz, Stephen led an initiative to set up a Centre of Excellence in Earth System Science entitled Geocycles, which he chaired for 8 years from 2005 to 2013. This Centre integrated the research of 10 departments of the University of Mainz with the Max Planck Institute for Chemistry, Mainz's Roman-Germanic Museum and the Mainz University of Applied Sciences.

Stephen is a Geochemistry Fellow of the European Association of Geochemistry and the Geochemical Society and was awarded Australian Laureate and Distinguished Professorships during his time at Macquarie as well as the Ringwood Medal of the Geological Society of Australia in 2018. During his 26 years in Germany he received the Gerhard Hess and Heinz-Maier-Leibnitz awards from the Deutsche Forschungsgemeinschaft and the Goldschmidt Prize of the German Mineralogical Society. He served as Chief Editor of the journal *Lithos* for 11 years and has been appointed for five years each to national grant selection panels in Germany and Australia. He is now Professor Emeritus at Macquarie University and an Honorary Professor at the Australian National University. Away from science, his main passion lies in track and field athletics, where he has acted as an official during events, as meet manager of a national championships, and as a compiler of state and national track statistics.