

An arc-related Phanerozoic anorthosite?

The petrogenesis of the Indian Rock Anorthosite in the Sierra Nevada Batholith

Katlego Matlou

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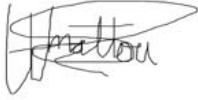
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Declaration

I declare that this dissertation is my own, unaided work. It is being submitted for the Degree of Master of Science at the University of the Witwatersrand, Johannesburg. It has not been submitted before for any degree or examination at any other University.



KW Matlou

23rd day of July 2021 at Johannesburg

Abstract

Massif-type anorthosites are batholithic bodies consisting primarily of plagioclase and minor mafic phases. Most importantly, they are known to be temporally restricted to the Proterozoic Eon (2.643 and 0.502 Ga). Several different tectonic settings for the massif-type anorthosite have been suggested (i.e. continental rift, plumes, catastrophic magmatic events), although recent studies suggest a more likely tectonic setting in a continental arc environment. However, no anorthositic bodies with both massif-type characteristics and cognate anorthosite inclusion have been found in Phanerozoic arc environments, such as the Andes, casting doubt on the continental arc hypothesis. In this project, we focus on a small body of anorthosite known as the Indian Rock Anorthosite (IRA), situated on the eastern flank of the Mesozoic Sierra Nevada Batholith. Although this body has been mapped by the USGS, almost no scientific data exist on the petrogenesis of the IRA. Investigating why this body exists in the subduction-related Mesozoic Sierra Nevada Batholith is crucial to not only understand the temporal restriction and tectonic setting of massif-type anorthosites but to also question if and how the subduction environment may have changed over the Proterozoic-Phanerozoic transition. The IRA, an approximately $\sim 1 \text{ km}^2$ body, is found near the headwaters of Thibaut Creek on the eastern scarp of the Sierra Nevada Mountain range and is surrounded by mafic granodiorites of the Black Canyon Pluton and the McDoogle Pluton.

The IRA is light grey and has an average grain size of about 4 mm. Over 90 % of the rock consists of euhedral plagioclase with an average composition of An_{84} and the remaining 10 % is made up of clinopyroxene (locally altered to hornblende and chlorite), with quartz and K-feldspar occurring as interstitial phases. Two samples from the IRA (sample IRA 4-2 and IRA 18-6) yield zircon U-Pb ages of 94.01 ± 0.05 and 95.29 ± 0.06 Ma, respectively, which rules out the possibility of the IRA being a fragment of an older Proterozoic crust trapped in a Mesozoic-dated arc. The surrounding granodiorites of the Black Canyon Pluton and the McDoogle Pluton were dated at 96 ± 2 Ma and 94 ± 4 Ma, respectively. The IRA shows similarities in geochemistry with its immediate surrounding rocks in terms of $\epsilon_{\text{Nd}}(95\text{Ma})$ and initial $^{87}\text{Sr}/^{86}\text{Sr}(95\text{Ma})$, indicating that they are coeval and comagmatic.

In its field structure and plutonic nature, the IRA resembles a typical, albeit small, massif-type anorthosite. However, the high An content in the IRA (analysed: An_{72} - An_{89} , calculated: An_{79} - An_{82}) is more typical of Archaean anorthosites (An_{61} - An_{94}) and cognate anorthosite inclusions in arcs (An_{63} - An_{100}) (Kristmannsdóttir, 1971; Ashwal, 2010; Ashwal and Bybee, 2017). The adcumulus texture, mineral content and whole-rock geochemistry resembles are cognate anorthosite inclusions in continental arcs. The major differences between the IRA and Proterozoic massif-type anorthosites in general, besides the age, is its small size, insignificant positive Eu anomalies and the high An content in plagioclase. The magnitude of Eu anomalies and the changes in An content are a function of pressure,

temperature, water content and oxygen fugacity during the formation of magmatic rocks. AlphaMELTS-based thermodynamic models show that conditions of formation of the IRA include a combination of hydrous magmas ($\text{H}_2\text{O} = 0.5 \text{ wt\%}$) at relatively high temperatures and pressures ($\sim 1213^\circ\text{C}$ and $<10 \text{ kbars}$) and under oxidising conditions. These conditions are ideal for the production of trivial amount of plagioclase production with less pronounced Eu anomalies due to plagioclase fractionation and increased An content. AlphaMELTS modelling also established that the IRA formed from a hydrous high-Al magma ($\text{Al}_2\text{O}_3 >17 \%$) (tholeiitic basalt) that was contaminated by granitic crustal material. These petrogenetic characteristics are typical of young Phanerozoic subduction zones, where the colder subducting slab is capable of transporting water deeper in the crust under high pressure and high temperatures, typified by Alpine or blueschist/eclogite metamorphic facies.

The IRA is unique in a sense that it occurs in a Phanerozoic continental arc setting and it relates to both cognate anorthosite in continental arcs in term of its adcumulus texture, mineral content and also shows similar plutonic nature of a typical massif-type anorthosite. The discovery of the IRA and the description of its petrogenesis in this work have shown that bodies of anorthosite, reminiscent of massif-type, can form in Phanerozoic arc environments. This is significant for the petrogenesis of anorthosites in general and our understanding of the $\sim 2.7 - 0.5 \text{ Ga}$ temporal restriction of massif-type anorthosites, which have been hypothesized to form in continental arc environments. The paucity of massif-type anorthosites in Phanerozoic arcs need no longer disprove the continental arc hypothesis, as we have shown that such types of anorthosite, albeit as smaller bodies with some important chemical differences, can form in these environments.

In addition, we can also use the presence of the IRA to understand changes that may have occurred in continental arc environments from the Proterozoic to Phanerozoic Eons. Various published temporal datasets along with numerical modelling have suggested that Proterozoic subduction zones were dry, producing magma at low pressure and high temperatures with relatively low oxygen fugacity, which allowed for the vast accumulation of plagioclase with large Eu anomalies and intermediate An contents. In contrast, young Phanerozoic subduction zones are hydrated, occurring at high pressure and high temperature, under oxidising conditions. These conditions will reduce the production of plagioclase in the differentiating magmas derived from the mantle wedge and increase the An content of this plagioclase. This would explain why so few discrete bodies of anorthosite are found in Phanerozoic arcs and where they are found as discrete bodies (e.g. the Indian Rock Anorthosite) or as cognate inclusions in arc volcanics, the An content ranges from 70 – 100 %. These results suggest that fundamental changes in continental arc dynamics between the Proterozoic and Phanerozoic may be responsible for the temporal restriction of Proterozoic anorthosites.

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List of Abbreviations

3D- 3 Dimension

An- Anorthite

BCP- Black Canyon Pluton

DM- Depleted Mantle

EM I- Enriched Mantle I

EPMA- Electron Probe Micro-analyser

HAOT- High-alumina olivine tholeiite

HF- Hydrofluoric acid

HFE- High Flux Event

HNO₃ - Nitric acid

HREE- Heavy Rare-earth elements

LA-ICPMS- Laser Inductively Coupled Plasma Mass Spectrometry

LREE- Light Rare-earth elements

IRA- Indian Rock Anorthosite

ICPMS- Inductively Coupled Plasma Mass Spectrometry

IDA- Isotope Dilution Analysis

MME- Mafic Magmatic Enclaves

MSWD- mean square of weighted deviates

NRF- National Research Foundation

NRF-NEP- National Research Foundation- National Equipment Programme

SNB- Sierra Nevada Batholith

REE- Rare-earth elements

RI- Radioisotope

USA- United State of America

WIGL- Wits Isotope Geosciences Lab

Table of Contents

Declaration	i
Abstract	ii
Acknowledgements	iv
List of Abbreviations	v
Table of Contents	vii
List of Figures	ix
List of Tables	xviii
1. Introduction	1
1.1 The Sierra Nevada Batholith and the Indian Rock Anorthosite (IRA)	4
1.2 Purpose of this Study	7
2. The Anorthosite Problem	8
2.1 Characteristics of Proterozoic Massif-type Anorthosites	9
2.1.1 The Parental Magma of Proterozoic Massif-type Anorthosite	9
2.1.2 Tectonic Setting of Proterozoic Massif-type Anorthosites	13
2.2 Characteristics of Anorthosite Inclusions in Arcs	15
3. Geological background and the Indian Rock Anorthosite (IRA)	19
3.1 The construction and parental magma source of the SNB	22
3.2 General Petrology of Sierra Nevada Batholith	23
3.2.1 Geochemistry of the Southeastern Sierra Nevada Batholith	24
3.2.2 Geochronology of the Southeastern Sierra Nevada Batholith	26
4. Sampling and Analytical methods	34
4.1 Field sampling and petrography	34
4.2 Whole-rock major and trace elements analysis	36
4.2.1 XRF for whole-rock major elements	37
4.2.2 ICP-MS for whole-rock trace elements	37
4.3 In-situ Mineral Chemistry	38
4.3.1 EPMA (major mineral assemblages)	38
4.3.2 LA-ICPMS U-Pb Zircon and Apatite Geochronology	39
4.3.3 LA-MC-ICPMS Lu-Hf Isotopic Analysis in Zircon	45
4.4 Whole-rock Rb-Sr and Sm-Nd Isotope Geochemistry	45
4.5 alphaMELTS modelling software	48
5. Field Observations and Petrology	49
5.1 Petrology of the Indian Rock Anorthosite	52
5.2 Granodiorites of the McDoole and Black Canyon Plutons	57
5.3 Alaskite Dykes	60
6. Geochemistry and alphaMELTS Results	64
6.1 Whole-rock Chemical Compositions	64
6.1.1 Major Elements Systematics	64

6.1.2	Trace Element Systematics.....	69
6.2	Mineral Chemistry (plagioclase).....	73
6.3	Radiogenic Isotopes.....	75
6.3.1	Zircon and Apatite U-Pb Geochronology.....	75
6.3.2	Rb-Sr and Sm-Nd whole-rock isotope geochemistry.....	79
6.3.3	<i>In-situ</i> Zircon Lu-Hf isotope ratios.....	83
6.4	alphaMELTS modelling	87
6.4.1	Isobaric baseline experiment (4 kbar)	89
6.4.2	The Effects of Change in Pressure	92
6.4.3	Addition of crustal contaminant and H ₂ O	95
7.	Discussion	112
7.1	Age relationship between the IRA and SNB.....	112
7.2	The Petrogenetic link between IRA and SNB.....	116
7.3	Classifying the IRA	122
7.4	The Formation of the IRA	127
7.5	Change in Subduction Zone Properties from Proterozoic to Phanerozoic Eon.....	138
8.	Conclusions, Recommendations, and Future work.....	143
	Recommendation and future work.....	144
	References	145
	Appendices	i

List of Figures

Figure 1.1: A ternary plagioclase-olivine-(ortho-) clinopyroxene showing different types of anorthositic lithologies. This classification of the anorthositic rocks follows the IUGS classification (Streckeisen, 1967; Ashwal, 1993).	1
Figure 1.2: The temporal restriction of different types of anorthosites through Earth history using An content of the plagioclase against age. After Ashwal and Bybee (2017).....	2
Figure 1.3: The global distribution of Archean megacrystic anorthosites (green) and Proterozoic massif-type anorthosites (red) with reference to major cratons (pink) (Ashwal and Bybee, 2017). The Proterozoic massif-type anorthosites occur mostly at the margins of major cratons.....	3
Figure 1.4: The Mesozoic Sierra Nevada Batholith located in California (USA) showing the general geology, pluton age, as well as coeval mafic/ultramafic units and wall rock pendants (after Irwin, 2003; Snelling and Gates, 2009). The 0.706 Sr isotope isopleth is also shown as described in Cecil et al. (2012). The study area is shown in the red square and is located in the Cretaceous Black Canyon Pluton. The lower left inset shows map location of California with major cities, the Sierra Nevada Batholith in green, and Mesozoic metamorphic rocks in pink.....	5
Figure 2.1: Schematic diagram showing a model for the generation of massif-type anorthosites. 1. Hot mantle-derived magma underplates the crust. 2. The crystallisation of the mafic minerals, which sink to the bottom, and partial melting of the crust above the magma. 3. The melt becomes saturated, and plagioclase minerals then rise to the top of the chamber whereas mafic minerals sink. 4. Plagioclase accumulation becomes buoyant and floats to the top. 5. Plagioclase crystal mush coalesces to form massif anorthosite, whereas granitoids crustal melts rise to the shallow levels. Modified from Ashwal (1993).	11
Figure 2.2: Global distribution of major Proterozoic massif-type anorthosites (Hill, 2017). HC, Horse Creek; HL: Harp Lake; KIC: Kunene; Ko: Korosten; La: Laramie; Lbv: Labrieville; Mi: Michikamau; MM: Mealy Mountains; N: Nain; Rog: Rogaland; Su: Suwalki. Associated mafic intrusions are also shown. Ki: Kiglapait; VB: Voisey's Bay.	15
Figure 2.3: Anorthositic inclusion in andesite, northern Mariana Island Arc, showing phase layering (Stern, 1979).....	17
Figure 2.4: The anorthosite from the Konkanhundi gabbro-anorthosite suite from Peninsular India (Santosh and Li, 2018).	18
Figure 3.1: Map of the southern Sierra Nevada region which shows domains interpreted to reflect rock uplift related to delamination of the root to the southern Sierra Nevada Batholith (Saleeby et al., 2012). The blue box shows the approximate location of the Indian Rock Anorthosite.	20
Figure 3.2: Geological map of the location of the noritic-anorthosite in Sierra Nevada Batholith (Loomis, 1963).	21

Figure 3.3: The change in ϵ_{Nd} , K_2O wt%, SiO_2 wt% and Fe oxidation across the SNB from the west to the east with the change in distance from the range crest (Chapman and Ducea, 2019). a. Generally the ϵ_{Nd} decreases from the west to the east. b. the K_2O increases from the west to east while the SiO_2 shows little change across the batholith. c. Fe oxidation increases across the batholith from west to east. RDC - mafic-ultra- mafic ring dike complex, HFE - High Flux Event.....	26
Figure 3.4: Map of the Sierra Nevada Batholith showing general age distribution across the batholith and the 0.706 isopleths (map after Snelling and Gates, (2009), isopleth and range crest from Cecil et al. (2012)), age data extracted from Kistler et al. (1986); Saleeby et al. (1987, 2007); Clemens-Knott (1992); Coleman et al. (1992); Olson (1992); Bradford (1995); Cousens (1996); Sisson et al. (1996); Truschel (1996); Ratajeski et al. (2001); Kylander-Clark et al. (2005); Glazner et al. (2008); Gray et al. (2008); Cecil et al. (2012); Nelson et al. (2013); and Chapman and Ducea (2019). The red dashed line showing the 0.706 isopleth and the blue dashed line showing the crest.	28
Figure 3.5: The change of ages of the Sierra Nevada Batholith with distance away from the range crest. The SNB shows a decrease in age (Ma) from the west to the east (Chapman and Ducea, 2019).	29
Figure 3.6: Geological map of the southeastern Sierra Nevada Batholith with ages of the plutons (Mahan et al., 2003).....	30
Figure 3.7: Timeline showing the ages of the Southern Sierra Nevada Batholith. The published U-Pb ages are from (Mahan et al. (2003): McDoogle Pluton, Sawmill Lake Pluton), (Carl (2000): Black Canyon Plutonic rock), (Chen and Moore (1982); Mahaffey (2016): Tinehama granodiorite, Wood Lake), (Coleman et al. (2000): Lamarck granodiorite).	32
Figure 3.8: Location of the study site and the Indian Rock Anorthosite. Black Canyon Plutonic rocks labelled Km (purple) surround the IRA (Kan-yellow). Km is made up of mafic granodiorite, quartz diorite, diorite, gabbro, and peridotite (modified from Moore, 1963). The published U-Pb ages are from Mahan et al. (2003): McDoogle Pluton, Sawmill Lake Pluton), (Carl (2000): Black Canyon Plutonic rock), (Chen and Moore (1982); Mahaffey (2016): Tinehama granodiorite, Wood Lake), (Coleman et al. (2000): Lamarck granodiorite).	33
Figure 4.1: Google Earth image showing sample locations (green points) in Thibaut Creek (blue dashed line) and the location of the Indian Rock Anorthosite (IRA) (Google Earth, earth.google.com).....	35
Figure 4.2: Comparison of the BCR-2 and BHVO-2 standards from this study and the reference values from the USGS (2000). Most of the data agree within 5% (shaded band) except for P_2O_5 and NiO ...	37
Figure 4.3: Comparison of the BCR-2 and BHVO-2 standards from this study and the reference values from the USGS (2000). Most of the data agree within 5% (shaded band), and the rest fall within 20%.	38
Figure 4.4: Wetherill Concordia plots and weighed mean Pb/Pb ages produced for standards are a. b. GJ1 ($^{207}Pb/^{206}Pb= 601.53 \pm 4.44$ Ma); c. d. 91500 ($^{207}Pb/^{206}Pb= 1067.47 \pm 8.27$ Ma); e. f. Temora ($^{207}Pb/^{206}Pb= 421.72 \pm 9.07$ Ma); g. h. Plešovice zircon ($^{207}Pb/^{206}Pb= 332.17 \pm 6.97$ Ma) and i. j. BB ($^{207}Pb/^{206}Pb= 535.27 \pm 8.57$ Ma).	43

Figure 4.5: Tera–Wasserburg Concordia plots and weighed mean Pb/Pb ages of the apatite standards. The blue bar was excluded from the calculation. a. b. Madagascar apatite (473.25 ± 0.82 Ma); c. d. Durango apatite (32.38 ± 0.11 Ma); and e. f. Mt McClure apatite (523.73 ± 1.12 Ma).	44
Figure 5.1: A geological map showing the plutons surrounding the Indian Rock Anorthosite (modified from Moore, 1963).	50
Figure 5.2: Field images showing an overview of the granodioritic plutons in the study area. a. Peaks of the McGann and the Black Canyon Plutons observed from the Sawmill Lake campsite b. The Black Canyon Pluton and the summit of Mount Baxter. c. An overview of the granodiorite plutons surrounding Sawmill Lake. d. Contact between the McGann Pluton and the Black Canyon Pluton granodiorites, with felsic alaskite dykes (10-60 cm thick) crosscutting the Black Canyon Pluton. e. The McGann Pluton with country-rock screens (brown coloured scree).	51
Figure 5.3: a. The anorthosite outcrop extending from the headwaters of the Thibaut Creek to the summit of Indian Rock. b. c. Enlargement of the anorthosite outcrop on the slopes of Indian rock, surrounded by the mafic granodiorite of the Black Canyon Pluton (Km unit).	53
Figure 5.4: Anorthosite boulders found in the outwash fan of Thibaut Creek, which is the only stream that drains the IRA outcrop. a. A mafic vein made up of clinopyroxene crosscutting the boulder. b. Undeformed, crystalline texture of the anorthosite made up of plagioclase and clinopyroxene grains.	53
Figure 5.5: A quartz-alkali-feldspar-plagioclase ternary showing modal mineral abundance and classification of the Indian Rock Anorthosite, McDoogle Pluton, Black Canyon Pluton and felsic dykes.	54
Figure 5.6: Photomicrographs of the Indian Rock Anorthosite showing medium-grained (0.2-0.4 cm) plagioclase grains with clinopyroxene, chlorite, hornblende, apatite and pyrite. a. Zoned cumulate plagioclase with interstitial mafic minerals (clinopyroxene and hornblende). b. Poikilitic texture with plagioclase inclusions and clinopyroxene oikocrysts. c. Altered grains of clinopyroxene to hornblende and chlorite. d. 4 mm grains of plagioclase with altered grains of clinopyroxene.....	55
Figure 5.7: PPL and XPL photomicrographs of the IRA showing different textures between the minerals and the relationships amongst them. a. b. show zoned plagioclase grains with clinopyroxene alter to chlorite at the edges. c. d. shows a plagioclase grain enclosed within an orthoclase grain. e. f. g. A plagioclase grain having fractures filled with clinopyroxene.	56
Figure 5.8: Field images of granodioritic McDoogle and Black Canyon Plutons. a. The granodiorite of the McDoogle Pluton showing foliation trending NW-SE. b. The Black Canyon Pluton, with crosscutting felsic dykes. c. Elongated mafic enclave (oriented NW-SE) in the granodiorite of McDoogle Pluton. d. Overview of the Black Canyon Pluton, which hosts the IRA.	58
Figure 5.9: a. Photomicrograph of the granodiorite of McDoogle pluton showing an interlocking texture having elongated cumulate plagioclase grains, intercumulus hornblende and clinopyroxene. b. Photomicrograph of the mafic granodiorite of the Black Canyon Pluton. c. Contact between a mafic	

enclave and granodiorite of the McDoogie pluton shown by the change in grain size. d. The mafic enclave consists of fine grains (~0.1 cm) of plagioclase, biotite, quartz, and a vein of clinopyroxene.	59
Figure 5.10: PPL and XPL thin-section images of the granodiorites (McDoogie and Black Canyon Pluton) showing different textures between the minerals and the relationships amongst them. a. b. The granodiorite of the McDoogie Pluton shows biotite grain with magnetite inclusion and euhedral plagioclase grains. c. d. The granodiorite of the Black Canyon Pluton shows biotite altering to hornblende and chlorite.	60
Figure 5.11: Field photographs of the crosscutting felsic dykes (alaskite). a. Felsic dykes crosscutting the granodiorite from the Black Canyon Pluton, which hosts the anorthosite. b. The sharp contact between the dykes and the granodiorite. c. The felsic dyke crosscutting the McDoogie granodiorite in a sub-vertical manor. d. The Black Canyon granodiorite having multiple felsic sills of alaskite.	61
Figure 5.12: Photomicrographs of alaskite dykes shown in xpl and ppl. a. The thin section shows a granitic texture with interlocking grains of quartz, plagioclase and microcline that is typical for granitic rocks. b. The thin section shows the sharp contact between the granodiorite and the dyke, which is characterised by a change in grain size and mineralogy.	62
Figure 5.13: Thin-section images of the granitic alaskite dykes shown in xpl and ppl. a. b. The rocks have a granophyric texture. c. d. Interlocking grains of quartz, plagioclase and microcline.	63
Figure 6.1: SiO ₂ vs CaO, TiO ₂ , Na ₂ O, Fe ₂ O ₃ , Al ₂ O ₃ , and MgO wt%. The granodiorite shows varying SiO ₂ wt% content compared to other rocks, anorthosites have high Al ₂ O ₃ and CaO wt% reflecting high modal plagioclase content, and felsic dykes (alaskite) have high SiO ₂ compared to other elements. a. There is a decrease in CaO wt% with increasing SiO ₂ wt% content. b. The granodiorites have high TiO ₂ wt% compared to other samples. c. The anorthosites show low Na ₂ O wt% relative to the granodiorites and the alaskite dykes. d. Granodiorite and alaskite dykes show a negative slope trend with the anorthosites being the outliers. e. A negative trend between SiO ₂ wt% and Al ₂ O ₃ . f. Granodiorite and granitic-alaskite dyke show a negative slope trend between SiO ₂ vs MgO wt% with the anorthosites being the outliers.	66
Figure 6.2: MgO vs TiO ₂ , Ni (ppm), Cr (ppm), CaO, Fe ₂ O ₃ , and Al ₂ O ₃ wt%. The anorthosites show high CaO and Al ₂ O ₃ , while the granodiorites have high MgO wt%. a. b. e. Positive trend between MgO vs TiO ₂ wt%, Ni (ppm), and Fe ₂ O ₃ wt%. d. f. Show similar patterns with anorthosites having high CaO wt% and Al ₂ O ₃ wt% against MgO wt%. c. Cr (ppm) vs MgO wt% do not show any observable trend.	67
Figure 6.3: Chondrites-normalised REE diagram using chondrite values from Anders and Grevesse, (1989). All the analysed samples show a negative slope indicating enrichment in LREE. The granodiorites of the McDoogie and Black Canyon Pluton are more enriched in REEs compared to the anorthosites.	72

Figure 6.4: Spider diagram showing primitive mantle normalised plots. The primitive mantle values were taken from McDonough and Sun (1995).....	73
Figure 6.5: An content of plagioclase grains from the core to the rim and plagioclase inclusions in clinopyroxene oikocrysts. Solid horizontal lines show mean compositions for the different analysis positions. Core compositions (sky blue) are slightly more An rich than the rims (dark blue). The inclusions (pink) on average have lower An content compared to the cumulates.	74
Figure 6.6: Classification of plagioclase minerals from in-situ data analysis of the Indian Rock Anorthosite.	75
Figure 6.7: Cathodoluminescence (CL) and back-scattered electron (BSE) images of zircon and apatite from the Indian Rock Anorthosite, with open circles showing representative laser ablation sites before ablation. a. Zircons from sample IRA 4-2 and IRA 18-6 show regular oscillatory magmatic zoning. Zircon 6ZR7 shows a core rimmed by continuous oscillatory zones. The lighter zones indicate high U content. b. Apatite grains from IRA 4-2 and IRA 18-6. Apatite 4.2AP2, 4.2AP have different domains highlighted by various shades of grey within the grains.	77
Figure 6.8: a. Zircon U-Pb Wetherill Concordia plots of the IRA showing the ^{238}U - ^{206}Pb ages and b. the weighed-mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages. A Stacey-Kramers common Pb correction (Chew et al., 2014) was used when plotting the Wetherill Concordia plots and weighted mean ages using IsoplotR.	78
Figure 6.9: Apatite U-Pb Tera-Wasserburg plots of the IRA showing the ^{238}U - ^{206}Pb ages and the weighed-mean $^{207}\text{Pb}/^{206}\text{Pb}$. Stacey-Kramers common Pb correction (Chew et al., 2014) was applied when plotting the Tera Wasserburg Concordia plots and weighted mean ages using IsoplotR.	79
Figure 6.10: I_{Sr} (95Ma) vs Samples, ϵ_{Nd} (95Ma) vs Samples, ϵ_{Nd} (95Ma) vs $^{87}\text{Sr}/^{86}\text{Sr}$ (95Ma), and $^{143}\text{Nd}/^{144}\text{Nd}$ vs $^{87}\text{Sr}/^{86}\text{Sr}$ (95Ma). a. McDoogle Plutons' granodiorite (IRA 6-3) shows I_{Sr} (95Ma) values that fall within the range of the anorthosites (grey band). b. McDoogle and Black Canyon Plutons' granodiorite (IRA 6-3 and IRA 18-2 respectively) shows ϵ_{Nd} (95Ma) values that fall within the range of the anorthosites (grey band). c, ϵ_{Nd} (95Ma) vs I_{Sr} (95Ma) plot. d, Initial Nd vs initial Sr isotope plot at 95 Ma. Values for lines labelled Bulk Earth (BE) and CHUR (chondritic uniform reservoir) are calculated at 95Ma, assuming BE: $^{87}\text{Sr}/^{86}\text{Sr}= 0.7045$ and $^{87}\text{Rb}/^{86}\text{Sr}= 0.083$, and CHUR: $^{143}\text{Nd}/^{144}\text{Nd}= 0.512638$ and $^{147}\text{Sm}/^{144}\text{Nd}= 0.1967$ (Workman and Hart, 2005). The plotted error bars are 2s and they are smaller than the symbols in the diagrams.	82
Figure 6.11: Whole rock ϵ_{Nd} (95Ma) vs crystallization age diagram. The evolution-depleted mantle (DM) is calculated back from present ratios of $^{87}\text{Sr}/^{86}\text{Sr} = 0.7045$ and $^{87}\text{Rb}/^{86}\text{Sr} = 0.083$ (Workman and Hart, 2005). The light blue regression line is the Indian Rock Anorthosite, the black regression line is the Black Canyon Pluton and the dark blue regression line is the McDoogle Pluton.	83
Figure 6.12: Lu-Hf isochron diagram for the Indian Rock Anorthosite. The error bars are 2s and all the lines in the diagram are determined using sample IRA 18-6 and IRA 4-2.	86
Figure 6.13: Zircon Lu-Hf model age of the IRA. The graph of ϵ_{Hf} (95Ma) vs crystallisation age (Ma). The evolution-depleted mantle (DM) is calculated back from present ratios of $^{176}\text{Lu}/^{177}\text{Hf}= 0.0524$ and	

$^{176}\text{Hf}/^{177}\text{Hf} = 0.28326$ (Workman and Hart, 2005). The light blue regression line is sample IRA 4-2, and the dark blue regression line is sample IRA 18-6.	87
Figure 6.14: A schematic diagram of the Cascade-Sierra Province showing relative locations of the two magmatic arcs (Cascade Batholith (purple) and the Sierra Nevada Batholith (orange)).	89
Figure 6.15: The graphs show mineral accumulation in a closed system and their abundances. Diagram a, b, and c, show mineral accumulation from the primitive arc melts under isobaric conditions (4 kbar) with a decrease in temperature.	91
Figure 6.16: The change in plagioclase An content with the change in temperature and addition of 0.5 wt% water. Generally, the An content decreases with a decrease in temperature. Melt 1 shows high An content compared to the other Melt 2 and 3. Melt 2 and 3 show a similar trend under similar conditions compared to Melt 1. The grey lines represent primitive arc melts crystallisation trend under an isobaric condition with a decrease in temperature, and the blue lines are the primitive arc melts with the addition of water.	92
Figure 6.17: Possible primitive melts of the southern SNB simulated under different pressure and temperature. Clinopyroxene phases form at high temperature and pressure, while the plagioclase, K-feldspar, pyrite forms at lower temperature and pressure.	94
Figure 6.18: The change in plagioclase An content with changing temperature and pressure. The An content of melt 1 shows high An content, while melts 2 and 3 have the least. The first crystals of plagioclase forming at high temperature and pressure have high An content compared to the crystals at low temperature and pressure.	95
Figure 6.19: The graphs show mineral crystallisation in a closed system and their abundance. Diagram a, b, and c show mineral accumulation from the primitive arc melts under isobaric conditions (4 kbar) with a decrease in temperature at 20 °C increments and with the addition of H ₂ O to the system.	97
Figure 6.20: Mineral crystallisation in the system with the addition of felsic contaminants in the form of a solid and water. The diagrams display different parental compositions and changing the amount of water in the system from 5 to 7 wt%, and the amount of contaminants was kept constant at a ratio of 1:2 to the parental melt.	100
Figure 6.21: The change in plagioclase An content with the decrease in temperature. Generally, the An content decreases with a decrease in temperature. Melt 1 shows high An content compared to the other Melts (2 and 3). Melt 1 with higher water content (0.7 wt%) shows a wide range of An content compared to the 0.5 wt% H ₂ O. The dark grey lines represent primitive arc melts+ crust and 0.5 wt% H ₂ O crystallisation trend under an isobaric condition with a decrease in temperature, and the blue lines are the primitive arc melts+ crust + 0.7 wt% H ₂ O.	101
Figure 6.22: Models of melt 1: tholeiitic basalt with crustal contamination and changing water content in the system. Clinopyroxene forms at high temperatures and pressures, while the plagioclase forms at lower temperatures and pressure.	104

Figure 6.23: The change in plagioclase An content with changing temperature and pressure of melt 1: tholeiitic basalt. The increasing water content shows a slight increase in the plagioclase An content.	105
Figure 6.24: Models of melt 2: basalt with crustal contamination and changing water content in the system. Clinopyroxene forms at high temperatures and pressures, while the plagioclase forms at lower temperatures and pressure exponentially.	107
Figure 6.25: The change in plagioclase An content with changing temperature and pressure of melt 2: basalt. The increasing water content shows a slight increase in the plagioclase An content.	108
Figure 6.26: Models of melt 3: Andesite with crustal contamination and changing water content in the system. Clinopyroxene forms at high temperatures and pressures while the plagioclase forms at lower temperatures and pressure exponentially.	110
Figure 6.27: The change in plagioclase An content with changing temperature and pressure of melt 3: andesite. The increasing water content shows a slight increase in the plagioclase An content.	111
Figure 7.1: Spatial distribution of samples in the Sierra Nevada Batholith. This diagram shows the age distribution of the Sierra Nevada Batholith in a 3D space with the arrow showing the direction from older to younger rocks. The map of Sierra Nevada Batholith shows the age position of samples. The age and latitude-longitude data are from Kistler et al. (1986); Saleeby et al. (1987, 2007); Clemens-Knott (1992); Coleman et al. (1992); Olson (1992); Bradford (1995); Cousens (1996); Sisson et al. (1996); Truschel (1996); Ratajeski et al. (2001); Kylander-Clark et al. (2005); Glazner et al. (2008); Gray et al. (2008); Cecil et al. (2012); Nelson et al. (2013); and Chapman and Ducea (2019).	114
Figure 7.2: Timeline showing the ages of the Southern SNB and the zircon and apatite U-Pb ages of the IRA. IRA apatite U-Pb ages are shown as white dots, while grey dots show zircon U-Pb ages. Error bars are smaller than the symbols. The published U-Pb ages are from Mahan et al. (2003); McDoogle Pluton, Sawmill Lake Pluton, Carl (2000); Black Canyon Plutonic rock, Chen and Moore (1982) and Mahaffey (2016); Tinehama granodiorite, Wood Lake, Coleman et al. (2000); Lamarck granodiorite.	116
Figure 7.3: Harker diagrams of samples from this research compared with the Sierra Nevada Batholith samples from the Navdat (http://earthchem.org/) database.	118
Figure 7.4: Chondrites normalised REE plots using chondrite values by Anders and Grevesse, (1989) of the samples of this study compared to Sierra Nevada Batholith samples extracted from the Navdat database.	120
Figure 7.5: Samples from this research compared with Mesozoic isotopic data from the Sierra Nevada Batholith (green polygon). The data used for the green-oval was extracted from DePaolo, (1981); Dodge and Bateman, (1988); Mukhopadhyay (1989); Mukhopadhyay and Manton (1994); Ducea and Saleeby (1998); and Ducea (2001). The data includes compositions ranging from granodiorites, granites, quartz monzonite, and quartz diorite. M-young mantle, C-older lower crust, and S-sedimentary.	121

Figure 7.6: Plot of Indian Rock Anorthosite, the granodiorites of Black Canyon and McDoogie pluton compared to plutons of the southern Sierra Nevada Batholith. The data of the plutons were extracted from Nelson et al. (2013) and references therein.	122
Figure 7.7: An content of cognate anorthosite inclusions and the IRA (data extracted from Ashwal (2010)). The grey areas show different types of cognate anorthosite inclusions in different settings. The red dots show anorthositic inclusions in island arcs, green dots are anorthosites in arc-root plutonic complexes and the sky-blue in the IRA.	124
Figure 7.8: Samples from this study compared to the Proterozoic massif-type anorthosites. The data was extracted from Navdat database. The IRA falls at the centre of the Proterozoic massif-type anorthosites indicating similar major element compositions.	126
Figure 7.9: Chondrite-normalised REE using CI chondrites values by Anders and Grevesse, (1989). Samples from this study compared to the Proterozoic massif-type anorthosites. The data was extracted from Navdat database.	127
Figure 7.10: A schematic phase diagram of the albite-anorthite system showing the effect of changing variables (e.g. temperature, pressure and water content) on plagioclase composition (modified from Bennet et al., (2019)).	129
Figure 7.11: The change of ϵ_{Nd} values with distance from the trench or convergence margin (Chapman et al., 2017). a, Shows the assimilation of the older crustal material to the mantle source. b, Shows increased crustal assimilation to the mantle source during crustal thickening.	131
Figure 7.12: Fractionation trend of Depleted Mantle and the Enriched Mantle I. a, the DM fractionation trends with addition of the crustal material and this shows the contamination of the mantle source with the older and younger crustal material. b, The EM I source trend shows lesser crustal contamination compared to the DM to obtain the observed isotopic signatures in our samples.	133
Figure 7.13: $\epsilon_{Nd(t)}$ vs crystallisation ages (Ma) plot. The crystallisation ages and $\epsilon_{Nd(t)}$ data are from Kistler et al. (1986); Saleeby et al. (1987, 2007); Clemens-Knott (1992); Coleman et al. (1992); Olson (1992); Bradford (1995); Cousens (1996); Sisson et al. (1996); Truschel (1996); Ratajeski et al. (2001); Kylander-Clark et al. (2005); Glazner et al. (2008); Gray et al. (2008); Cecil et al. (2012); Nelson et al. (2013); and Chapman and Ducea (2019).	135
Figure 7.14: Zircon Lu-Hf model age of the eastern Sierra Nevada Batholith and the IRA. The eastern Sierra Nevada Batholith data was extracted from Attia et al. (2020) The graph of $\epsilon_{Hf(t)}$ vs crystallisation age (Ma). The evolution-depleted mantle (DM) is calculated back from present ratios of $^{176}\text{Lu}/^{177}\text{Hf} = 0.0524$ and $^{176}\text{Hf}/^{177}\text{Hf} = 0.28326$ (Workman and Hart, 2005). The second figure is the zoom out area of the red box.	137
Figure 7.15: The graph showing maximum pressure in metamorphic terranes through the Earth's history, crustal reworking curve and .the ages of supercontinents shown by grey areas (Hawkesworth et al., 2020). The red dots are high T/P, purple dots are inter T/P and the blue dots are low T/P.	139

Figure 7.16: Variations in peak metamorphic conditions through time. a. Shows difference in temperature and pressure from the peak metamorphism estimates (Hawkesworth et al., 2020). b, Metamorphic age with thermal gradient grouped by metamorphic facies series (Zheng and Zhao, 2020). The facies series are from Zheng and Zhao (2020), peak metamorphism conditions estimates are from Brown (2007). The dashed lines show boundaries between the different facies series.....	140
Figure 7.17: Different rates of water loss in oceanic subduction zones as a function of depth. Cold subduction zones lose little water even at 230 km depth (blue-dashed line), while the hot subductions lose water at a faster rate (red-dashed line) (Zheng et al., 2016).	142

List of Tables

Table 4.1: A sample list with rock names and analysis carried out on each one.	36
Table 4.2: Summarised age data of the zircon standards from the LA-ICPMS.....	41
Table 4.3: Summarized age data of the apatite standards from the LA-ICPMS.....	43
Table 4.4: Summarised Lu-Hf isotope data for zircon standards.....	45
Table 4.5: Weight of isotope dilution samples and spikes for column chemistry.	47
Table 4.6: Parent/daughter ratios for standards used for Rb/Sr analyses and the published, accepted values.....	48
Table 4.7: Parent/daughter ratios for standards used for Sm/Nd analyses and the published, accepted values.....	48
Table 6.1: Whole-rock major element oxide data for a representative selection of 10 samples from the study area.....	65
Table 6.2: CIPW norm calculations of the samples using whole rock major element chemistry.....	69
Table 6.3: Whole-rock trace elements data in ppm, including REE data in ppm from ICPMS. Element ratios of trace elements normalised using Anders and Grevesse (1989). $[Eu/Eu^*]_N = Eu_N(Sm_N)(Gd_N)$	71
Table 6.4: Mean major element compositions of plagioclase and calculated An content data from the EPMA.....	74
Table 6.5: Summarised LA-ICPMS U-Pb and Pb-Pb zircon and apatite age data of the Indian Rock Anorthosite.	76
Table 6.6: Rb/Sr ratio results from the MC-ICPMS analyses. The I_{Sr} values are calculated at 95 Ma. 80	
Table 6.7: Sm/Nd ratio results from the MC-ICPMS. The initial Nd and ϵ_{Nd} values are calculated at 95 Ma.....	81
Table 6.8: $^{176}Lu/^{177}Hf$, $^{176}Hf/^{177}Hf$, I_{Hf} and $\epsilon_{Hf}(95\text{ Ma})$ values including $T_{CHUR}(95\text{ Ma})$ ages of sample IRA 4-2 and IRA 18-6.....	84
Table 6.9: $^{176}Lu/^{177}Hf$, $^{176}Hf/^{177}Hf$, I_{Hf} and $\epsilon_{Hf}(95\text{ Ma})$ values including $T_{CHUR}(95\text{ Ma})$ ages of sample IRA4-2 and IRA 18-6.....	85
Table 6.10: Primitive arc melts from the Cascades batholiths from the GEOROC database (Schmidts and Jagoutz, 2017).....	89

1. Introduction

Anorthosites are classified as igneous plutonic rocks, that are dominated by plagioclase (generally over 90%), with the remainder being pyroxenes, Fe-Ti-oxide phases and/or olivine, with trace amounts of biotite, and amphibolite (Streckeisen, 1967, 1976; Ashwal, 1993; Maji *et al.*, 2010). The term anorthosite was derived from the crystallographic division of feldspars into orthose (monoclinic) and anorthose (triclinic) varieties (Ashwal, 1993; Frost *et al.*, 2010). Anorthositic intrusions, when defined in the broadest sense, contain anorthosite (*sensu stricto*), leucogabbro (norite), leuconorite and leucotroctolite (see Figure 1.1). The latter three terms define anorthositic varieties where the proportions of clinopyroxene, orthopyroxene and olivine, respectively, range between 10-35% (Figure 1.1). The term anorthosite (*sensu stricto*) refers to rocks with more than 90% plagioclase (Ashwal, 1993; Scoates and Mitchell, 2000).

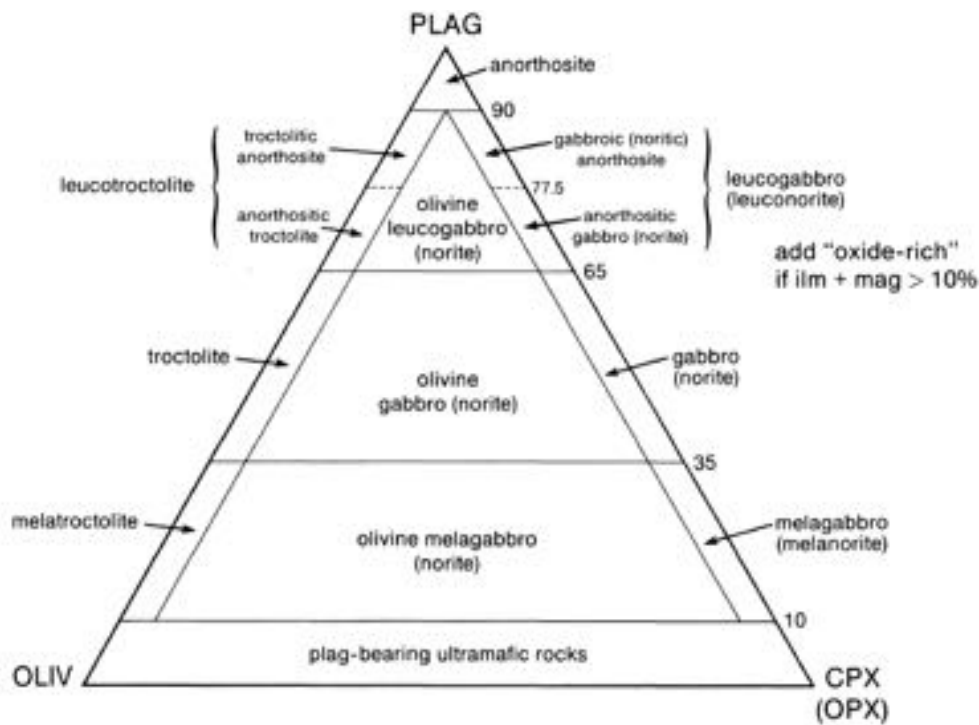


Figure 1.1: A ternary plagioclase-olivine-(ortho-) clinopyroxene showing different types of anorthositic lithologies. This classification of the anorthositic rocks follows the IUGS classification (Streckeisen, 1967; Ashwal, 1993).

Ashwal (1993) divided the occurrences of anorthosite into six categories: 1) Archean megacrystic anorthosites, 2) Proterozoic massif-type anorthosites, 3) anorthosites of layered mafic intrusions, 4) cognate anorthosite inclusions in arc related settings, 5) anorthosites of oceanic settings and 6) Primordial or lunar anorthosites. Of the six basic types of anorthosites, three varieties show puzzling

temporal restriction (Figure 1.2): Archean megacrystic anorthosite, Proterozoic massif-type anorthosite and extra-terrestrial or lunar anorthosites (also known as primordial anorthosites) (Morse, 1982; Ashwal, 2010). The temporally restricted anorthosites occur in the Hadean (on Earth's moon), the Archean (2.49-3.73 Ga) and Proterozoic Eons (~2.6- 0.5 Ga) (Ashwal, 1993, 2010; Ashwal and Bybee, 2017). Lunar samples returned from Apollo missions indicated that anorthosite is a significant constituent of the lunar crust (Ashwal and Bybee, 2017). Analyses of the samples show that the outer ~500 km of the Moon was initially molten due to high temperatures associated with accretion processes (Wood *et al.*, 1970; Warren, 1985; Nishi *et al.*, 2018). Primordial anorthosite possess the most calcic plagioclase with an An-content of An₉₃₋₉₈ (Ashwal, 2010; Arai and Maruyama, 2017). This anorthosite is known to be the purest of the three temporally restricted anorthosites, with over 95% plagioclase (Arai and Maruyama, 2017). Early geochronology of the samples returned from Apollo missions yielded ages between 4.29 and 4.57 Ga (Taylor *et al.*, 2009; Borg *et al.*, 2015; Barboni *et al.*, 2017).

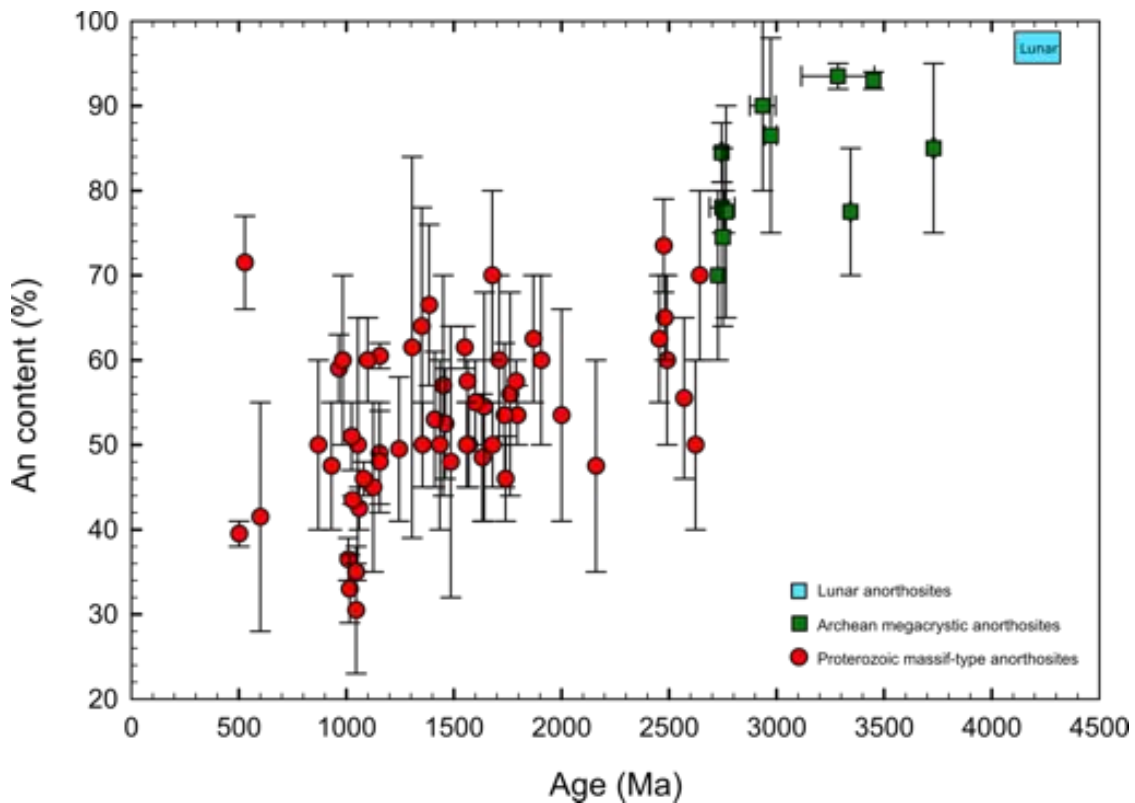


Figure 1.2: The temporal restriction of different types of anorthosites through Earth history using An content of the plagioclase against age. After Ashwal and Bybee (2017).

Figure 1.3 shows the global occurrence of the Archean and Proterozoic anorthosites. Archean megacrystic anorthosites (2.49-3.73 Ga) (Ashwal, 2010) are the oldest known terrestrial anorthosites and contain calcic plagioclase (An₆₁₋₉₄, average An₈₀) (Ashwal and Bybee, 2017). They are associated with mafic magmatic rocks such as gabbro and leucogabbro in Archean greenstone belts (Ashwal, 1993, 2010; Ashwal and Bybee, 2017; Polat *et al.*, 2018; Santosh and Shan-Shan, 2018). Besides the An-

content of the calcic plagioclase, this anorthosite variety is known for its unique texture: equidimensional to rounded feldspars megacrysts ranging from ~2 cm to >30 cm in a finer-grained basaltic/gabbroic groundmass (Ashwal and Bybee, 2017; Polat *et al.*, 2018). They occur as complexes ranging from 1-10 km², and the largest well-preserved examples are the ~560 km² Shawmere Anorthosite Complex of Ontario and the ~500 km² Fiskenæsset complex of West Greenland (Peck and Valley, 1996; Polat *et al.*, 2018). Proterozoic massif-type anorthosites are known to be the most volumetrically abundant type of anorthosite on Earth and are restricted to the Proterozoic Eon, between 2.643 and 0.502 Ga (Ashwal, 2010; Ashwal and Bybee, 2017). They occur as intrusive bodies ranging from ~1-10 km² to massive batholithic bodies up to 18 000 km². The largest known complexes are the Kunene Complex of southern Angola and northern Namibia (>18 000 km², based on interpretation of satellite imagery) (Bybee *et al.*, 2019) and the Lac-Saint-Jean Complex of Quebec (~17 000 km²) (Higgins, 1998; Ashwal, 2010). The plagioclase An-content in these bodies ranges from An₃₀-An₇₀, with an average of An₅₃ (Ashwal and Bybee, 2017).

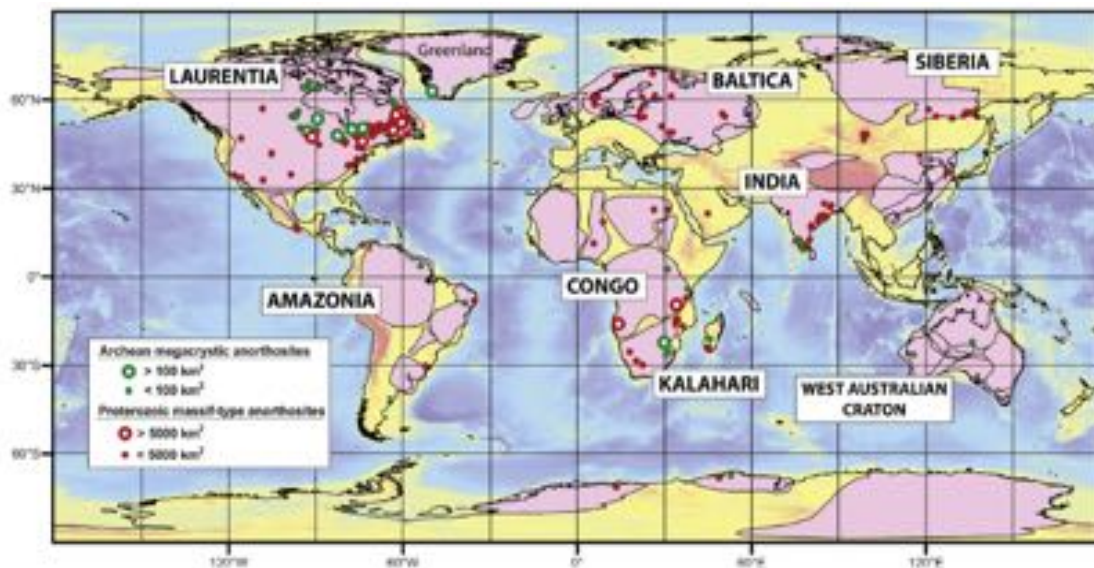


Figure 1.3: The global distribution of Archean megacrystic anorthosites (green) and Proterozoic massif-type anorthosites (red) with reference to major cratons (pink) (Ashwal and Bybee, 2017). The Proterozoic massif-type anorthosites occur mostly at the margins of major cratons.

The temporal restriction of Proterozoic anorthosites is still a puzzling and contentious issue, which has important implications for understanding magmatic processes operating in the Proterozoic Eon and is the subject of this dissertation. The temporal restriction is one of several controversial issues surrounding the petrogenesis of Proterozoic anorthosites (see Section 2.1). Other problems include the source of the magmas, the processes of plagioclase concentration, the ascent mechanisms of these magmas and the tectonic setting in which these intrusive form (see Section 2.1.1). Many different tectonic settings have been proposed for Proterozoic anorthosite formation (see Section 2.1.2; (Herz,

1969; Bridgewater and Windley, 1973; Berg, 1977; Emslie, 1978), with arc environments having recently gained traction as the likely environments of formation (Wiebe, 1993; McLelland and Selleck, 2010; Bybee *et al.*, 2014, 2019; Lehmann *et al.*, 2020). An arc setting would allow delivery of a large volume of basaltic magma to the base of the crust, where magmas can fractionate and interact with the lower crust and subsequently rise as plagioclase-rich diapirs, commonly in elongate belts (Hamilton, 1981; Ashwal, 2010). However, if this is the case, an important question arises: if massif-type anorthosites form in arc environments in the Proterozoic Eon, why do we not see any coeval and comagmatic anorthosite intrusions in Phanerozoic arc environments? Proterozoic (and specifically Grenvillian) anorthosites are found as xenolithic terranes/fragments in arcs (Martignole *et al.*, 2005; Casquet *et al.*, 2010; Cisneros de León *et al.*, 2017), but these occurrences are not useful in assessing the arc-related nature of massif-type anorthosites, as they are not comagmatic, nor coeval, with the arc terranes.

However, a small, unique anorthositic body ($\sim 1 \text{ km}^2$) known as the Indian Rock Anorthosite (IRA), occurs on the eastern flank of the Mesozoic Sierra Nevada Batholith (SNB). This body has only been mentioned briefly by Moore (1963), where he mapped the body and partially described the petrology (see Section 1.1). However, the lack of detailed petrography, geochemistry and geochronological constraints raises doubts over the age and origin of the anorthosite. It may represent a xenolithic, Proterozoic fragment caught up in Mesozoic, arc-related granitoid magmatism or it may be comagmatic with these intrusives, representing one of the only discrete bodies of Phanerozoic arc-related anorthosite. This dissertation focusses on describing and understanding the petrogenesis of the IRA and its link to the Sierra Nevada Batholith, in addition to testing the genetic link between massif-type anorthosite and arc environments and constraining geodynamic changes across the Proterozoic-Phanerozoic transition in subduction zone environments.

1.1 The Sierra Nevada Batholith and the Indian Rock Anorthosite (IRA)

The Mesozoic Sierra Nevada Batholith (SNB) occurs on the eastern escarpment of the Sierra Nevada Mountain range in California (United States of America) (Figure 1.4). It represents one of the best-studied convergent margin batholiths on Earth, and it is made up of numerous distinct plutons grouped into comagmatic suites based on similarities in terms of age and composition (Dorais *et al.*, 1990; Lackey *et al.*, 2005; Ducea *et al.*, 2015). It is known as one of the largest batholiths in North America, extending for $\sim 600 \text{ km}$ from southeastern Alaska to Baja, California, with a varying width of ~ 80 to 120 km (Tobisch *et al.*, 1995; Cecil *et al.*, 2012). It also represents a segment of the bigger, continuous circum-Pacific granitoid belt along the western margin of North America, which was emplaced in a continental magmatic arc between 220 and 80 Ma (Chen and Moore, 1982; Ducea and Saleeby, 1998;

Ducea, 2001; Mahan *et al.*, 2003; Barbarin, 2005; Gray *et al.*, 2008). The batholith covers ~90% of the southern and central Sierra Nevada mountain range, and ~70% of the northern part of this range (Figure 1.4) (Barbarin, 2005).

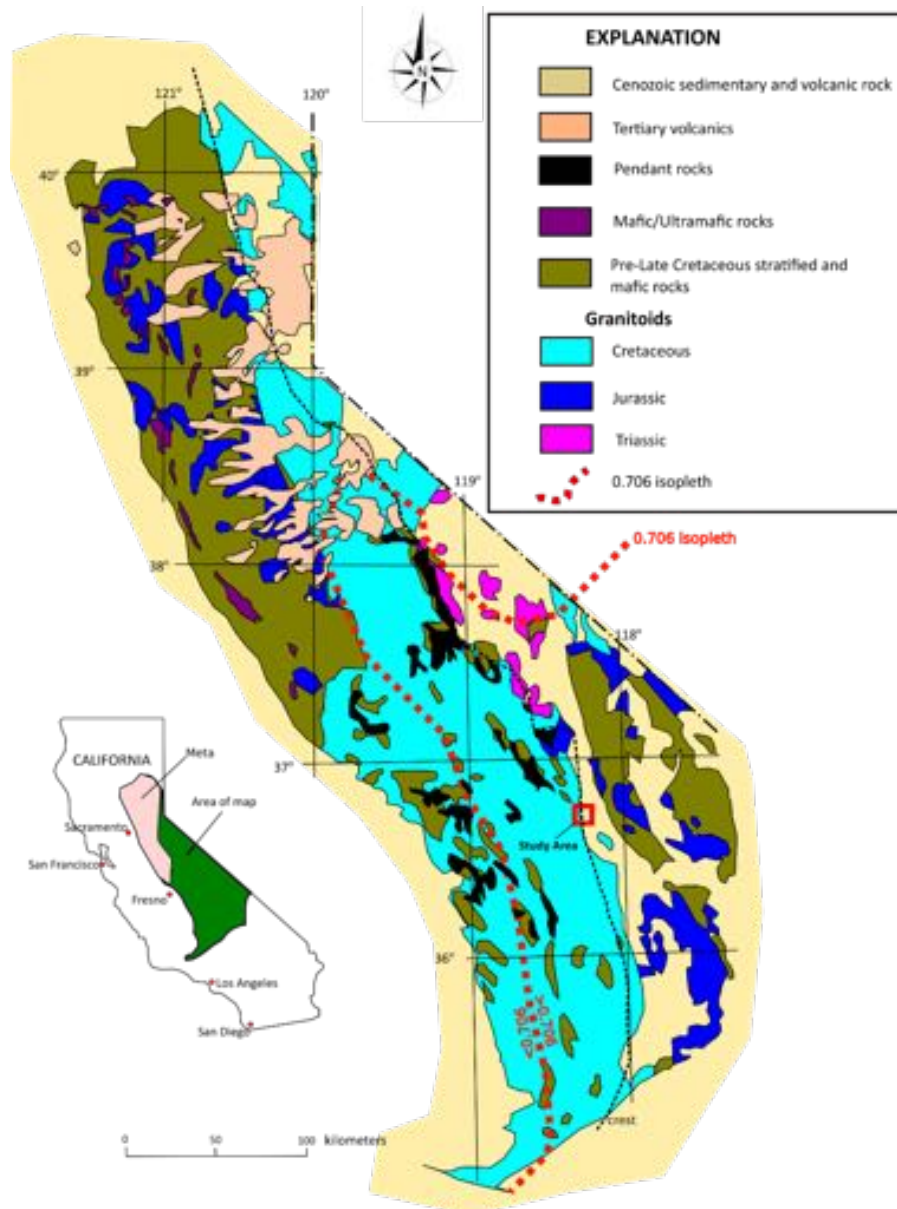


Figure 1.4: The Mesozoic Sierra Nevada Batholith located in California (USA) showing the general geology, pluton age, as well as coeval mafic/ultramafic units and wall rock pendants (after Irwin, 2003; Snelling and Gates, 2009). The 0.706 Sr isotope isopleth is also shown as described in Cecil *et al.* (2012). The study area is shown in the red square and is located in the Cretaceous Black Canyon Pluton. The lower left inset shows map location of California with major cities, the Sierra Nevada Batholith in green, and Mesozoic metamorphic rocks in pink.

It formed as a result of the Farallon plate subducting beneath the southern edge of the North American plate, leading to widespread magmatism intruding the Precambrian, Palaeozoic and early Mesozoic

oceanic crust (Chen and Moore, 1982; Bateman, 1992; Ducea and Saleeby, 1998; Ducea, 2001; Lee *et al.*, 2007; Cecil *et al.*, 2012; Ducea *et al.*, 2015; Martin and Clemens-Knott, 2015). The batholith is mainly made up of subduction-related, intermediate to felsic intrusive rocks and metamorphic screens (wall-rocks found in-between plutons) (Memeti *et al.*, 2010) and pendants (Ducea, 2001; Cecil *et al.*, 2012). The batholith consists of several plutons ranging from less than 1 km² to over 100 km², composed of rocks ranging from gabbro to granite, including tonalite, quartz diorite, granodiorite, quartz monzonite and alaskite (Chen and Moore, 1982; Ducea and Saleeby, 1998; Barbarin, 2005). Those rocks are associated with metamorphic screens which are made up of schistose to moderately gneissose domains of hornfels and pendants composed predominantly of quartzite, carbonate-pelite and silicic metavolcanics (Saleeby and Busby, 1993; Cecil *et al.*, 2012). This continental arc can be divided into several plutons based on similarities in texture, intrusive relations, colour index, and composition (Chen and Moore, 1982; Cecil *et al.*, 2012). The batholiths were emplaced by multiple injections of magma over a period of time between 248 to 80 Ma, resulting in sharp contacts between the plutons, which are separated by the thin septa of metamorphic rocks (Chen and Moore, 1982; Cecil *et al.*, 2012). The anorthosite body of interest, here named the Indian Rock Anorthosite (IRA), is situated in the southern part of the Sierra Nevada Batholith, 15 km north-west of Independence, near the headwaters of Thibaut Creek, on the eastern scarp of the Sierra Nevada mountain range (Moore, 1963). As the name suggests, the anorthosite is coincident with a summit peak known as Indian Rock (Moore, 1963). This anorthosite is hosted within the Black Canyon Pluton, which is made up of mafic granodiorites dated at 96 ± 2 Ma (Carl, 2000).

This IRA is ~2 km long and ~0.5 km wide with a surface area of approximately 1 km². No geochronological or geochemical data are available, but Moore (1963) briefly describes the characteristics and field relationships between the anorthosite and the surrounding granitoids, speculating that they are coeval based on gradational relationships between the enveloping mafic granitoids of the Black Canyon Pluton and the anorthosite. The IRA is surrounded by gabbros to the east, and diorites and mafic granodiorites of the Black Canyon Pluton to the west (Figure 3.8) (Moore, 1963). In the area, the plutons were subjected to recurring metamorphism by intrusive bodies (Holland *et al.*, 2013). The McDoogle Pluton, which occurs to the east of the anorthosite, has elongated mafic enclaves showing preferred orientation. The IRA is light grey and has an average grain size of ~4 mm (Moore, 1963). The IRA is composed of over 90% euhedral plagioclase with an average composition of An₇₀, with the remainder being clinopyroxene (altered to hornblende and chlorite), quartz and orthoclase occurring as interstitial oikocrystic phases. Most of the Black Canyon Plutonic rocks (gabbro, diorite, peridotite, and mafic granodiorite) have similar mineralogical composition to the surrounding granitoid rocks, with varying mineral modal abundances. The granodiorites of the Black Canyon Pluton in the area have a higher percentage of mafic minerals and lower modal abundances of orthoclase and quartz to plagioclase (Kistler and Ross, 1990). The gabbro contains interstitial K-

feldspar, magmatic hornblende instead of pyroxene, but in other areas, the pyroxenes are surrounded by interstitial hornblende (Kistler, 1990; Wernicke *et al.*, 1996). Given the paucity of data available on the Indian Rock Anorthosite and the scarcity of known Phanerozoic anorthosites globally, this dissertation aims to characterise the body in detail and compare it with the broader SNB in the region.

1.2 Purpose of this Study

It is not clear why the formation of massif-type anorthosite ceased towards the end of the Proterozoic Eon, nor why it began towards the end of the Neoarchean. Although Proterozoic anorthosites are thought to be products of arc-related magmatism (Wiebe, 1993; McLelland and Selleck, 2010; Ashwal and Bybee, 2017), no anorthositic body, typical of massif-type anorthosites has been characterized as coeval to a host Phanerozoic arc environment. Massif-type anorthosites, however, have been identified as exogenous, Proterozoic terranes in Phanerozoic arc environments. This is encompassed in the classification of ‘anorthosite inclusions’, as defined by Ashwal (1993), where the anorthosites are either derived from the same magma as the host (i.e. a cognate inclusion) or as a fragment of an older, unrelated crust (i.e. a xenolith/xenolithic terrane). As an example of the latter case, there are two small bodies of anorthosites situated in the Western Sierras Pampeanas of Argentina, with U-Pb zircon ages of 1070 ± 41 Ma, which have been interpreted as a fragment of the Proterozoic crust of a Precordillera terrane (Casquet *et al.*, 2005). Another body is found in southern Peru, near Ilo, which yields a Nd model age of ~ 1.42 Ga, but the U-Pb zircon ages yielded 464 ± 5 Ma (Casquet *et al.*, 2010). If massif-type anorthosites are produced by typical petrogenetic mechanisms in arc environments (i.e., mantle wedge melting, underplating, MASH zone processes and subsequent magma ascent etc.), why are they restricted to the Proterozoic Eon? And why are cognate anorthosite inclusions in continental arcs seemingly restricted to the Phanerozoic? This conundrum, and its implications for magmatic changes from the Proterozoic to Phanerozoic, form the focus of this work. The Indian Rock Anorthosite (IRA) is located in the Mesozoic Sierra Nevada Batholith, and it is a good candidate for understanding what processes involved in the formation of massif-type anorthosites can be applied to the formation of the IRA, and how have they changed over time? This project also aims to determine the petrogenesis of the Indian Rock Anorthosite (IRA), including the age relationships and the genetic link to the southern Mesozoic Sierra Nevada Batholith (SNB). Furthermore, a better understanding of the IRA and its comparison to Proterozoic massif-type anorthosites may provide insight into changes that occurred in arc-related magmatism between the Proterozoic and Phanerozoic Eons. The techniques and methods used to describe the IRA, compare it with the SNB and understand the petrologic and geodynamic implications of this body are described in Chapter 4 and include petrographic and textural analysis, major and trace element compositional data, radiogenic isotope data, including high-precision U-Pb zircon and apatite geochronology, and alphaMELTS modelling.

2. The Anorthosite Problem

Despite the remarkably simple anorthositic mineral assemblage, the petrogenesis of anorthosite remains controversial (Philpotts, 1981). Bowen (1917) defined and discussed this problem when attempting to reconstruct the formation, occurrence and source of anorthosites. He further addressed the enigma surrounding the processes involved in concentrating plagioclase crystals to form anorthosites. The principal puzzling characteristic of anorthosite is the nature of the parental magma that is capable of producing plagioclase-rich assemblages (Dempster *et al.*, 1999). Due to their near-monomineralic nature, the composition of the parental magma was not well known. Suggestions of anorthositic liquids were proposed (e.g., Loewinson-Lessing, 1923), but this would imply magmatic temperatures well over 1225 °C (Longhi, 2005). However, Bowen (1917) realised that anorthosites do not exert significant contact metamorphic influence on the surrounding rocks, disproving high magmatic temperatures expected for hyperfeldspathic melts/magmas. The confusion was centred on whether an anorthositic parental liquid (hyperfeldspathic magma) exists (Morse, 1982), or if the anorthosite could result from fractionation and accumulation of plagioclase in basaltic magma (Emslie, 1978; Longhi and Ashwal, 1985; Ashwal, 1993; Scoates and Frost, 1996). The existence of a hyperfeldspathic magma was supported by the occurrence of anorthositic dykes in anorthosite complexes (Simmons and Hanson, 1978; Wiebe, 1980). The evidence for the basaltic parental magma is the presence of high-Al gabbros and ferrogabbros, which are coeval with the anorthosite (Markl and Frost, 1999). They have been suggested to represent parental or residual magma from the anorthosite (Olson, 1992; Mitchell *et al.*, 1996). Chill margins occurring in anorthosite complexes range from high-Al to high-Fe gabbros (Robins *et al.*, 1997), which can be inferred to represent the parental magma (Markl and Frost, 1999). A high-alumina basaltic melt has been identified to be the parental composition magma of anorthosite (Ashwal, 1993) since the high-Al level would be in equilibrium with plagioclase cumulates (Emslie *et al.*, 1994). High-Al basaltic magmas can be produced by partial melting of a plagioclase/spinel peridotite source, fractional crystallisation of mafic cumulates like olivine and pyroxene, or assimilation of aluminous crustal material in a mantle-derived magma (Emslie *et al.*, 1994; McLelland and Selleck, 2010). The problem with the above-mentioned processes is that none can crystallise a large quantity of anorthosite in a closed system.

The hyperfeldspathic liquid hypothesis has fallen out of favour because of the high temperatures required (Green, 1969; Emslie, 1970). The favoured view is that these rocks are formed by fractionation and floatation of plagioclase grains from a high-Al basaltic magma, a notion which is supported by many studies over the years (Morse, 1972; Emslie, 1978; Hamilton, 1981; Beard and Barker, 1989; Ashwal, 1993; Scoates and Mitchell, 2000). Hietanen (1963), Morse (1972, 2006), Ashwal (1993), Dempster *et al.* (1999), Scoates and Mitchell (2000), and Duchesne (2001) amongst others, reached a

consensus that the plagioclase crystals firstly fractionated from a basaltic magma at Moho or lower crustal levels, accumulated and migrated away from the source through diapiric or conduit ascent, intruding into the country rock. The plagioclase grains equilibrated with the liquids, and due to the high sodic content, they float rather than sank in mafic liquids (Hamilton, 1981).

2.1 Characteristics of Proterozoic Massif-type Anorthosites

Massif-type anorthosites are a deep crustal and upper mantle boundary plutonic phenomenon, and they depict a unique period in Earth's history (Duchesne and Demaiffe, 1978; Dempster *et al.*, 1999). They occur as large monomineralic intrusive bodies with subordinate olivine, pyroxene or Fe-Ti oxides (Ashwal, 1993, 2010; Maji *et al.*, 2010). The term “massif-type” is derived from the French word for “massive”, and it connotes a large block of mountain ground or a plutonic mass of moderate size (Ashwal and Bybee, 2017). They also have a massive texture, which is mainly composed of coarse (1 cm to rarely 1 m), euhedral, lath-shaped plagioclase, sometimes containing adcumulus growth textures (Philpotts, 1981; Wiebe, 1993; Ashwal and Bybee, 2017). Chemically, they are characterized by high CaO, Al₂O₃, Na₂O, Sr, K/Rb, Sr/Ba, LREE and positive Eu anomalies and low Fe-Ti oxides (Simmons and Hanson, 1978; Ashwal, 1993; Maji *et al.*, 2010). Most large anorthosite complexes have a domical morphology that appears to have risen diapirically into the country rocks (Philpotts, 1981; Menuge, 1988). The internal structures of the complexes show deformation and negative Bouguer gravity anomalies characterize typical geophysical signatures (Philpotts, 1981). The youngest known massif-type anorthosite is the 1 km² Wilmington anorthosite in Delaware, dated around 0.502 Ga (Ashwal, 2010) and the oldest is the 95 km² Pare Mountains body in Tanzania dated around 2.643 Ga (Tenczer *et al.*, 2006). This type of anorthosite is commonly associated with suites of coeval granitoid rocks including hypersthene monzonite, mangerite, charnockite, and quartz mangerite, and are therefore collectively called anorthosite-mangerite-charnockite-granite (AMCG) complexes (Duchesne and Demaiffe, 1978; Ashwal, 1993; Scoates and Mitchell, 2000; Bolle *et al.*, 2003; Maji *et al.*, 2010; Vander Auwera *et al.*, 2011; Heinonen *et al.*, 2014; Hill, 2017). An classic example of AMCG complexes occur in eastern North America forming a broad Proterozoic age (1.7- 0.9 Ga) belt that is part of the Grenville Province (Bolle *et al.*, 2003).

2.1.1 The Parental Magma of Proterozoic Massif-type Anorthosite

The cumulate nature and composition of anorthosites makes the determination of the parental magma challenging. Therefore, previously scientists have proposed a wide range of parental magma compositions and sources. These controversial questions were spurred by the non-cotectic nature of anorthosites. Experimental work (e.g., Longhi *et al.*, 1999; Vander Auwera *et al.*, 2011) in conjunction

with field observations (e.g., Mitchell *et al.*, 1996; Charlier *et al.*, 2010) suggest that the composition of the parental melt is a high-Al basalt magma. However, this leads to important questions regarding the location of the complementary, but generally missing mafic to ultramafic component of basalt melt fractionation. To address this question, scientists have suggested that during the fractionation process, the felsic crystals floated to the tops of magma chambers, and the dense mafic components sequestered at the Moho or sank back into the mantle (e.g., Ashwal, 1993; McLelland and Selleck, 2010). This contradicts another school of thought, which suggests that a magma capable of producing rocks like the anorthosite has to be a more felsic parental magma with 70- 80% plagioclase (Buddington, 1969; Mitchell *et al.*, 1996). Experimental work by Mysen and Boettcher (1976) show that the hydrous melting of the mafic material (garnet-websterite) is not capable of producing plagioclase crystals observed in the massif-type anorthosites, and that a plagioclase-rich parental magma at high pressure (15- 23 kbar) is a more plausible parental melt.

Despite the other scientists suggesting a more felsic parental magma, there is field evidence supporting a basaltic parental magma composition. The field evidence includes dykes and “chill margins” which have been documented in many anorthosite complexes like the Nain Plutonic Suite, Harp Lake, Kiglapait, Laramie, Skattøra migmatite complex and Rogaland (Emslie, 1973; Berg, 1980; Nolan and Morse, 1986; Demaiffe *et al.*, 1990; Wiebe and Snyder, 1993; Mitchell *et al.*, 1996; Robins *et al.*, 1997; Selbekk *et al.*, 2000). Those dykes and “chill margins” have An content that relates to the study done by Morse (1972) showing that the An content of the anorthosites is closely related to the norites in the area and the parental melt of the Nain complex is a low K, high-alumina basalt with FeO, CaO, and MgO nearer to 1, which is similar to an oceanic tholeiite. In terms of formation, many petrologists (e.g., Emslie, 1978, 1985; Ashwal, 1993; Wiebe, 1993; Longhi *et al.*, 1999; Bolle *et al.*, 2003; Charlier *et al.*, 2010; Bybee *et al.*, 2014) agree that the petrogenesis involves two principal stages; 1) extensive crystallisation (with crustal assimilation) of a mantle-derived parental basaltic magma at or near the base of the crust, 2) ascent (by either diking or diapirism) and intrusion of the plagioclase mush into the mid-to-upper crust where final crystallization occurs.

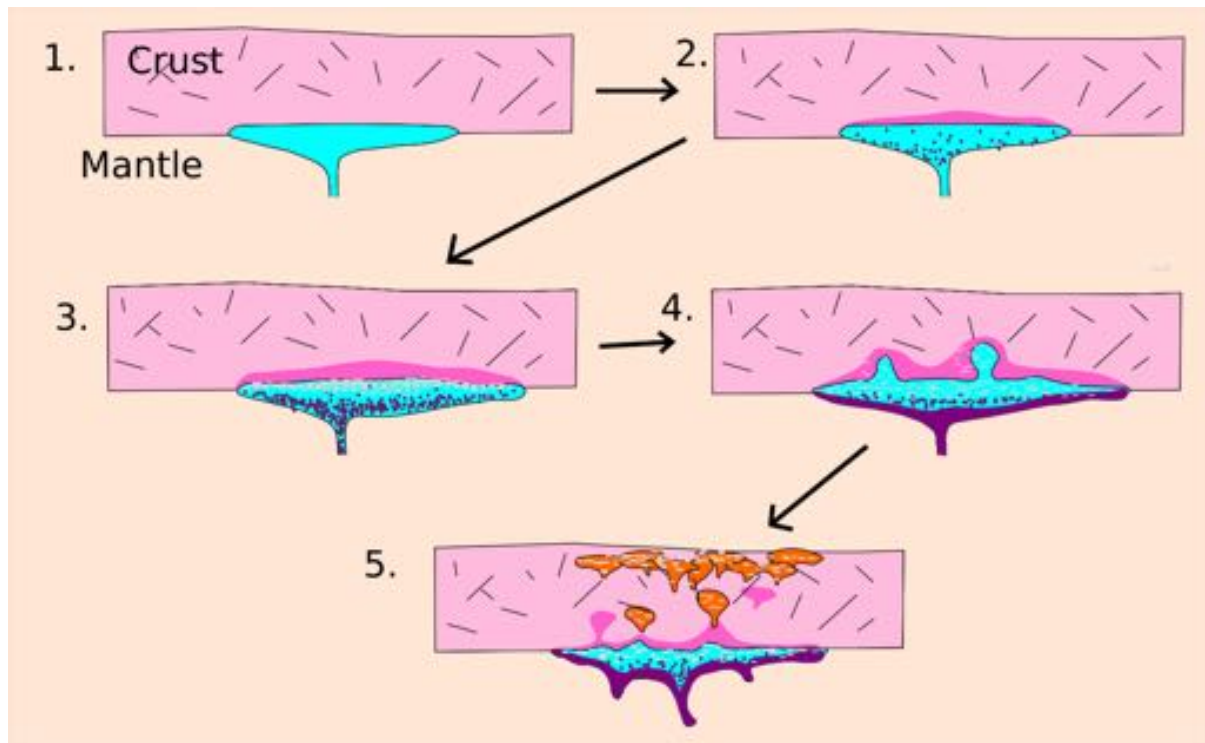


Figure 2.1: Schematic diagram showing a model for the generation of massif-type anorthosites. 1. Hot mantle-derived magma underplates the crust. 2. The crystallisation of the mafic minerals, which sink to the bottom, and partial melting of the crust above the magma. 3. The melt becomes saturated, and plagioclase minerals then rise to the top of the chamber whereas mafic minerals sink. 4. Plagioclase accumulation becomes buoyant and floats to the top. 5. Plagioclase crystal mush coalesces to form massif anorthosite, whereas granitoids crustal melts rise to the shallow levels. Modified from Ashwal (1993).

Determination of the parental magma source has divided researchers for many decades, debating between mantle and lower crustal sources. This source has caused confusion due to the lack of understanding of the relationship of the granitoids of the AMGC suite to the anorthosite. Field relations, mineralogy, trace element studies and radiogenic and stable isotope signatures suggest that the parental magma was derived from a basaltic composition melt (Bowen, 1922; Buddington, 1969; Emslie, 1978, 1985; Simmons and Hanson, 1978; Wiebe, 1980; Morse, 1982; Olson, 1992; Emslie and Hegner, 1993; Mitchell *et al.*, 1996; Scoates and Frost, 1996; Scoates and Mitchell, 2000; Ashwal and Bybee, 2017; Hill, 2017). One group suggests that the basaltic melt originated from the mafic lower crust (Michot, 1964; Philpotts, 1968; Taylor *et al.*, 1984; Longhi *et al.*, 1999; Selbekk *et al.*, 2000; Bédard, 2010). This group argues that the rocks of the AMGC suite are related and derived from the fractionation of the mafic lower crust (monzodiorite or granodiorite). The anatexis of the Al-enriched residue results in the anorthosite magma and the mafic component of the anorthosite massif complexes being generated from the mafic portion of the lower crust (Taylor *et al.*, 1984). The other group argues that the anorthosite and the granitoids are chemically independent and they consider the felsic rocks to be the result of the partial crustal partial melting induced by the intrusion and emplacement of the anorthositic mushes

(Simmons and Hanson, 1978; Ashwal and Bybee, 2017). Heinonen *et al.* (2014) suggested that the granitoids that occur with the anorthosites originate from continental arcs. For example, the ferrodiorites associated with the Proterozoic massif type anorthosites are interpreted as residual melts from the deep crustal fractionation of basaltic magma (Mitchell *et al.*, 1996). The parental magma of the ferrodiorites originates from the deep crustal anatexis of basic to intermediate rocks (Duchesne *et al.*, 1999), or immiscible liquids from a mangeritic magma with no genetic link to the anorthosite (Owens and Dymek, 1995). In most complexes, the major rock types of the AMCG suite have gradational contacts (Philpotts, 1981). The granitoids in the AMCG suite have been suggested to form by anatexis at the base of the crust, and the mafic phases are the residual liquids extracted after the crystallisation of anorthosite (Figure 2.1) (Ashwal, 1993; Bolle *et al.*, 2003).

Isotopic work done by Menuge (1988), Emslie and Hegner (1993), and Dempster *et al.* (1999) has shown the mantle to be the more plausible source of the anorthosites. The work by Morse (1982) showed that the tholeiitic signature of the anorthosites supports the hypothesis of mantle derivation. Trace element data presented by Simmons and Hanson (1978) supports the generation of anorthositic rocks from a tholeiitic basalt at a depth equivalent to 15 to 20 kbar, which leaves a pyroxenitic residue. The magma of such composition would become supersaturated with plagioclase during the upward movement through the crust because of its buoyancy (Emslie, 1973; Simmons and Hanson, 1978). According to Mitchell *et al.* (1996), the possible parental magma for anorthosites should be high-Al gabbro characterised by high Al_2O_3 (15- 19 wt%), highly enriched in REE with a positive Eu anomaly, high ϵ_{Nd} (up to +2), and low initial $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7033). Oxygen, Nd and Lu-Hf isotope data (Menuge, 1988; Heinonen *et al.*, 2014) show that the source of the magma is a depleted mantle, with the magmas having been contaminated by the assimilation of crust. This is supported by a recent study done by Hill (2017) using experimental petrology, which showed that it is possible to produce anorthosites from a mantle-derived magma by crustal assimilation, supporting conclusions reached by Morse (1972, 2006), Olson (1992), Markl and Frost (1999), Scoates and Mitchell (2000), Heinonen *et al.* (2014), and Ashwal and Bybee (2017). In support of the assimilation of crustal material model, Michot (1964) and Philpotts (1968) suggested that the contaminants can be either pelitic sediments within a basaltic host or mixing of basaltic magmas with granitic melts, which can enhance the crystallization of plagioclase.

It has been argued that the depleted-mantle Nd and Pb isotopic signatures do not necessarily mean that the source is a depleted upper mantle but was consistent with the origin of the partial melting of a basic rocks (i.e. pyroxene-rich mafic cumulates) with short crustal residence times (Bolle *et al.*, 2003). Simmons and Hanson (1978), Taylor *et al.* (1984), Longhi *et al.* (1999), Duchesne (2001), Wiszniewska *et al.* (2002), Druppel *et al.* (2007), and Vander Auwera *et al.* (2011) amongst others suggested that the source of the parental magma is lower mafic crust, where the melt is generated by the process of underplating by a mantle plume and also increasing the Al-content in the residue after the melt extraction.

This is supported by geophysical evidence, Re-Os isotope data and experimental work done by Duchesne *et al.* (1999) and Schiellerup *et al.* (2000). Taylor *et al.* (1984) showed that a magma capable of crystallising Al-rich rocks could be generated from the lower crust after the removal of granitic melts, which in turn depletes the lower crust in K, U, and Th, and enriches the residue (crust) in Al, Sr, and Eu. Petrological experimental work done by Longhi *et al.* (1999) showed that at a depth of formation (10-13 kbar) of the Proterozoic anorthosites and associated mafic intrusions that it is not possible to form these bodies through the process of fractionation from a mantle-derived mafic magma, with or without contamination. But this is possible if the parental magma is a gabbro-noritic rock, which can be found in the lower continental crust or mafic plutons sunk out of the crust (Longhi *et al.*, 1999; Bolle *et al.*, 2003).

2.1.2 Tectonic Setting of Proterozoic Massif-type Anorthosites

The event/s, source of heat or tectonic setting capable of restricting massif-type anorthosite to the Proterozoic Eon have been an active topic of discussion for many years. The criteria for a suitable model of tectonic setting suggested by Ashwal and Bybee (2017) include the requirement that large volumes of basaltic magma be supplied to the deep continental crust, commonly in linear belts that can extend for more than 1500 km, with each system existing temporally for 80-100 million years. Massif-type anorthosites are known to be a Proterozoic phenomenon and for decades they were thought to have been generated in continental rift environments to form large coarse-grained intrusive bodies (Bridgewater and Windley, 1973; Emslie, 1973; Berg, 1977; Morse, 1982). This was supported by the mineral assemblages study by Berg (1977) in the contact aureoles on the Nain Plutonic Suite, which suggest an emplacement depth corresponding to 2.5-4.2 kbar with ambient temperatures less than 250 °C. In the Nain Plutonic Suite, the garnet porphyroblasts have partially reacted to produce symplectites of cordierite and hypersthene, resulting in a corona-like texture which indicate that some of the rocks in the area have experienced a polymetamorphic history (Berg, 1977). The regional distribution of the pressures appears systematic and suggests an arch (which trends WNW-ESE) in the ancient pressure surfaces (paleo-isobars), which coincides with a proposed graben structure in the Conrad discontinuity based on gravity studies (Berg, 1977). The association of the mentioned features is rationalised with a mechanism of incipient continental rifting. The rift environment idea is more likely to create enormous bodies of massifs, but it fails to account for anorthosites that were emplaced without having initiated major continental break up (e.g., Grenville Provinces), Emslie (1978) pointed out that rifting locally post-dates the emplacement of the anorthosites, and this makes the rift environment to be less likely.

Then the idea of a convergent margin setting being a possible tectonic setting for massif-type anorthosites gained more traction. Corrigan and Hanmer (1997) proposed that massif-type anorthosites are a direct result of collisional orogeny. This model involves the post-collisional replacement of mantle

lithosphere by asthenosphere by either delamination or convective thinning. The model is based on the U-Pb ages and baddeleyite ages of the AMCG complexes of the central Grenville province, which coincide, with the timing of thrusting along first order contractional shear zones. The model is also plausible in terms of explaining the episodic nature of the AMCG-type magmatism with synchronous with contractional tectonics. Scoates and Chamberlain (1997) proposed that the 1.76 Ga Horse Creek anorthosite complex was emplaced several million years after a collision along the suture in a late- to post-orogenic transtensional environment. McLelland *et al.* (2010) proposed a model, which shows late- to post-collisional delamination of mantle lithosphere that is replaced by asthenosphere and production of basaltic magmas by adiabatic decompression. The model was based on a zircon geochronological study, which showed that the Adirondack anorthosite-mangerite-charnockite-granite suite was emplaced during the terminal Shawinigan orogenesis, and the timing coincides with ages of other large anorthosite-mangerite-charnockite-granite massifs in southeast Laurentia. A more recent model proposed by Ashwal and Bybee (2017) indicates that the most probable tectonic environment for Proterozoic massif-type anorthosite formation is a convergent margin setting. This model suggests an Andean-type continental arc, which agrees with the observation of massif-type anorthosite and related rocks, and most importantly it accounts for the lengthy and episodic mantle-derived magmatism, the emplacement into the deep continental crust and the spatial arrangement in linear belts of appropriate scale. The challenges faced by the model are that there is no known massif-type anorthosite found in modern Andean-type arcs today.

Thermobarometric studies of unmetamorphosed anorthosite massifs illustrate that they were emplaced at shallow to intermediate crustal level (Berg, 1977). According to Morse (1982), massif-type anorthosites crystallised at depths of 7 km, where the temperatures of the country-rock were probably less than 250 °C. Frost *et al.* (2010) showed that massif-type anorthosites are formed by fractionation processes in a magma chamber *in situ* or close to the present level of exposure followed by doming and expulsion of residual liquid. The general trend of occurrence of massif-type anorthosite has been observed in two broad tectonic settings which are an active or post-collisional margin and a within plate association (Figure 2.2) (Herz, 1969; Heinonen, 2012). Most of them are found in paleogeodynamic reconstructions before the break-up of continents (Duchesne and Demaiffe, 1978). They occur dominantly in the Mid-Proterozoic (1800-1000 Ma) in the region of the Proterozoic belts characterised by composite batholiths consisting of multiple, coalesced plutons of different compositions dominated by anorthosite and leuconorite (Wiebe, 1993; McLelland and Selleck, 2010). The Mid-Proterozoic Era is geologically dominated by the breakup of the Columbia Supercontinent (Nuna) (Roberts *et al.*, 2015). The two broad Proterozoic belts include [1] one in Laurasia and the other in Gondwana, and [2] the great circle belt found in the supercontinent (Figure 2.2) (Duchesne and Demaiffe, 1978). The great circle belt known as Grenvillian mobile belt was formed as the result of the formation of Rodinia supercontinent at the end of Mesoproterozoic into the Neoproterozoic (Casquet *et al.*, 2010; Roberts *et*

al., 2015). During the breakup of Rodinia, the anorthosite clusters were dispersed in several continents (Casquet *et al.*, 2010). Dempster *et al.* (1999) suggested that the temporal restriction of the massif-type anorthosite might be related to the nature of the crust at that time. This idea was based on the conclusion that the crust plays a considerable role in crystallising anorthosites. The majority of the Earth's crust (50-70%) was formed towards the end of the Archean (Dempster *et al.*, 1999), which gave rise to the first supercontinent during the Proterozoic Eon. Proterozoic anorthosites are restricted through the geological timeline, and they are known to occur in the Proterozoic Eon between 2.7 Ga to 0.5 Ga (Ashwal and Bybee, 2017).). The reason behind the temporal restriction of this type of anorthosite is still unknown. Ashwal (2010) showed that the An-content of plagioclase could be used as a geochronometer, but it was not enough to explain the temporal restriction.

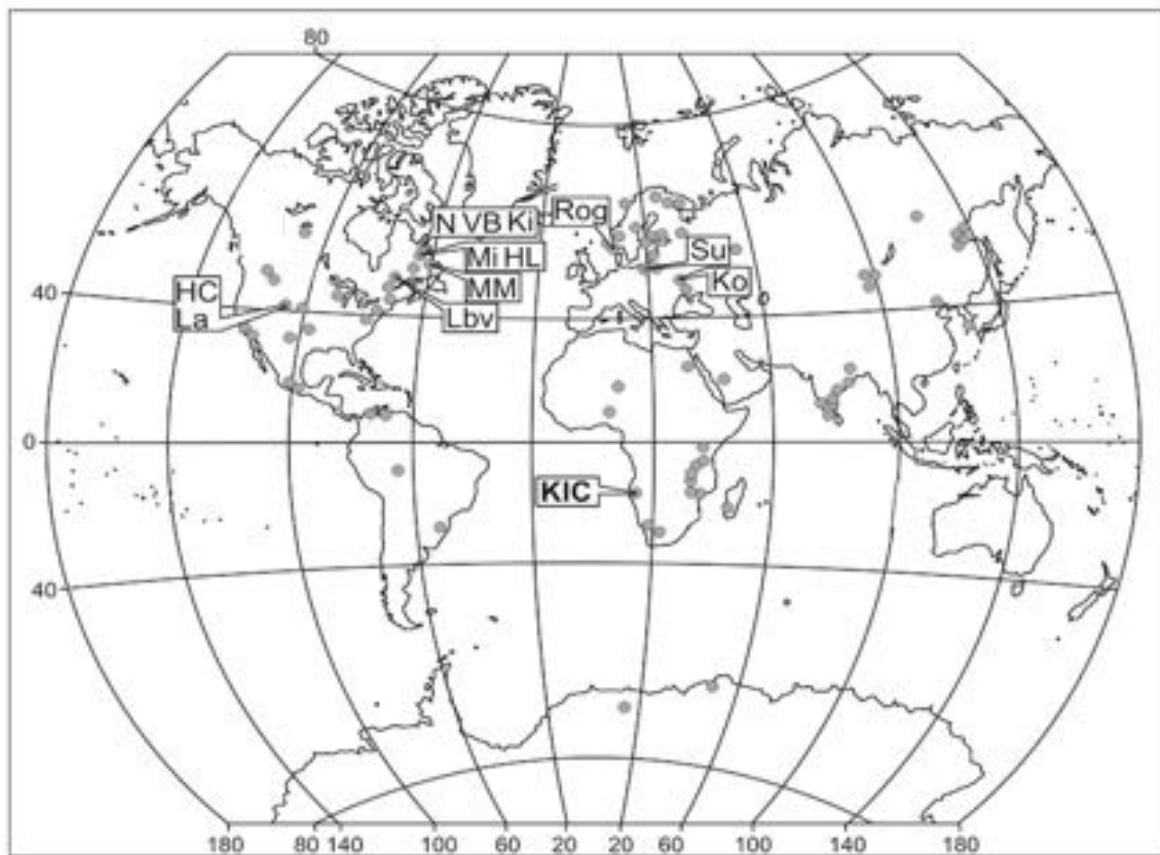


Figure 2.2: Global distribution of major Proterozoic massif-type anorthosites (Hill, 2017). HC, Horse Creek; HL: Harp Lake; KIC: Kunene; Ko: Korosten; La: Laramie; Lbv: Labrieville; Mi: Michikamau; MM: Mealy Mountains; N: Nain; Rog: Rogaland; Su: Suwalki. Associated mafic intrusions are also shown. Ki: Kiglapait; VB: Voisey's Bay.

2.2 Characteristics of Anorthosite Inclusions in Arcs

Anorthosite inclusions represent the accumulation of plagioclase from the same magma as the host (cognate inclusion) or as a fragment of an older, unrelated crust (xenolith) (Ashwal, 1993). Several types of cognate inclusions have been identified: granular anorthosite with > 90 % unzoned, coarse plagioclase (An₅₉₋₆₂), laminated anorthosite with tabular zoned plagioclase (An₆₃₋₆₅ to An₄₀₋₄₅), or megacrysts with euhedral plagioclase (An₃₅₋₆₅) (Ashwal, 1993). Cognate anorthosite inclusions represent the accumulation of plagioclase that crystallised from their host magmas. Ashwal (1993) noted that the sizes and the plagioclase An content of the megacrysts correlate to the composition of the host. For example, small megacrysts (An₅₈₋₆₅) are situated in minimally fractionated gabbros, larger megacrysts (An₅₀₋₅₈) in an evolved host, or coarse glomeroporphyritic aggregates of sodic labradorite, and giant megacrysts in alkali gabbros and trachydolerites. The plagioclase compositions of the anorthosite match with the host dykes rocks, and geochemical and isotopic signature of the inclusions demonstrate that they crystallised from the same magma as the host (Ashwal, 2021). A good example of this type is the anorthosite inclusions in a Tertiary dolerite sill which occurs on the island of Hrapsey and Purkey in western Iceland. Anorthosite inclusions in volcanic rocks of island arcs are common and their composition ranges from anorthosite to ultramafic lithologies (Ashwal, 1993). Examples include Kurile-Kamchatka region, U.S.S.R, Lesser Antilles, and Hakone and associated volcanoes, Japan (Ashwal, 1993; Kuno, 19760; Lewis, 1973). These bodies are characterized by high calcic plagioclase (An₈₀₋₁₀₀) and Mg-rich olivine (Fo₇₀₋₈₅), and other phases include pyroxenes, Fe – Ti oxides, and glass (Ashwal, 1993). Some of the larger inclusions have phase layering and other small-scale primary igneous heterogeneities (Figure 2.3). In some island arc volcanic rocks are megacrysts of calcic plagioclase having anorthite content close to An₁₀₀ (Ashwal, 1993).



Figure 2.3: Anorthositic inclusion in andesite, northern Mariana Island Arc, showing phase layering (Stern, 1979).

The anorthosites in the Konkanhundi gabbro-anorthosite suite from Peninsular India, adjacent to the major collisional suture between the Western and Eastern Dharwar cratonic blocks is believed to have formed in a collision environment (Figure 2.4). The anorthosites display high Al_2O_3 contents with negative correlation against MgO, high CaO and Sr contents, and positive Eu anomalies indicating plagioclase flotation in the magma chamber (Santosh and Li, 2018). This anorthosite has been interpreted as an anorthosite in an Archean continental arc (Santosh and Li, 2018). The pronounced positive Eu anomalies have been interpreted as plagioclase accumulation from a fractionated magma (Zhou *et al.*, 2016). The negative Nb anomalies suggest the magma was derived from a depleted mantle or within a subduction zone contaminated with old continental crust (Mohan *et al.*, 2013; Polat *et al.*, 2009). The anorthosites in the Konkanhundi gabbro-anorthosite suite has magmatic zircon $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 2594 ± 16 Ma, which is consistent with the ages of the lithologies around it. The cause of the unusual high An content in the plagioclase has been a debate amongst petrologists. High H_2O content in the parental melts has been proposed to be the crucial factor in accountability of the abnormally calcic plagioclase (Stern, 1979; Arculus and Wills, 1980). This is supported by the observation of primary amphiboles in the inclusions (Ashwal, 1993), and it will be further investigated in this research.

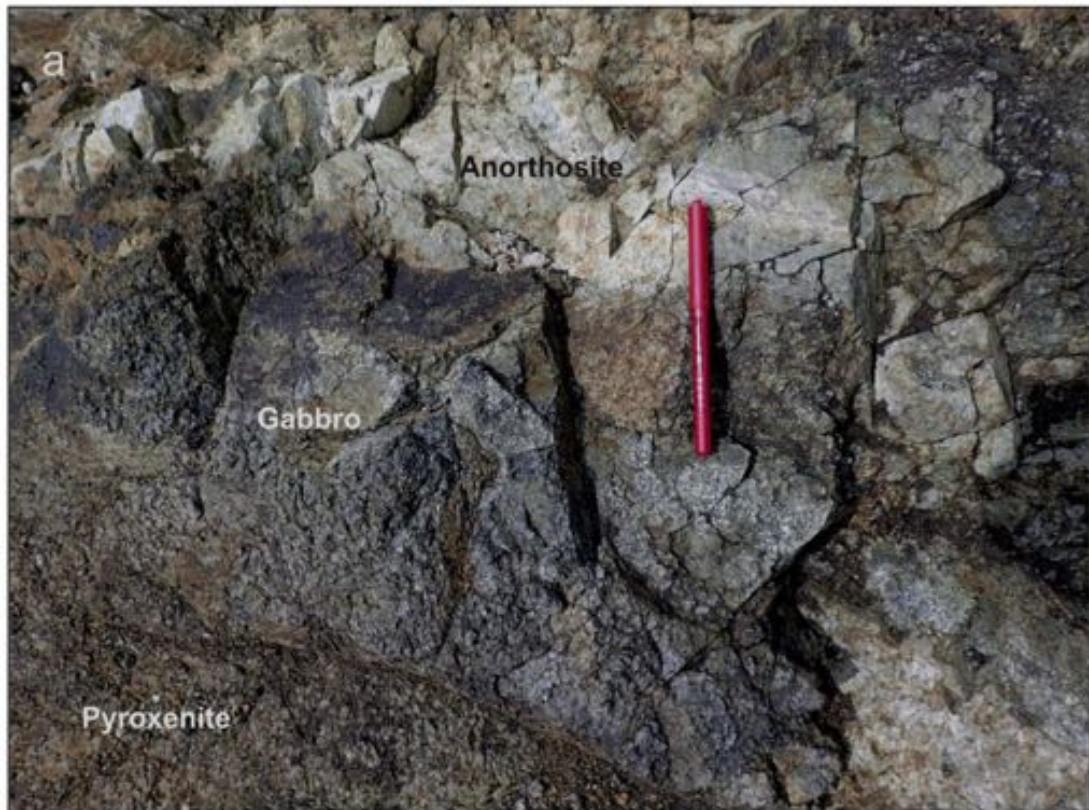


Figure 2.4: The anorthosite from the Konkanhundi gabbro-anorthosite suite from Peninsular India (Santosh and Li, 2018).

Xenolithic anorthosite inclusions in arc terranes range in size from xenocrysts <1 cm to several hundred meters in a Proterozoic diabase of the Beaver Bay Complex (Morrison et al., 1983; Ashwal, 2010). The inclusions are mineralogically, geochemically and isotopically distinct from the host (Ashwal, 2020). Examples of this type are two massif anorthosite bodies situated in the western part of the South American continent. One is situated in the Western Sierras Pampeanas of Argentina and yield U-Pb zircon ages of 1070 ± 41 Ma (Casquet *et al.*, 2010). It has been interpreted as a fragment of the Proterozoic crust of Precordillera terrane (Casquet *et al.*, 2008, 2010). Another body called the Arequipa Massif is found in southern Peru near Ilo. The Arequipa Massif is a Paleoproterozoic inlier in South American Grenville-age orogen (Casquet *et al.*, 2010). The Grenville-age metamorphism took place in three stages, at 1040 Ma in the Quilca and Camaná areas of the Camaná–Mollendo domain, and in the San Juan Marcona domain, 940 ± 6 Ma in the Mollendo area of the Camaná–Mollendo domain, and between 1000 and 850 Ma in the Atico domain. Therefore, the ages of the Arequipa Massif domains are associated with different Grenvillian metamorphic histories. Those two anorthositic bodies occur as inclusions in a Phanerozoic and modern-day Andean arc, but little is known about the Indian Rock Anorthosite body situated in the Sierra Nevada batholith, which may be one of its kind.

3. Geological background and the Indian Rock Anorthosite (IRA)

The Mesozoic Sierra Nevada Batholith is made up of plutonic rocks that were emplaced from 248 to 80 Ma by westward subduction beneath the North American plate (Kistler, 1990; Bateman, 1992; Memeti *et al.*, 2010; Cecil *et al.*, 2012). The batholith is mostly dominated by products of the continental arc produced by the convergence of south-western Laurentia and the Farallon plate (Bateman, 1992; Martin and Clemens-Knott, 2015). Paleozoic sedimentary country rocks crop out in the eastern part of the batholith (in the White and Inyo Mountains), and metamorphosed sediments with volcanic strata crop out in the western part (Bateman, 1992). The plutonic rocks are volumetrically greater in the central and southern Sierra Nevada (between $\sim 35^\circ$ N and 37.5° N) (Bateman, 1992; Tobisch *et al.*, 1995) with an estimated granitic volume of $\sim 4 \times 10^5$ km³. The thermal axis of the magmatic arc migrated eastwards at an average of ~ 2.7 mm/year (Chen and Moore, 1982; Tobisch *et al.*, 1995). The depth of pluton emplacement becomes shallower towards the north (Ducea and Saleeby, 1998; Cecil *et al.*, 2012). The southern and central Sierra Nevada has been documented to have a dense crustal root which delaminated in the late Neogene-Quaternary time from the base of the crust (Figure 3.1) (Ducea and Saleeby, 1998; Ducea, 2000; Cecil *et al.*, 2012; Saleeby *et al.*, 2012). Petrologic and isotopic studies of the SNB indicate that there are at least two generalised magma sources (depleted mantle and continental crustal material) during the generation of the batholith (DePaolo, 1981; Pickett and Saleeby, 1994). Kistler (1990) and Chen and Tilton (1991) showed that the nature of the crustal basement to the pluton changes across the batholith from the continental-derived metasedimentary rocks in the west to a crystalline basement in the east. The basement underneath the batholith changes from mainly Proterozoic continental and transitional lithosphere in the south to accreted Phanerozoic belts in the north (Cecil *et al.*, 2012).

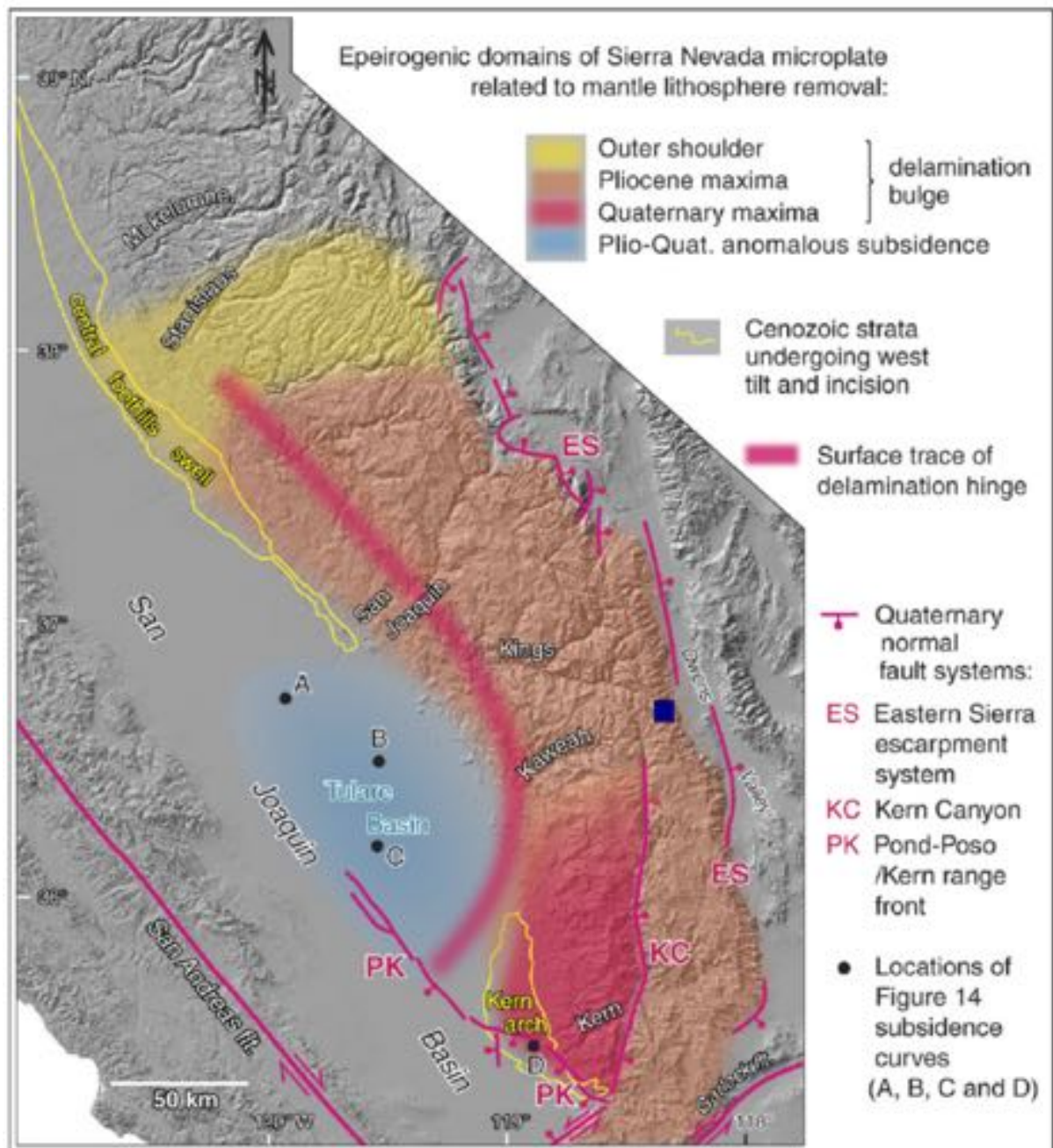


Figure 3.1: Map of the southern Sierra Nevada region which shows domains interpreted to reflect rock uplift related to delamination of the root to the southern Sierra Nevada Batholith (Saleeby *et al.*, 2012). The blue box shows the approximate location of the Indian Rock Anorthosite.

The batholith comprises granodioritic plutons with subordinate Black Canyon Pluton made up of mafic granodiorites which hosts the Indian Rock Anorthosite (Moore, 1963; Coleman *et al.*, 2000). Anorthositic bodies in the Sierra Nevada are not entirely unknown. One example is documented by Nishimori (1974), who described a range of gabbroic rocks, including anorthositic gabbros which occur as components of the Peninsular Ranges Batholith in San Diego, California, USA. The rocks show

layering with thickness ranging between a few mm to 1 m. The rocks are made up of varying proportions of plagioclase, olivine, orthopyroxene, clinopyroxene and hornblende. The plagioclase An content and olivine forsterite content ranges from An_{80-100} and Fo_{78-81} respectively (Nishimori, 1974). Another intrusive body was described by Loomis (1963), which occurs in the Sierra Nevada Batholith near Lake Tahoe (Figure 3.2). The intrusions described by Loomis (1963) are noritic-anorthosite, which has 80 % plagioclase grains with plagioclase An content ranging from An_{53} to An_{70} and for a rock to be classified as a noritic anorthosite, the plagioclase modal abundance must be more than 77.5 % but less than 90% (Streckeisen, 1967). The noritic anorthosite bodies are composed 80% plagioclase, 0.7% orthoclase, 9.5 % hypersthene, 2.7 % augite, 1 % hornblende, 0.3 % biotite, 4% magnetite, 0.1 % apatite (Loomis, 1963). The IRA has been described as a rock that has more than 90% plagioclase by Moore (1963) (see Section 1.1 for details), which indicates that according to the IUGS classification it can be classified as an anorthosite. This type of intrusive anorthosite with over 90% plagioclase and massive texture has never been described in a Phanerozoic arc environment as a comagmatic/coeval body, but only occurred as an older inclusion instead, for example, the two small bodies of anorthosites situated in the Western Sierras Pampeanas of Argentina and another in southern Peru near Ilo. Therefore, a broad understanding of the formation and mechanisms involved in the formation of the batholith can help us paint a picture of how the IRA is related to the batholith.

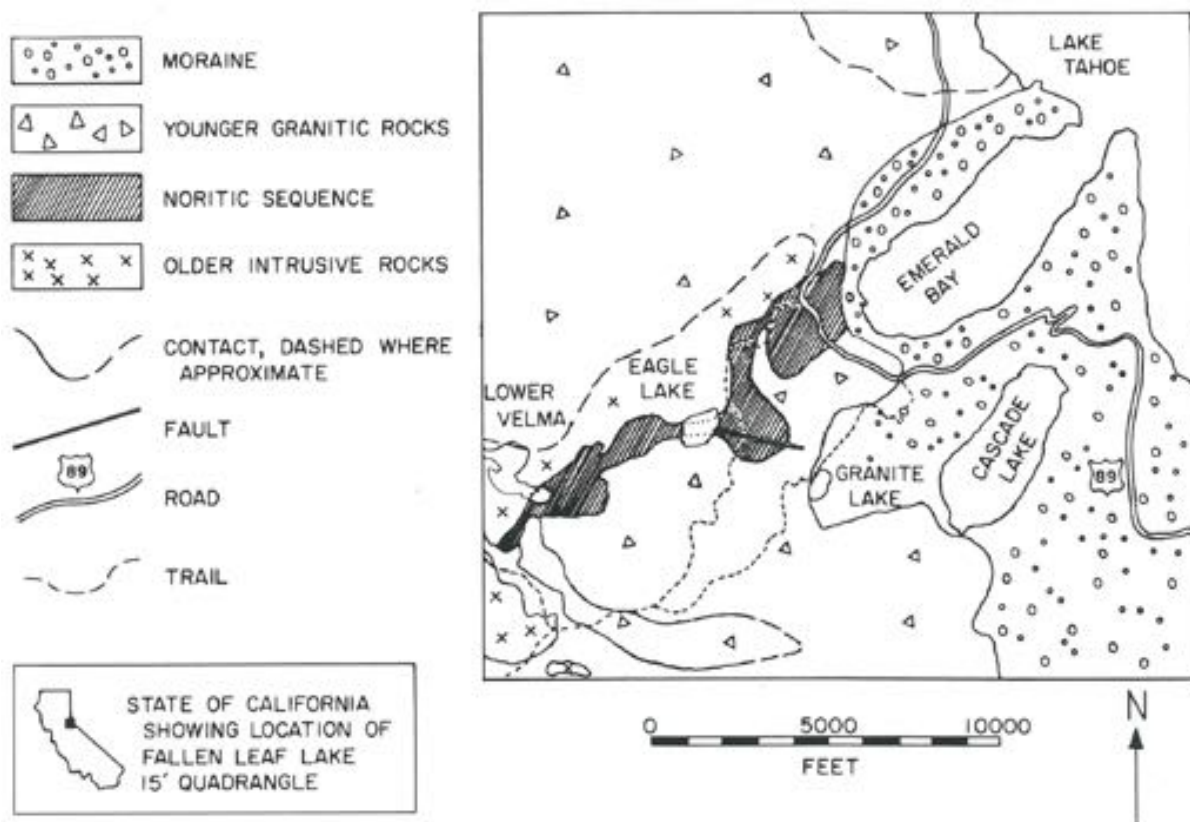


Figure 3.2: Geological map of the location of the noritic-anorthosite in Sierra Nevada Batholith (Loomis, 1963).

3.1 The construction and parental magma source of the SNB

Continental margin batholiths provide an opportunity to study the evolution and growth of the continental crust as well as the interaction between the continental and oceanic crust. During the formation of young oceanic crust at the mid-ocean ridges, the older oceanic crust subducts under continental crust, forming a compressional/convergent margin known as subduction zones (Paterson and Ducea, 2015). This process creates long arcuate belts of volcanoes parallel to the subduction zone that overlie subsurface magmatic systems above subduction margins (Ducea *et al.*, 2015; Paterson and Ducea, 2015). When a belt of volcanoes builds on oceanic crust, they are called oceanic arcs, whereas continental, Andean-type or Cordilleran-type arcs form at the margins of the older continental crust (Ducea *et al.*, 2015; Paterson and Ducea, 2015). In continental arc settings, there is relatively constant mafic melt production in the mantle wedge resulting in thickening events forcing filtration of mafic magmas into the lower crustal reservoirs that eventually flare-up in large batholith forming episodes (Ducea *et al.*, 2015). These flare-up events usually leave behind a large amount of residual material, mostly dominated by eclogites and plagioclase-poor pyroxene rocks (Lee *et al.*, 2007; Ducea *et al.*, 2015). Subduction magmatism occurring in the upper plate is characterized by dehydration reaction of the downgoing slab which adds fluids to the continental arcs resulting in hydrated minerals assemblages in the rocks (Ducea *et al.*, 2015). During the collision between the oceanic and continental crust, processes like erosion and tectonics result in the upper crustal rocks in the arc being subducted into the mantle wedge, where they are partially melted and recycled into new magmas (Paterson and Ducea, 2015). The deformation environment for magmatic arcs occurs from varying extensional to neutral to contractional regimes, mostly in response to the downgoing slab between 0 and 100 km depth, slab age and convergence rate (Tobisch *et al.*, 1995).

The granitoids were generated in the lower crust as the results of underplating by the rising basaltic melt and heat from the mantle (Bateman, 1992). The basaltic magma caused the generation of anatectic magmas by partial melting the overlying crust or mixed with crustal materials of variable bulk compositions to produce parental magmas of a wide range of compositions and isotopic signatures (Bateman, 1992). Low seismic velocities in the upper mantle and high electrical conductivities have led the conclusion that the asthenosphere beneath the southern SNB is partially molten with the presence of sulphides and fluid inclusions at depths of 35 - 40 km (Ducea, 2000). The feldspathic rocks (granulites) have been interpreted as cumulates of the SNB, whereas the olivine-free and garnet-rich rocks (typically showing higher equilibration pressures) have been interpreted as fragments of subducted oceanic crust (Dodge and Bateman, 1988; Ducea and Saleeby, 1998). There is another possibility that feldspathic rocks could be tertiary cumulates related to the magmas that carried the

xenoliths, or lower crustal fragments that are not associated with lower pressure material (Ducea and Saleeby, 1998).

The general depth of exposure throughout the SNB is 7 - 13 km (2 - 4 kbar); in the southern SNB, the exposure of Lake Isabella region is 10-17 km (3-5 kbar) and Tehachapi Mountains is 23 - 30 km (7 - 9 kbar) (Lackey *et al.*, 2005). Two xenolith observations suggest that the Sierra Nevada crust had a thickness of ~70 km and the mantle lithosphere had a thickness of at least 120 km during the Mesozoic and thinned out in the late Cenozoic due to the processes of extension and delamination (Ducea, 2001). In the first observation, the metasedimentary xenoliths found in the San Joaquin volcanic field are eclogite facies quartzites that equilibrated at pressures above 15 kbar and the most profound measured pressure of 25 kbar, indicating a depth of at least 70 km (Ducea, 2001; Lackey *et al.*, 2005). The second observation is the garnet peridotite xenoliths, showing Northern American lithospheric isotopic signatures that equilibrated at ~120 km depth (Ducea, 2001). This indicates that the mantle was thicker in the Mesozoic than presently, and the angle of subduction decreased during the Laramide orogeny (Ducea, 2001). Concordant wall-rock inclusions and screens separate the plutons from Jurassic dated plutons (Mahan *et al.*, 2003). The microstructural fabric in the McDoogle Pluton has preserved a northeast-southwest foliation indicating a horizontal shortening, sub-vertical extension and a northeast-side-up shear (Mahan *et al.*, 2003). The thermobarometric values from the McDoogle Pluton range from 0.19 to 0.26 GPa (depths of 7 to 9.5 km) and the samples from Sawmill Lake Pluton gave 0.25 to 0.33 GPa (Mahan *et al.*, 2003). The broad exposure in the area is estimated to be ~10 km with the thermobarometric values of the wall rock bearing garnet-andalusite (6 km northeast of the McDoogle Pluton) yielding peak metamorphic conditions of 700 °C and 0.2 GPa (Mahan *et al.*, 2003). The McDoogle Pluton comprises a central quartz monzodiorite phase, bounded by mafic granodiorite phase on the eastern and western margins. Major mineral phases of the granodiorites are plagioclase, hornblende, biotite and K-feldspar. These phases contain disk- or rod-shaped fine-grained mafic enclaves that are aligned in a magmatic fabric (Mahan *et al.*, 2003).

3.2 General Petrology of Sierra Nevada Batholith

The Mesozoic plutons on the Cordilleran fold belts represent the roots of a magmatic arc related to a paleosubduction zone (Windley and Smith, 1976). The plutons tend to become younger, more silicic, and more voluminous towards the east (Cecil *et al.*, 2012). In the western part of the batholith, tonalite is the most common rock type, along with the biotite granodiorite, trondjemite and granite, while hornblende-biotite granodiorite is the most common rock type, along with biotite granite, while the eastern part is dominated by granodiorite and granite (Bateman, 1992; Lackey *et al.*, 2005). In terms of modal abundance, the K-feldspar increases eastward from an average of 4 % in the west to 30 % in the eastern part (the White Mountains) (Bateman, 1992). Plagioclase decreases from an average of 60

% in the west to 30 % on the east of the batholith (Bateman, 1992). The mafic minerals reduce from an average of 14 % in the west to 4 % on the eastern part of the batholith, then increase to 10 % in the Jurassic rocks of the White Mountains and decreases again in the Cretaceous rocks. The modal abundance percentage of quartz minerals range from 20-30 % across the batholith and the White Mountains, and then rises to 35 % in the eastern Sierra Nevada (Bateman, 1992). Mafic magmatic enclaves (MME) are most common in calc-alkaline granitoids of Sierra Nevada and are abundant in Cordilleran granitoids (Bateman, 1992). Many of the MME in the northern and southern SNB are fine-grained and dark, containing abundant plagioclase grains and scarce, small hornblende which usually includes a finely grained margin (Barbarin, 2005). The composition of the MMEs ranges from diorite to quartz diorite with 45-55% plagioclase and 40-50% mafic minerals.

3.2.1 Geochemistry of the Southeastern Sierra Nevada Batholith

The changes in composition and isotopic ratios eastward across the batholith are mainly caused by the: (1) increase in the thickness of the country-rock crust and a larger crustal component in the parental magmas, (2) increase in the sedimentary component within the crust, and (3) increase in K, U, and Th in the mantle with distance from the subduction zone (Bateman, 1992). The variations in Sr and Nd isotopes in the intrusions indicate that the parental magmas were isotopically heterogeneous, numerous local structures (such as schlieren, troughs, local diapirs and magmatic fabrics characterising flow instabilities and strain heterogeneities at batholith scale), or there was variable assimilation of crustal material in the system as they were being emplaced (Bateman, 1992; Memeti *et al.*, 2010). A common feature of Cordilleran orogens such as the Sierra Nevada Batholith is that the arc rocks are located further from the trench are more radiogenically evolved (Kistler and Peterman, 1973; Farmer and DePaolo, 1983; Chapman *et al.*, 2017). This trend is controlled by crustal assimilation (DePaolo, 1981), and the nature of the mantle: asthenospheric near the trench, and lithospheric away from the trench (Chapman *et al.*, 2017; Chapman and Ducea, 2019). Sr, Nd and O isotopic data show that the plutonic rocks of the Sierra Nevada represent mixed materials from the mantle and the crustal source materials from the sediments of the Mesozoic, Proterozoic crust and mafic Precambrian lower crust (Kistler and Peterman, 1978; DePaolo, 1981; Kistler *et al.*, 1986; Reid and Hamilton, 1987). Samples studied by Kistler and Peterman (1978) and DePaolo (1981a) show a negative correlation between Nd and Sr isotope ratios. The ϵ_{Nd} values of SNB rocks range from +5 to -12 and most plutons have negative values with an average of -6 (Figure 3.3 a) (Ducea, 2001). The spatial distribution of ϵ_{Nd} shows increasingly evolved values towards the east (Figure 3.3 a) (Chapman and Ducea, 2019). The oceanic lithosphere has a constant ϵ_{Nd} of +10, and the SNB exhibits evidence of complex petrogenetic processes shown by variable values of ϵ_{Nd} . The Nd ratios are more related to geographic positioning than to ages (Kistler and Peterman, 1978; DePaolo, 1981). DePaolo (1981) suggested that 50% of the material in the

batholith came from the mantle, and it was added to the continental crust during the Mesozoic. Based on the geochemical data, Ducea (2001) concluded that the second stage of melting, fractionation and generation of residual material must have occurred in the lower lithosphere beneath the arc in a transition zone from the lower crust to upper mantle in a thickened continental crust (as a result of crustal thickening). The resultant residual material could be a restite (i.e., solid left after partial melting and extraction of melt from the deep crust) or cumulate (i.e., from crystal fractionation in the stagnant magma chamber), or both (Ducea, 2001).

In the batholith, K₂O wt% increases eastward, and the SiO₂ wt% shows a little variation (Figure 3.3 b), while the CaO, FeO wt% and MgO wt% decrease at a slower rate, trace elements like U, Th, Rb, Be, Ta, Ba, and the REEs increases eastwards (Bateman, 1992). The initial ⁸⁷Sr/⁸⁶Sr increases eastward along with the oxidation ratio $[mol \frac{(2Fe2O3 \times 100)}{(2Fe2O3+FeO)}]$ (Figure 3.3 c), while the initial ¹⁴³Nd/¹⁴⁴Nd decreases eastward (Bateman, 1992). In the southern region of the batholith, the Precambrian continental margin is marked by a classic initial ⁸⁷Sr/⁸⁶Sr = 0.706 isopleth, which bisects the western and central part of the batholith (Figure 3.4) (Cecil *et al.*, 2012). The initial ⁸⁷Sr/⁸⁶Sr = 0.706 isopleth is a geographic marker of paleogeographic, paleotectonic, and geochemical significance (Kistler and Peterman, 1978; Kistler, 1990). The 0.706 isopleth roughly divides the accreted exotic terrane and the western margin of the Laurentian craton (Hannon and Huff, 2019). The positioning of the initial ⁸⁷Sr/⁸⁶Sr = 0.7060 can be related to the edge of the Precambrian continental crust in the western United States (Kistler and Peterman, 1978; Kistler, 1990; Hannon and Huff, 2019). To the west of the 0.706 boundaries, the batholith is underlain by a predominantly oceanic crust of Phanerozoic age and mafic granitoids characterized by unradiogenic Sr values (low initial ⁸⁷Sr/⁸⁶Sr) (Kistler, 1990; Chin *et al.*, 2012). To the east, is the Precambrian basement (1700 Ma) (Kistler, 1990) and felsic, granitic plutons which interacted extensively with the Precambrian sediments, responsible for their radiogenic Sr values (high initial ⁸⁷Sr/⁸⁶Sr) (Kistler and Ross, 1990; Chen and Tilton, 1991; Chin *et al.*, 2012). Along the eastern margin of the Sierra Nevada, the isopleth is abruptly offset in a right-lateral sense for ~300 km by faulting from southwest trend to a northwest trend (Kistler, 1990). The initial ⁸⁷Sr/⁸⁶Sr values, K₂O content, quartz diorite boundary line, REE patterns and lead isotopic content studied in this batholith indicate that the source region was laterally variable, which applies in many batholiths of similar scale (Chen and Moore, 1982). Three reservoirs, namely the craton-derived sedimentary rocks, young, depleted crust and ancient lower continental crust are thought to have influenced the chemistry of the arc (Lackey *et al.*, 2005).

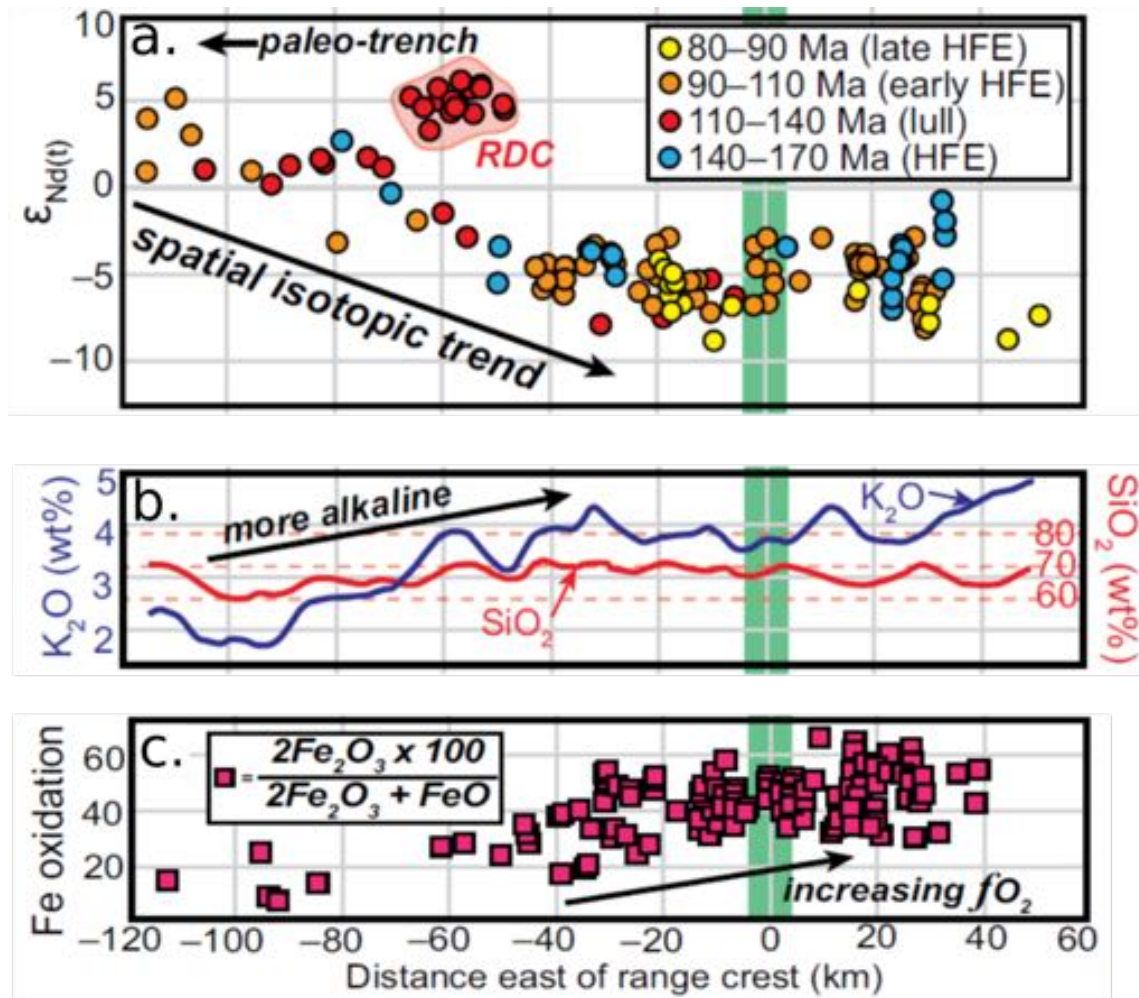


Figure 3.3: The change in ϵ_{Nd} , K_2O wt%, SiO_2 wt% and Fe oxidation across the SNB from the west to the east with the change in distance from the range crest (Chapman and Ducea, 2019). a. Generally the ϵ_{Nd} decreases from the west to the east. b. the K_2O increases from the west to east while the SiO_2 shows little change across the batholith. c. Fe oxidation increases across the batholith from west to east. RDC - mafic-ultra- mafic ring dike complex, HFE - High Flux Event.

3.2.2 Geochronology of the Southeastern Sierra Nevada Batholith

Through many studies of zircon U-Pb geochronology, Memeti *et al.* (2010) and Paterson and Ducea (2015) observed that magmatism in continental arcs results in magmatic provinces of different ages (Figure 3.4). The Sierra Nevada batholith becomes younger, moving eastwards from the coast (Figure 3.5 a) (Chapman and Ducea, 2019). The change in the lithology through the batholith relates to the change in ages (early Cretaceous in the west and late Cretaceous in the east), major and trace element composition, and Sr, Nd and Pb radiogenic isotope signature. These changes are characterized by the flare-up and migration of magmatism across the heterogeneous crustal and upper mantle domains (Cecil *et al.*, 2012). The Sierra Nevada had three high flux event during the Mesozoic, around 225 Ma, 160

Ma and 95 Ma (Ducea, 2001; Cecil *et al.*, 2012; Paterson and Ducea, 2015; Chapman and Ducea, 2019). U-Pb zircon ages of plutons in the northern Sierra Nevada show the presence of a late Jurassic arc (167 and 145 Ma) and a Mid-Cretaceous arc (125 and 90 Ma) (Cecil *et al.*, 2012). Eocene-Miocene volcanic rocks and clastic sedimentary are preserved in the northern Sierra Nevada (Cecil *et al.*, 2012). In the White and Inyo Mountains, in the eastern part of the batholith contains Jurassic rock types which include monzonite, quartz monzonite, granodiorite, and granite, and the rest of the outcrops are Cretaceous granite rocks (Bateman, 1992). In the central Sierra Nevada batholith, the oldest rocks (ca. 120 Ma) occur in the west-central part with a regular trend towards the eastern crest having younger ages ca. 80 Ma (Chen and Tilton, 1991). Triassic dated plutons are exposed on the east part of the batholith, which shows that arc magmatism began ca. 250 Ma near the Permian-Triassic boundary (Martin and Clemens-Knott, 2015). Large influxes of magma came in the Late Jurassic (165-145 Ma) and Late Cretaceous (100- 85 Ma) (indicated by the yellow band HFE (High Flux Event) in Figure 3.5) resulting in a thick Mesozoic crustal column (Memeti *et al.*, 2010; Cecil *et al.*, 2012).

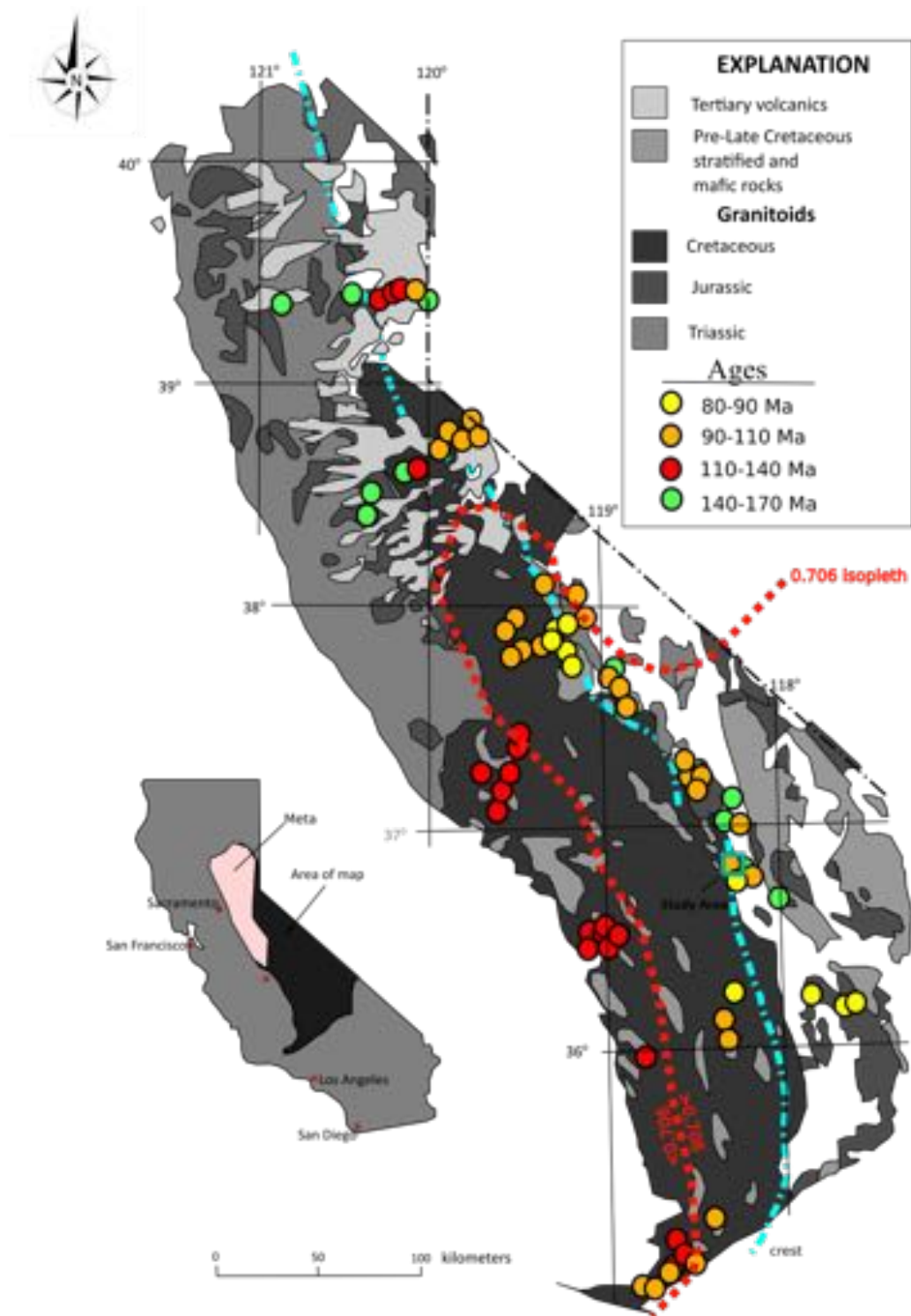


Figure 3.4: Map of the Sierra Nevada Batholith showing general age distribution across the batholith and the 0.706 isopleths (map after Snelling and Gates, (2009), isopleth and range crest from Cecil et al. (2012)), age data extracted from Kistler et al. (1986); Saleeby et al. (1987, 2007); Clemens-Knott (1992); Coleman et al. (1992); Olson (1992); Bradford (1995); Cousens (1996); Sisson et al. (1996); Truschel (1996); Ratajeski et al. (2001); Kylander-Clark et al. (2005); Glazner et al. (2008); Gray et al. (2008); Cecil et al. (2012); Nelson et al. (2013); and Chapman and Ducea (2019). The red dashed line showing the 0.706 isopleth and the blue dashed line showing the crest.

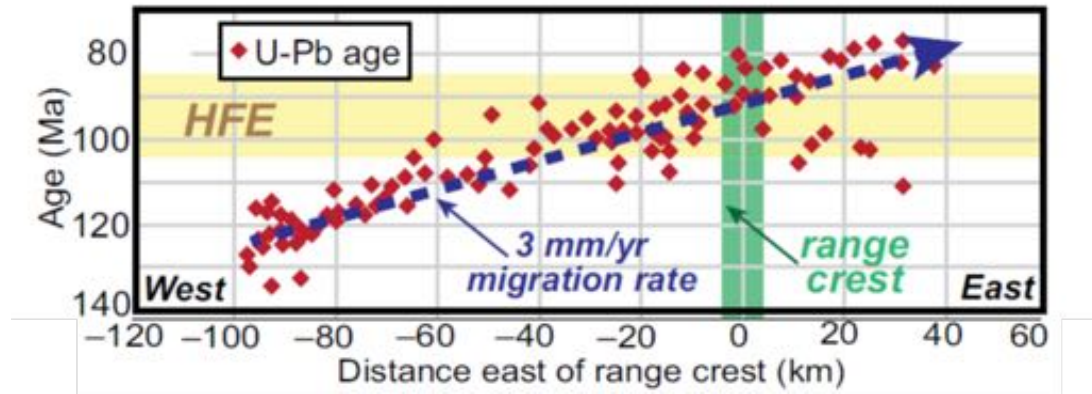


Figure 3.5: The change of ages of the Sierra Nevada Batholith with distance away from the range crest. The SNB shows a decrease in age (Ma) from the west to the east (Chapman and Ducea, 2019).

The southern SNB has more than 20 plutons with compositions ranging from gabbro to leucogranite and ages ranging from 165 to 92 Ma. The rocks along the eastern side (south of the Mount Whitney Pluton) are made up of four significant units (1) the Jurassic Sacatar quartz diorite, (2) Summit gabbro and (3) the Cretaceous granite of Owens Peak, and (4) Morris Peak's granodiorite (Bartley *et al.*, 2007). The McDoogie Pluton is a northwest-trending sub vertically sheeted complex with dimensions of approximately 12 km long and 2 km in width. It is located on the eastern margin of the Sierra Nevada batholith, and it was emplaced as sub vertically sheeted granitoids complex into the steep reverse-sense Sawmill Lake shear zone by Independence dyke swarms (Figure 3.6) (Mahan *et al.*, 2003). Field and geochronology evidence support this observation. The Independence dyke swarm extends for 500 km from the central Sierra Nevada batholith, and it has been dated to be 148 Ma (Moore and Hopson, 1961; Chen and Moore, 1982; Hourigan *et al.*, 2003; Hopson *et al.*, 2008) and they extensively intrude pre-Cretaceous rocks (Mahan *et al.*, 2003). The swarms intruded during a brief interval at 148 Ma, and they serve as an important temporal and structural marker in the plutons (Carl and Glazner, 2002).

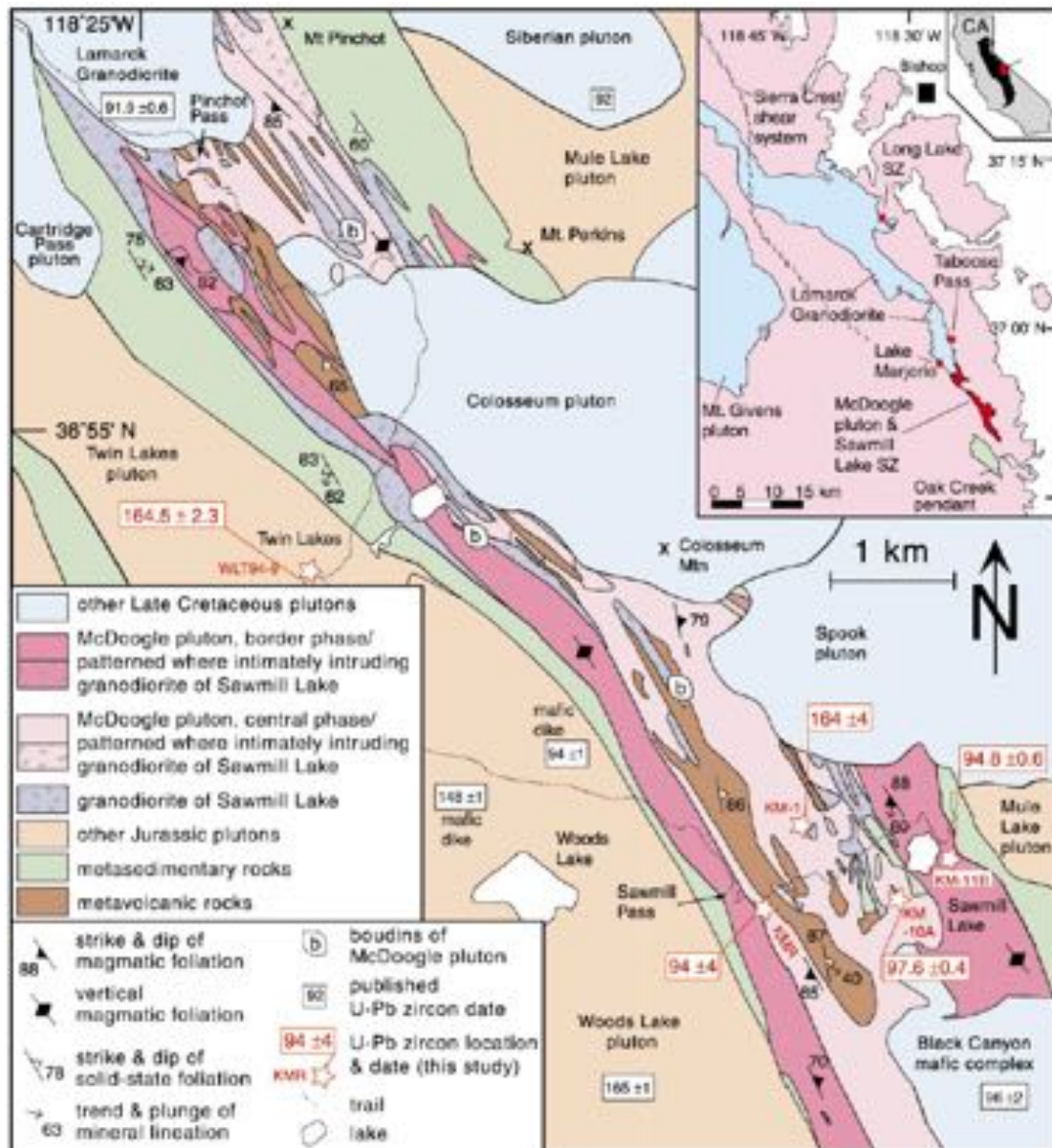


Figure 3.6: Geological map of the southeastern Sierra Nevada Batholith with ages of the plutons (Mahan *et al.*, 2003).

Figure 3.7 shows the age range of plutons in the southern Sierra Nevada Batholith showing (Chen and Moore, 1982; Carl, 2000; Coleman *et al.*, 2000; Mahan *et al.*, 2003; Mahaffey, 2016). The oldest rocks in the southern SNB are the Wood lake granodiorite orthogneiss on the northwest of the McDoogie Pluton which yielded U-Pb zircon ages of 165 ± 1 Ma (Figure 3.8) (Chen and Moore, 1982). Then the granodiorite in the wall rocks (which consist largely of metasedimentary and metavolcanic rocks of probable early to middle Paleozoic and Jurassic to Cretaceous ages) near the Twin Lakes Pluton yielded an age of $164 \pm 2.1/-6.8$ Ma been interpreted to be the crystallisation age of the pluton, which has an age of 164.5 ± 2.3 Ma. (Mahan *et al.*, 2003). The Independence mafic dyke swarms have been suggested to have a Cretaceous age, but the U-Pb zircon of felsic dykes yielded an age of 148 Ma (Moore and

Hopson, 1961; Chen and Moore, 1982). The felsic dykes have been correlated with the Independence dyke swarms (Bartley *et al.*, 2007). The McDoogie Pluton is hosted within the Sawmill Lake shear zone, which is bracketed between 148 Ma, which is the Independence dyke swarm age (Mahan *et al.*, 2003). The Lamarck granodiorites (Late Cretaceous) are exposed between the Independence dyke swarms and the Bishop Pluton (Figure 3.8) (Bateman, 1992; Coleman *et al.*, 1995). The McDoogie Pluton intrudes Jurassic plutons, and it is crosscut by Late Cretaceous plutons with the central quartz monzodiorite phase dated at 94 ± 4 Ma (U-Pb titanite geochronology), the border mafic granodiorite phase (94.8 ± 0.6 Ma), and the hornblende granodiorite are considered to be older yielding 97.6 ± 0.4 Ma (Mahan *et al.*, 2003). The mafic granodiorite from the Black Canyon complex intruded by the McDoogie Pluton yielded a discordant U-Pb zircon age of 96 ± 2 Ma from a diorite sample (Carl, 2000). Mule Lake Pluton dykes densely intrude outcrops of the McDoogie Pluton and they have been thought to be Jurassic (Mahan *et al.*, 2003). The granite of Owens Peak yielded 91.7 ± 1.4 Ma and Morris Peak's granodiorite yielded 93.2 ± 1.9 Ma (Bartley *et al.*, 2007). Lamarck Pluton (91.9 ± 0.6 Ma: Coleman *et al.*, (2000)) and the Colosseum Pluton and the Spook Pluton cross-cuts the shear zone and the McDoogie Pluton (Moore, 1963; Mahan *et al.*, 2003). The ages of these rocks dated around Cretaceous period are typical of the eastern Sierra Nevada.

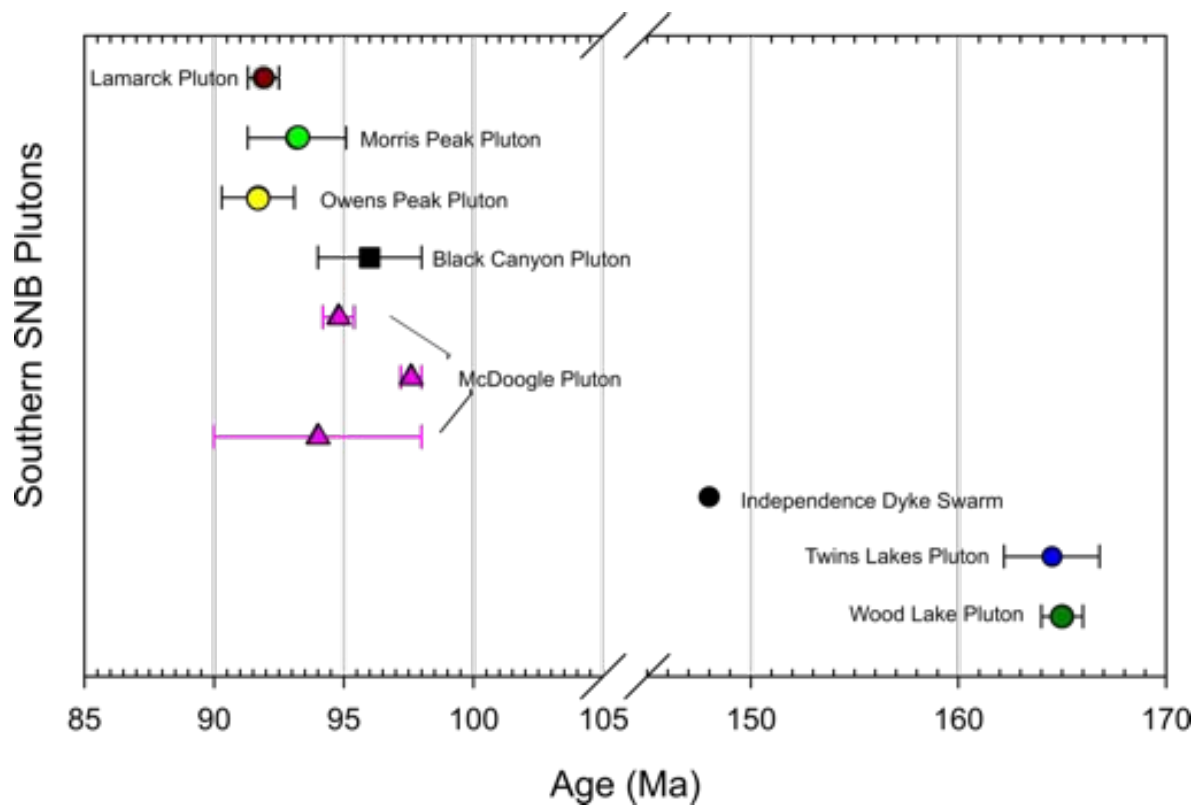


Figure 3.7: Timeline showing the ages of the Southern Sierra Nevada Batholith. The published U-Pb ages are from (Mahan et al. (2003): McDoogle Pluton, Sawmill Lake Pluton), (Carl (2000): Black Canyon Plutonic rock), (Chen and Moore (1982); Mahaffey (2016): Tinehama granodiorite, Wood Lake), (Coleman et al. (2000): Lamarck granodiorite).

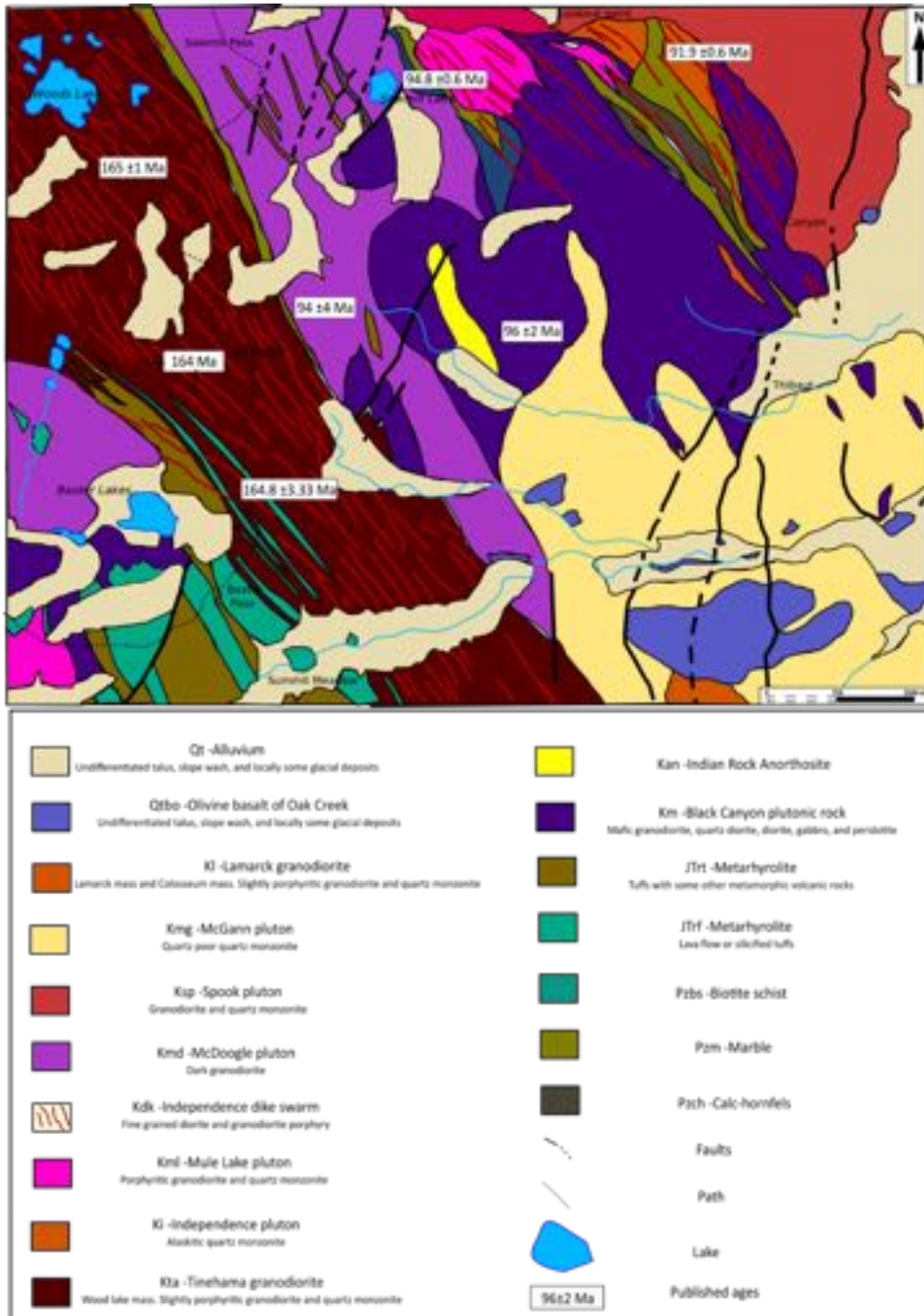


Figure 3.8: Location of the study site and the Indian Rock Anorthosite. Black Canyon Plutonic rocks labelled Km (purple) surround the IRA (Kan-yellow). Km is made up of mafic granodiorite, quartz diorite, diorite, gabbro, and peridotite (modified from Moore, 1963). The published U-Pb ages are from Mahan et al. (2003): McDoyle Pluton, Sawmill Lake Pluton), (Carl (2000): Black Canyon Plutonic rock), (Chen and Moore (1982); Mahaffey (2016): Tinehama granodiorite, Wood Lake), (Coleman et al. (2000): Lamarck granodiorite).

4. Sampling and Analytical methods

This work will provide an insight into how the IRA formed in a Phanerozoic terrane, even though the occurrence of discrete coeval and comagmatic anorthosite bodies in arc terranes is rare. A combination of petrography, textural analysis, whole rock, as well as mineral composition data (both for major and trace elements and radiogenic isotopes) and U-Pb geochronology will be used to achieve the aforementioned goals of this work.

4.1 Field sampling and petrography

Samples were collected during two field excursions to the eastern Sierra Nevada scarp in October 2017 and September 2018. The focus of sampling in the southeastern Sierra Nevada Batholith was to collect samples of the IRA itself, as well as samples of the surrounding plutons (including the host granodiorite), and dykes crosscutting the plutons. The area is characterised by steep terrane, and the anorthosite is situated on the summit of Indian Rock peak. Access to the anorthosite was therefore challenging, and the team was not able to access *in-situ* outcrop on these two excursions due to safety concerns. The anorthositic samples were consequently collected from outwash fans of the Thibaut Creek (Figure 4.1). Given the unique nature of the anorthosite body in the region and the single tributary draining its outcrop area, these samples must be directly derived from the mapped IRA outcrop. The granodiorites (IRA6-3, IRA18-1, and IRA18-2) and felsic dykes (granitic-alaskite; IRA7-4, and IRA18-3) were collected *in situ* from accessible outcrops surrounding the anorthosite. Twelve samples were chosen for detailed petrographic study and ten samples for geochemical analyses (Table 4.1).



Figure 4.1: Google Earth image showing sample locations (green points) in Thibaut Creek (blue dashed line) and the location of the Indian Rock Anorthosite (IRA) (Google Earth, earth.google.com).

Thin sections were prepared and studied using petrographic microscopy at the University of the Witwatersrand, and most photomicrographs were produced with a full thin section scanning facility connected to a BX63 microscope. These provide an assessment of the textural features and mineral assemblages of the lithologies in the area. This method is used to test whether there is a textural and compositional relationship between the anorthosite and the surrounding rocks and help to constrain the fractionation patterns that occurred during its emplacement. Homogenous areas that typified each sample, and which showed interesting textures, were selected for the sectioning.

Table 4.1: A sample list with rock names and analysis carried out on each one.

Sample ID	Lithology Field name	TS	ICPMS	XRF	EPMA	ID/MC-ICPMS	LA-ICPMS	Final Rock name
IRA3-1	Anorthosite	√	√	√	-	√	-	Anorthosite
IRA4-2	Anorthosite	√	√	√	-	-	√	Anorthosite
IRA8-5	Anorthosite	√	√	√	-	√	√	Anorthosite
IRA18-4a	Anorthosite with the mafic interstitial vein	√	-	-	√	-	-	Anorthosite
IRA18-4b	Anorthosite with the mafic interstitial vein	√	-	-	√	-	-	Anorthosite
IRA18-5	Anorthosite	√	√	√	√	√	√	Anorthosite
IRA18-6	Anorthosite	-	-	-	-	-	√	Anorthosite
IRA18-7	Anorthosite	√	√	√	√	√	√	Anorthosite
IRA6-3	McDoogie Pluton (granodiorite)	√	√	√		√	-	Granodiorite
IRA18-1	McDoogie Pluton (granodiorite)	√	√	√	-	√	-	Granodiorite
IRA18-2	Black Canyon Pluton (granodiorite)	√	√	√	-	√	√	Granodiorite
IRA7-4	Felsic dyke (Alaskite)	√	√	√	-	-	√	Granitic-alaskite
IRA18-3	Felsic dyke (Alaskite)	√	√	√	-	-	-	Granitic-alaskite

4.2 Whole-rock major and trace elements analysis

The whole-rock geochemical analyses (XRF and ICPMS) will, together with textural and mineral assemblage analysis, allow us to assess the magmatic evolution/processes that operated during the formation of the anorthosite and surrounding rocks. Ten samples were selected for whole-rock major and trace elements analyses (Table 4.1). The samples were crushed using a carbide steel jaw crusher and milled to less than 50 µm powder using a carbide steel swing mill at the School of Geosciences, University of the Witwatersrand. Major elements analyses included SiO₂, TiO₂, Al₂O₃, Fe₂O₃, FeO, MnO, MgO, CaO, Na₂O, K₂O, P₂O₅ and Loss on Ignition (LOI), while Li, P, Sc, Ti, V, Cr, Co, Ni, Cu, Zn, Ga, Rb, Sr, Y, Zr, Nb, Ba, Sn, Cs, La, Ce, Pr, Nd, Pm, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, Lu, Hf, Ta, W, Pb, Th, and U were analysed as trace elements. Major elements analyses have values that agree within 5% of the published reference standard value (USGS, 2000) with the exception of P₂O₅ and NiO. Trace elements analyses has values that agree within 20% of the reported standard values. Major element and trace element data appears in Section 7.2 where it is used to understand petrogenetic

link between the IRA and SNB and Section 7.3 to classify and compare the IRA with other types of anorthosites.

4.2.1 XRF for whole-rock major elements

Major elements were used to characterise the IRA and surrounding rocks and to contribute to testing whether the IRA was comagmatic with surrounding granitoids and derived from the same liquid line of descent. The whole-rock major element data were obtained using the PANalytical PW2404 spectrometer XRF machine from the Earth Lab, University of the Witwatersrand. The standards used were BHVO-2 and BCR-2 (Figure 4.2). The values from this work agree within 5% of the reported values of the reference standard from USGS (2000), except for P_2O_5 and NiO. The standards data and detailed methodology are listed in Appendix B.

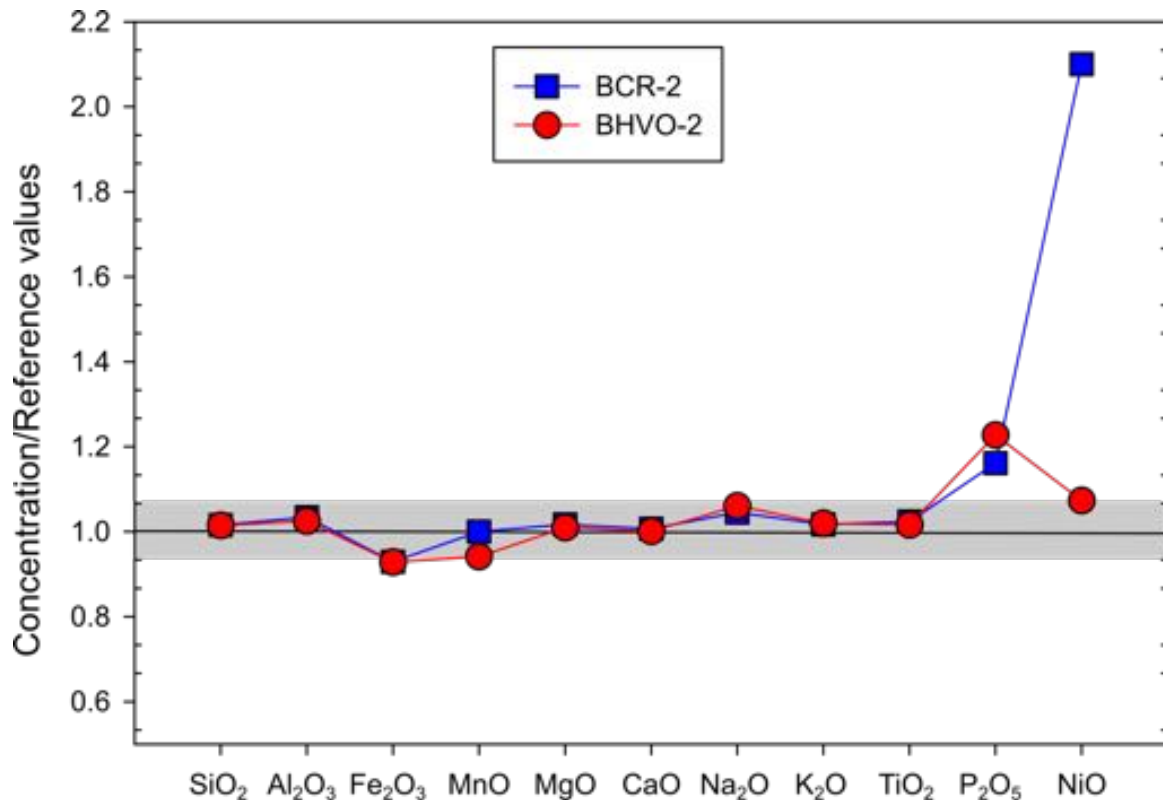


Figure 4.2: Comparison of the BCR-2 and BHVO-2 standards from this study and the reference values from the USGS (2000). Most of the data agree within 5% (shaded band) except for P_2O_5 and NiO.

4.2.2 ICP-MS for whole-rock trace elements

Trace element data were used to assess magmatic processes (fractionation process) that operated during the formation and emplacement of the anorthosite. Trace element data were used to plot spider diagrams and chondrite normalised rare earth element (REE) plots to understand the genetic link and geochemical variation of the anorthosite and the surrounding rocks (granodiorite). Whole-rock trace element data were obtained using a Thermo-Fisher iCAP RQ ICPMS instrument at the EarthLab, University of the Witwatersrand. The standards used were BHVO-2 and BCR-2 (Figure 4.3). Most of the values from this work agree within 5% of the reported values for the reference standard from USGS (2000), and the rest can be fitted within 20%. The standards data and detailed methodology are listed in Appendix B.

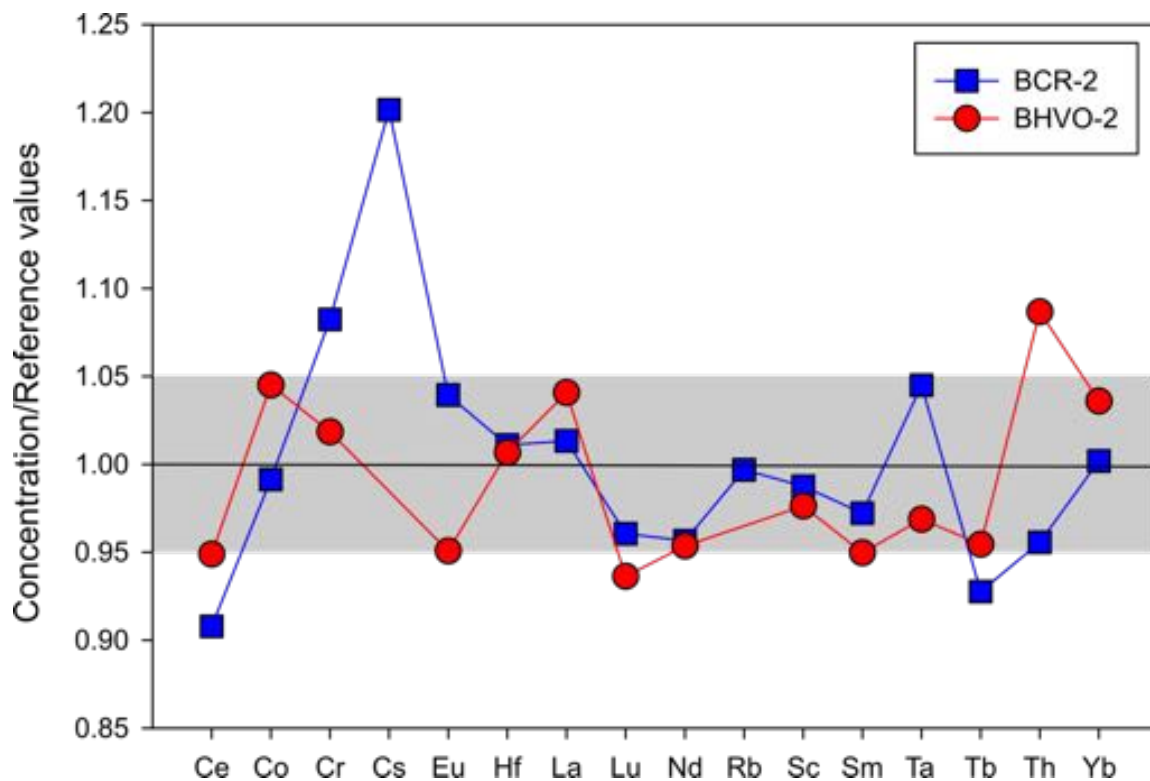


Figure 4.3: Comparison of the BCR-2 and BHVO-2 standards from this study and the reference values from the USGS (2000). Most of the data agree within 5% (shaded band), and the rest fall within 20%.

4.3 *In-situ* Mineral Chemistry

4.3.1 EPMA (major mineral assemblages)

This *in-situ* method was used to analyse the mineral chemistry of the major rock-forming mineral in the IRA (e.g., plagioclase) and importantly the An-content of the plagioclase crystals. *In-situ* major element mineral compositions for SiO₂, TiO₂, Al₂O₃, FeO*, MnO, MgO, CaO, K₂O, Cr₂O₃ and BaO in plagioclase were measured with a beam current of 20 nA, 15 kV accelerating voltage, and a 10 μm

beam diameter. For pyroxenes, SiO₂, TiO₂, Al₂O₃, FeO*, MnO, MgO, CaO, K₂O, Cr₂O₃, NiO and ZnO were measured using beam current of 20 nA, accelerating voltage of 20 kV, and a <1 µm beam diameter. The microprobe analyses were used to analyse *in situ* major elements of rock-forming minerals from the core to the rim of selected mineral grains, using a CAMECA SX5-FE EPMA instrument housed in the Microscopy and Microanalysis Unit (MMU), University of the Witwatersrand. The plagioclase major mineral data appears in Section 7.3 where it is used to classify and compare IRA with other types anorthosites.

4.3.2 LA-ICPMS U-Pb Zircon and Apatite Geochronology

Zircon U-Pb ages were used to determine the crystallization age of the samples. Apatite U-Pb dating was used to calculate cooling ages through a closure temperature of 350 - 550 °C (Cochrane *et al.*, 2012; Gawęda *et al.*, 2014; Hall *et al.*, 2018). Apatite, with the general formula Ca₅(PO₄)₃(F, Cl, OH) (Kirkland *et al.*, 2018), occurs mostly as an accessory phase in igneous rocks due to the low solubility of P₂O₅ in silicate melts and the limited amount of phosphorus incorporated into the crystal lattices of major rock-forming minerals (Gawęda *et al.*, 2014). This means that the apatite U-Pb exhibits a low closure temperature; therefore the ages cannot be interpreted in terms of dating magmatic activities, but it gives valuable information on the timing of thermal events and cooling history of crystalline basement terranes (Gawęda *et al.*, 2014; Kirkland *et al.*, 2018). This method can be used as a thermochronometer of minerals to date mafic rocks (Pochon *et al.*, 2016; Hall *et al.*, 2018). See Appendix E for detailed sample preparation methodology. Zircon and apatite geochronology appears in Section 7.1 where it is used to understand the age relationship between the IRA and SNB.

4.3.2.1 Zircon Geochronology

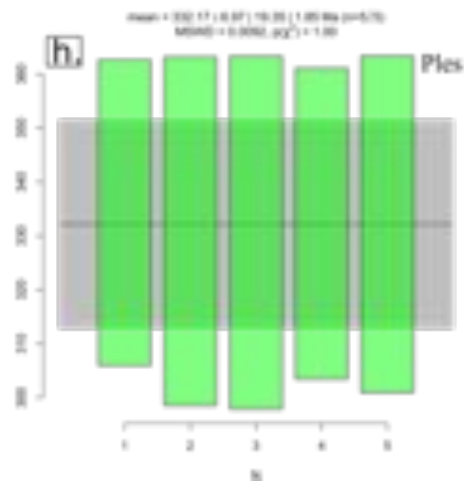
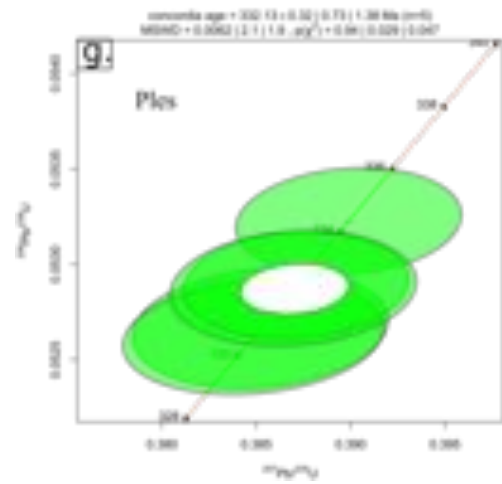
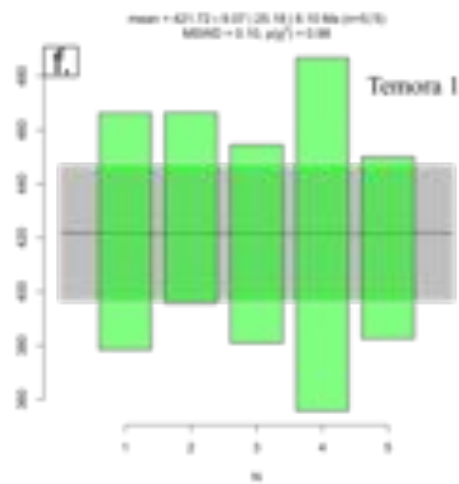
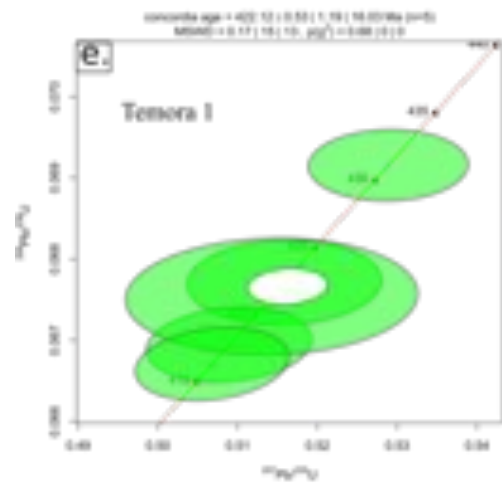
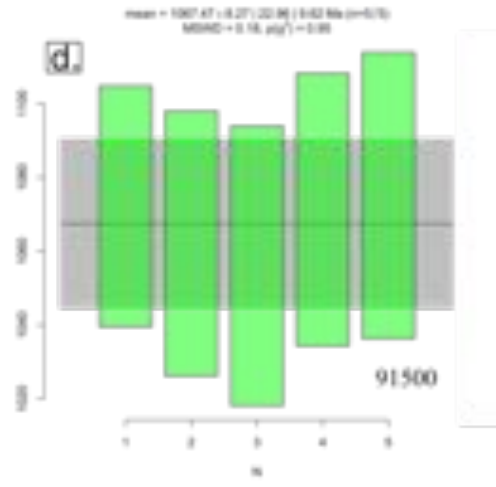
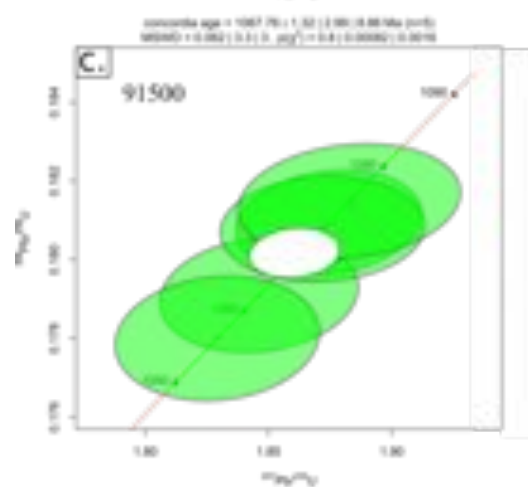
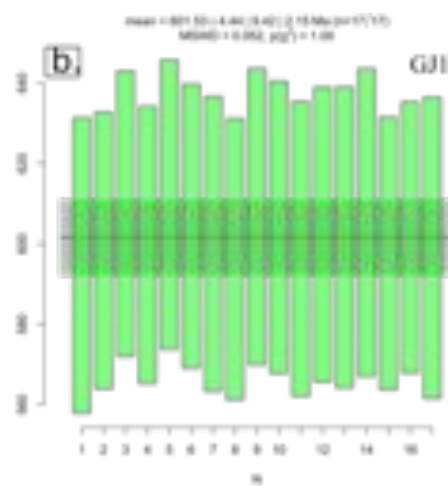
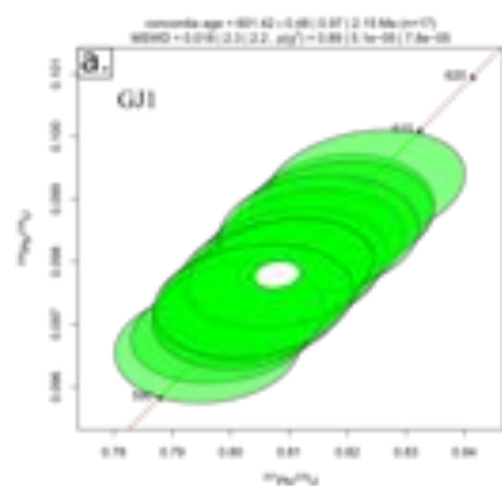
The zircon U-Pb ages were analysed using a single 30-µm diameter spot on apatite grains mounted in resin. Before the analyses, the grains were imaged on an SEM (scanning electron microscope) under a CL (cathodoluminescence) and BSE (backscattered electron) source at the University of Johannesburg. The laser was set at 4 mJ with a frequency of 2 Hz and an attenuator value of 25%. The resulting fluence was 1.2 Jcm⁻². Measurement was made over 70 s with a background measurement of 25 s. The instrument used for these analyses was a 193 nm ArF RESolution SE excimer laser-ICPMS (Australian Scientific Instruments, Fyshwick) at the University of the Witwatersrand. The standards used were GJ1 (as the primary standard), 91500, Temora-2, Plešovice zircon, and BB zircon (Table 4.2). The GJ1 has been dated to be ²⁰⁷Pb/²⁰⁶Pb = 609 ± 1 Ma (Belousova *et al.*, 2006), 91500 (²⁰⁷Pb/²⁰⁶Pb = 1065 ± 1 Ma (Wiedenbeck *et al.*, 1995)), Temora (U²³⁸/Pb²⁰⁶ = 416.17 ± 0.24 Ma (Black *et al.*, 2003)), Plešovice

zircon ($U^{238}/Pb^{206} = 337.13 \pm 0.37$ Ma (Sláma and Košler, 2007) and BB ($^{207}Pb/^{206}Pb = 564 \pm 33$ Ma (Santos *et al.*, 2017)). The standards data are listed in Appendix E.

Table 4.2: Summarised age data of the zircon standards from the LA-ICPMS

STD	U-Pb age (Ma)	MSWD	Weighed mean Pb/Pb age (Ma)	MSWD	<i>n</i>	Published values
GJ1	601.42 ± 0.48	0.018	601.53 ± 4.44	0.052	17	Pb/Pb= 609 ± 1 Ma; Belousova <i>et al.</i> , 2006
91500	1067.76 ± 1.32	0.062	1067.47 ± 8.27	0.18	5	Pb/Pb= 1065 ± 1 Ma; Wiedenbeck <i>et al.</i> , 1995
Temora	422.12 ± 0.53	0.17	421.72 ± 9.07	0.1	5	U/Pb= 416.17 ± 0.24 Ma; Black <i>et al.</i> , 2003
Plešovice	332.13 ± 0.32	0.0062	332.17 ± 6.97	0.0092	5	U/Pb= 337.13 ± 0.37 Ma; Sláma and Košler, 2007
BB	534.13 ± 0.59	0.012	535.27 ± 8.57	0.35	5	Pb/Pb= 564 ± 33 Ma; Santos <i>et al.</i> , 2017

In this study, the $^{207}\text{Pb}/^{206}\text{Pb}$ ages obtained for GJ1 = 601.53 ± 4.44 Ma (Figure 4.4 a, b), 91500 = 1067.47 ± 8.27 Ma (Figure 4.4 c, d), Temora = 421.72 ± 9.07 Ma (Figure 4.4 e, f), Plešovice zircon = 332.17 ± 6.97 Ma (Figure 4.4 g, h), and BB = 535.27 ± 8.57 Ma (Figure 3.1 i, j). The values from this study fall within the error of the published literature values for these standards (Table 4.2 and Figure 4.2). Ninety-seven spots were measured on the samples (IRA 4-2; *n* = 53; IRA 18-6; *n* = 44) with five spots of 91500, Temora-1, Plešovice zircon, BB zircon each and 17 spots on GJ1. Weighed mean $^{207}\text{Pb}/^{206}\text{Pb}$ and Concordia ages were calculated using the IsoplotR open-source software (Vermeesch, 2018). The age outputs from the programme are reported as $t \pm x|y|z$ (t = age, x = standard error, y = (95%) confidence interval and z = overdispersion) (Vermeesch, 2018). The data reduction from the LA-ICPMS was done using Iolite software (Paton *et al.*, 2011). Stacey-Kramers common Pb correction was used when plotting the Wetherill Concordia plots and producing weighted mean ages.



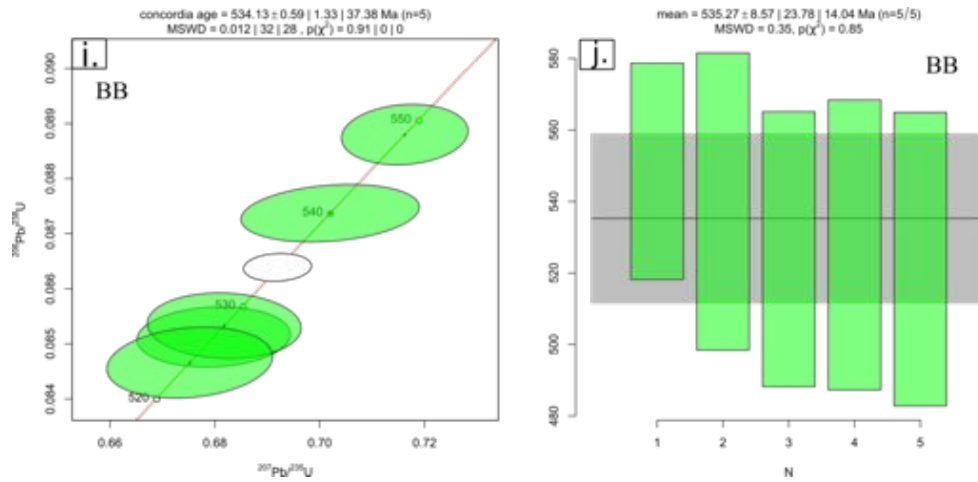


Figure 4.4: Wetherill Concordia plots and weighed mean Pb/Pb ages produced for standards are a. b. GJ1 ($^{207}\text{Pb}/^{206}\text{Pb} = 601.53 \pm 4.44 \text{ Ma}$); c. d. 91500 ($^{207}\text{Pb}/^{206}\text{Pb} = 1067.47 \pm 8.27 \text{ Ma}$); e. f. Temora ($^{207}\text{Pb}/^{206}\text{Pb} = 421.72 \pm 9.07 \text{ Ma}$); g. h. Plešovice zircon ($^{207}\text{Pb}/^{206}\text{Pb} = 332.17 \pm 6.97 \text{ Ma}$) and i. j. BB ($^{207}\text{Pb}/^{206}\text{Pb} = 535.27 \pm 8.57 \text{ Ma}$).

4.3.2.2 Apatite Geochronology

The apatite U-Pb ages were analysed using a single 30 μm diameter spot on zircon grains mounted on a resin. Before the analyses, the grains were imaged using a SEM under a CL and BSE source at the University of Johannesburg. The cores and rims of apatite grains were ablated for 30 s using 30 μm spot at 2 J/cm², after a 30 s laser warm-up and a 20 s dwell time between samples. The ^{238}U , ^{208}Pb , and ^{207}Pb masses were measured. The standards used were Madagascar apatite (as primary standard), Durango apatite and Mt McClure apatite. Madagascar apatite has been dated to be $474.25 \pm 0.41 \text{ Ma}$ (Thomson *et al.*, 2012), Durango apatite ($31.44 \pm 0.18 \text{ Ma}$ (Mcdowell *et al.*, 2005)), and Mt McClure apatite ($518 \pm 11 \text{ Ma}$ (Thomson *et al.*, 2012)) (Table 4.3). The standards data are listed in Appendix E.

Table 4.3: Summarized age data of the apatite standards from the LA-ICPMS.

STD	U-Pb age (Ma)	MSWD	Weighed mean Pb/Pb age (Ma)	MSWD	<i>n</i>	Published values
Madagascar apatite	473.25 ± 0.82	0.019	470.02 ± 0.87	5.62	18	$474.25 \pm 0.41 \text{ Ma}$; Thomson <i>et al.</i> , 2012
Durango apatite	32.38 ± 0.11	0	32.38 ± 0.11	3.81	18	$31.44 \pm 0.18 \text{ Ma}$; Mcdowell <i>et al.</i> , 2005
McClure apatite	523.73 ± 1.12	0.0095	523.77 ± 1.12	11.7	11	$518 \pm 11 \text{ Ma}$, Thomson <i>et al.</i> , 2012

In this study, the U-Pb ages obtained for Madagascar apatite = $473.25 \pm 0.82 \text{ Ma}$, Durango apatite = $32.38 \pm 0.11 \text{ Ma}$, and Mt McClure apatite = $523.73 \pm 1.12 \text{ Ma}$ (Figure 4.5). All data reduction was processed using Iolite software (Paton *et al.*, 2011). This was done using data reduction VizualAge UcomPbine method outlined by Chew *et al.* (2011, 2014). Stacey-Kramers common Pb correction (Chew *et al.*, 2014) was used when plotting the Tera–Wasserburg Concordia plots and weighted mean

ages (Figure 4.5). The values from this study fall within the error of the published values in the literature for the standards (Table 4.3).

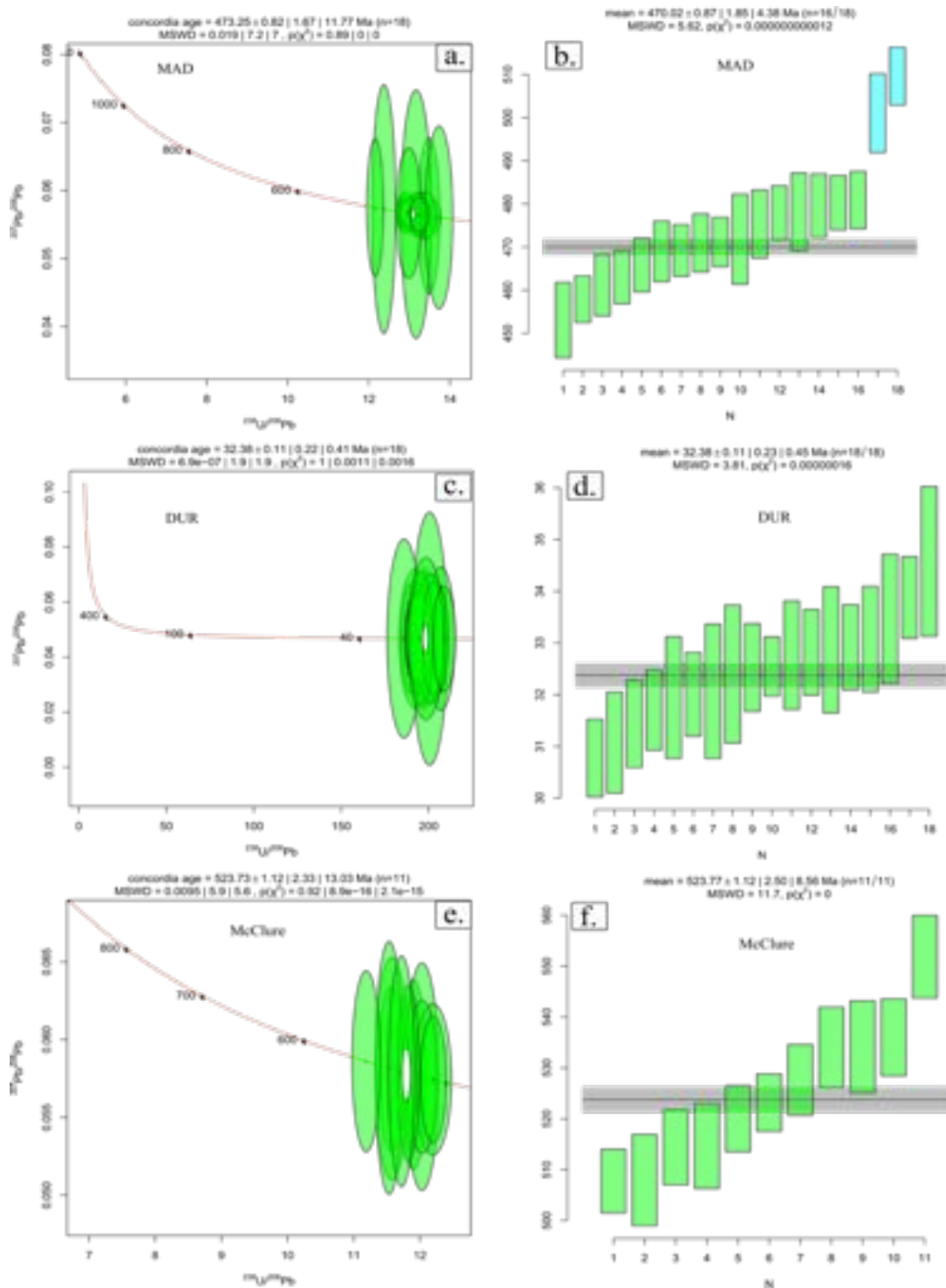


Figure 4.5: Tera-Wasserburg Concordia plots and weighed mean Pb/Pb ages of the apatite standards. The blue bar was excluded from the calculation. a. b. Madagascar apatite (473.25 ± 0.82 Ma); c. d. Durango apatite (32.38 ± 0.11 Ma); and e. f. Mt McClure apatite (523.73 ± 1.12 Ma).

4.3.3 LA-MC-ICPMS Lu-Hf Isotopic Analysis in Zircon

The Lu-Hf isotopes in zircon were analysed using a single 70 µm diameter spot on zircon grains mounted in resin. Before the analyses, images of the grains were taken using SEM using a BSE and CL source at the University of Johannesburg. The laser was set at 6 mJ with a frequency of 7 Hz and an attenuator value of 50%. The resulting fluence was 5.6 J/cm². Measurement was made over 70 s with a background measurement of 25 s. The instrument used for these analyses was a 193 nm ArF RESOLUTION SE excimer laser (Australian Scientific Instruments, Fyshwick), Nu Plasma II MC-ICPMS (Nu Instruments, Wrexham) with 16 Faraday detectors, and five ion-counting detectors at the University of Johannesburg. The detector array was as follows: H4 = ¹⁸⁰Hf; H3 = ¹⁷⁹Hf; H2 = ¹⁷⁸Hf; H1 = ¹⁷⁷Hf; Ax = ¹⁷⁶Hf + ¹⁷⁶Yb + ¹⁷⁶Lu; L1 = ¹⁷⁵Lu; L2 = ¹⁷⁴Hf + ¹⁷⁴Yb; L3 = ¹⁷³Yb; L4 = ¹⁷²Yb; and L5 = ¹⁷¹Yb. For mass bias correction, ¹⁷¹Yb and ¹⁷³Yb were used to calculate Yb mass fractionation and to correct Yb and Lu mass fractionation; ¹⁷⁹Hf and ¹⁷⁷Hf were used to calculate Hf mass fractionation. For interference corrections: ¹⁷¹Yb was used to calculate and correct ¹⁷⁴Yb interference on ¹⁷⁴Hf and the ¹⁷⁶Yb interference on ¹⁷⁶Hf; ¹⁷⁵Lu was used to correct the ¹⁷⁶Lu interference on ¹⁷⁶Hf. Background correction is done by measuring background signal on mass before ablation and subtracting from the signal. The standards used were all-natural zircons known as Mud Tank zircon, LV11 zircon and Plešovice zircon (Table 4.4). The published values for Mud Tank zircon Lu-Hf isotopes is 0.282507 ± 6 (Woodhead and Hergt, 2005), LV11 zircon is 0.282830 ± 28 (Heinonen *et al.*, 2010) and Plešovice zircon is 0.282476 ± 11 (Cecil *et al.*, 2011). For this study, the values obtained for Mud Tank zircon (0.282492, n=3), LV11 zircon (0.282832, n=3), and Plešovice zircon (0.282455, n=7).

Table 4.4: Summarised Lu-Hf isotope data for zircon standards.

STD	Hf-Hf	2SE	n	Published values
Mud Tank zircon	0.282492	0.000011	3	0.282507 ± 6; Woodhead and Hergt, 2005
LV11 zircon	0.282832	0.000018	3	0.282830 ± 28; Heinonen <i>et al.</i> , 2010
Plešovice zircon	0.282455	0.000010	7	0.282476 ± 11; Cecil <i>et al.</i> , 2011

4.4 Whole-rock Rb-Sr and Sm-Nd Isotope Geochemistry

Isotope dilution and ion-exchange chromatography were used to separate elements of interest (Sm-Nd, and Rb-Sr) (Tessalina *et al.*, 2014). All the sample preparation and purification experiments were carried out in the Wits Isotope Geosciences Lab (WIGL) at the University of the Witwatersrand, and isotopic analysis on the multi-collector inductively coupled plasma mass spectrometer (MC-ICPMS) at the University of Johannesburg. Seven samples were analysed to determine Sr and Nd isotopic compositions. The samples shown in Table 4.5 were weighed (in mg) according to different concentrations of trace element in ppm in Table 6.3. The Rb-Sr and Sm-Nd isotope data appears in

Section 7.2 were they are used to understand the petrogenetic link of the IRA and the SNB and Section 7.4 they are used to understand the formation of the IRA.

Sample digestion

Whole-rock powders were weighed and decanted into a 15 mL Savillex beaker along with weighed mixed Rb-Sr and Sm-Nd spikes. 3 ml of HF and 1.5 ml HNO₃ were added to the sample-spike mixture and put on the hot plate for 18 days to break down silicates. The beakers were opened and left on the hot plate to dry down. Subsequently, 2 ml HNO₃ were added to the beakers and left on the hot plate for two days. After two days the beakers were opened to dry after the addition of 3 ml HNO₃ and 1 ml HCl, then capped and stored on the hot plate. Seven days later, the samples were dried down and stored in a 3 ml 6 M HCl solution on the hot plate. After those steps, the samples were ready for column chemistry.

Purification of Sr and Nd

For Rb-Sr and LREE separation, the columns were packed with an AG50W-X12 resin and they were conditioned with three rounds of 2 ml 1.5 M HCl followed by addition of 1 ml of sample in 1.5 M HCl to the columns. Two rounds of 1 ml were passed through the columns to rinse the column, and 4 ml 1.5 M HCl was used for matrix elution. For Rb elution, 4 ml 1.5 M HCl was passed through the columns and collected in Savillex beakers, followed by 10 ml 1.5 M HCl for Ca clean up. Sr elution with 15 ml 1.5 M HCl was collected in Savillex beakers. To get rid of the HREEs, 2 ml 6 M HCl was discarded, and the LREE elution was collected after passing 7 ml 6 M HCl. All the collected elutions were dried down, and 2-3 drops of HNO₃ was added to dissolve any trace of resin residue. For Sm-Nd separation, the LREE elute was used. Three rounds of 2 ml 0.17 M HCl were used for column conditioning which were packed with an Ln resin before loading the samples in 0.3 ml 0.17 M HCl. The LN resin is composed of polymeric beads coated with the organic extractant di-(2-ethylhexyl)-phosphoric acid (HDEHP) and is commonly used in REE mining applications (Huang *et al.*, 2017). Two rounds of 0.4 ml were passed through the columns to rinse and 11 ml 0.17 M HCl was used for La-Ce elution. Nd was eluted in 10 ml 0.17 M HCl while Sm was eluted in 4 ml 0.4 M HCl.

Table 4.5: Weight of isotope dilution samples and spikes for column chemistry.

Sample No:	Lithology Field name:	Sample weight (mg)	Sr spike (mg)	Nd spike (mg)
IRA6-3	McDoogie Pluton (granodiorite)	70.43	104.55	142.49
IRA18-1	McDoogie Pluton (granodiorite)	60.02	166.79	212.15
AGV-2	Standard	60.68	176.59	178.75
IRA18-2	Black Canyon Pluton (granodiorite)	101.04	128.39	312.83
IRA3-1	Anorthosite	159.8	265.51	52.49
IRA8-5 (Nd)	Anorthosite	300.98	-	52.55
IRA18-5	Anorthosite	101.13	308.94	47.53
IRA18-7	Anorthosite	102.54	322.75	44.25
IRA8-5 (Sr)	Anorthosite	69.96	77.51	-

Determination of Sr and Nd isotope ratios by MC-ICPMS

In preparation for isotopic analyses on the Nu Plasma II MC-ICPMS at the University of Johannesburg, the samples were suspended in 2 ml HNO₃. The MC-ICPMS was used to measure parent/daughter ratios as well as isotopic ratios. The MC-ICPMS has the ability to analyse a broader range of elements, and it has a multi-collector which capable of measuring different isotopes of an element at the same time. Rb and Sm analyses were measured in a static mode during the collection of 60 ratios using masses 85 and 87 and masses 147, 148, 149 and 152 collectors respectively. Rb interference was monitored by measuring mass 85. Sm interference was monitored by measuring mass 147. Spike corrections were accomplished using an Excel spreadsheet. Replicate analyses of the standards yielded an average of $^{143}\text{Nd}/^{144}\text{Nd} = 0.512108 \pm 0.000015$ (2SE; n=11) for JNdi-1 with an accepted value of 0.512115 (Tanaka *et al.*, 2000), and $^{87}\text{Sr}/^{86}\text{Sr} = 0.710211 \pm 0.000014$ (2SE; n=15) for SRM987 with an accepted value of 0.710244 (Ma *et al.*, 2013). For Rb/Sr, the standard used for these analyses was an andesite standard called AGV-2, which yielded $^{87}\text{Sr}/^{86}\text{Sr} = 0.703934 \pm 0.000006$ (2SE; n=1), while the accepted value is 0.703992 (Ma *et al.*, 2013). For Sm/Nd, the standards used for these analyses were AGV-2 and BHVO-2. The analyses yielded $^{143}\text{Nd}/^{144}\text{Nd} = 0.512778 \pm 0.000010$ (2SE; n=1) for AGV-2 and 0.512980 ± 0.000010 (2SE; n=2) for BHVO-2. The accepted values for AGV-2 and BHVO-2 are 0.512786 and 0.512979, respectively. All the data are presented in Appendix D: Rb-Sr and Sm-N. The isotopic ratios were initialised to 95 Ma.

Table 4.6: Parent/daughter ratios for standards used for Rb/Sr analyses and the published, accepted values.

Standards	$^{87}\text{Rb}/^{86}\text{Sr}$	2σ	$^{87}\text{Sr}/^{86}\text{Sr}$	2σ	Accepted values
AGV-2	0.302671	0.000145	0.703934	0.000006	0.703992
BHVO-2	0.068048	0.000058	0.703422	0.000007	0.703478

Table 4.7: Parent/daughter ratios for standards used for Sm/Nd analyses and the published, accepted values.

Standards	$^{147}\text{Sm}/^{144}\text{Nd}$	2σ	$^{143}\text{Nd}/^{144}\text{Nd}$	2σ	Accepted Nd/Nd values
AGV-2	0.108505	0.000075	0.512778	0.000010	0.512786
BHVO-2	0.149503	0.000176	0.512979	0.000013	0.512979
BHVO-2	0.149510	0.000096	0.512981	0.000007	0.512979

4.5 alphaMELTS modelling software

alphaMELTS is a numerical modelling package which gives insight into the controls on magma composition, first-order physical principles controlling magma evolution and provides a quantitative assessment of phase relationships and liquid lines of descent in magmatic systems. The alphaMELTS software contains a package of simple text menus for subroutine versions of the algorithms MELTS, pMELTS and pHMELTS. This software was used to calculate equilibrium phase assemblages along a thermodynamic path and calculate major element distributions, including phase stability during the liquid line of descent of a magma. This program is suitable for multi-component systems (which may be anhydrous, hydrous, or water under saturated, with a change in $f\text{O}_2$ and/or $a\text{H}_2\text{O}$ (Antoshechkina and Asimow, 2016). In this program, different parental melt compositions (appropriate for arc environments) were simulated from calculated liquidus temperatures, with variable water contents and degrees of crustal contamination, to understand how these variables may affect the composition of anorthosites forming in arc environments. The identified possible parental magma are tholeiitic basalt, basalt and andesite for the SNB, which were identified by Schmidts and Jagoutz (2017) (discussed in Section 6.4). The modelling is used to understand the conditions involved in the formation of the IRA.

5. Field Observations and Petrology

Figure 5.1 illustrates the geological relationships between the lithologies under consideration in this dissertation, which includes anorthosites (IRA, map unit = Kan), granodiorites of the McDoogle (map unit = Kmd) and Black Canyon Plutons (BCP, map unit = Km) and crosscutting felsic dykes (alaskites). The IRA unit is made up of anorthositic rocks (IRA 3-1, IRA 4-2, IRA 8-5, IRA 18-4a, b, IRA 18-5, IRA 18-6, and IRA 18-7). The McDoogle Pluton (IRA 6-3 and IRA 18-1) and the Black Canyon Pluton (IRA 18-2) samples are granitoids made up of granodioritic rocks and granitic-alaskite (IRA 7-4, and IRA 18-3). The anorthosite of the IRA has a brownish grey colour and contains clumps of clinopyroxene giving it a distinctive “spotted” appearance. The weathered surface is pockmarked from weathering out of mafic minerals. The granodiorite of the McDoogle Pluton has a light grey colour, with equigranular texture and it contains mafic magmatic enclaves with a strong parallel foliation trending in an NW-SE direction. The granodiorite of the BCP has a dark grey colour compared to the granodiorite of the McDoogle Pluton. The alaskite dyke has a light grey to pinkish grey to tan granitoids with minor, disseminated mafic grains of clinopyroxene and biotite. Figure 5.2 illustrates the spatial relationships between the major plutons in the area of interest surrounding Sawmill Lake. Near Sawmill Lake, the McGann Pluton crops out and displays a lighter shade of grey compared to the adjacent Black Canyon Pluton (Figure 5.2 a. b. c). The felsic dykes (alaskite) crosscut the granodiorite of the BCP and the contact between the mafic granodiorites of the McGann and Black Canyon Pluton can be observed in Figure 5.2 d. The area has many country-rock screens that occur within the granodiorite plutons (Figure 5.2 e). The IRA, which is surrounded by the BCP, has a white expression in satellite imagery (Figure 4.1) and is also clearly seen in Figure 5.3 against the darker backdrop of the BCP. The characteristics of each of these units will be discussed in more detail in the subsequent sections.

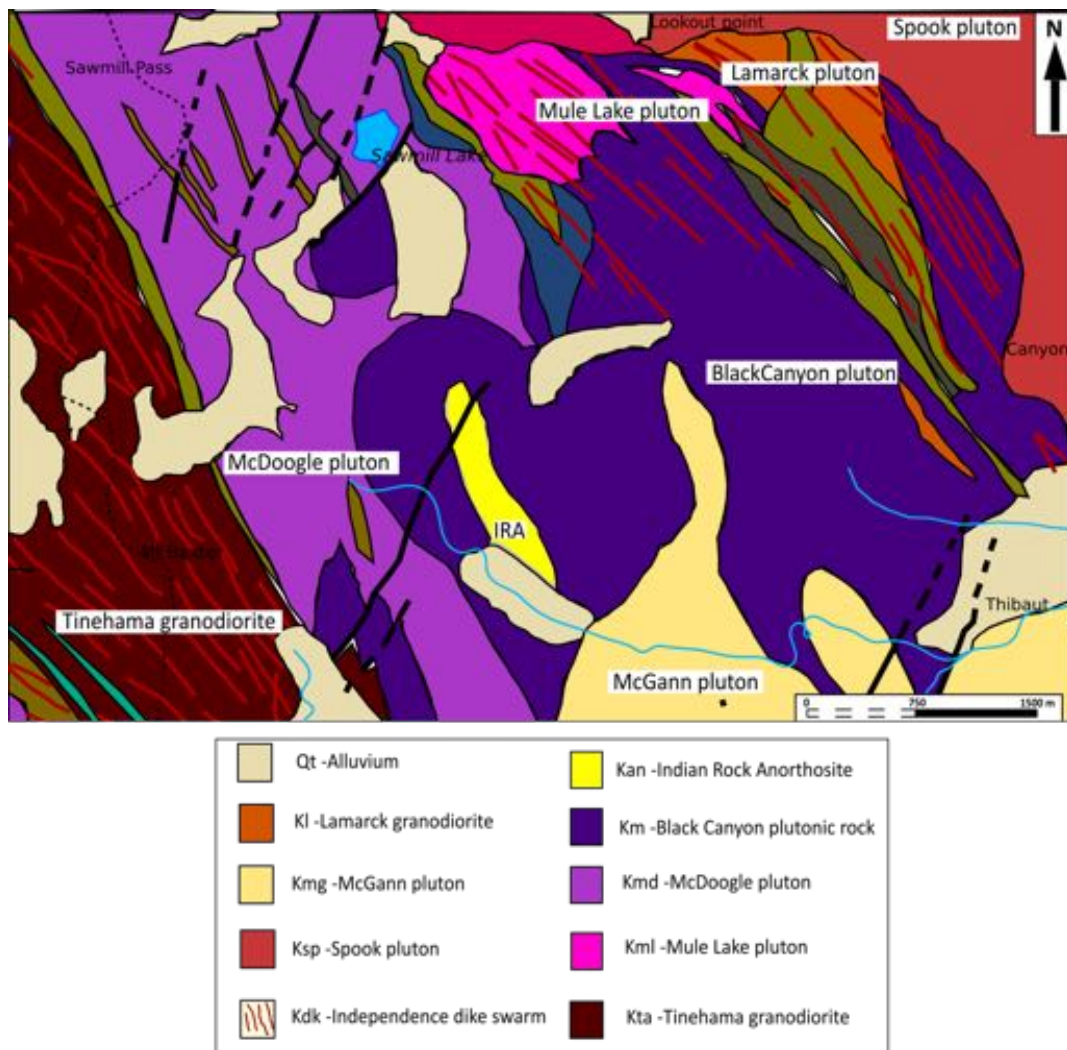


Figure 5.1: A geological map showing the plutons surrounding the Indian Rock Anorthosite (modified from Moore, 1963).



Figure 5.2: Field images showing an overview of the granodioritic plutons in the study area. a. Peaks of the McGann and the Black Canyon Plutons observed from the Sawmill Lake campsite b. The Black Canyon Pluton and the summit of Mount Baxter. c. An overview of the granodiorite plutons surrounding Sawmill Lake. d. Contact between the McGann Pluton and the Black Canyon Pluton granodiorites, with felsic alaskite dykes (10-60 cm thick) crosscutting the Black Canyon Pluton. e. The McGann Pluton with country-rock screens (brown coloured scree).

5.1 Petrology of the Indian Rock Anorthosite

This anorthositic unit is located in steep terrain extending from the summit of Indian Rock into the headwaters of the Thibaut Creek (Figure 5.1 and 5.3). This terrain is encompassed by the granodiorites of Black Canyon, McDoogle and McGann Plutons (Figure 5.3). Given the inherent safety concerns in accessing the IRA, boulders were sampled from the outwash fan of the Thibaut Creek, which is the only stream accessing the anorthosite. Figure 5.3 b, c shows the anorthosite body extending from the headwater of the Thibaut Creek up towards the summit of Indian Rock. Figure 4.1 and Figure 5.1 shows the elongate nature of the IRA (2 km long and 0.5 km wide, 1 km²). The weathered surface of the rock is typically brownish grey, and the fresh surface is grey. Weathered surfaces are sometimes pockmarked from the weathering out of small mafic minerals. The rock has a distinctive “spotted” appearance caused by aggregations of clinopyroxene (Figure 5.4). The grain size ranges from 0.2-0.4 cm (average ~0.3 cm) (Figure 5.4). Anorthosite boulders typically show a homogenous crystalline texture with only minor changes in grain size. The samples have crosscutting mafic veins composed predominantly of clinopyroxene (Figure 5.4 a). The samples show undeformed crystalline texture made up of plagioclase and clinopyroxene grains. The clinopyroxene blebs (2 cm) appear interstitial in hand samples.

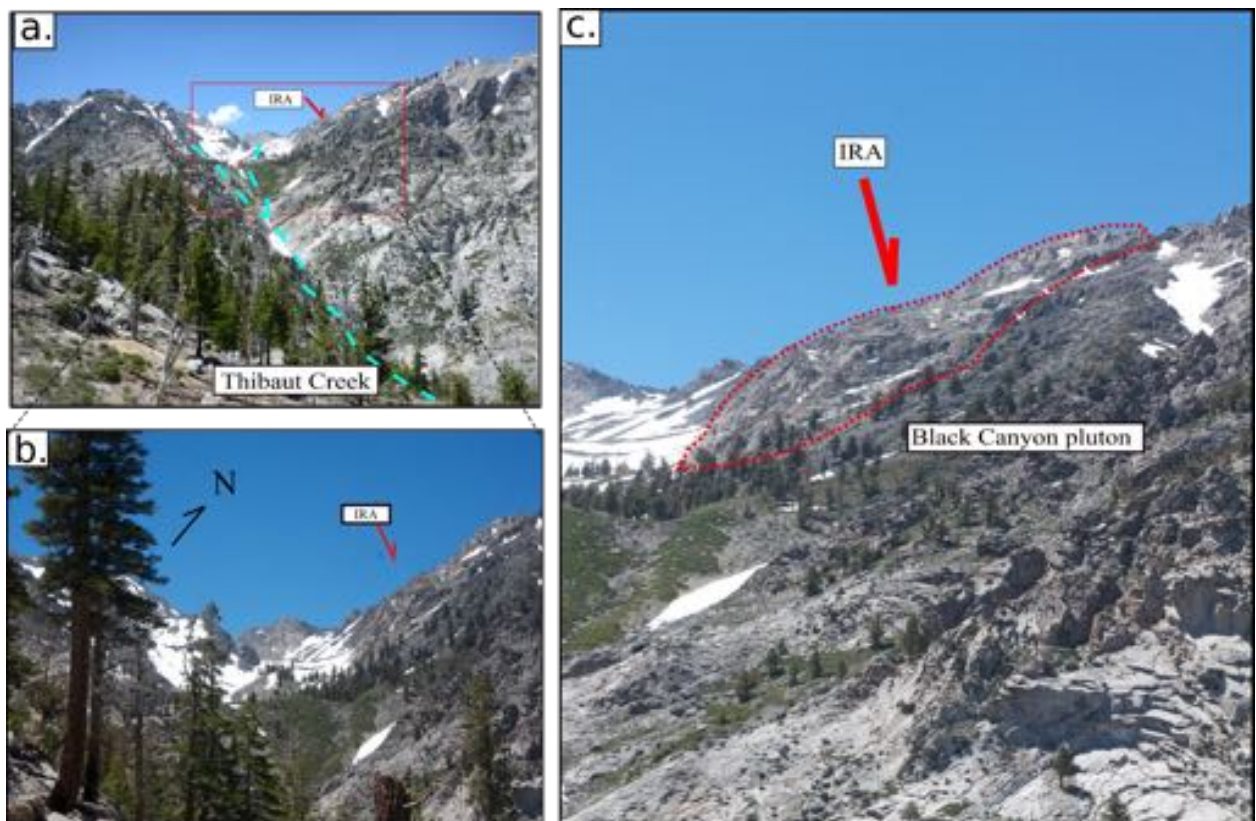


Figure 5.3: a. The anorthosite outcrop extending from the headwaters of the Thibaut Creek to the summit of Indian Rock. b. c. Enlargement of the anorthosite outcrop on the slopes of Indian rock, surrounded by the mafic granodiorite of the Black Canyon Pluton (Km unit).

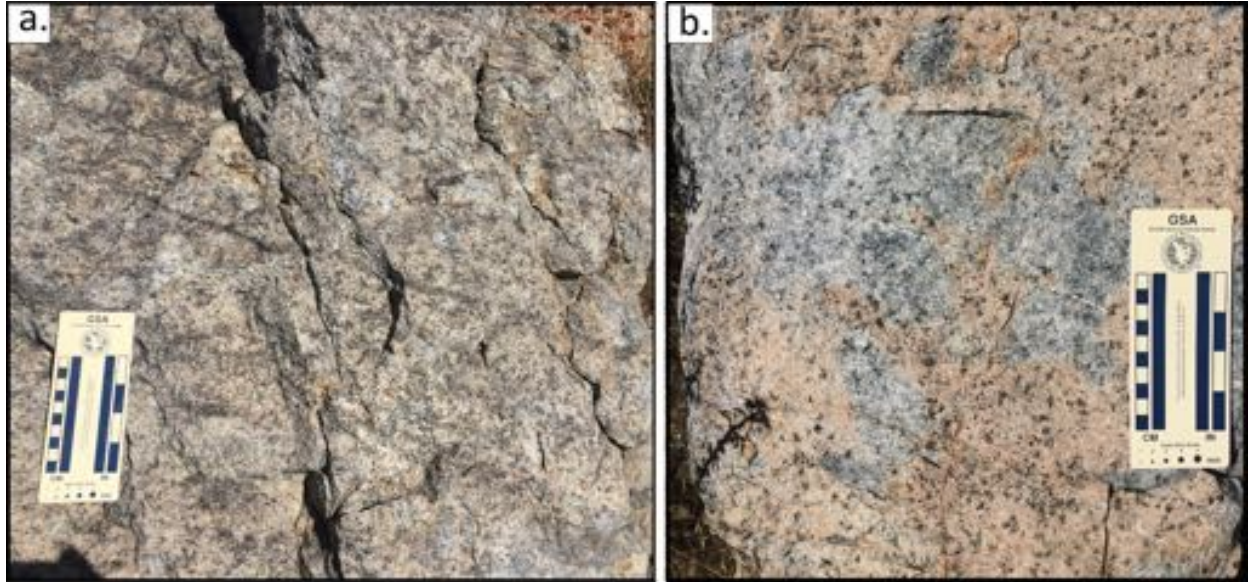


Figure 5.4: Anorthosite boulders found in the outwash fan of Thibaut Creek, which is the only stream that drains the IRA outcrop. a. A mafic vein made up of clinopyroxene crosscutting the boulder. b. Undeformed, crystalline texture of the anorthosite made up of plagioclase and clinopyroxene grains.

In thin section, the samples contain subhedral, zoned plagioclase laths make up over 90% of the rock, with the remainder comprising 5-6 % clinopyroxene, 2-3 % chlorite, 1-2 % hornblende, < 1 % apatite and pyrite, <1 % orthoclase (Figure 5.6). The clinopyroxene occurs as both cumulus and intercumulus phases. Some of the plagioclase grains are enclosed in interstitial clinopyroxene, forming a poikilitic texture (Figure 5.6 a. b. c). Figure 5.7 e. f. g shows clinopyroxene occurring in the cracks of the plagioclase grains forming symplectic intergrowths (Figure 5.7 e. f. g). In other areas, the samples have variable adcumulate texture shown by plagioclase grains forming polygonal boundaries with little or no intercumulus minerals (Figure 5.7). The main cumulus phase is zoned plagioclase, and the intercumulus phases include primary clinopyroxene, and secondary chlorite and hornblende, while apatite and pyrite occur as accessory minerals. Orthoclase also occurs interstitially to plagioclase. These relationships suggest a sequential crystallisation of the minerals with plagioclase crystallising first followed by the clinopyroxene, orthoclase, apatite and then pyrite. However, some of the plagioclase grains are included within orthoclase (e.g., Figure 5.7 d).

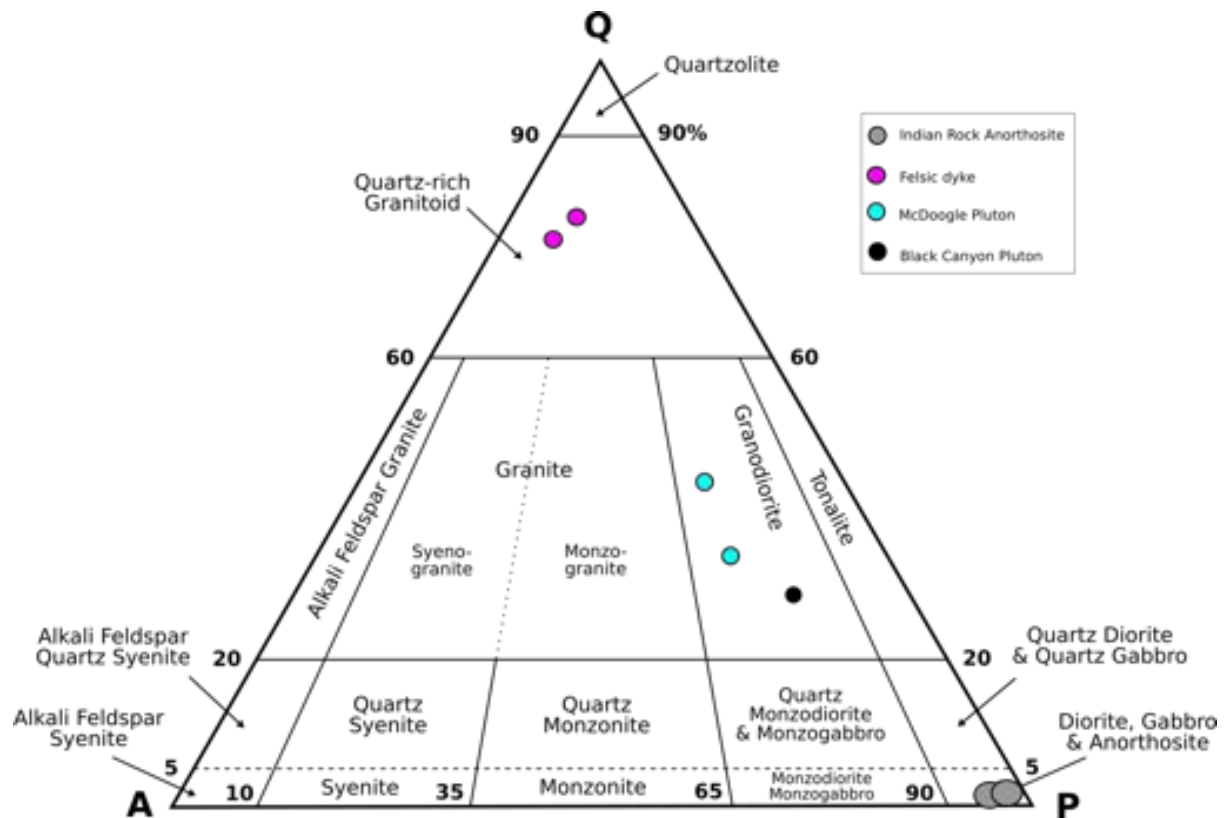


Figure 5.5: A quartz-alkali-feldspar-plagioclase ternary showing modal mineral abundance and classification of the Indian Rock Anorthosite, McDoogie Pluton, Black Canyon Pluton and felsic dykes.

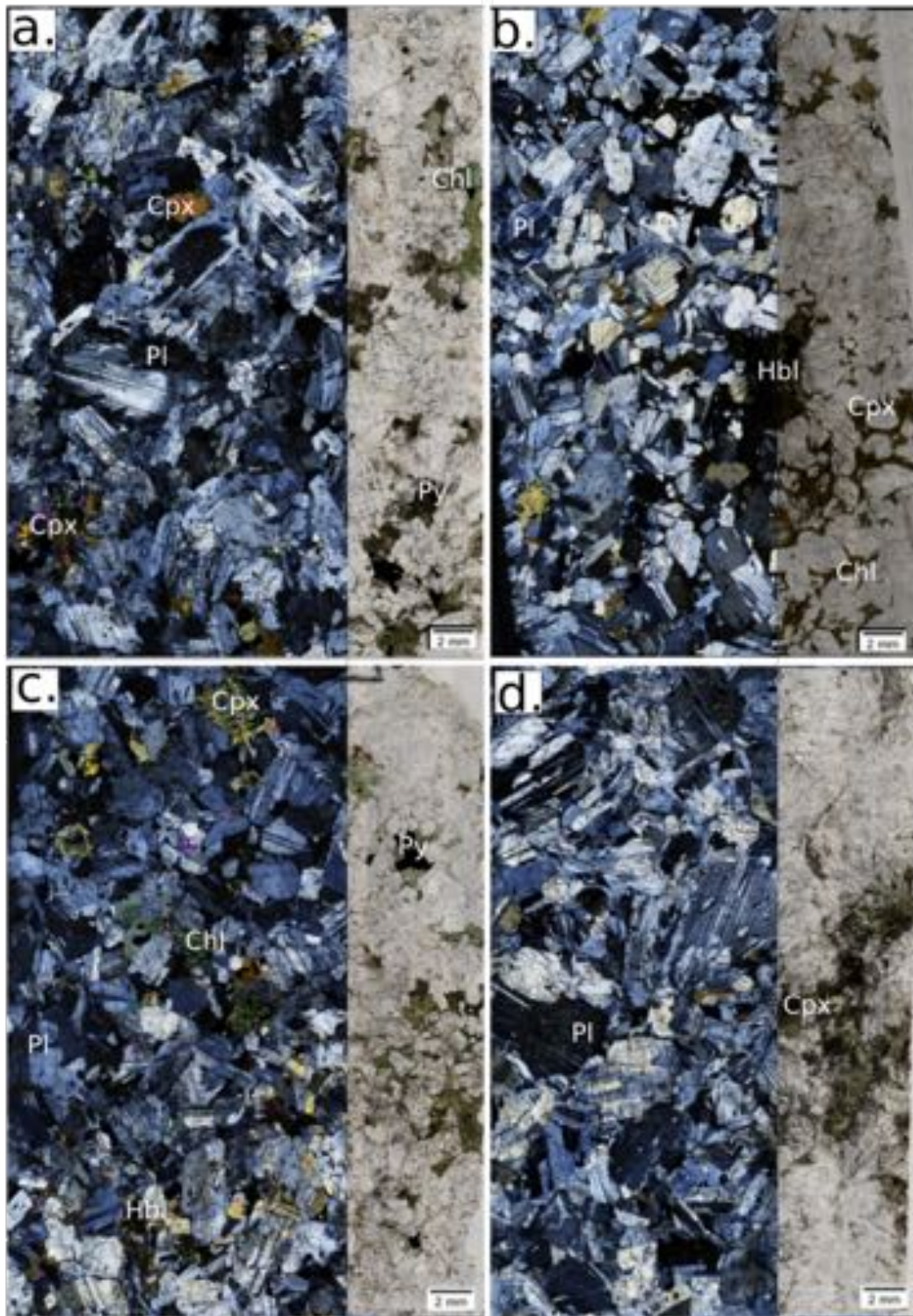


Figure 5.6: Photomicrographs of the Indian Rock Anorthosite showing medium-grained (0.2-0.4 cm) plagioclase grains with clinopyroxene, chlorite, hornblende, apatite and pyrite. a. Zoned cumulate plagioclase with interstitial mafic minerals (clinopyroxene and hornblende). b. Poikilitic texture with plagioclase inclusions and clinopyroxene oikocrysts. c. Altered grains of clinopyroxene to hornblende and chlorite. d. 4 mm grains of plagioclase with altered grains of clinopyroxene.

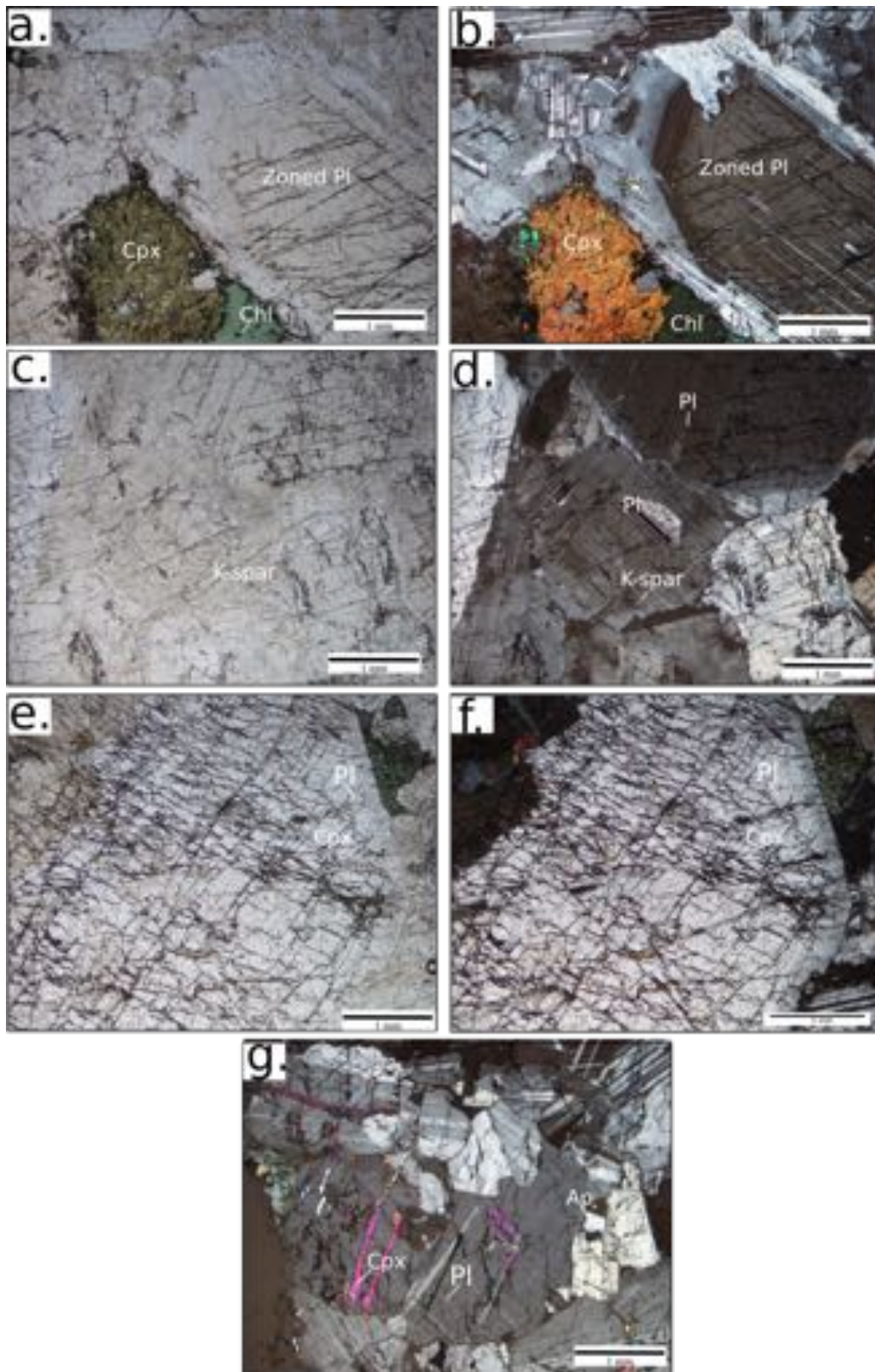


Figure 5.7: PPL and XPL photomicrographs of the IRA showing different textures between the minerals and the relationships amongst them. a. b. show zoned plagioclase grains with clinopyroxene alter to chlorite at the edges. c. d. shows a plagioclase grain enclosed within an orthoclase grain. e. f. g. A plagioclase grain having fractures filled with clinopyroxene.

5.2 Granodiorites of the McDoogle and Black Canyon Plutons

McDoogle Pluton

This predominantly granodioritic pluton occurs on the western side of the Black Canyon Pluton (Figure 5.1). The colour on the weathered and fresh surfaces of these granodiorites are similar (Figure 5.8 a). The pluton has multiple near-horizontal crosscutting felsic dykes (Figure 5.8 b). The grain size ranges from 0.3-0.5 cm and generally displays an equigranular texture. The granodiorite contains mafic magmatic enclaves, which have a strong parallel foliation trending in an NW-SE direction (Figure 5.8 c) and this is shown by the flattened preferred orientation of the enclaves and the alignment of mafic minerals (hornblende). The mafic enclaves are ellipse-shaped with $L = 8\text{-}30\text{ cm}$, $W = 2\text{-}8\text{ cm}$, extended in an NW-SE trend (Figure 5.8 c). The McDoogle Pluton granodiorites show modal abundances ranging between 35- 40 % quartz, 45-55 % plagioclase, 5-10 % microcline, 15-20 % clinopyroxene, 1 % hornblende and 1-2 % biotite (Appendix A). The plagioclase grains are generally fresh and subhedral to euhedral, while the clinopyroxenes are interstitial (Figure 5.9 a). The thin section has fresh grains of plagioclase, quartz, biotite with clear grain boundaries, altered grains of clinopyroxenes, and secondary hornblende and chlorite (Figure 5.10 a, b). The mafic enclaves in the McDoogle Pluton are fine-grained and they have more mafic minerals such as clinopyroxene, hornblende, and biotite than the host does. The enclaves have a gradational contact with the granodiorite shown by the gradual change in grain sizes at the contact zone (Figure 5.9 c). The enclaves have randomly orientated veins of clinopyroxene (~0.2 mm thick) (Figure 5.9 d) and oikocrysts of clinopyroxene hosting fine-grained quartz, plagioclase and K-feldspars, in a well-developed poikilitic texture.

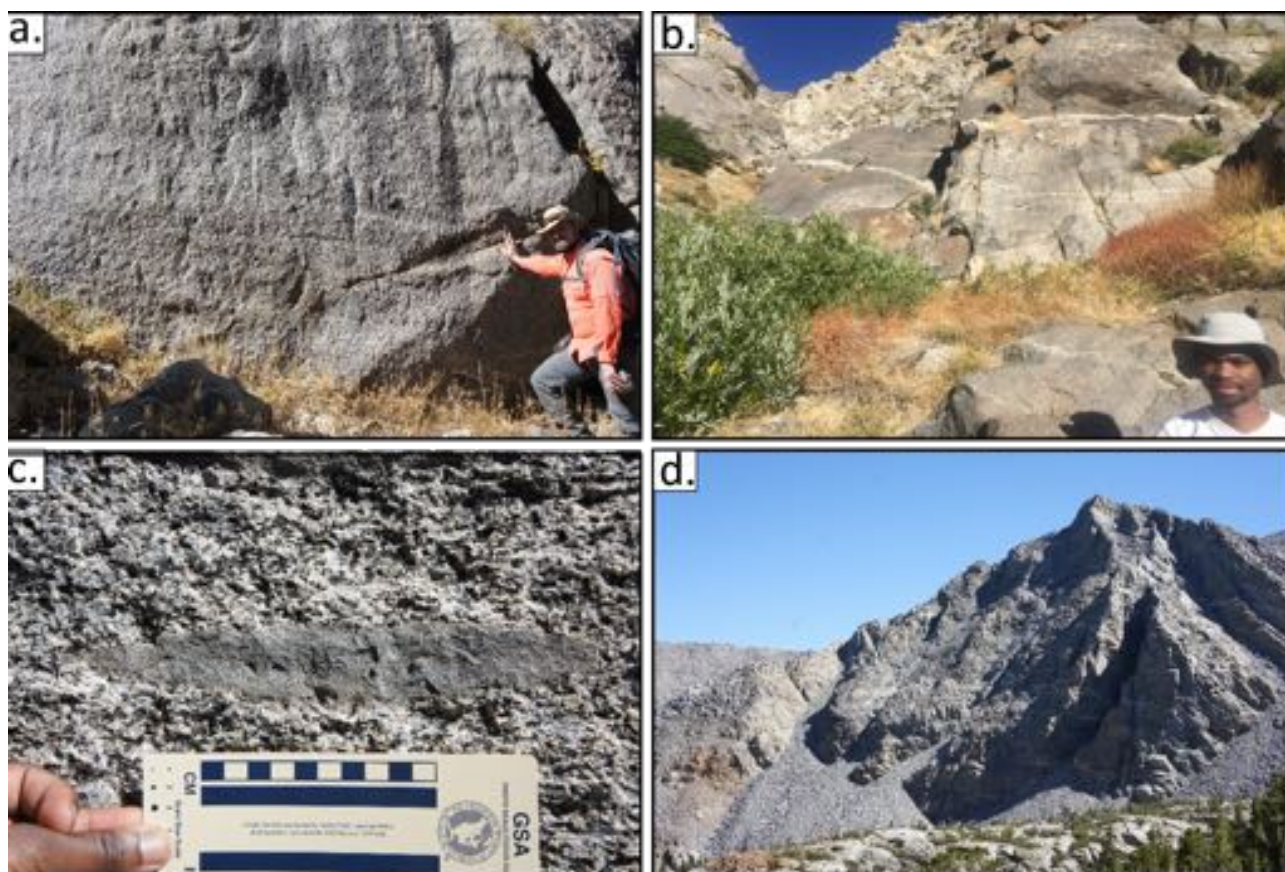


Figure 5.8: Field images of granodioritic McDoogle and Black Canyon Plutons. a. The granodiorite of the McDoogle Pluton showing foliation trending NW-SE. b. The Black Canyon Pluton, with crosscutting felsic dykes. c. Elongated mafic enclave (oriented NW-SE) in the granodiorite of McDoogle Pluton. d. Overview of the Black Canyon Pluton, which hosts the IRA.

Black Canyon Pluton

The Black Canyon Pluton hosts the Indian Rock Anorthosite, and it occurs on the east side of the McDoogle Pluton (Figure 5.1). The granodiorite has a dark grey colour indicating higher mafic content (making over 5 % of the rock) compared to the McDoogle Pluton (Figure 5.5). The grain size ranges from 0.3-0.5 cm and it contains multiple felsic dykes crosscutting the pluton in a near-horizontal orientation (Figure 5.8 b). In thin section, the rocks are generally composed of 20 % quartz, 50 % plagioclase, 10 % microcline, 10 % clinopyroxene, 6 % hornblende and 4 % biotite (Figure 5.9 b). The plagioclase grains are altered to sericite and some clinopyroxene grains are altered to chlorite (Figure 5.10 c, d). The mafic minerals (clinopyroxene, hornblende, and biotite) occur as intercumulus phases. This granodiorite has a high mafic minerals compared to the granodiorite of the McDoogle Pluton, which makes up over 5 % of the rock (Figure 5.9 b).

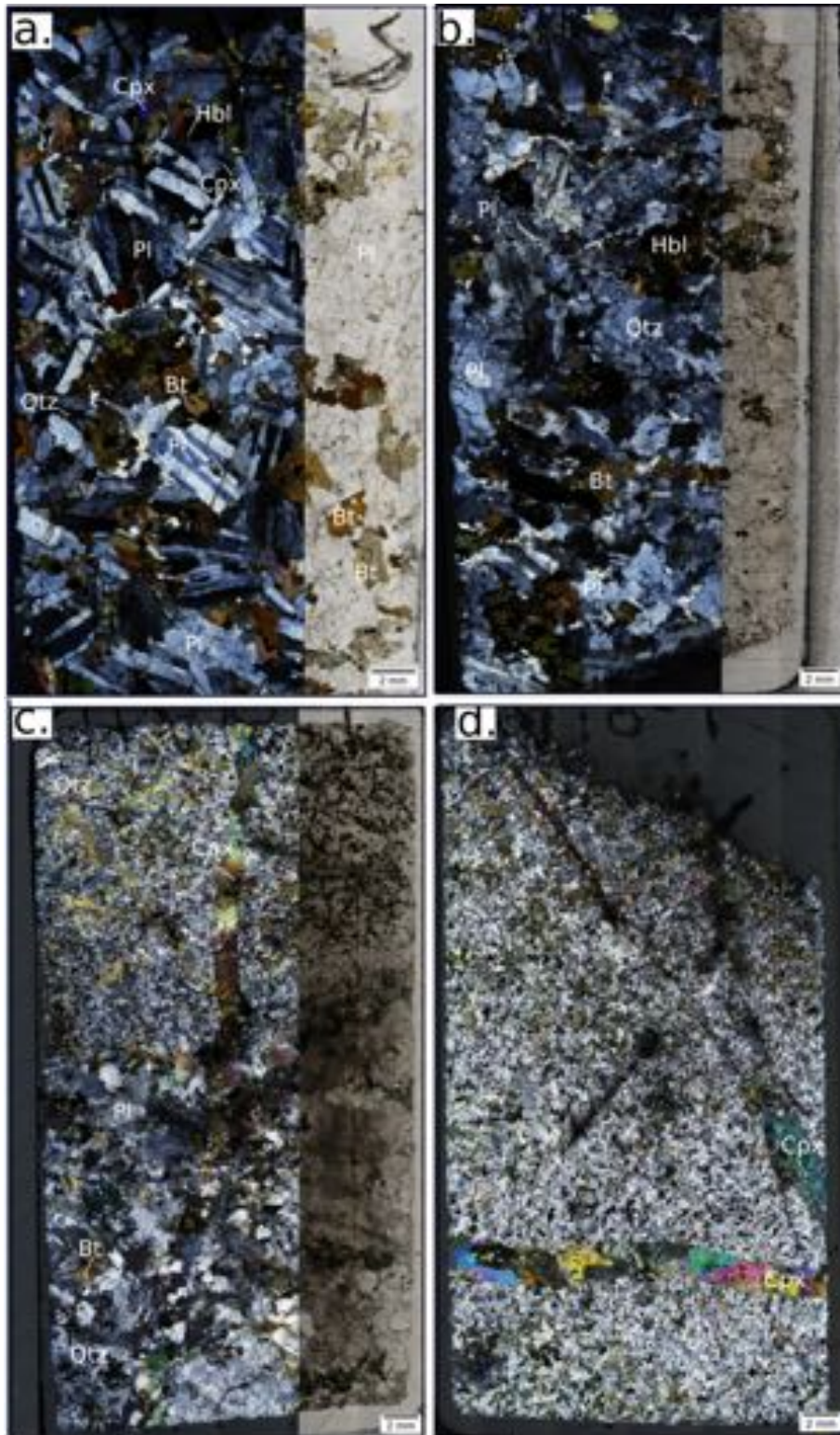


Figure 5.9: a. Photomicrograph of the granodiorite of McDoogle pluton showing an interlocking texture having elongated cumulate plagioclase grains, intercumulus hornblende and clinopyroxene. b. Photomicrograph of the mafic granodiorite of the Black Canyon Pluton. c. Contact between a mafic enclave and granodiorite of the McDoogle pluton shown by the change in grain size. d. The mafic enclave consists of fine grains (~0.1 cm) of plagioclase, biotite, quartz, and a vein of clinopyroxene.

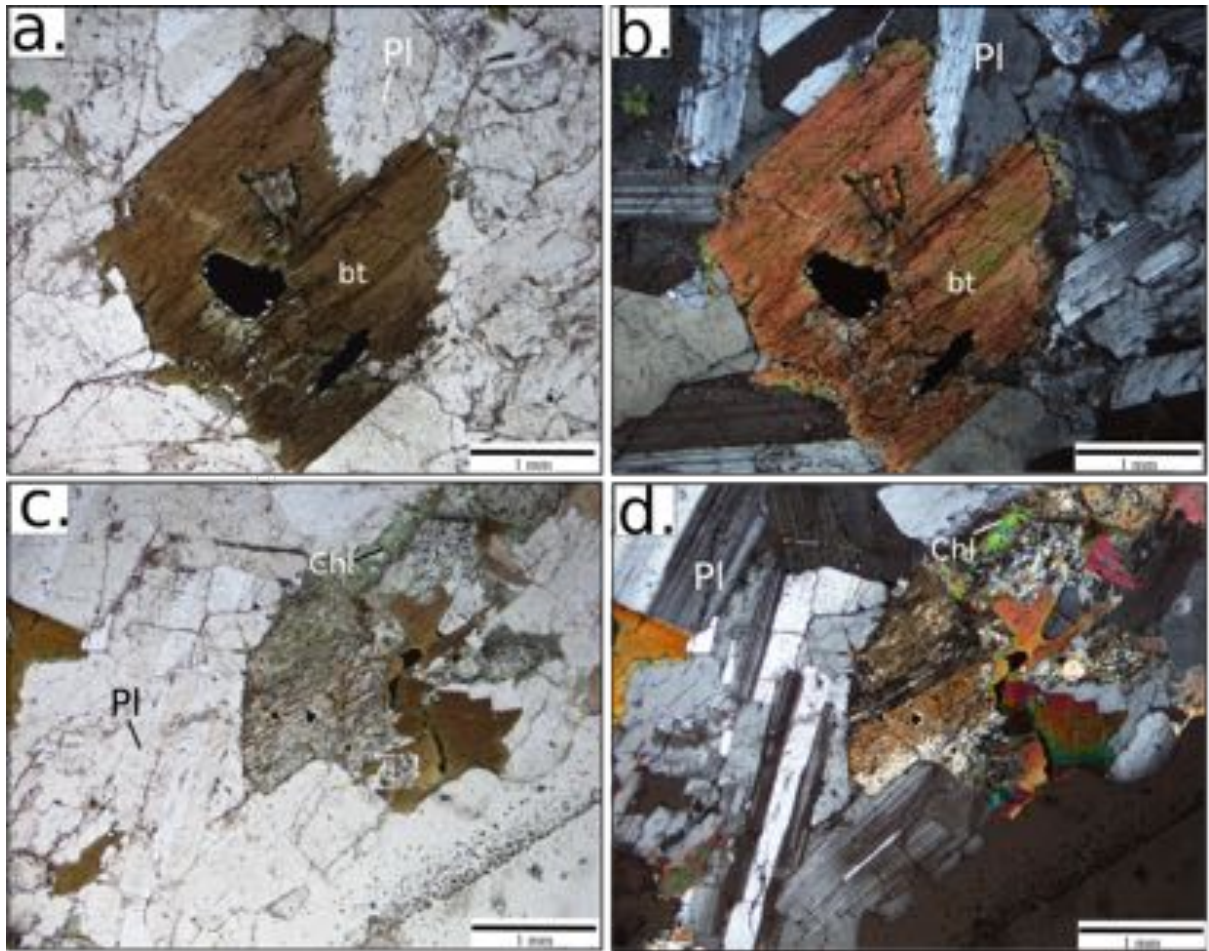


Figure 5.10: PPL and XPL thin-section images of the granodiorites (McDoogle and Black Canyon Pluton) showing different textures between the minerals and the relationships amongst them. a. b. The granodiorite of the McDoogle Pluton shows biotite grain with magnetite inclusion and euhedral plagioclase grains. c. d. The granodiorite of the Black Canyon Pluton shows biotite altering to hornblende and chlorite.

5.3 Alaskite Dykes

The characteristic crosscutting dykes of the region are light grey to pinkish-grey to tan granitoids with a minor, disseminated mafic grains of clinopyroxene and biotite (Figure 5.11 a, b). These dykes crosscut the granodiorites of Black Canyon and McDoogle Plutons in varying orientations as observed in Figure 5.11 c, d. The dykes are ~4-60 cm thick and display sharp contacts with the host (Figure 5.11).

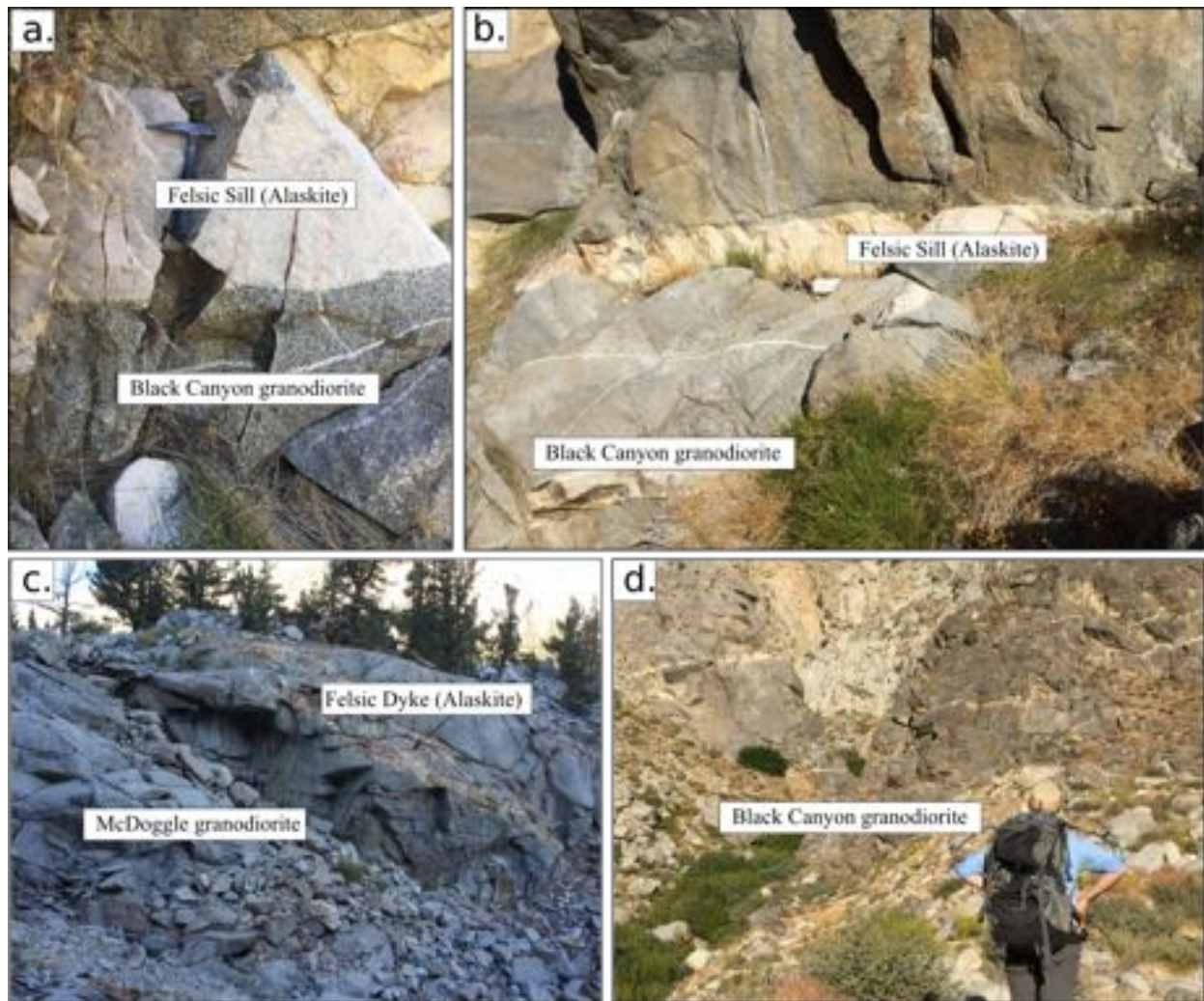


Figure 5.11: Field photographs of the crosscutting felsic dykes (alaskite). a. Felsic dykes crosscutting the granodiorite from the Black Canyon Pluton, which hosts the anorthosite. b. The sharp contact between the dykes and the granodiorite. c. The felsic dyke crosscutting the McDoogle granodiorite in a sub-vertical manner. d. The Black Canyon granodiorite having multiple felsic sills of alaskite.

In thin section, the samples are fine- to medium-grained (~0.05- 0.1 cm). The samples have an equigranular texture with interlocking grains of quartz, plagioclase, and microcline (Figure 5.12 a). The rock is composed of 60-70 % quartz, 10-15 % plagioclase, 4-5 % clinopyroxene, 2-3 % biotite, 1-2 % hornblende and 1-2 % microcline (Appendix A). Figure 5.12 b shows a sharp contact between the dyke and the mafic granodiorite. The contact is shown by a sudden change in grain size and mineral assemblage with a chill margin. The dykes display localized granophyric textures (Figure 5.13 a, b), together with hornblende grains having inclusions of quartz and microcline. Moreover, in other areas, the clinopyroxenes occur as inclusions in the microcline. Some grains of plagioclase are zoned, and biotite is altered to hornblende (Figure 5.13 c, d).

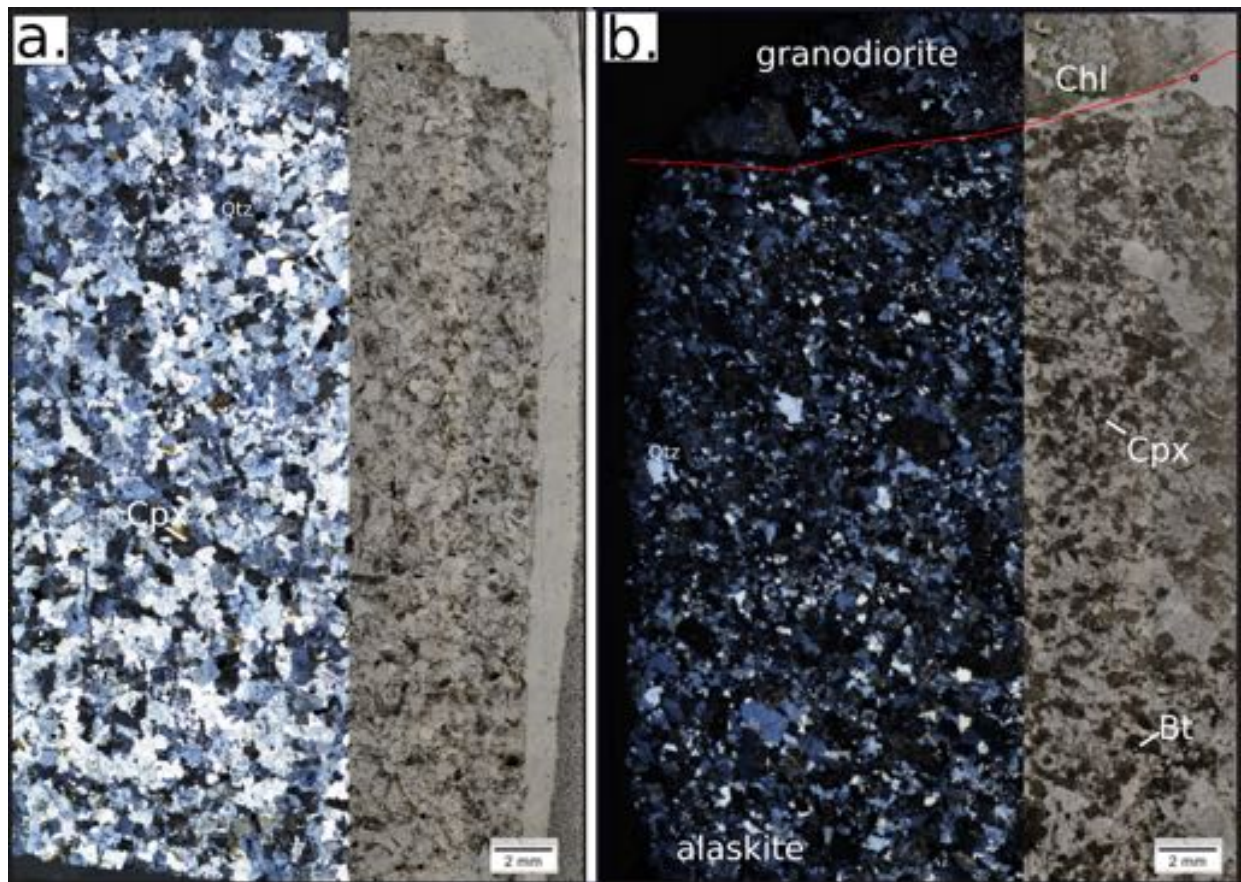


Figure 5.12: Photomicrographs of alaskite dykes shown in xpl and ppl. a. The thin section shows a granitic texture with interlocking grains of quartz, plagioclase and microcline that is typical for granitic rocks. b. The thin section shows the sharp contact between the granodiorite and the dyke, which is characterised by a change in grain size and mineralogy.

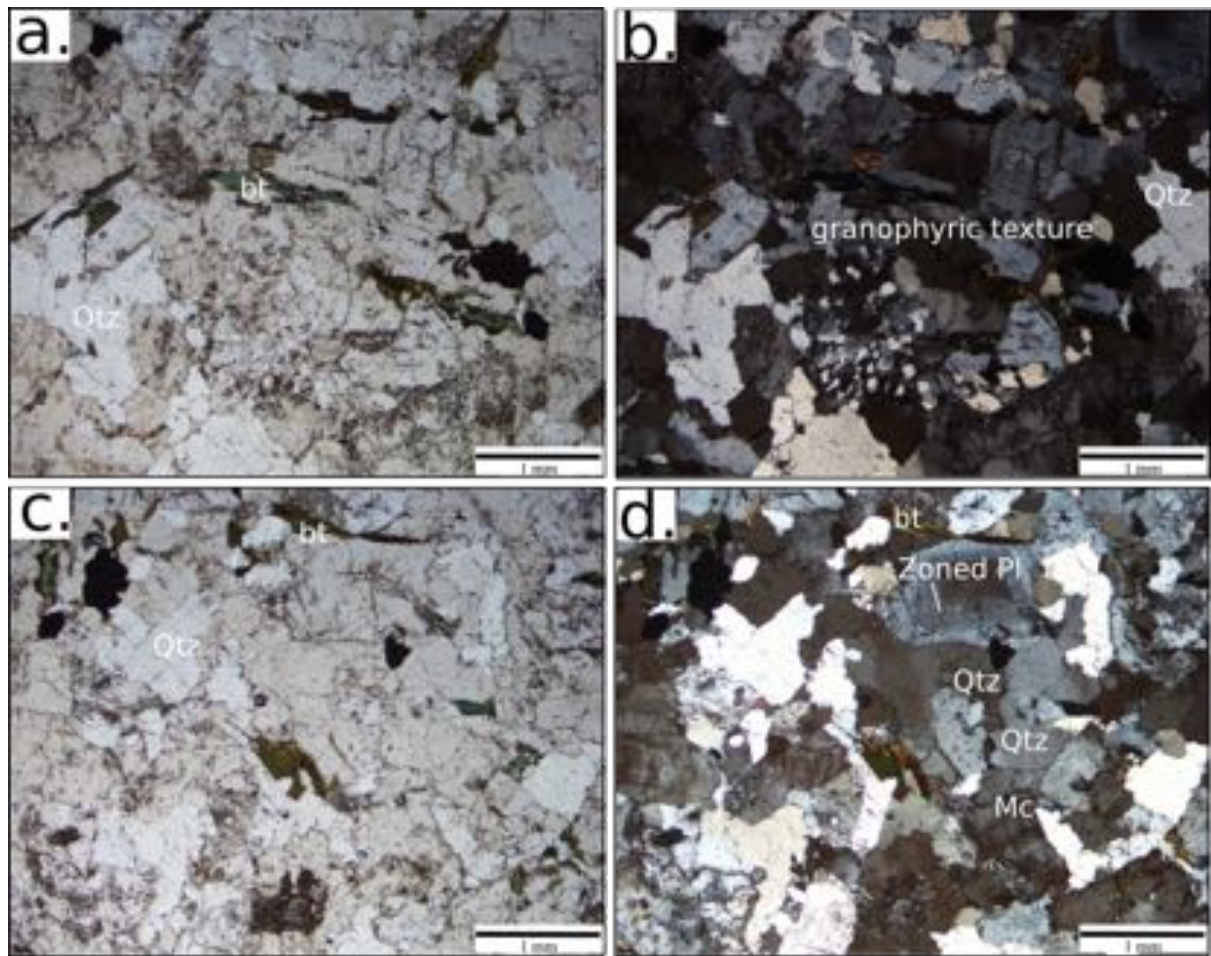


Figure 5.13: Thin-section images of the granitic alaskite dykes shown in xpl and ppl. a. b. The rocks have a granophyric texture. c. d. Interlocking grains of quartz, plagioclase and microcline.

6. Geochemistry and alphaMELTS Results

6.1 Whole-rock Chemical Compositions

This chapter presents the whole-rock major and trace element data of the IRA, the surrounding granodiorites and the crosscutting alaskite dykes. Ten samples were selected for whole-rock geochemical analyses and the data are presented in Table 6.1 and 6.2 below.

6.1.1 Major Elements Systematics

The anorthosites are characterised by moderate variations in SiO_2 (46.7- 48.76 wt%), CaO (13.02- 15.67 wt%), Al_2O_3 (27.33- 30.66 wt%), TiO_2 (0.20- 0.51 wt%) and Na_2O (1.89- 3.01 wt%) (Figure 6.1). The low mafic mineral content is reflected by the low MgO and $\text{Fe}_2\text{O}_3\text{T}$ concentrations (both ≤ 3.76 wt%) (Figure 6.1 d, f). The anorthosites have low K_2O (0.25- 0.62 wt%) and TiO_2 (0.20- 0.51 wt%) concentrations relative to the granodiorite and the felsic dykes (Table 6.1). The granodiorites from the two plutons are characterised by large variations in chemical compositions due to the varying modal mineral abundances shown by SiO_2 (60.76- 56.43 wt%), CaO (8.28- 3.92 wt%), and Na_2O (3.44- 4.98 wt%), while the felsic dykes show no or little variation in the major elements. Granodiorite of McDoogle Pluton has high CaO , $\text{Fe}_2\text{O}_3\text{T}$ and low SiO_2 compared to granodiorite of the Black Canyon Pluton.

Table 6.1: Whole-rock major element oxide data for a representative selection of 10 samples from the study area.

Rock Type	Anorthosite	Anorthosite	Anorthosite	Anorthosite	Anorthosite	McDoogle Pluton (granodiorite)	McDoogle Pluton (granodiorite)	Black Canyon pluton (granodiorite)	Felsic dyke (Alaskite)	Felsic dyke (Alaskite)
Sample ID	IRA3-1	IRA4-2	IRA8-5	IRA18-5	IRA18-7	IRA6-3	IRA18-1	IRA18-2	IRA7-4	IRA18-3
elements (oxides)										
SiO ₂	46.7	48.76	46.75	48.1	47.83	56.43	52.37	60.76	72.97	72.66
Al ₂ O ₃	29.11	27.33	30.66	28.81	29.63	17.23	17.56	17.05	14.32	14.13
TiO ₂	0.3	0.51	0.2	0.42	0.38	1	1.22	1.04	0.11	0.16
Fe ₂ O ₃ T	2.81	3.76	1.88	3.31	3.46	7.89	8.87	5.89	1.32	1.5
MgO	1.29	1.25	0.95	1.33	1.33	3.93	3.62	1.8	0.28	0.31
CaO	14.62	13.02	15.67	14.38	14.7	6.37	8.28	3.92	1.45	1.28
Na ₂ O	2.14	3.01	1.89	2.01	1.81	3.44	4.98	3.93	2.84	3.28
K ₂ O	0.39	0.43	0.25	0.49	0.62	2.88	1.06	4.7	6.05	5.71
P ₂ O ₅	0.09	0.13	0.07	0.07	0.08	0.31	0.34	0.27	0.04	0.04
MnO	0.03	0.04	0.02	0.03	0.03	0.12	0.23	0.08	0.01	0.02
NiO	0	0	0	0.01	0.01	0.01	0.01	0	0	0.01
LOI	1.51	1	0.91	0.76	1.09	0.61	0.61	0.63	0.35	0.62
Total	98.99	99.24	99.25	99.72	100.96	100.22	99.17	100.05	99.74	99.71

Three groupings of samples are can be observed in Figure 6.1 and Figure 6.2, which geochemically corresponds with the three rock types (i.e. anorthosite, granodiorite and the alaskite dyke. The first group is defined by anorthosite which has low SiO₂ (46.7- 48.76 wt%), high CaO (13.02- 15.67 wt%) and Al₂O₃ (27.33- 30.66 wt%). The second group is composed of granodiorites of the McDoogle and the Black Canyon Plutons. This group shows lower CaO (3.92- 8.28 wt%) and Fe₂O₃T (5.89- 8.87 wt%) concentrations, with increased SiO₂ (52.37- 60.76 wt%), TiO₂ (1- 1.22 wt%), NiO (11.65- 33.78 wt%) and MgO (1.8- 3.93 wt%). The third group is composed of alaskite dykes with higher SiO₂ (72.66- 72.97 wt%), low MgO (0.28- 0.31 wt%) and Fe₂O₃T (1.32- 1.5 wt%).

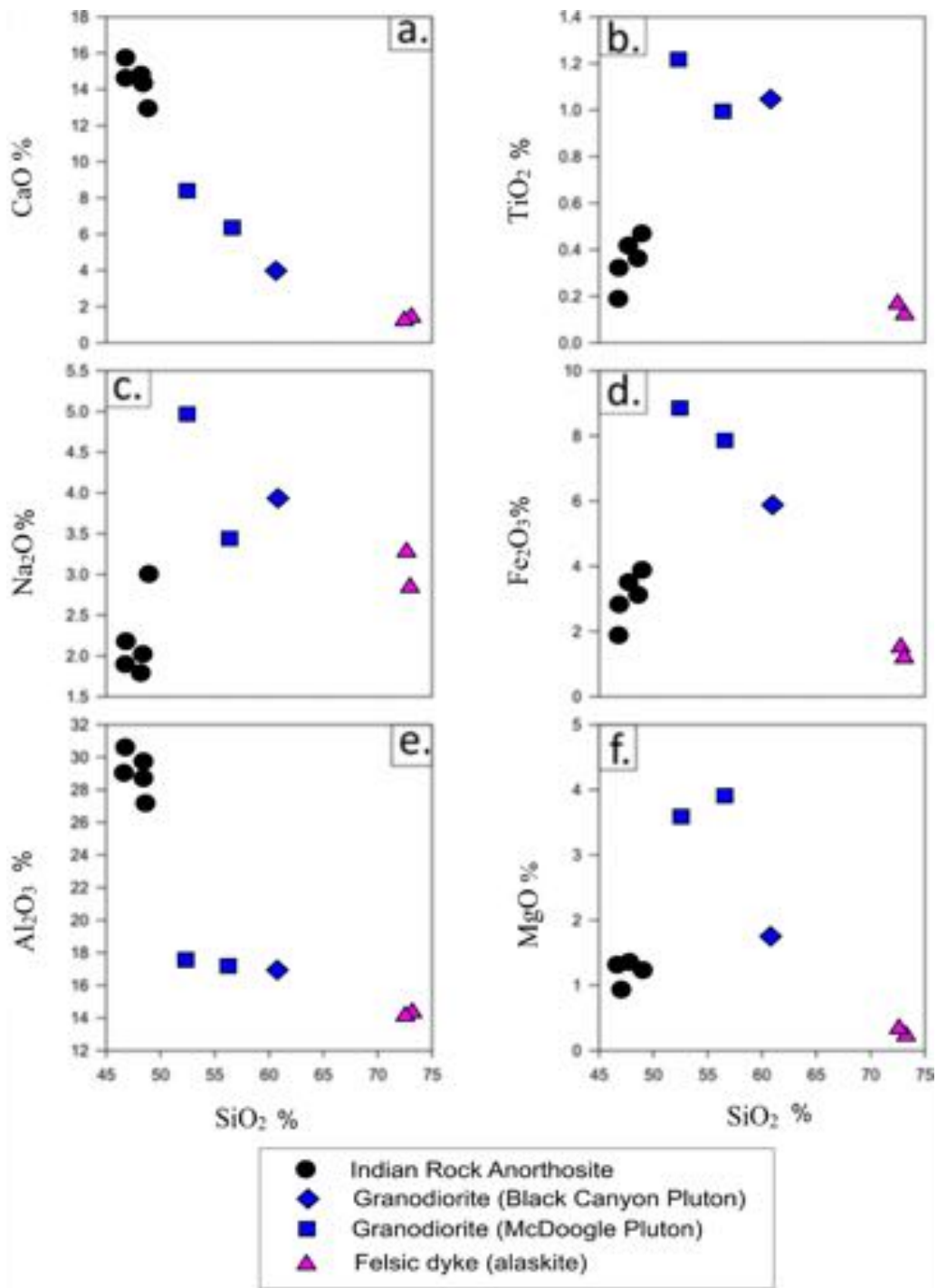


Figure 6.1: SiO_2 vs CaO , TiO_2 , Na_2O , Fe_2O_3 , Al_2O_3 , and MgO wt%. The granodiorite shows varying SiO_2 wt% content compared to other rocks, anorthosites have high Al_2O_3 and CaO wt% reflecting high modal plagioclase content, and felsic dykes (alaskite) have high SiO_2 compared to other elements. a. There is a decrease in CaO wt% with increasing SiO_2 wt% content. b. The granodiorites have high TiO_2 wt% compared to other samples. c. The anorthosites show low Na_2O wt% relative to the granodiorites and the alaskite dykes. d. Granodiorite and alaskite dykes show a negative slope trend with the anorthosites being the outliers. e. A negative trend between SiO_2 wt% and Al_2O_3 . f. Granodiorite and granitic-alaskite dyke show a negative slope trend between SiO_2 vs MgO wt% with the anorthosites being the outliers.

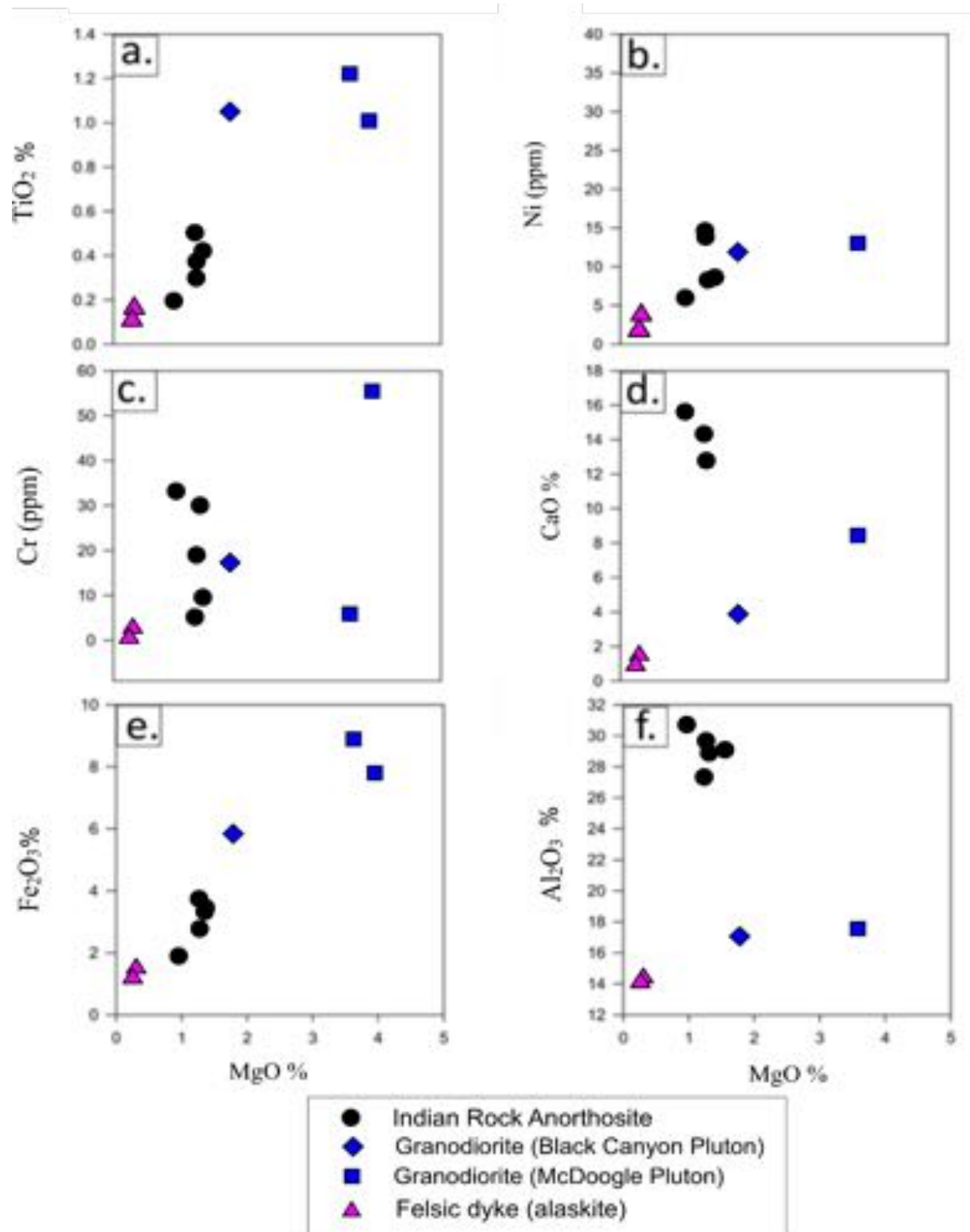


Figure 6.2: MgO vs TiO₂, Ni (ppm), Cr (ppm), CaO, Fe₂O₃, and Al₂O₃ wt%. The anorthosites show high CaO and Al₂O₃, while the granodiorites have high MgO wt%. a. b. e. Positive trend between MgO vs TiO₂ wt%, Ni (ppm), and Fe₂O₃ wt%. d. f. Show similar patterns with anorthosites having high CaO wt% and Al₂O₃ wt% against MgO wt%. c. Cr (ppm) vs MgO wt% do not show any observable trend.

6.1.1.1 CIPW norm

The CIPW normative calculations of the analysed samples are presented in Table 6.2 . The anorthosites contain between 88.23 - 93.37 volume % of plagioclase with the rest made up of olivine (1.25 - 3.82 %), orthoclase (1.63 - 4.01 %), diopside (1.05 - 2.77 %), ilmenite (0.34 - 0.58 %), magnetite (0.15 - 0.48 %) and apatite (0.14 - 0.27 %). The calculated anorthite content ranges from An₇₀ to An₈₂. The granodiorites of the McDoogie pluton have plagioclase making up between 55.87 to 68.22 volume % of the rocks and the rest is orthoclase (7.18 - 19.08 %), hypersthene (14.56 %), diopside (4.60 – 13.74 %) quartz (3.46 %), ilmenite (1.15 – 1.43 %), magnetite (0.64 – 0.74 %), and apatite (0.64 – 0.72 %). The calculated anorthite content ranges from An₃₆ to An₄₄. The granodiorite of the Black Canyon pluton has plagioclase making about 50.40 volume %. The rest of the rock is made up of orthoclase (30.17 %), hypersthene (8.43 %), quartz (7 %), diopside (1.84 %), ilmenite (1.16 %), apatite (0.54 %), and magnetite (0.46 %). The calculated anorthite content is An₄₁. The normative calculations show that the Black Canyon pluton has more orthoclase compared to the McDoogie pluton. The felsic dykes have high volume % of orthoclase ranging from 35.23 - 37.19 %, followed by plagioclase (34.21 - 31.12 %), quartz (29.16- 27.90 %), hypersthene (2.08 - 1.87 %), corundum (0.37 - 0.22 %), and ilmenite (0.11- 0.10 %), magnetite (0.10 – 0.11 %), and apatite (0.08 %). The calculated anorthite content ranges from An₁₈ to An₂₂.

Table 6.2: CIPW norm calculations of the samples using whole rock major element chemistry.

Rock Type	Anorthosite		Anorthosite		Anorthosite		Anorthosite		Anorthosite	
Sample ID	IRA3-1		IRA4-2		IRA8-5		IRA18-5		IRA18-7	
	Weight %	Volume %	Weight %	Volume %	Weight %	Volume %	Weight %	Volume %	Weight %	Volume %
	Norm	Norm	Norm	Norm	Norm	Norm	Norm	Norm	Norm	Norm
Quartz	0,00	0,00	0,00	0,00	0,00	0,00	0,45	0,48	0,00	0,00
Plagioclase	86,78	90,95	85,26	89,28	90,43	93,37	85,15	88,23	86,21	88,42
Orthoclase	2,30	2,58	2,54	2,82	1,48	1,63	2,90	3,20	3,66	4,01
Corundum	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00	0,00
Diopside	2,73	2,30	3,33	2,77	2,36	1,98	2,29	1,91	1,27	1,05
Hyperssthene	0,66	0,53	1,58	1,24	1,58	1,26	6,73	5,30	6,71	5,22
Olivine	3,82	2,89	3,72	2,73	1,67	1,25	0,00	0,00	0,61	0,44
Ilmenite	0,57	0,34	0,97	0,58	0,38	0,23	0,80	0,48	0,72	0,43
Magnetite	0,41	0,22	0,55	0,30	0,28	0,15	0,48	0,26	0,51	0,27
Apatite	0,21	0,19	0,30	0,27	0,16	0,14	0,16	0,14	0,19	0,16
Total	97,48	100,00	98,25	99,99	98,34	100,01	98,96	100,00	99,88	100,00
Anorthite	68,67	79	59,79	70	74,44	82	68,14	80	70,89	82
Albite	18,11	21	25,47	30	15,99	18	17,01	20	15,32	18
Rock Type	McDoggie Pluton (granodiorite)		McDoggie Pluton (granodiorite)		Black Canyon pluton (granodiorite)		Felsic dyke (Alaskite)		Felsic dyke (Alaskite)	
Sample ID	IRA6-3		IRA18-1		IRA18-2		IRA7-4		IRA18-3	
	Weight %	Volume %	Weight %	Volume %	Weight %	Volume %	Weight %	Volume %	Weight %	Volume %
	Norm	Norm	Norm	Norm	Norm	Norm	Norm	Norm	Norm	Norm
Quartz	3,19	3,46	0,00	0,00	6,68	7,00	29,01	29,16	27,67	27,90
Plagioclase	52,17	55,87	62,03	68,22	48,25	50,40	30,96	31,12	33,84	34,21
Orthoclase	17,02	19,08	6,26	7,18	27,78	30,17	35,75	37,19	33,74	35,23
Corundum	0,00	0,00	0,00	0,00	0,00	0,00	0,56	0,37	0,32	0,22
Diopside	5,41	4,60	13,74	11,88	2,26	1,84	0,00	0,00	0,00	0,00
Hyperssthene	18,03	14,56	0,00	0,00	11,02	8,43	2,61	1,87	2,90	2,08
Olivine	0,00	0,00	10,72	8,25	0,00	0,00	0,00	0,00	0,00	0,00
Ilmenite	1,90	1,15	2,32	1,43	1,98	1,16	0,21	0,12	0,30	0,17
Magnetite	1,16	0,64	1,30	0,74	0,86	0,46	0,19	0,10	0,22	0,11
Apatite	0,72	0,64	0,79	0,72	0,63	0,54	0,09	0,08	0,09	0,08
Total	99,60	100,00	97,16	98,42	99,46	100,00	99,38	100,01	99,08	100,00
Anorthite	23,07	44	22,43	36	15,00	31	6,93	22	6,09	18
Albite	29,11	56	39,60	64	33,25	69	24,03	78	27,75	82

6.1.2 Trace Element Systematics

The rare earth element (REE) and trace element compositions discussed below are normalised using solar systems abundance of the elements based on meteorites compositions from Anders and Grevesse (1989) (Figure 6.3). The samples show similar patterns with varying amount of REE enrichment compared to the solar system composition. They are depleted in HREEs relative to the LREEs, and have negative slopes, indicating enrichment in LREEs relative to the HREEs. The anorthosites are generally enriched ($10\times$ chondrite) in REE and enriched in LREE with $[La/Sm]_N = 2.99-3.19\times$ chondrite and $[Dy/Yb]_N = 1.37-1.76\times$ chondrite, with a slightly flat concave-upward profile. The anorthosites have positive Eu anomalies ($[Eu/Eu^*]_N = 1.15-1.28$), with one exception (sample IRA 3-1) having a negative

anomaly ($[Eu/Eu^*]_N = 0.81$). The granodiorites are enriched in REEs relative to the anorthosites. The granodiorites (IRA6-3; IRA18-1; and IRA18-2) are enriched in LREE with $[La/Sm]_N = 3.18; 5.42; 3.64$ and $[Dy/Yb]_N = 1.41; 1.01; 1.47$, with $[Eu/Eu^*]_N = 0.64; 0.65; 0.75$ showing negative anomalies, respectively. The felsic dykes characterised as granitic-alaskites (IRA7-4 and IRA18-3) have concave-upward LREE profile with varying amount of HREE enrichment (Figure 6.3). They are enriched in LREE with $[La/Sm]_N = 12.07; 7.83$ and $[Dy/Yb]_N = 1.00; 1.00$, with the $[Eu/Eu^*]_N = 1.43; 1.46$, showing a positive anomaly, respectively. These chondrite-normalised ratios are summarised in Table 6.3.

Table 6.3: Whole-rock trace elements data in ppm, including REE data in ppm from ICPMS. Element ratios of trace elements normalised using Anders and Grevesse (1989). $[Eu/Eu^*]_N = \frac{Eu_N}{\sqrt{(Sm_N)(Gd_N)}}$

Rock Type	Anorthosite	Anorthosite	Anorthosite	Anorthosite	Anorthosite	McDoogle Pluton (granodiorite)	McDoogle Pluton (granodiorite)	Black Canyon pluton (granodiorite)	Felsic dyke (Alaskite)	Felsic dyke (Alaskite)
Sample ID	IRA3-1	IRA4-2	IRA8-5	IRA18-5	IRA18-7	IRA6-3	IRA18-1	IRA18-2	IRA7-4	IRA18-3
Trace elements										
Li	4.25	4.59	1.59	3.96	6.46	13.86	14.81	23.76	5.13	5.12
P	415.72	510.61	238.45	348.02	398.45	1113.25	1657.01	1318.55	132.92	183.85
Sc	1.75	1.73	1.08	3.66	3.03	7.80	14.13	7.94	0.57	1.31
Ti	1510.35	2423.52	947.41	2490.69	2255.72	4379.73	7453.09	6148.59	526.93	924.30
V	69.76	76.60	30.61	92.44	87.93	134.65	184.66	90.79	13.29	13.70
Cr	18.36	5.23	32.96	30.06	8.96	55.30	5.59	17.17	0.95	1.69
Co	5.29	7.83	3.29	10.06	10.75	20.77	21.03	12.32	1.65	2.92
Ni	13.43	8.33	6.03	14.63	8.45	33.78	13.04	11.65	1.74	3.83
Cu	6.61	20.25	4.03	37.57	36.67	57.38	8.13	33.23	2.79	7.13
Zn	35.33	27.47	10.05	30.70	21.67	93.61	128.71	82.80	12.15	18.37
Ga	12.84	17.75	16.48	19.59	19.50	16.62	24.98	19.78	13.12	14.98
Rb	10.61	6.57	5.64	9.10	14.50	78.34	66.73	91.62	172.81	213.95
Sr	383.78	342.44	402.01	716.84	742.85	354.03	585.01	297.77	180.05	384.12
Y	1.91	1.64	0.68	2.60	2.39	12.68	22.30	20.06	0.65	5.53
Zr	12.29	9.26	7.63	13.07	13.97	38.85	50.64	59.36	85.45	96.83
Nb	8.45	1.28	1.10	3.27	2.61	7.46	35.41	21.28	1.35	8.31
Ba	41.53	81.35	32.02	113.15	125.25	431.96	199.33	667.10	747.61	*bdl
Sn	0.90	0.48	0.30	1.13	0.67	1.33	6.58	4.17	0.28	0.64
Cs	0.48	0.43	0.27	1.44	2.36	4.35	1.59	6.23	3.84	3.52
Hf	0.45	0.32	0.28	0.49	0.43	1.30	2.30	1.75	2.57	2.76
Ta	3.36	0.55	0.06	0.20	0.14	0.40	1.84	1.10	0.10	1.24
W	10.68	0.58	0.44	0.74	0.19	0.68	0.41	0.49	0.12	0.24
Pb	2.09	3.55	2.32	2.90	3.11	11.71	10.87	13.72	20.99	24.42
Th	0.49	0.20	0.26	1.10	0.83	9.87	9.40	10.04	14.14	23.95
U	0.41	0.13	0.28	0.68	0.50	2.76	9.58	3.31	3.72	4.46
La	3.44	3.45	1.48	5.20	4.73	21.40	56.56	38.03	20.36	38.76
Ce	8.59	10.25	4.49	12.49	12.05	56.98	134.60	87.51	43.06	62.32
Pr	0.82	0.80	0.35	1.07	1.05	5.01	10.55	8.15	2.45	5.13
Nd	3.43	3.46	1.43	4.37	4.30	20.65	36.63	31.94	7.15	15.84
Pm	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Sm	0.66	0.67	0.27	0.87	0.84	3.90	6.04	6.05	0.98	2.86
Eu	0.18	0.27	0.11	0.33	0.31	0.83	1.34	1.48	0.42	1.21
Gd	0.63	0.66	0.24	0.82	0.79	3.84	6.38	5.79	0.79	2.16
Tb	0.08	0.08	0.03	0.10	0.09	0.47	0.70	0.71	0.05	0.20
Dy	0.42	0.40	0.15	0.53	0.49	2.59	3.90	3.94	0.15	0.99
Ho	0.08	0.07	0.03	0.10	0.09	0.49	0.75	0.73	0.02	0.18
Er	0.21	0.18	0.08	0.27	0.24	1.33	2.29	2.01	0.09	0.59
Tm	0.03	0.02	0.01	0.04	0.03	0.18	0.35	0.28	0.01	0.09
Yb	0.17	0.14	0.07	0.22	0.20	1.15	2.43	1.68	0.09	0.62
Lu	0.03	0.02	0.01	0.03	0.03	0.17	0.44	0.26	0.02	0.11
Elements ratios										
(Eu/Eu*) _N	0.81	1.22	1.28	1.15	1.15	0.64	0.65	0.75	1.43	1.46
(Sm/Gd) _N	2.21	2.27	0.88	2.90	2.79	13.25	21.27	20.27	3.01	8.51
(La/Sm) _N	3.01	2.99	3.19	3.45	3.25	3.18	5.42	3.64	12.07	7.83
(Dy/Yb) _N	1.54	1.75	1.37	1.52	1.58	1.41	1.01	1.47	1.00	1.00
*bdl=below detection limit										

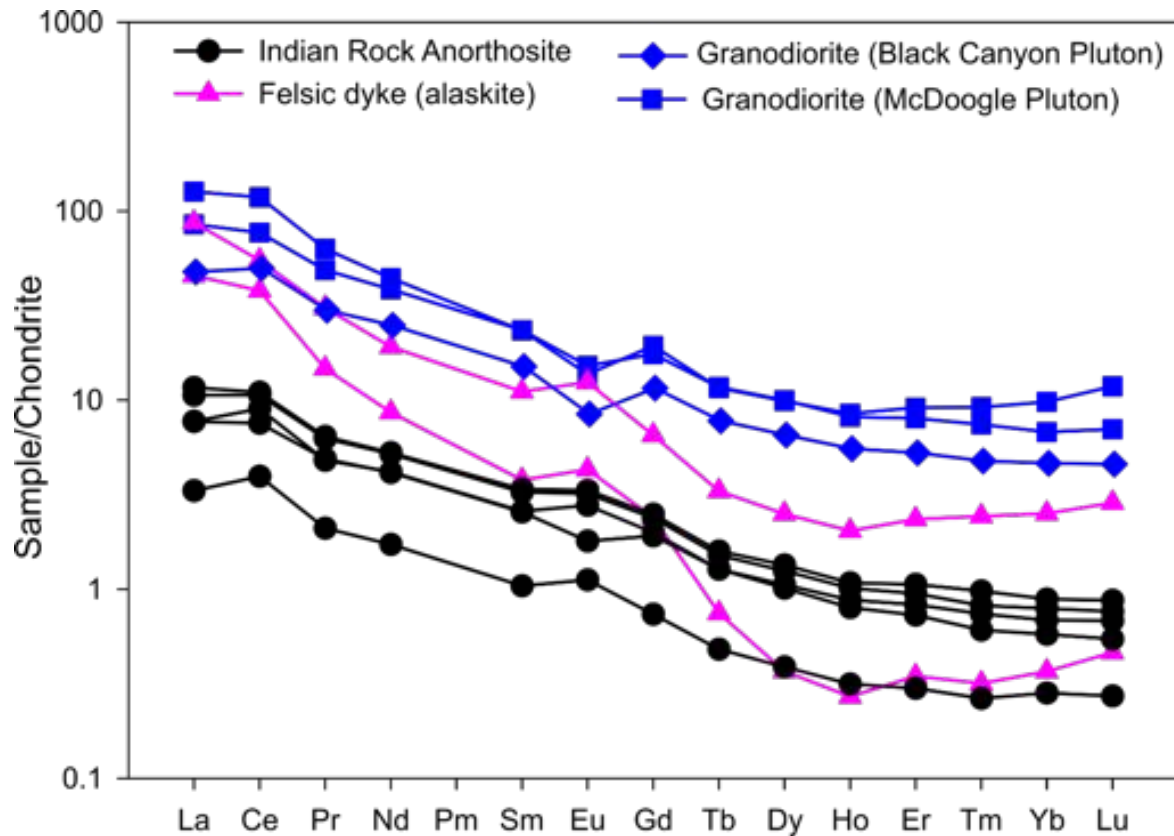


Figure 6.3: Chondrites-normalised REE diagram using chondrite values from Anders and Grevesse, (1989). All the analysed samples show a negative slope indicating enrichment in LREE. The granodiorites of the McDoogle and Black Canyon Pluton are more enriched in REEs compared to the anorthosites.

REE concentrations in the IRA are lower than the other rocks (granodiorites and felsic dykes) in the area. The anorthosite samples are relatively depleted in large ion lithophile elements (LILEs) such as Ba (32.02-125.25 ppm) and Rb (5.64-14.50 ppm) compared to the high concentration of Sr (342.44-742.85 ppm) (Table 6.1). The granodiorites have high HREE elements (Dy, Er and Yb) compared to other rock types. The high field-strength elements (HFSEs) such as Nd (1.43- 4.37 ppm), Ta (0.06- 3.36 ppm), and Hf (0.28- 0.49 ppm) also show depleted concentrations. Low Zr (7.63- 13.97 ppm) concentrations can be observed in the anorthosites compared to the granodiorite and the felsic dykes. Multi-element diagrams shown below are normalised using primitive mantle values from McDonough and Sun (1995). The anorthosites show enrichment in mobile elements and low La/Lu. They show enrichment in LILE and depletion in HFSE (e.g., Nb-Ta, P and Ti) (Figure 6.4). The anorthosites have a similar trend to the granodiorites. The anorthosite and granodiorite samples show positive U, Pb and Sr anomalies and negative Nb, Ta and P anomalies.

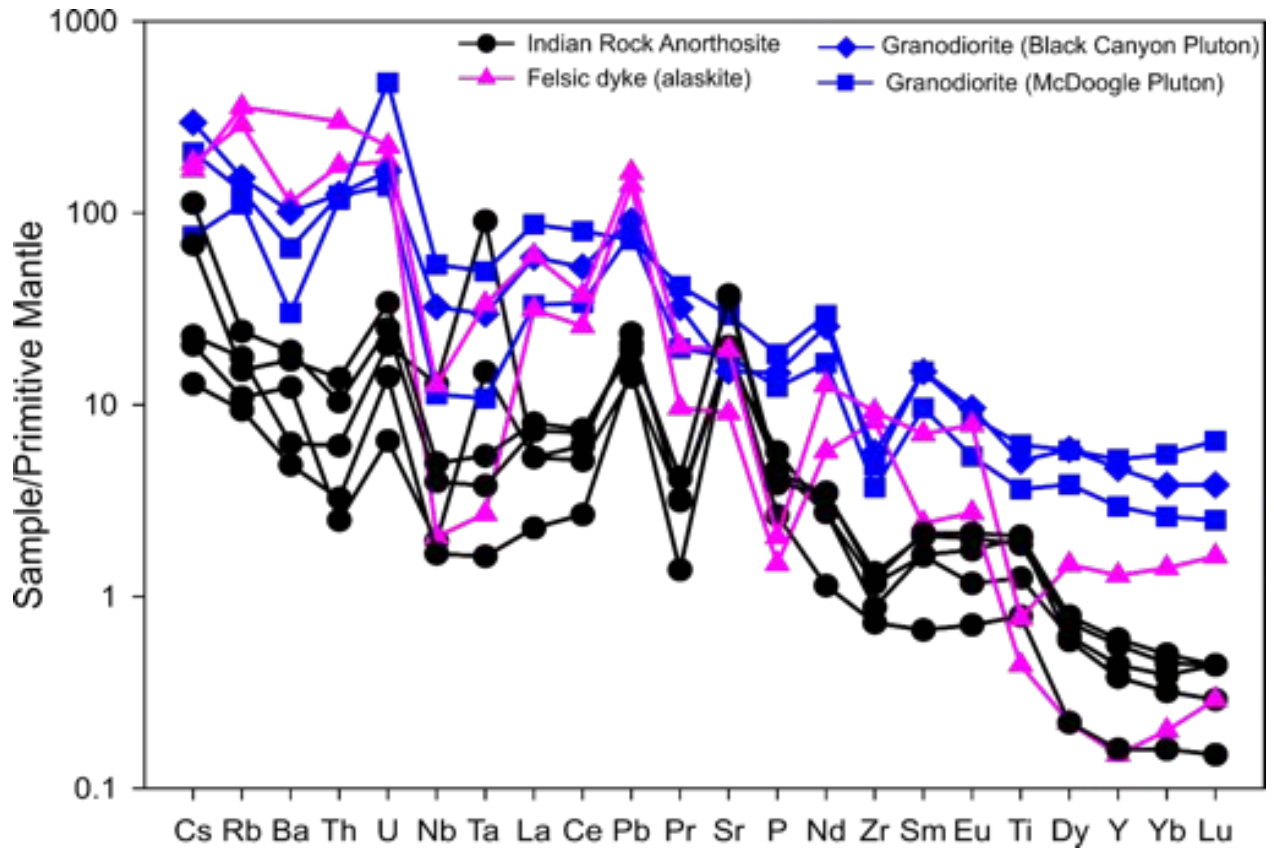


Figure 6.4: Spider diagram showing primitive mantle normalised plots. The primitive mantle values were taken from McDonough and Sun (1995).

6.2 Mineral Chemistry (plagioclase)

In-situ major element data for plagioclase grains are derived from both zoned (core and rim) and unzoned grains. The *in-situ* major element data were used to calculate the plagioclase anorthite content (An) using the mineral formula ($An\ molar = \frac{Ca}{(Ca+Na)} \times 100$). The minerals major element data is presented in Appendix B: Major and Trace Element data, with a summary in Table 6.4. *In-situ* major element composition of the major rock-forming mineral of the IRA (plagioclase) was obtained using the electron microprobe on selected samples (summarised in Table 6.4). Due to instrumental constraints, only plagioclase was analysed due to the importance of classifying anorthosite type. Plagioclase from the IRA shows high An content ranging from An₇₂ to An₈₉ (Figure 6.5). All textural varieties of anorthosite display similar normal/oscillatory zoned plagioclase grains patterns. The plagioclase grains occurring as inclusions in clinopyroxene oikocrysts have similar An content compared to the cumulates. The inclusions composition ranges from An₈₀ to An₈₈ in sample IRA18-5 and An₇₆ to An₈₅ in sample IRA18-7, while the cumulative plagioclase range from An₇₂ to An₈₉ generally. The core compositions of the cumulates have high An values ranging from An₈₂ to An₈₉, and the rims have values ranging from

An₇₂ to An₈₆. The plagioclase minerals are classified as bytownite, which has An content ranging from An₇₀₋₉₀ according to the feldspar classification (Figure 6.6). All the data are presented in Appendix C: EPMA data.

Table 6.4: Mean major element compositions of plagioclase and calculated An content data from the EPMA.

	n	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	Cr ₂ O ₃	BaO	Total	An	An range
IRA18-4a	8	46.98	0.01	33.94	0.29	0.01	0.00	17.73	1.68	0.02	0.01	0.01	100.68	85	75-88
IRA18-5	44	47.69	0.02	32.66	0.43	0.01	0.00	17.14	1.74	0.04	0.00	0.02	99.77	84	72-89
IRA18-7	49	47.36	0.03	34.01	0.16	0.01	0.00	15.90	1.78	0.04	0.01	0.23	99.52	83	79-88
IRA3-1	17	48.75	0.03	32.98	0.87	0.01	0.00	15.81	1.69	0.15	0.00	0.01	100.29	83	82-89
Values shown are averages of n point analyses															

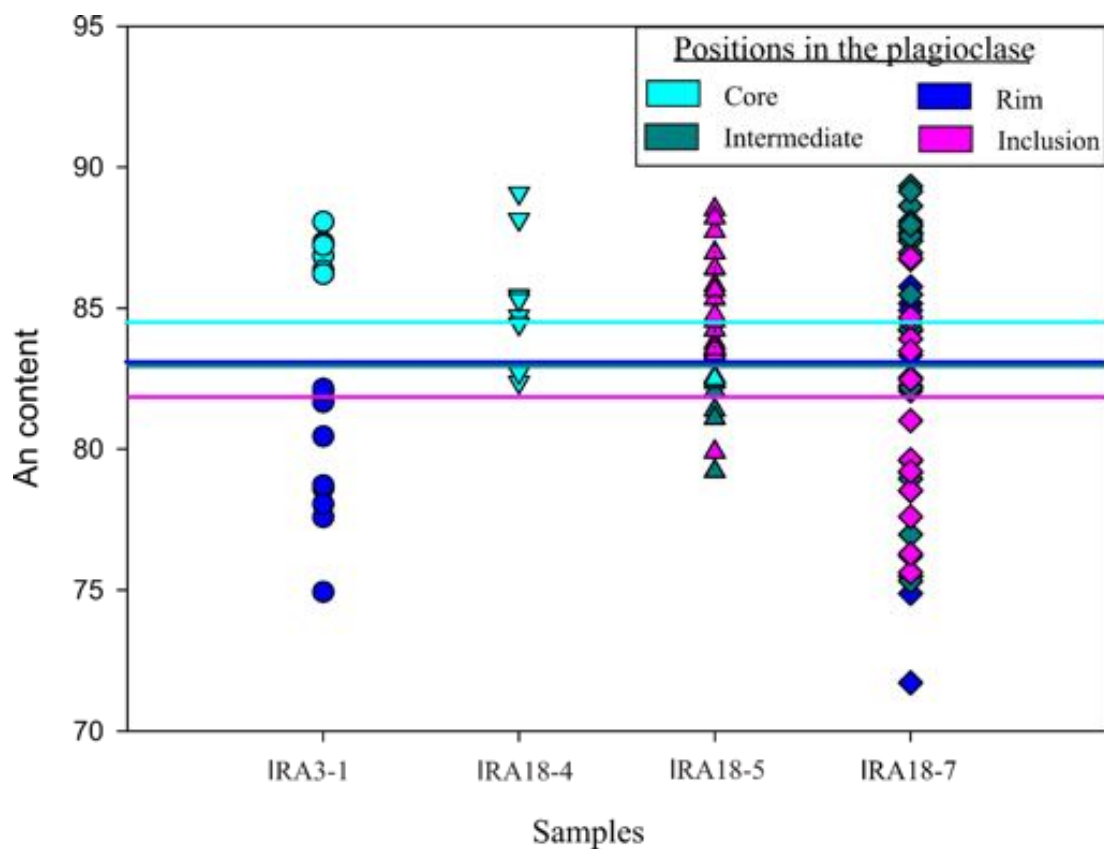


Figure 6.5: An content of plagioclase grains from the core to the rim and plagioclase inclusions in clinopyroxene oikocrysts. Solid horizontal lines show mean compositions for the different analysis positions. Core compositions (sky blue) are slightly more An rich than the rims (dark blue). The inclusions (pink) on average have lower An content compared to the cumulates.

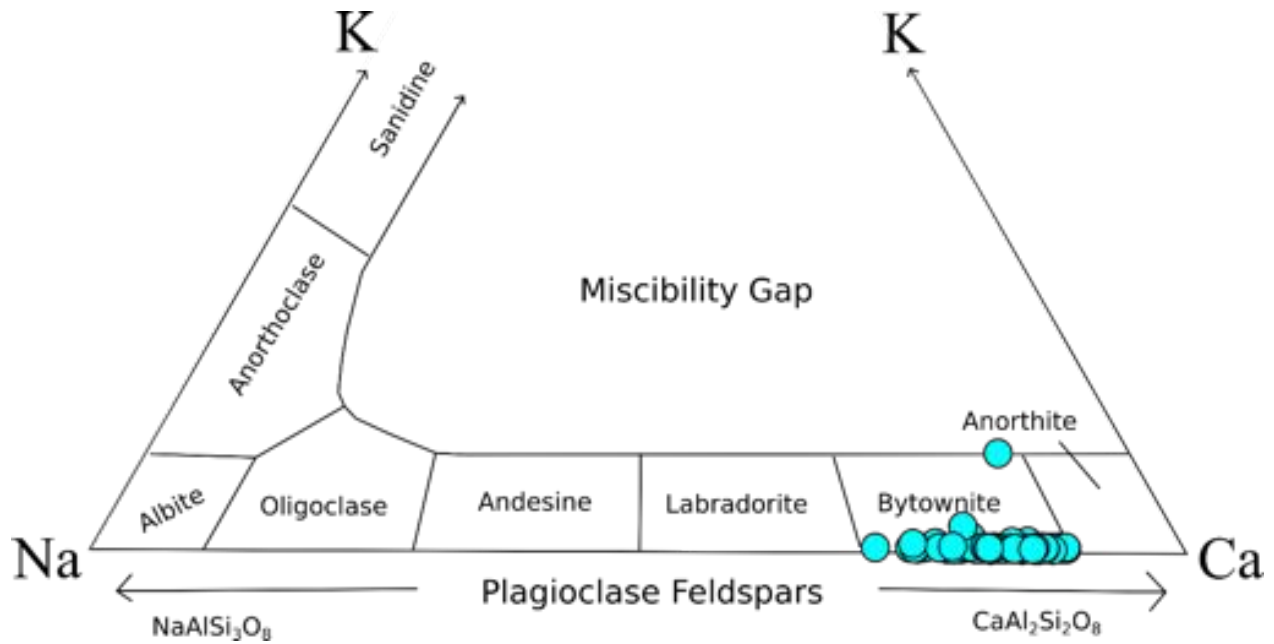


Figure 6.6: Classification of plagioclase minerals from in-situ data analysis of the Indian Rock Anorthosite.

6.3 Radiogenic Isotopes

6.3.1 Zircon and Apatite U-Pb Geochronology

Two anorthositic samples (IRA 4-2 and IRA 18-6) were selected for zircon U-Pb and apatite U-Pb geochronology. For both samples, the core and rims of the zircon, as well as different domains within apatite, yielded ages within error of each other. Weighted mean ^{238}U - ^{206}Pb ages are mostly used for young concordant zircons because this ratio is least sensitive to common Pb, therefore providing more reliable age determinations (Moore and Hopson, 1961). SEM-CL images of zircons from both samples exhibit magmatic growth textures with broad zones in the centre (Figure 6.7). The SEM-BSE images of the apatites show different domains, which can be differentiated by their reflectance and luminescence (Figure 6.7). Sample IRA 18-6 yielded a concordant ^{238}U - ^{206}Pb zircon age of 95.291 ± 0.056 Ma ($n=44$, MSWD= 0.052) and a weighted-mean $^{207}\text{Pb}/^{206}\text{Pb}$ of 95.36 ± 5.6 Ma ($n=44$, MSWD= 0.18) (Figure 6.8 a. c). Sample IRA 4-2 yielded a concordant ^{238}U - ^{206}Pb zircon age of 94.014 ± 0.051 Ma ($n=53$, MSWD= 0.2) with a weighted-mean $^{207}\text{Pb}/^{206}\text{Pb}$ of 95.12 ± 4.21 Ma ($n=53$, MSWD= 0.011) (Figure 6.8 b. d). Anorthosite IRA 18-6 yielded a ^{238}U - ^{206}Pb apatite age of 107.17 ± 0.19 Ma ($n=19$, MSWD= 0.86), with weighted-mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 107.21 ± 0.38 Ma ($n=19$, MSWD= 38.8) (Figure 6.9 a. c). Anorthosite IRA 4-2 yielded a ^{238}U - ^{206}Pb apatite age of 90.73 ± 0.19 Ma ($n=16$, MSWD= 0.46) with a weighted-mean $^{207}\text{Pb}/^{206}\text{Pb}$ of 91.30 ± 0.43 Ma ($n=16$, MSWD= 03.87) (Figure 6.9 b. d). The age data are summarised in Table 6.5 below.

Table 6.5: Summarised LA-ICPMS U-Pb and Pb-Pb zircon and apatite age data of the Indian Rock Anorthosite.

Sample ID	Zircon U-Pb age (Ma)	MSWD	Zircon weighed mean Pb/Pb age (Ma)	MSWD	<i>n</i>
IRA 4-2	94.014 ±0.051	0.2	95.12 ±4.21	0.011	53
IRA 18-6	95.291 ±0.056	0.052	95.36 ±5.60	0.18	44
Sample ID	Apatite U-Pb age (Ma)	MSWD	Apatite weighed mean Pb/Pb age (Ma)	MSWD	<i>n</i>
IRA 4-2	90.73 ±0.19	0.46	91.30 ±0.43	3.87	16
IRA 18-6	107.17 ±0.19	0.86	107.21 ±0.38	38.8	19

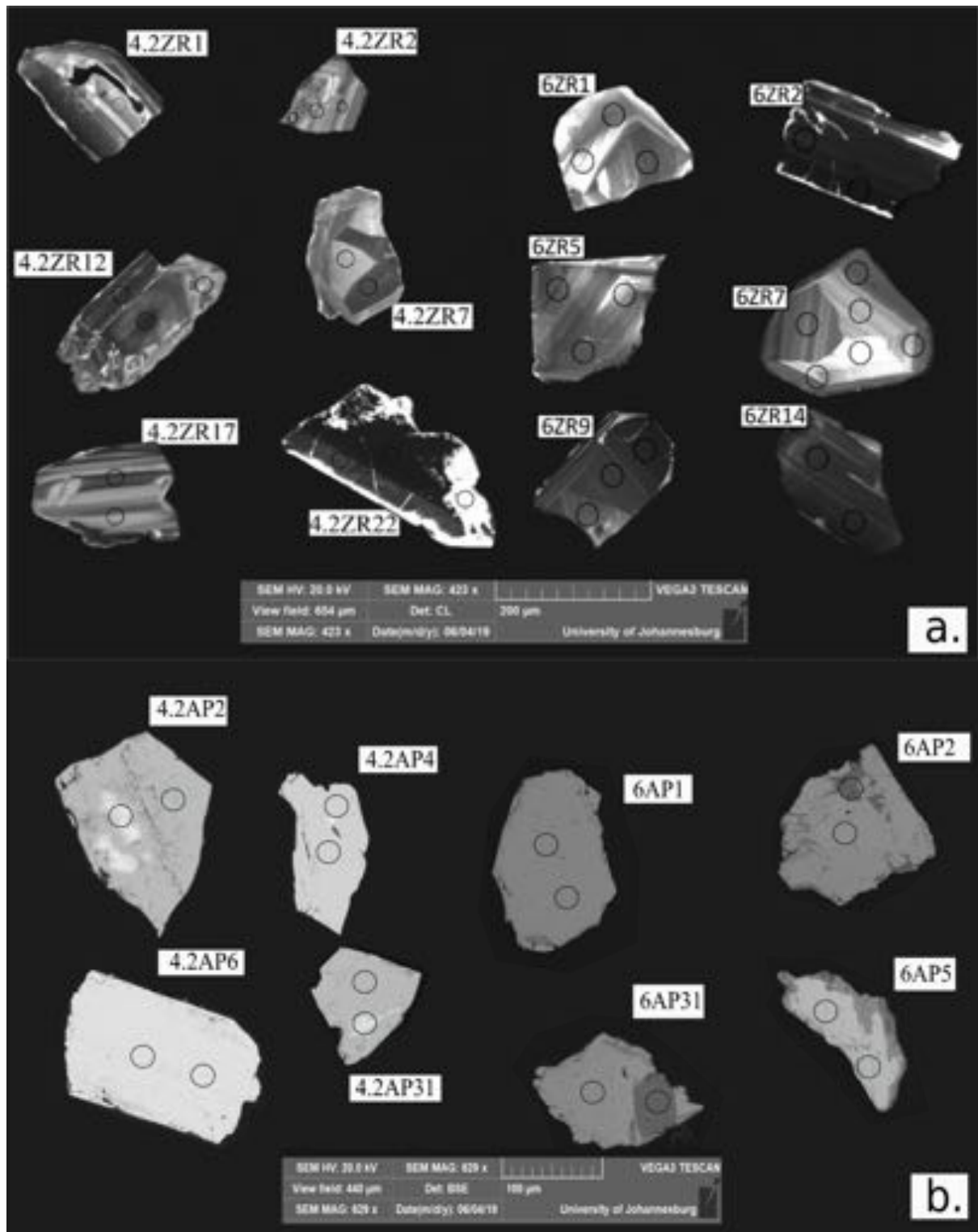


Figure 6.7: Cathodoluminescence (CL) and back-scattered electron (BSE) images of zircon and apatite from the Indian Rock Anorthosite, with open circles showing representative laser ablation sites before ablation. a. Zircons from sample IRA 4-2 and IRA 18-6 show regular oscillatory magmatic zoning. Zircon 6ZR7 shows a core rimmed by continuous oscillatory zones. The lighter zones indicate high U content. b. Apatite grains from IRA 4-2 and IRA 18-6. Apatite 4.2AP2, 4.2AP have different domains highlighted by various shades of grey within the grains.

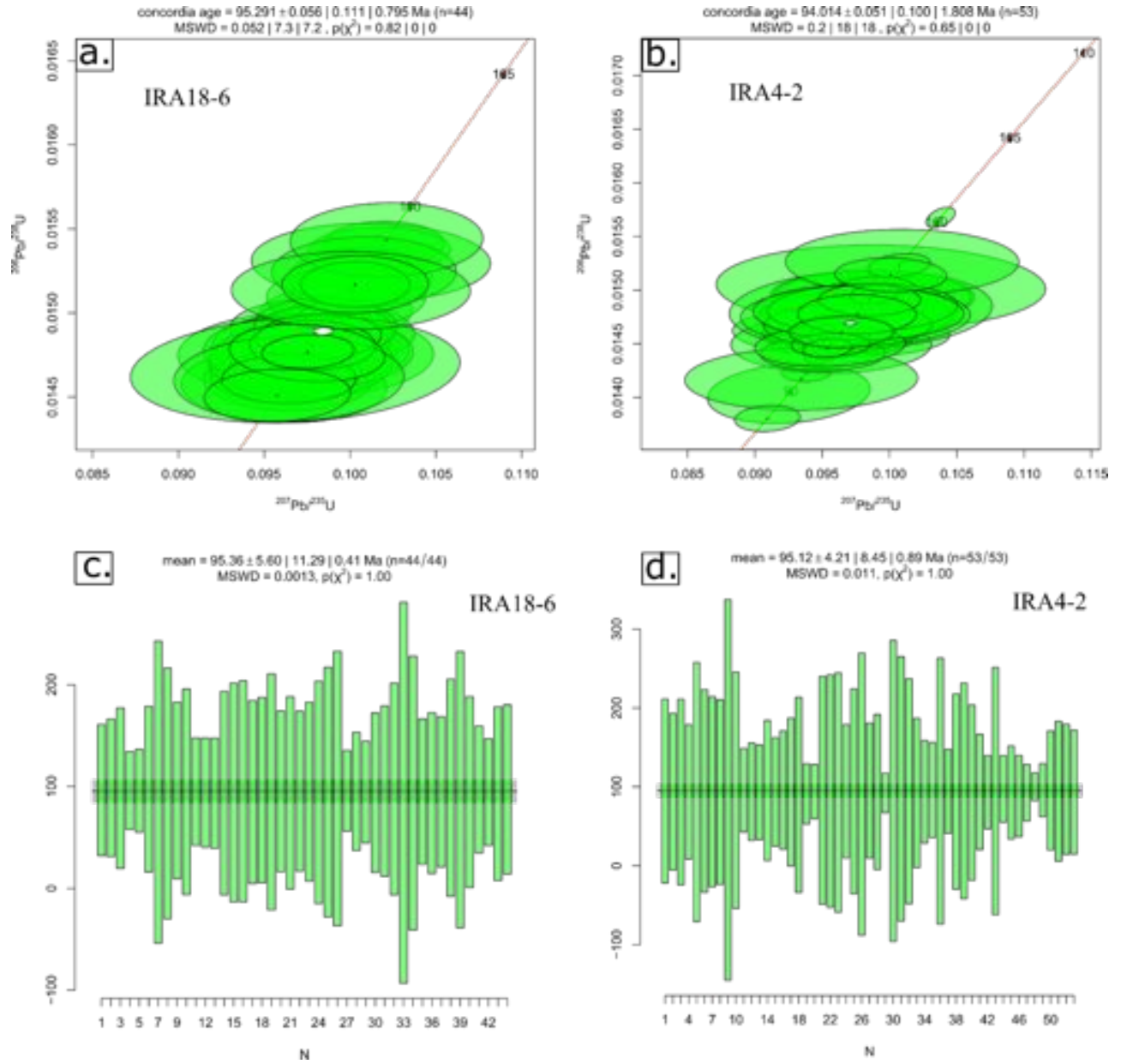


Figure 6.8: a. Zircon U-Pb Wetherill Concordia plots of the IRA showing the ^{238}U - ^{206}Pb ages and b. the weighed-mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages. A Stacey-Kramers common Pb correction (Chew et al., 2014) was used when plotting the Wetherill Concordia plots and weighted mean ages using IsoplotR.

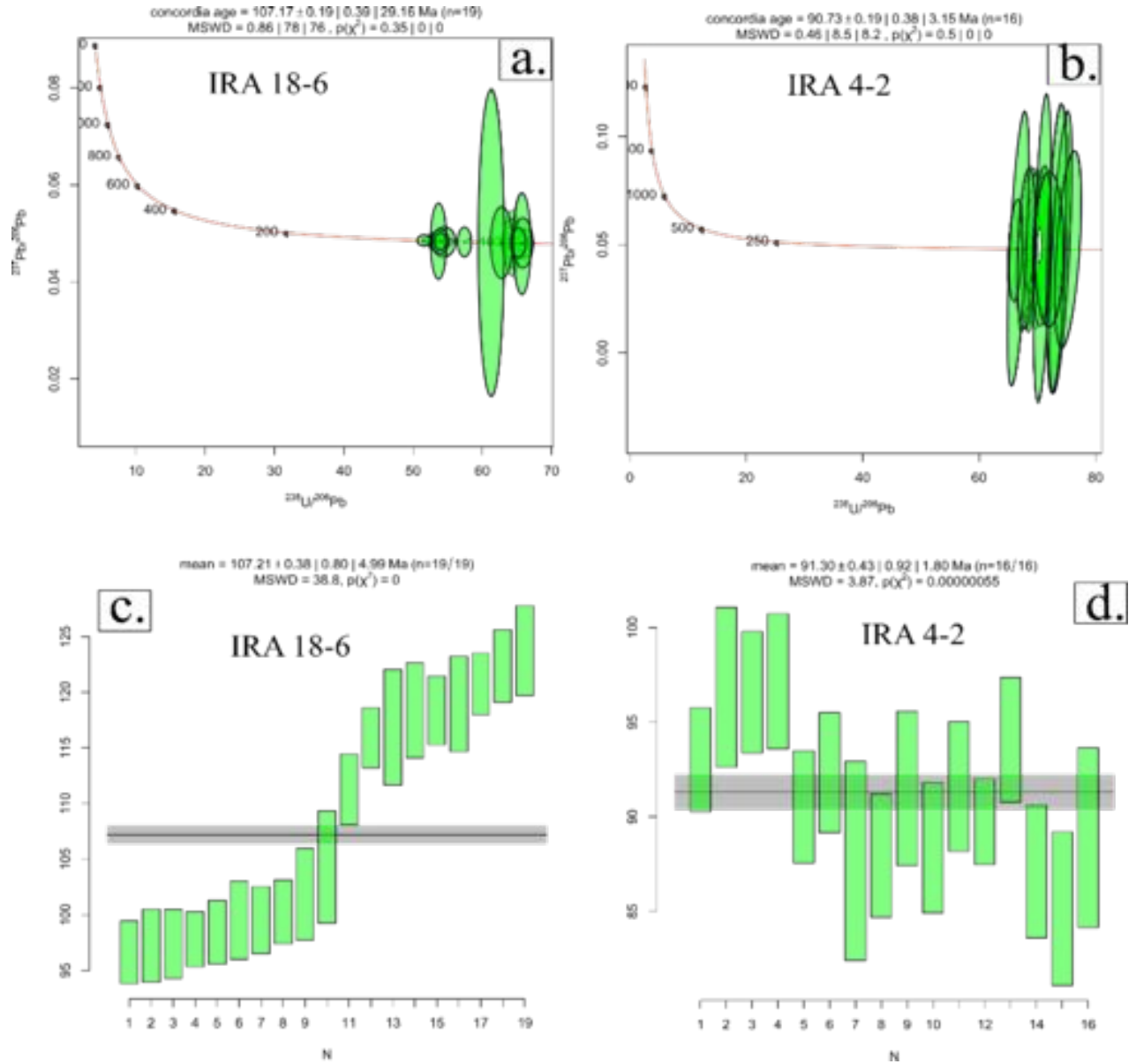


Figure 6.9: Apatite U-Pb Tera-Wasserburg plots of the IRA showing the ^{238}U - ^{206}Pb ages and the weighed-mean $^{207}\text{Pb}/^{206}\text{Pb}$. Stacey-Kramers common Pb correction (Chew et al., 2014) was applied when plotting the Tera Wasserburg Concordia plots and weighted mean ages using IsoplotR.

6.3.2 Rb-Sr and Sm-Nd whole-rock isotope geochemistry

Seven samples (three granodiorites and four anorthosites) were selected for Rb-Sr and Sm-Nd isotope analyses (Table 6.6 and Table 6.7). Whole-rock isotope ratios of the samples are plotted in Figure 6.10. The initial $^{87}\text{Sr}/^{86}\text{Sr}$ (I_{Sr}) ratios of the samples were calculated at 95 Ma, which is based on the newly determined ages of the IRA (previous section) and the range is 0.705926 to 0.707925 (Figure 6.10 a). The anorthosites have $I_{\text{Sr}}(95\text{Ma})$ of 0.705926 to 0.706306 and the granodiorites of the McDoogie Pluton have $I_{\text{Sr}}(95\text{Ma})$ ranging from 0.705926 to 0.707925 with an average of 0.706926. Granodiorite of the

Black Canyon Plutons (BCP) has $I_{\text{Sr}}(95\text{Ma})$ value of 0.706006. The granodiorite of the McDoogle Pluton (IRA 6-3) shows $I_{\text{Sr}}(95\text{Ma})$ value that falls within the range of the anorthosites (grey band in Figure 6.10 a).

Table 6.6: Rb/Sr ratio results from the MC-ICPMS analyses. The I_{Sr} values are calculated at 95 Ma.

Sample No:	Lithology Field name:	Rb (ppm)	%RSD	2 σ	Sr (ppm)	%RSD	2 σ	$^{87}\text{Rb}/^{86}\text{Sr}$	2 σ	$^{87}\text{Sr}/^{86}\text{Sr}$	2 σ	$I_{\text{Sr}}(95\text{Ma})$	2 σ
IRA3-1	Anorthosite	22.18782	0.002263	0.001004	1428.057	0.963054	27.50593	0.0449	0.00003	0.706366	0.00001	0.706306	0.00001
IRA8-5	Anorthosite	11.89048	0.001293	0.000308	1466.303	2.031792	59.58447	0.0235	0.00003	0.706341	0.00001	0.706310	0.00001
IRA18-5	Anorthosite	16.75251	0.003132	0.001049	1334.194	0.701397	18.71598	0.0363	0.00002	0.706298	0.00000	0.706250	0.00000
IRA18-7	Anorthosite	26.86415	0.002721	0.001462	1337.741	0.741883	19.84893	0.0581	0.00003	0.706349	0.00000	0.706271	0.00000
IRA6-3	McDoogle Pluton (granodiorite)	114.2864	0.132882	0.303733	633.6859	0.468116	5.932765	0.5216	0.00072	0.706617	0.00001	0.705926	0.00001
IRA18-1	McDoogle Pluton (granodiorite)	62.02181	0.058788	0.072923	648.6358	0.293927	3.813029	0.2766	0.00029	0.708292	0.00001	0.707925	0.00001
IRA18-2	Black Canyon Pluton (granodiorite)	181.8718	0.269889	0.981705	399.1887	0.205573	1.641244	1.3178	0.00207	0.707754	0.00001	0.706006	0.00001

The initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of all samples, calculated at 95 Ma and expressed as $\epsilon_{\text{Nd}}(95\text{Ma})$ values range from -7.07 to -3.66 (Table 6.7) (sample IRA8-5 being an outlier with 148.61). The anorthosites have $\epsilon_{\text{Nd}}(95\text{Ma})$ values ranging from -3.86 to -4.05 with an average of -3.96. The granodiorite of McDoogle Pluton values ranges from -3.66 to -7.06, with an average of -5.36 and the granodiorite of the BCP has a value of -3.93. The anorthosites and the granodiorites of McDoogle and Black Canyon Plutons (excluding Sample IRA 18-1) have $\epsilon_{\text{Nd}}(95\text{Ma})$ values that fall within the -4.05 to -3.66 range. Sample IRA 18-1 granodiorite has an $\epsilon_{\text{Nd}}(95\text{Ma})$ value of -7.07 (Figure 6.10 b). The granodiorite of McDoogle Pluton and BCP (IRA 6-3 and IRA 18-2 respectively) shows $\epsilon_{\text{Nd}}(95\text{Ma})$ values that fall within the range of the anorthosites (grey band) (Figure 6.10 b). The anorthosites have $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of 0.512377 to 0.512391. The granodiorites have values ranging from 0.512398 to 0.512384 for Black Canyon Pluton and McDoogle Pluton (samples IRA 6-3 and IRA 18-2 respectively). Granodiorite of McDoogle Pluton (IRA 18-2) shows low $^{143}\text{Nd}/^{144}\text{Nd}$ ratio compared to the other samples. Nd model ages (T_{DM}) are calculated from Workman and Hart (2005) depleted mantle evolution (Table 6.7). The depleted mantle evolution is expressed as present-day MORB ratios of $^{143}\text{Nd}/^{144}\text{Nd} = 0.51313$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.2485$. The ages range from 1135.56 to 1280.24 Ma, with an average of 1191.98 Ma (Figure 6.11). The T_{CHUR} ages range from 436.32 to 683.49 Ma. Sample IRA 18-1 has high T_{DM} (1280.24 Ma) and T_{CHUR} (683.49 Ma) compared to the other samples. $\epsilon_{\text{Nd}}(95\text{Ma})$ vs $^{87}\text{Sr}/^{86}\text{Sr}(95\text{Ma})$ and $^{143}\text{Nd}/^{144}\text{Nd}$ vs $^{87}\text{Sr}/^{86}\text{Sr}(95\text{Ma})$ diagrams (Figure 6.10 c. d) show the anorthosites and the granodiorite of McDoogle Pluton (IRA 6-3) values that do not depart from each other.

Table 6.7: Sm/Nd ratio results from the MC-ICPMS. The initial Nd and ϵ_{Nd} values are calculated at 95 Ma.

Sample No:	Lithology Field name:	$^{147}\text{Sm}/^{144}\text{Nd}$	2σ	$^{143}\text{Nd}/^{144}\text{Nd}$	2σ	I_{Nd} (95Ma)	2σ	ϵ_{Nd} (95Ma)	2σ	T_{CHUR} (Ma)	T_{DM} (Ma)
IRA3-1	Anorthosite	0.1183	0.00011	0.512391	0.00001	0.512318	0.00001	-3.86	0.262	480.67	1211.58
*IRA8-5	Anorthosite	0.0973	0.00059	0.520193	0.00027	0.520132	0.00027	148.61	5.334	-12081.33	-9539.56
IRA18-5	Anorthosite	0.1200	0.00004	0.512390	0.00002	0.512316	0.00002	-3.90	0.309	492.53	1234.14
IRA18-7	Anorthosite	0.1114	0.00006	0.512377	0.00002	0.512308	0.00002	-4.05	0.445	466.49	1150.40
IRA6-3	McDoogle Pluton (granodiorite)	0.1129	0.00007	0.512398	0.00001	0.512328	0.00001	-3.66	0.144	436.32	1135.56
IRA18-1	McDoogle Pluton (granodiorite)	0.1027	0.00008	0.512217	0.00001	0.512153	0.00001	-7.07	0.147	683.49	1280.24
IRA18-2	Black Canyon Pluton (granodiorite)	0.1113	0.00005	0.512384	0.00001	0.512314	0.00001	-3.93	0.114	454.78	1139.98

* Sample IRA8-5 values deviates from the rest of the data, it's a possible analytical error with problematic Sm/Nd ratio. It will be excluded in the plots.

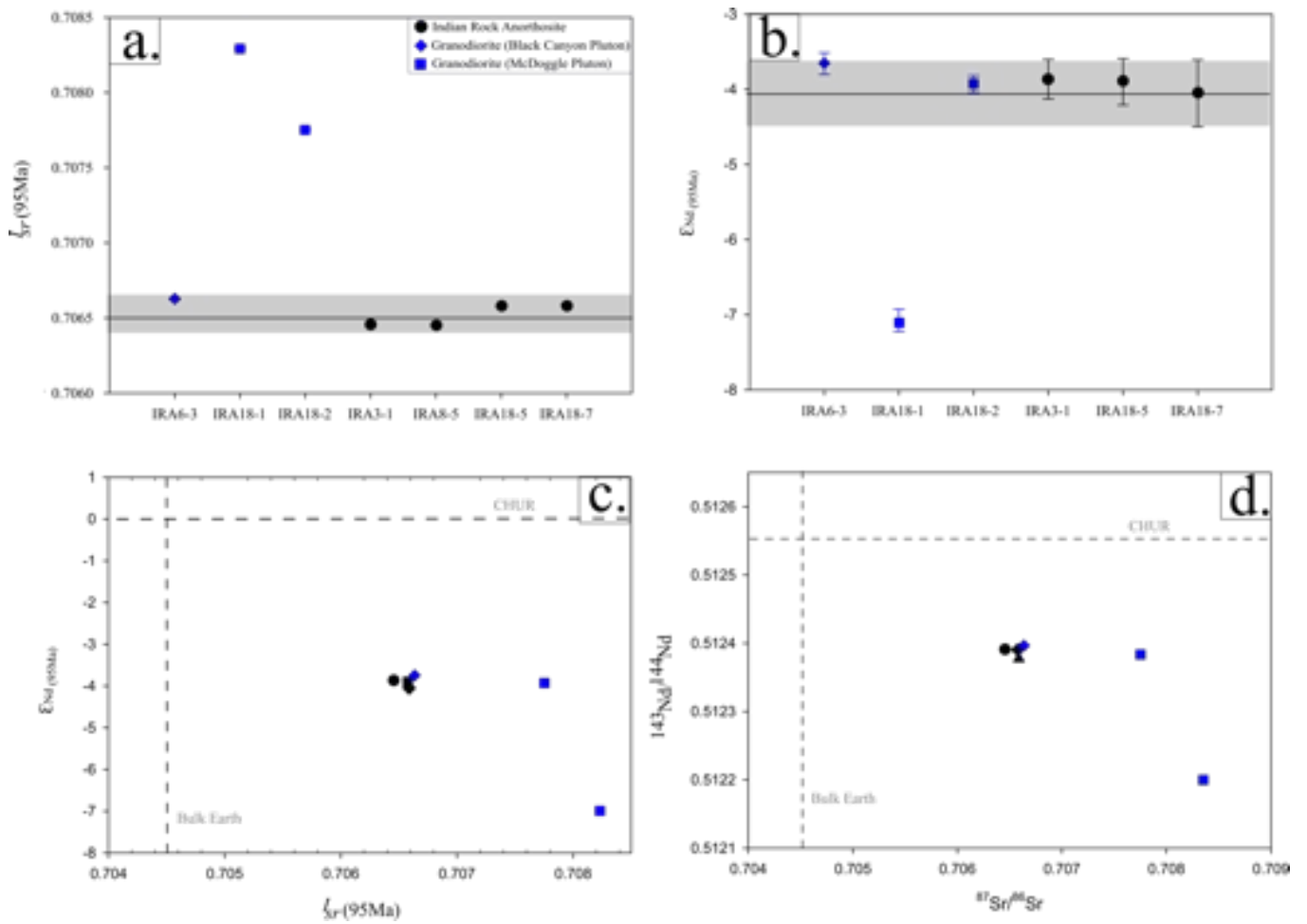


Figure 6.10: $I_{Sr}(95Ma)$ vs Samples, $\epsilon_{Nd}(95Ma)$ vs Samples, $\epsilon_{Nd}(95Ma)$ vs $^{87}Sr/^{86}Sr(95Ma)$, and $^{143}Nd/^{144}Nd$ vs $^{87}Sr/^{86}Sr(95Ma)$. a. McDoogle Plutons' granodiorite (IRA 6-3) shows $I_{Sr}(95Ma)$ values that fall within the range of the anorthosites (grey band). b. McDoogle and Black Canyon Plutons' granodiorite (IRA 6-3 and IRA 18-2 respectively) shows $\epsilon_{Nd}(95Ma)$ values that fall within the range of the anorthosites (grey band). c. $\epsilon_{Nd}(95Ma)$ vs $I_{Sr}(95Ma)$ plot. d. Initial Nd vs initial Sr isotope plot at 95 Ma. Values for lines labelled Bulk Earth (BE) and CHUR (chondritic uniform reservoir) are calculated at 95Ma, assuming BE: $^{87}Sr/^{86}Sr = 0.7045$ and $^{87}Rb/^{86}Sr = 0.083$, and CHUR: $^{143}Nd/^{144}Nd = 0.512638$ and $^{147}Sm/^{144}Nd = 0.1967$ (Workman and Hart, 2005). The plotted error bars are 2s and they are smaller than the symbols in the diagrams.

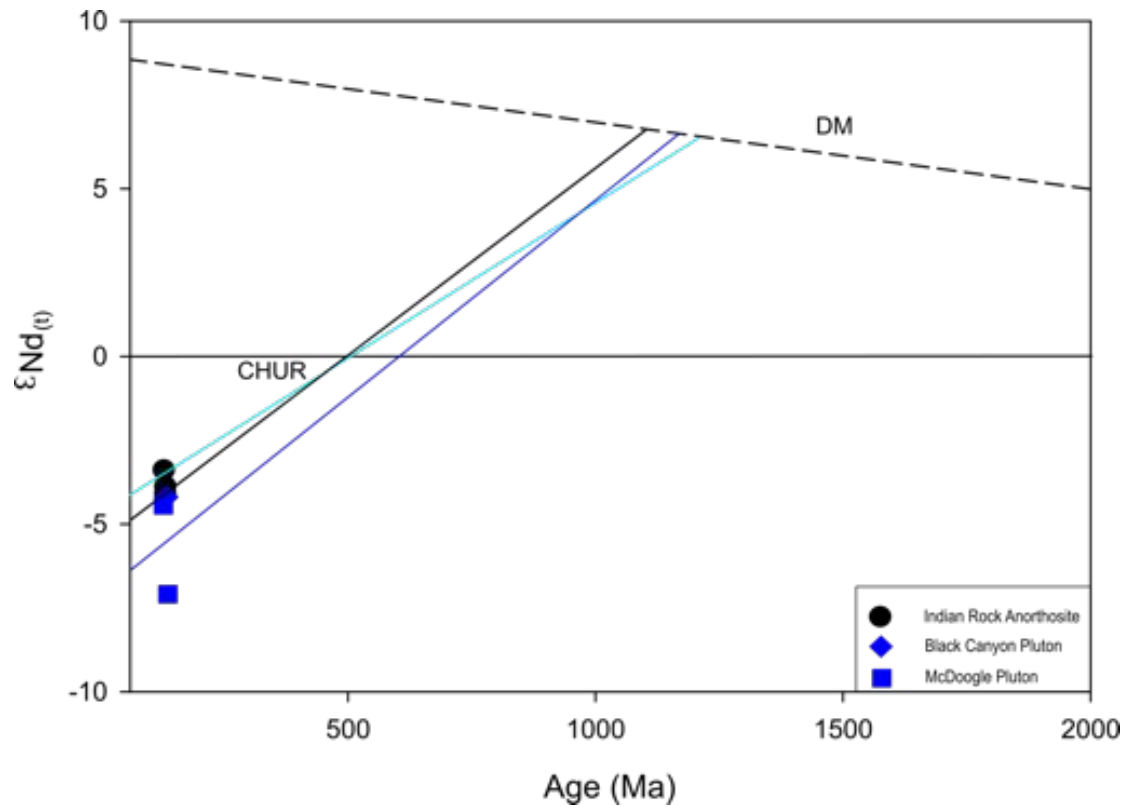


Figure 6.11: Whole rock $\epsilon_{Nd(95Ma)}$ vs crystallization age diagram. The evolution-depleted mantle (DM) is calculated back from present ratios of $^{87}Sr/^{86}Sr = 0.7045$ and $^{87}Rb/^{86}Sr = 0.083$ (Workman and Hart, 2005). The light blue regression line is the Indian Rock Anorthosite, the black regression line is the Black Canyon Pluton and the dark blue regression line is the McDoogie Pluton.

6.3.3 In-situ Zircon Lu-Hf isotope ratios

Zircon Lu-Hf isotopic compositions were measured using LA-ICPMS directly on top of the previously excavated U-Pb spots. The analysed *in-situ* zircon Lu-Hf isotopic compositions are listed in Table 6.8 for two anorthosite samples (IRA 4-2 and IRA 18-6). Sixty-seven spots were analysed for both samples (38 for IRA 4-2 and 29 for IRA 18-6). The $^{176}Lu/^{177}Hf$ ratios are relatively homogenous with value ranging between 0.0004 and 0.0053 (average = 0.0014, $2\sigma = 0.0001$) for sample IRA 4-2 and for sample IRA 18-6, they range between 0.0004 and 0.002 (average = 0.0014, $2\sigma = 0.0001$). The $^{176}Hf/^{177}Hf$ ratio has an average of 0.282594, $2\sigma = 0.0001$ and 0.282596, $2\sigma = 0.0001$ for sample IRA 4-2 and IRA 18-6 respectively (Figure 6.12). The calculated $\epsilon_{Hf(95Ma)}$ values are negative and they range from -8.05 to -4.12 (average = -4.74, $2\sigma = 0.298$) for sample IRA 4-2 and for sample IRA 18-6, the values range from -5.21 to -4.03 (average = -4.67, $2\sigma = 0.502$). Although the $\epsilon_{Hf(95Ma)}$ values of the two samples broadly overlap within errors, sample IRA 4-2 has a higher range, while sample IRA 18-6 has a narrower range. The Hf depleted mantle ($T_{DM(95Ma)}$) model ages are between 897.67 and 1134.50 Ma for sample IRA 4-

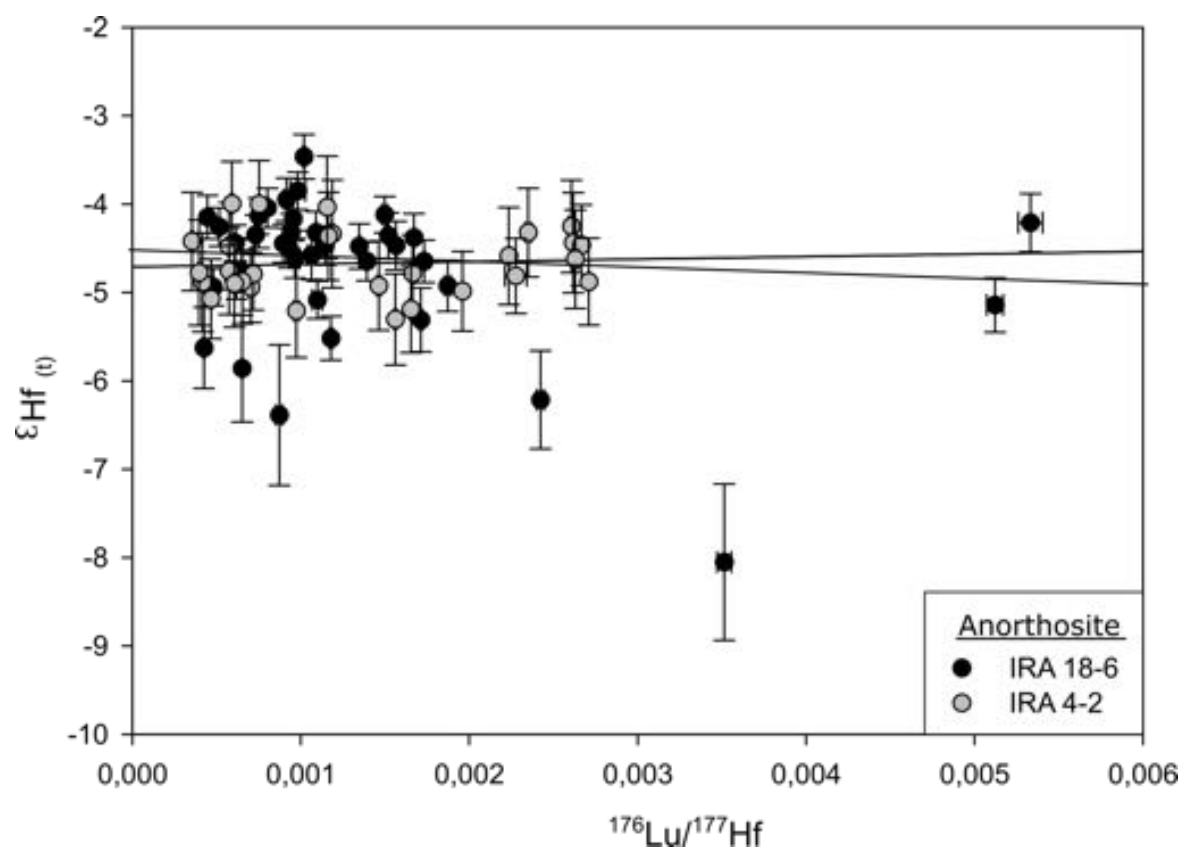
2, and for sample IRA 18-6 range between 898.82 and 979.12 Ma. The chondritic uniform reservoir ($T_{\text{CHUR}}(95\text{Ma})$) model ages are between 255.64 and 498.47 Ma for sample IRA 4-2, and for sample IRA 18-6 range between 277.86 and 340.41 Ma. Sample IRA 4-2 has a large amount of evolved zircon with lower $\epsilon_{\text{Hf}}(95\text{Ma})$ values than sample IRA 18-6.

Table 6.8: $^{176}\text{Lu}/^{177}\text{Hf}$, $^{176}\text{Hf}/^{177}\text{Hf}$, I_{Hf} and $\epsilon_{\text{Hf}}(95\text{ Ma})$ values including $T_{\text{CHUR}}(95\text{ Ma})$ ages of sample IRA 4-2 and IRA 18-6.

Sample	$^{176}\text{Lu}/^{177}\text{Hf}$	2σ	$^{176}\text{Hf}/^{177}\text{Hf}$	2σ	I_{Hf} at 95	2σ	ϵ_{Hf} at 95	2σ	T_{CHUR} (Ma)	T_{DM} (Ma)
4.2ZR1	0.000612	0.00001	0.282601	0.00001	0.282600	0.00001	-4.44	0.206	298.41	913.22
4.2ZR1-2	0.001098	0.00003	0.282584	0.00001	0.282582	0.00001	-5.08	0.212	331.23	949.46
4.2ZR2	0.001498	0.00002	0.282612	0.00001	0.282609	0.00001	-4.12	0.206	288.84	919.67
4.2ZR3	0.000516	0.00000	0.282606	0.00001	0.282605	0.00001	-4.26	0.217	289.49	903.97
4.2ZR3-2	0.000446	0.00000	0.282609	0.00001	0.282608	0.00001	-4.15	0.245	284.05	898.14
4.2ZR3-3	0.000618	0.00000	0.282589	0.00001	0.282588	0.00001	-4.86	0.216	317.54	929.83
4.2ZR4	0.001519	0.00001	0.282605	0.00001	0.282602	0.00001	-4.36	0.260	300.18	929.78
4.2ZR5	0.000982	0.00005	0.282618	0.00001	0.282617	0.00001	-3.85	0.213	273.34	897.67
4.2ZR5-2	0.000964	0.00003	0.282597	0.00001	0.282595	0.00001	-4.61	0.224	308.48	927.65
4.2ZR6	0.000803	0.00000	0.282613	0.00001	0.282611	0.00001	-4.04	0.223	281.28	901.55
4.2ZR7	0.001392	0.00000	0.282597	0.00001	0.282594	0.00001	-4.65	0.219	312.90	938.57
4.2ZR8	0.001732	0.00000	0.282597	0.00001	0.282594	0.00001	-4.65	0.240	315.23	946.34
4.2ZR9-2	0.001064	0.00002	0.282598	0.00001	0.282596	0.00001	-4.58	0.276	307.46	928.42
4.2ZR9-3	0.000652	0.00000	0.282561	0.00002	0.282560	0.00002	-5.86	0.605	363.46	970.00
4.2ZR9-4	0.000425	0.00001	0.282567	0.00001	0.282566	0.00001	-5.63	0.460	351.03	955.67
4.2ZR11-1	0.000648	0.00000	0.282592	0.00001	0.282591	0.00001	-4.76	0.239	313.14	926.52
4.2ZR11-2	0.000896	0.00001	0.282601	0.00001	0.282600	0.00001	-4.45	0.215	300.34	919.51
4.2ZR12	0.001347	0.00001	0.282601	0.00001	0.282599	0.00001	-4.48	0.254	304.70	930.76
4.2ZR13	0.000915	0.00002	0.282615	0.00001	0.282614	0.00001	-3.95	0.242	277.50	900.14
4.2ZR14	0.000929	0.00001	0.282600	0.00001	0.282598	0.00001	-4.50	0.223	302.93	922.28
4.2ZR14-2	0.000751	0.00001	0.282610	0.00001	0.282608	0.00001	-4.14	0.206	285.21	904.09
4.2ZR15	0.000482	0.00000	0.282587	0.00001	0.282586	0.00001	-4.94	0.210	320.26	930.00
4.2ZR16	0.001020	0.00001	0.282629	0.00001	0.282627	0.00001	-3.46	0.251	255.64	883.04
4.2ZR16-2	0.001566	0.00001	0.282602	0.00001	0.282599	0.00001	-4.47	0.274	305.77	935.38
4.2ZR17	0.001873	0.00002	0.282590	0.00001	0.282586	0.00001	-4.92	0.289	329.38	960.90
4.2ZR17-2	0.000873	0.00002	0.282546	0.00002	0.282545	0.00002	-6.39	0.796	389.50	996.00
4.2ZR18-1	0.001091	0.00001	0.282605	0.00001	0.282603	0.00001	-4.32	0.240	295.90	918.90
4.2ZR18-2	0.001179	0.00001	0.282572	0.00001	0.282569	0.00001	-5.52	0.248	351.94	968.62
4.2ZR19-1	0.001145	0.00000	0.282601	0.00001	0.282599	0.00001	-4.47	0.221	302.97	925.89
4.2ZR19-2	0.000731	0.00001	0.282604	0.00001	0.282603	0.00001	-4.35	0.224	294.78	912.03
4.2ZR19-3	0.002423	0.00002	0.282554	0.00002	0.282550	0.00002	-6.21	0.555	395.82	1027.40
4.2ZR19-4	0.003514	0.00004	0.282504	0.00003	0.282498	0.00003	-8.05	0.884	498.47	1134.50
4.2ZR20	0.005332	0.00007	0.282616	0.00001	0.282606	0.00001	-4.21	0.329	319.99	1018.61
4.2ZR20-1	0.005122	0.00005	0.282589	0.00001	0.282580	0.00001	-5.14	0.307	367.59	1054.45
4.2ZR21	0.001672	0.00000	0.282605	0.00001	0.282602	0.00001	-4.38	0.273	302.27	934.18
4.2ZR21-1	0.000956	0.00002	0.282609	0.00001	0.282608	0.00001	-4.17	0.248	287.82	909.72
4.2ZR22	0.001713	0.00001	0.282578	0.00001	0.282575	0.00001	-5.31	0.361	346.45	972.84
4.2ZR23	0.000943	0.00001	0.282605	0.00001	0.282603	0.00001	-4.33	0.217	295.06	915.73
Average	0.001354	0.00001	0.282594	0.00001	0.282591	0.00001	-4.74	0.298	317.53	942.41

Table 6.9: $^{176}\text{Lu}/^{177}\text{Hf}$, $^{176}\text{Hf}/^{177}\text{Hf}$, I_{Hf} and ϵ_{Hf} (95 Ma) values including T_{CHUR} (95 Ma) ages of sample IRA4-2 and IRA 18-6.

Sample	$^{176}\text{Lu}/^{177}\text{Hf}$	2σ	$^{176}\text{Hf}/^{177}\text{Hf}$	2σ	I_{Hf} at 95	2σ	ϵ_{Hf} at 95	2σ	T_{CHUR} (Ma)	T_{DM} (Ma)
6ZR1	0.002607	0.00002	0.2826	0.00001	0.282605	0.00001	-4.25	0.523	302.25	950.57
6ZR1-1	0.002350	0.00002	0.282607	0.00001	0.282603	0.00001	-4.32	0.501	303.87	947.36
6ZR2	0.002618	0.00000	0.282605	0.00002	0.282600	0.00002	-4.44	0.567	311.30	958.51
6ZR3	0.002667	0.00000	0.282604	0.00001	0.282599	0.00001	-4.47	0.457	313.00	960.85
6ZR3-1	0.000574	0.00000	0.282600	0.00001	0.282599	0.00001	-4.48	0.492	299.68	913.70
6ZR4	0.000591	0.00001	0.282614	0.00001	0.282612	0.00001	-4.00	0.476	277.86	895.13
6ZR4-2	0.001664	0.00001	0.282593	0.00001	0.282590	0.00001	-4.78	0.435	321.22	950.32
6ZR5	0.001187	0.00001	0.282605	0.00002	0.282603	0.00002	-4.34	0.611	297.11	921.55
6ZR5-1	0.000418	0.00000	0.282588	0.00002	0.282587	0.00002	-4.89	0.558	317.39	926.50
6ZR6	0.000468	0.00001	0.282583	0.00001	0.282582	0.00001	-5.07	0.450	326.23	934.92
6ZR6-1	0.000354	0.00001	0.282601	0.00002	0.282600	0.00002	-4.42	0.557	295.94	906.94
6ZR7	0.000574	0.00002	0.282592	0.00001	0.282591	0.00001	-4.76	0.491	312.59	924.85
6ZR7-1	0.002710	0.00000	0.282592	0.00001	0.282587	0.00001	-4.88	0.493	333.46	979.12
6ZR8	0.002630	0.00002	0.282599	0.00002	0.282595	0.00002	-4.62	0.558	320.38	966.49
6ZR9	0.001165	0.00002	0.282604	0.00001	0.282602	0.00001	-4.36	0.497	298.22	922.13
6ZR9-1	0.001463	0.00001	0.282589	0.00001	0.282586	0.00001	-4.92	0.504	326.36	951.34
6ZR10	0.000704	0.00001	0.282587	0.00001	0.282586	0.00001	-4.94	0.396	321.92	934.99
6ZR11	0.000718	0.00000	0.282591	0.00001	0.282590	0.00001	-4.79	0.407	315.10	929.34
6ZR11-1	0.001960	0.00000	0.282588	0.00001	0.282584	0.00001	-4.99	0.450	332.98	965.49
6ZR12	0.002235	0.00000	0.282600	0.00002	0.282596	0.00002	-4.59	0.550	315.84	955.59
6ZR12-1	0.000653	0.00001	0.282589	0.00001	0.282588	0.00001	-4.88	0.477	318.52	931.25
6ZR13	0.000753	0.00000	0.282614	0.00001	0.282612	0.00001	-4.00	0.496	279.06	898.82
6ZR13-1	0.001654	0.00001	0.282582	0.00001	0.282579	0.00001	-5.19	0.490	340.31	966.56
6ZR13-2	0.001562	0.00001	0.282578	0.00001	0.282575	0.00001	-5.31	0.516	344.99	969.03
6ZR14	0.002277	0.00006	0.282593	0.00001	0.282589	0.00001	-4.81	0.425	326.99	965.88
6ZR14-1	0.000397	0.00001	0.282591	0.00002	0.282590	0.00002	-4.78	0.596	312.22	921.69
6ZR15	0.000974	0.00002	0.282580	0.00001	0.282578	0.00001	-5.21	0.527	336.02	951.54
6ZR15-1	0.000607	0.00001	0.282588	0.00001	0.282587	0.00001	-4.90	0.495	319.21	931.09
6ZR16	0.001156	0.00000	0.282613	0.00002	0.282611	0.00002	-4.03	0.579	282.85	908.76
Average	0.001369	0.000011	0.282596	0.000014	0.282593	0.000014	-4.67	0.502	313.89	939.32



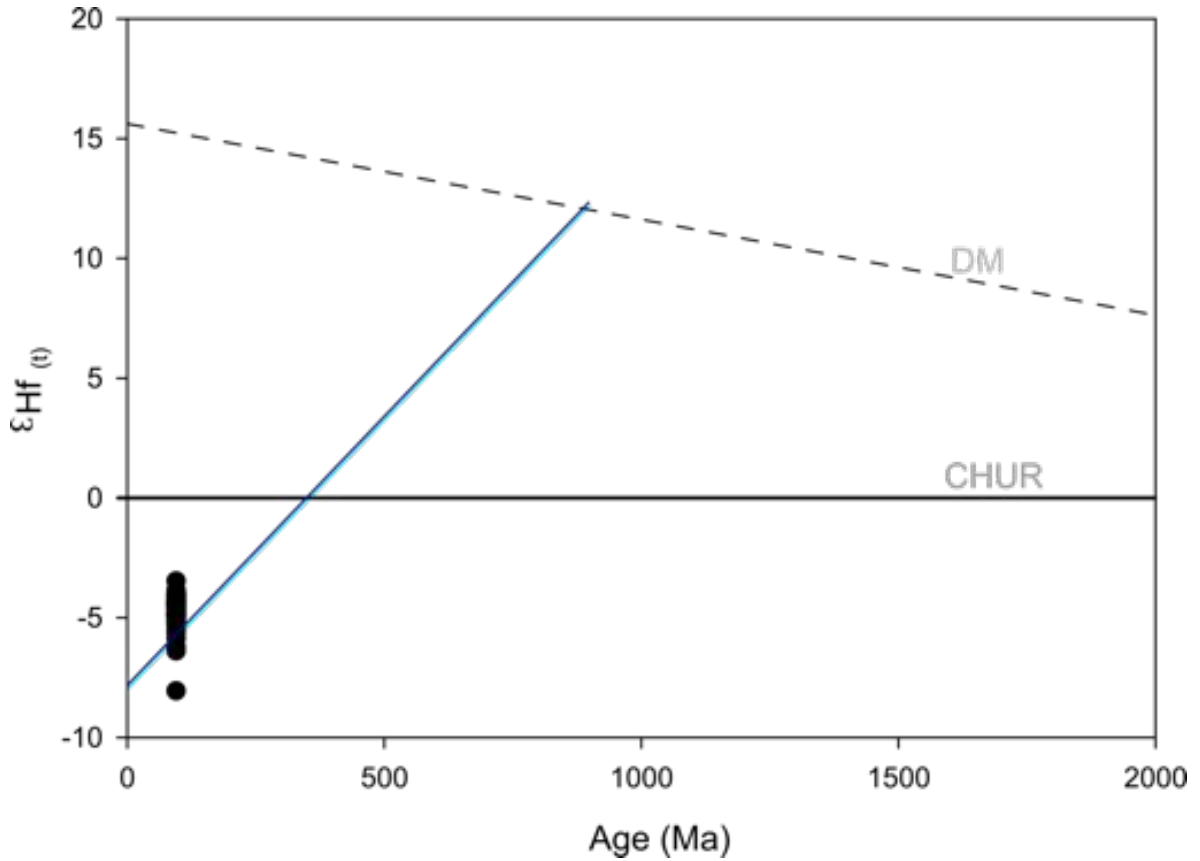


Figure 6.13: Zircon Lu-Hf model age of the IRA. The graph of $\epsilon_{\text{Hf}(95\text{Ma})}$ vs crystallisation age (Ma). The evolution-depleted mantle (DM) is calculated back from present ratios of $^{176}\text{Lu}/^{177}\text{Hf} = 0.0524$ and $^{176}\text{Hf}/^{177}\text{Hf} = 0.28326$ (Workman and Hart, 2005). The light blue regression line is sample IRA 4-2, and the dark blue regression line is sample IRA 18-6.

6.4 alphaMELTS modelling

The model presented by Ashwal and Bybee (2017) suggests that the possible tectonic setting for massif-type anorthosite is an Andean-type continental arc, but this model falls short in terms of addressing the question of why there is no known massif-type anorthosite in modern Andean-type arc today? Here, we attempt to address this dilemma using modelling software called alphaMELTS. The alphaMELTS modelling for this research aims to assess the possibility and the quantification of the parameters required to produce massif-type anorthosite from parental magmas in an Andean-type arc environment like the Sierra Nevada Batholith. Also looking at what the IRA tells us about the change in subduction zones from Proterozoic to Phanerozoic Eon. Given the lack of evidence of primitive parental magmas in the IRA region, suitable primitive arc magma compositions are derived from the literature. Primitive arc-related magmas display significant compositional difference, controlled by melting of the peridotitic mantle wedge above the subducting slab, slab geometries, crustal thickness, water content etc. (Kelemen *et al.*, 2003; Müntener *et al.*, 2018). Schmidts and Jagoutz (2017) documented possible

primitive arc melts by extracting all the plutonic arc rock analyses from the global *GEOROC* database. The database contains igneous rocks from arcs around the world, including 124 samples from the Cascade Arc, which occurs north-west of the Sierra Nevada Batholith (Figure 6.14). The Cascade magmatic arc is a belt of Quaternary volcanoes that extends, for 1250 km, from Lassen Peak in the Northern California to Meager Mountain in Canada, above the subduction zone where the Juan de Fuca Plate plunges beneath the North American Plate (Hildreth, 2007). The identified possible primitive melts are the tholeiitic basalt, basalt, and andesite from work done by Schmidts and Jagoutz (2017) (Table 6.10). Major elements systematics of these parental melts indicate that mantle equilibration occurred at 10-25 kbar and 1220-1350 °C for the basalt and tholeiitic basalt and 5 - 12 kbar and 1200 °C for the andesite (Schmidts and Jagoutz, 2017). These primitive melts are modelled under different conditions, such as temperature, pressure and varying water content to observe the phases they produce and their compositions to better determine a possible parental melt for the IRA.

In this section, polybaric calculations are performed to understand the behaviour of the melts at different pressure, temperature and water content. The general depth of exposure throughout the SNB is 7 - 13 km, which equates to 2 - 4 kbar. In the southern SNB, exposure depth is 10-17 km equating to 3-5 kbar (average 4 kbar) (Lackey et al., 2005). Isobaric calculations over a range of temperatures from superliquidus (1300 °C) to subsolidus (900 °C) at 4 kbar were performed using alphaMELTS version 1.9 (Smith and Asimow, 2005). The calculations involve different parental composition identified from Schmidts and Jagoutz (2017) under different conditions which include variable water content and proportion of upper crustal contaminant in the system. Phase proportions were determined at every 50 °C of temperature decrease, and the oxygen fugacity was set to the fayalite–magnetite–quartz (FMQ) buffer. All of the calculations were performed under the assumption that processes happened in an open system under fractional crystallisation, where there was an interaction between the melt and the country-rock. Early formed crystals were separated from the evolving magma, therefore depleting certain elements in the system. The baseline experiments (isobaric and polybaric calculations) were conducted using uncontaminated anhydrous melts. The isobaric calculations were modelled according to the general depth of the study area, which is 10 - 17 km which equates to 3 - 5 kbar (~4 kbar) (Lackey *et al.*, 2005). Polybaric calculations with a temperature change were modelled to see the change in the phase proportions and composition. The baseline experiments were used to observe the effects of change in pressure, temperature, water content, and crustal assimilation.

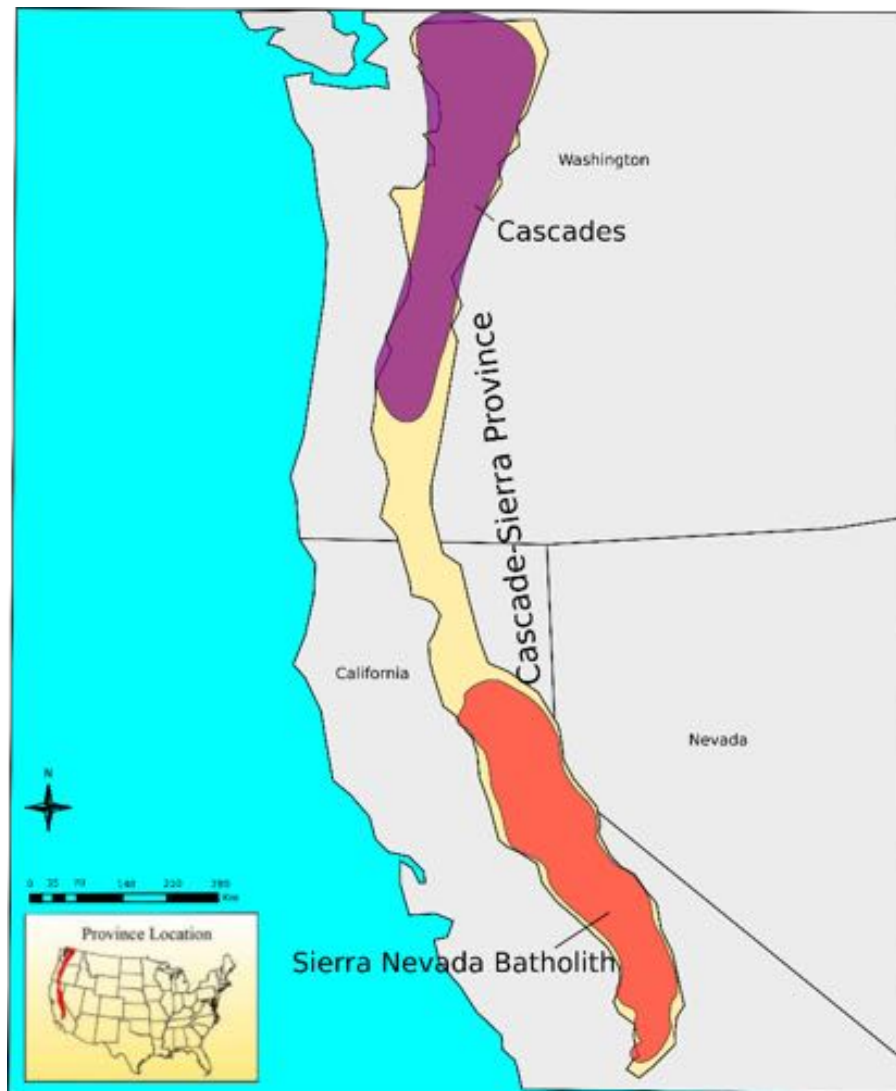


Figure 6.14: A schematic diagram of the Cascade-Sierra Province showing relative locations of the two magmatic arcs (Cascade Batholith (purple) and the Sierra Nevada Batholith (orange)).

Table 6.10: Primitive arc melts from the Cascades batholiths from the GEOROC database (Schmidts and Jagoutz, 2017).

Primitive Melt	Composition	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Total	n
Melt 1	Thol. Basalt	48.6	0.71	17.7	8.55	0.16	10.03	11.54	2.45	0.16	0.08	99.98	52
Melt 2	Basalt	50.6	1.3	16.3	8.25	0.14	9.09	9.35	3.29	1.21	0.42	99.95	34
Melt 3	Andesite	53.3	0.96	16.5	7.29	0.13	8.62	8.57	3.35	1.02	0.28	100.02	38

6.4.1 Isobaric baseline experiment (4 kbar)

Melt 1- Tholeiitic Basalt (Figure 6.15 a): The first phase to crystallise is the plagioclase at 1246.77 °C (the liquidus temperature). The second phase to crystallise is the clinopyroxene at 1206.77 °C, followed by K-feldspar at 1006.77 °C. As the temperature drops, oxide phases like hematite/ilmenite and

corundum starts to crystallise at 886.77 °C. The melt crystallised ~71 % plagioclase, ~28 % clinopyroxene, and ~<1 % K-feldspar, hematite/ilmenite and corundum. Throughout the simulation, the plagioclase An content decreased with decreasing temperature (Figure 6.16). The first plagioclase phases crystallised with An₇₈ and the last plagioclase phase had An₄₉. The presence of hematite/ilmenite and corundum also caused the plagioclase An content to decrease again from An₄₉ to An₄₀. K-feldspar in the melt causes the plagioclase An content to increase from An₄₅ to An₄₉ at 906.77 °C.

Melt 2- Basalt (Figure 6.15 b): The first phase to crystallise is plagioclase at 1194.63 °C (the liquidus temperature). The second phase to crystallise is clinopyroxene at 1174.63 °C. The plagioclase An content decreased with decreasing temperature (Figure 6.16). The melt crystallised ~75 % plagioclase and ~25 % clinopyroxene. The first plagioclase phases crystallised with An₆₃ and the last plagioclase phase had An₅₈. Until the subsolidus, we only see two phases forming and this melt produces a basaltic rock.

Melt 3- Andesite (Figure 6.15 c): Plagioclase and clinopyroxene crystallise simultaneously at 1178.47 °C (the liquidus temperature). The two phases continue to form simultaneously as the temperature decreases until at 998.47 °C, where the plagioclase stops forming and K-feldspar starts to form at 978.47 °C alongside the clinopyroxene. The melt crystallised ~72.5 % plagioclase, ~27 % clinopyroxene, and ~<0.5 % K-feldspar. There is a steady decrease in An content with a decrease in temperature towards the subsolidus from the first plagioclase phases crystallised with An₅₉, and the last plagioclase phase had An₅₈ (Figure 6.16).

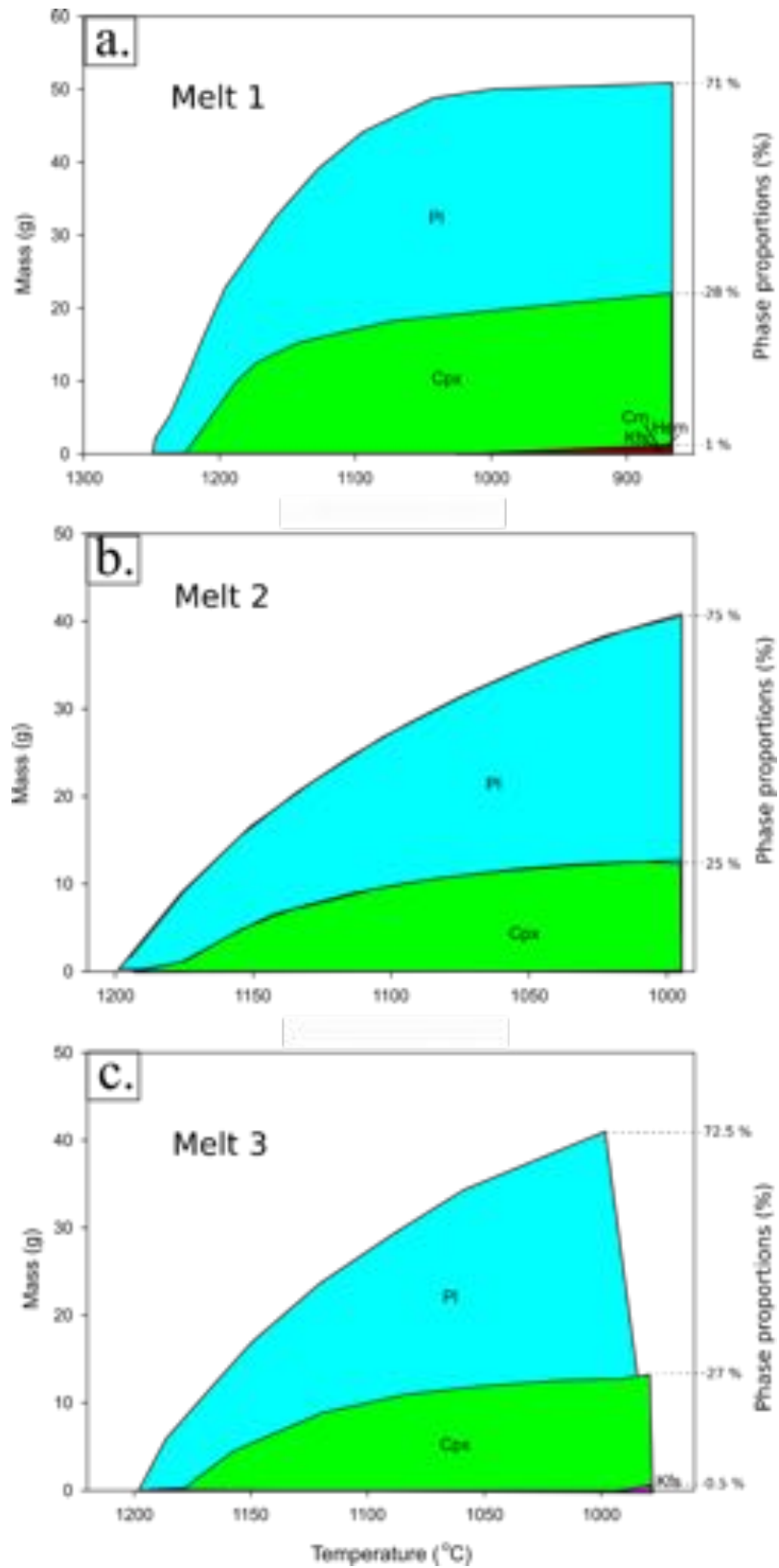


Figure 6.15: The graphs show mineral accumulation in a closed system and their abundances. Diagram a. b. and c. show mineral accumulation from the primitive arc melts under isobaric conditions (4 kbar) with a decrease in temperature.

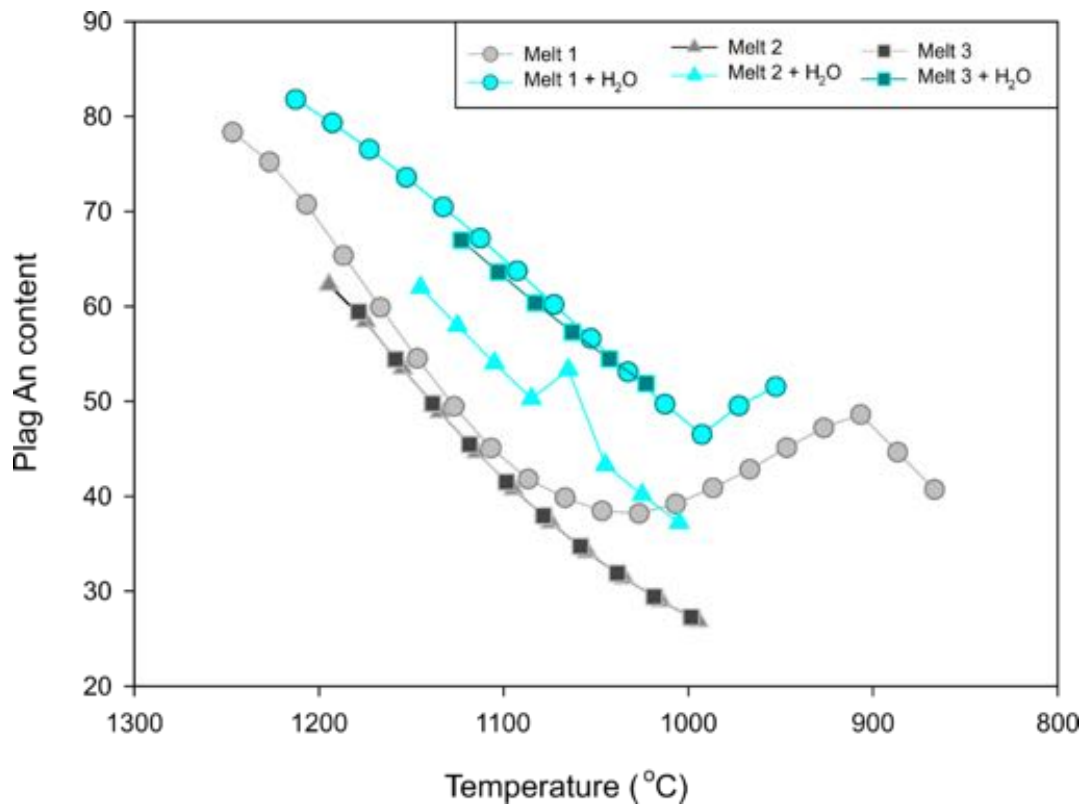


Figure 6.16: The change in plagioclase An content with the change in temperature and addition of 0.5 wt% water. Generally, the An content decreases with a decrease in temperature. Melt 1 shows high An content compared to the other Melt 2 and 3. Melt 2 and 3 show a similar trend under similar conditions compared to Melt 1. The grey lines represent primitive arc melts crystallisation trend under an isobaric condition with a decrease in temperature, and the blue lines are the primitive arc melts with the addition of water.

6.4.2 The Effects of Change in Pressure

The parental melts of massif-type anorthosites have been thought to originate from either the upper mantle or lower crust. A group of scientists have favoured the crustal origin source that is either inherently aluminous or it becomes so as a result of prior anatexis (i.e. Michot, 1964; Philpotts, 1968; Simmons and Hanson, 1978; Taylor *et al.*, 1984; Longhi *et al.*, 1999; Selbekk *et al.*, 2000; Bédard, 2010). The other group supports the mantle origin theory, which suggests that plagioclase are concentrated as floatation cumulates from the fractionation of Al-rich basalt in the crust or by assimilation of crustal material in the melt (Bowen, 1922; Buddington, 1969; Wiebe, 1980; Morse, 1982; Emslie, 1985; Ashwal, 1993; Emslie and Hegner, 1993; Scoates and Frost, 1996; Dempster *et al.*, 1999; Scoates and Mitchell, 2000; Ashwal and Bybee, 2017; Hill, 2017). It is believed that the parental melts started to crystallised at depth prior to and during diapirism (Hamilton, 1981; Philpotts, 1981; Menuge, 1988), which means it continued to crystallise polybarically during the rise to the

shallow crust. Therefore, the melts were simulated under different temperatures and pressures with an initial pressure of 25 kbar.

Melt 1- Tholeiitic Basalt (Figure 6.17 a, b): As the temperature and pressure decrease, the first phase to crystallise is the clinopyroxene at 1433.52 °C, 22.87 kbar. The second phase to form is plagioclase at 1253.52 °C, 16.25 kbar. The two phases crystallise simultaneously for a short temperature and pressure range (1253.52- 1153.52 °C, 16.25- 16.16 kbar respectively), then the plagioclase continued to crystallise until 813.52 °C, 13.27 kbar. The tholeiitic basalt melt crystallised about ~58% plagioclase and ~42 % clinopyroxene. The plagioclase An content decreases with decreasing temperature and pressure (Figure 6.18). The first plagioclase has an An₆₈ and then it rapidly decreases as the temperature and pressure decreases until the last crystals forming with An₃₆.

Melt 2- Basalt (Figure 6.17 c, d): The first phase to form is clinopyroxene at 1408.58 °C, 24.11 kbar. The second phase to form is the plagioclase at 1148.58 °C, 18.55 kbar. The third phase is pyrite at 1008.58 °C, 19.43 kbar, followed by the K-feldspar at 848.58 °C, 19.09 kbar. The basaltic melt crystallised about ~54% plagioclase, ~41 % clinopyroxene, ~3% K-feldspar, and ~2% pyrite. The plagioclase An content of the first phases are low, starting at An₂₉ and the last crystal having An₁₂ (Figure 6.18).

Melt 3- Andesite (Figure 6.17 e, f): The first phase to form is clinopyroxene at 1399.42 °C, 22.79 kbar. The second phase to form is the plagioclase at 1159.42 °C, 16.79 kbar. The third phase is K-feldspar at 939.42 °C, 17.02 kbar. The clinopyroxene and plagioclase crystallise simultaneously for a short temperature and pressure range (1159.42-1119.42 °C, 16.79-16.89 kbar respectively). The andesitic melt crystallised about ~53% plagioclase, ~38 % clinopyroxene, 8% K-feldspar, and 1% pyrite. The plagioclase An content of the first phases starts at An₃₆ and the last crystal having An₁₇ (Figure 6.18). The presence of K-feldspar causes a slight increase in the plagioclase An content.

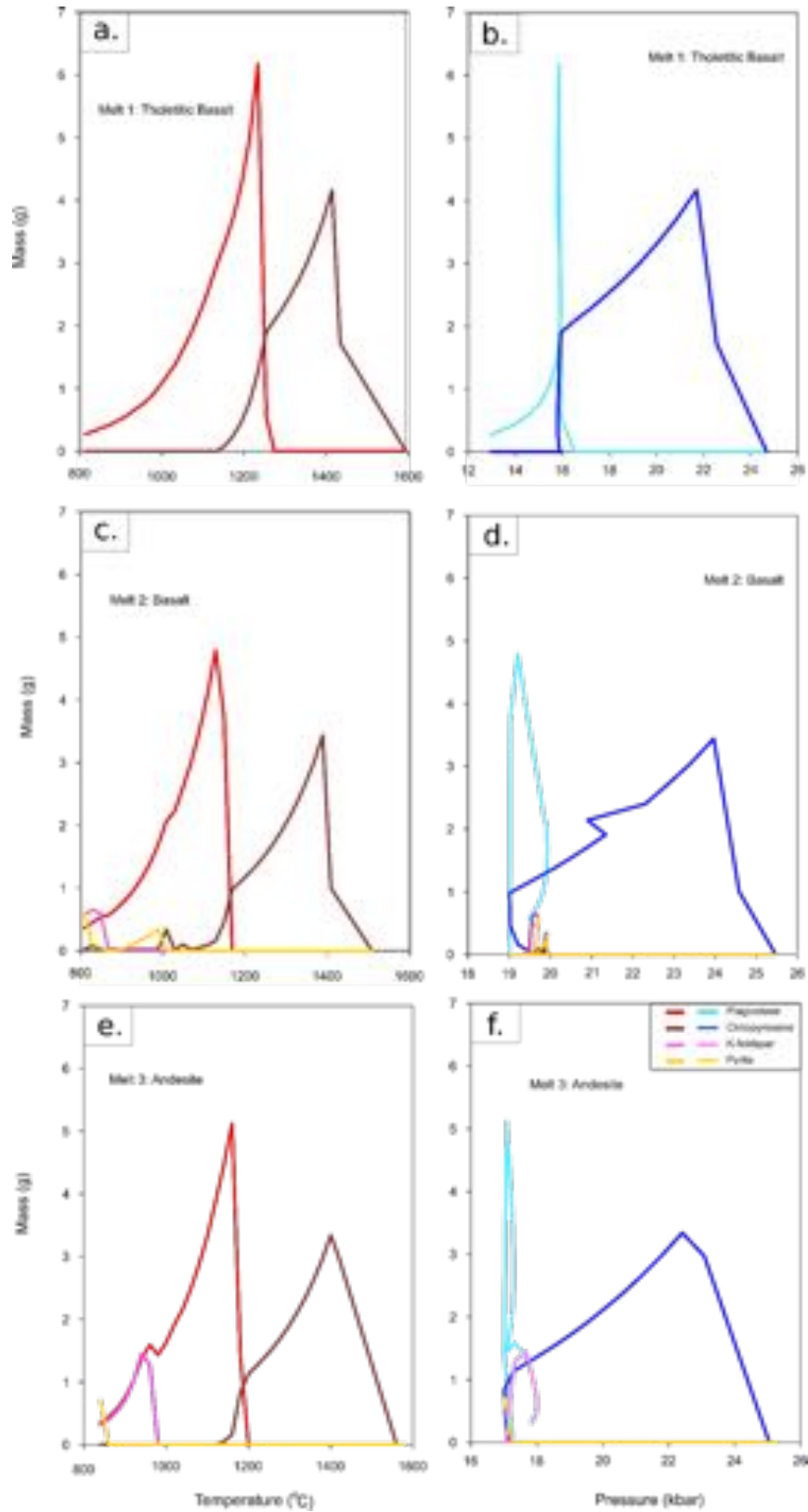


Figure 6.17: Possible primitive melts of the southern SNB simulated under different pressure and temperature. Clinopyroxene phases form at high temperature and pressure, while the plagioclase, K-feldspar, pyrite forms at lower temperature and pressure.

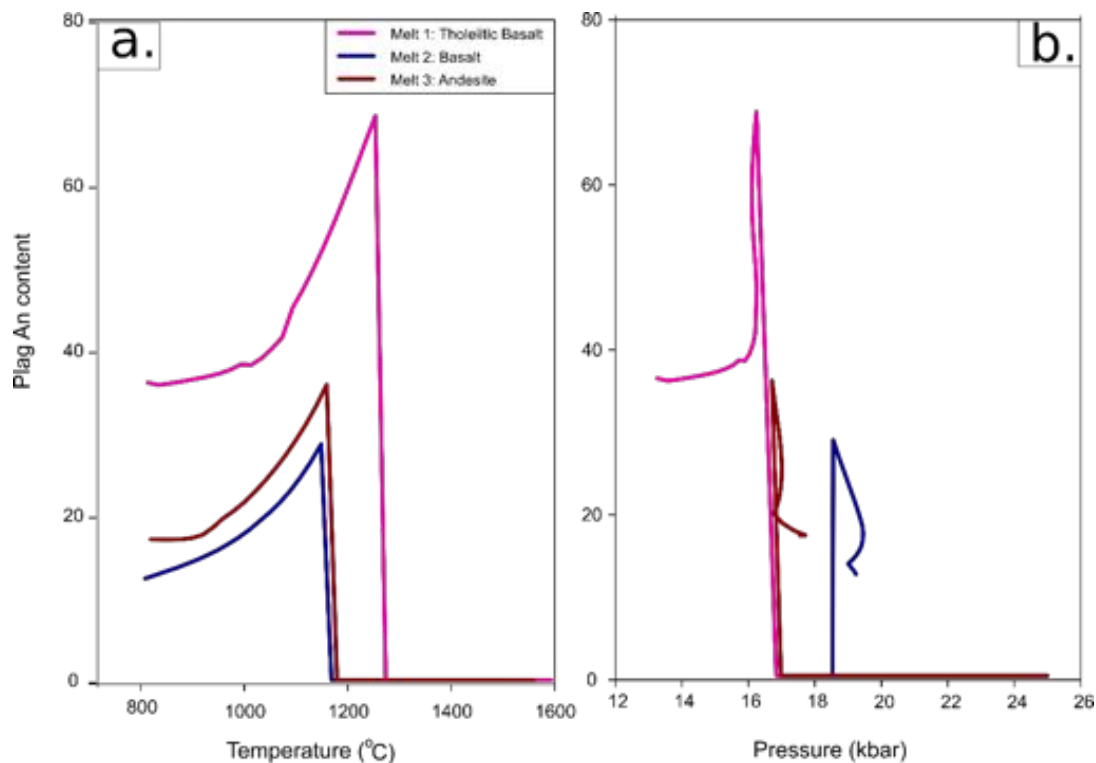


Figure 6.18: The change in plagioclase An content with changing temperature and pressure. The An content of melt 1 shows high An content, while melts 2 and 3 have the least. The first crystals of plagioclase forming at high temperature and pressure have high An content compared to the crystals at low temperature and pressure.

6.4.3 Addition of crustal contaminant and H₂O

Addition of 0.5 wt% H₂O at 4 kbar

For Melt 1 + 0.5 wt% H₂O (Figure 6.19 a), the liquidus temperature dropped to 1213 °C and the first phase to form is plagioclase and the second phase to form is clinopyroxene at 1193 °C. The third phase to form is biotite, alongside hornblende at 973 °C, followed by hematite at 953 °C, and lastly, corundum, apatite, perovskite and pyrite at 933 °C. The mineral proportions are ~58 % plagioclase, ~37 % clinopyroxene, ~3 % apatite, ~2% hornblende, ~<1 % biotite, corundum, pyrite and perovskite. The plagioclase An content of the first crystals are An₈₂ to the last crystals with An₅₂. For Melt 2 + 0.5 wt% H₂O (Figure 6.19 b), the liquidus temperature dropped to 1144.96 °C and the first phases to crystallise are plagioclase and clinopyroxene simultaneously until 1004.96 °C. Hematite/ilmenite, apatite, melilite and perovskite are the last phases to form at 985 °C. The mineral proportions are ~61 % plagioclase, ~38 % clinopyroxene, and ~<1 % hematite/ilmenite, apatite, melilite and perovskite. The plagioclase An content of the first crystals are An₆₂ with the final plagioclase phase showing An₃₇. For Melt 3 + 0.5 wt% H₂O (Figure 6.19 c), the liquidus temperature dropped to 1142.83 °C and the first phase to

crystallise is clinopyroxene with plagioclase as the second phase at 1122.83 °C. The mineral proportions are ~60 % clinopyroxene and ~40 % plagioclase. The plagioclase An content of the first crystals are An₆₇ while last crystals show An₅₂. The melt crystallised only clinopyroxene and plagioclase until the subsolidus and the amount of clinopyroxene increased compared to the previous simulation. Generally, the addition of water lowers the liquidus temperature and increases the plagioclase An content.

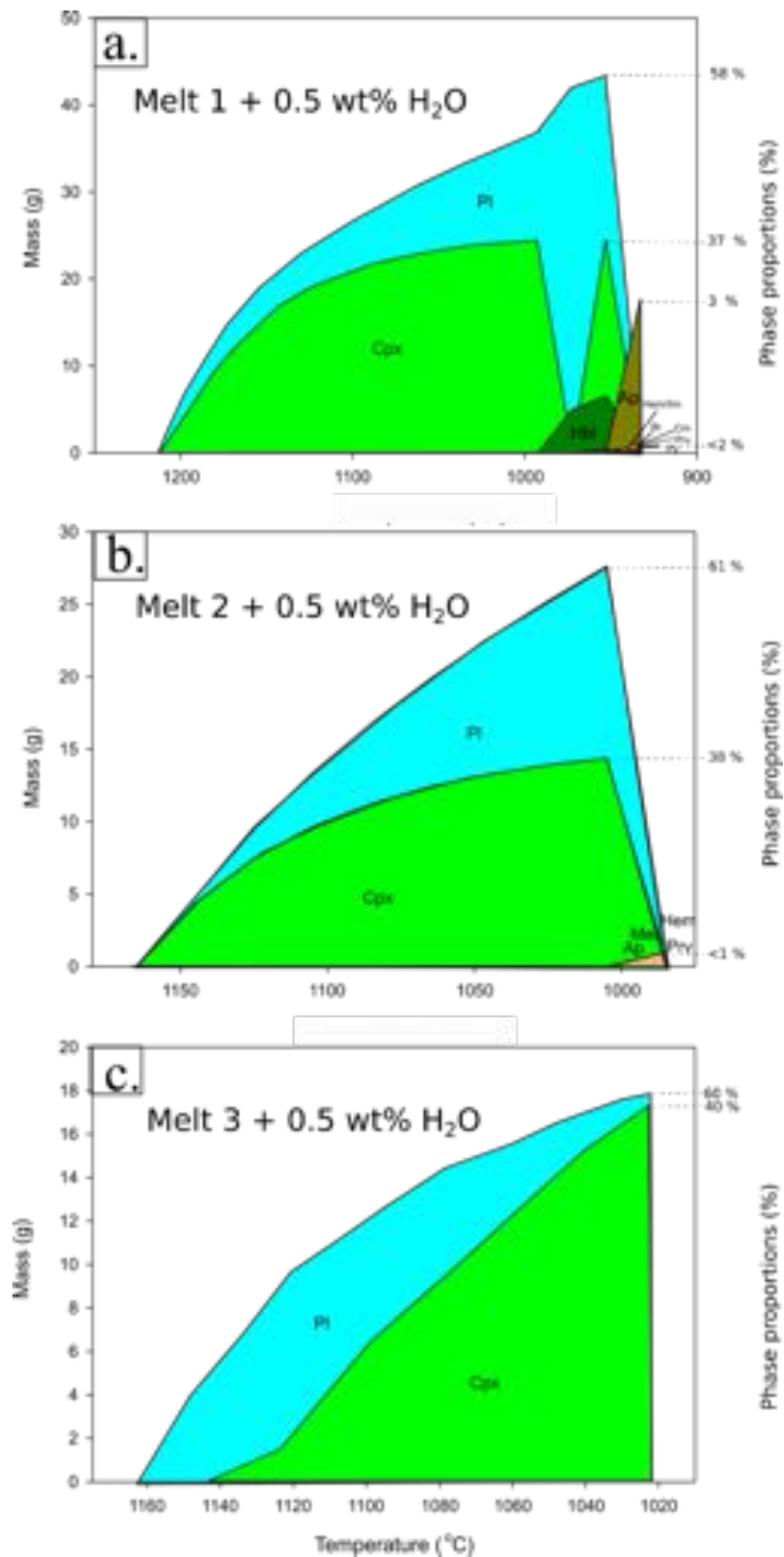


Figure 6.19: The graphs show mineral crystallisation in a closed system and their abundance. Diagram a, b, and c show mineral accumulation from the primitive arc melts under isobaric conditions (4 kbar) with a decrease in temperature at 20 °C increments and with the addition of H₂O to the system.

Compared to the isobaric baseline experiments, the proportions of plagioclase decreased with the introduction of water in the system, while the proportions of clinopyroxenes increased. In Melt 1, the amount of plagioclase decreased from ~71 % to ~58 %, and the clinopyroxene increased from ~28 % to ~37 %. New phases appeared with the introduction of water, for example, phases like apatite, hornblende, biotite, corundum, pyrite and perovskite appeared in the system. The maximum plagioclase An content increased from An₇₈ to An₈₂ (Figure 6.16). In Melt 2 and Melt 3, a similar effect can be seen with the proportions of the plagioclase and clinopyroxenes. In Melt 2, the plagioclase decreased from ~75 % to 68%, the clinopyroxene increased from ~25 % to 38%, and new phases, i.e. hematite/ilmenite, apatite, melilite and perovskite appeared in the system. The water in the system barely affected the maximum plagioclase An content which changed from An₆₃ to An₆₂. In Melt 3, the clinopyroxene abundance decreased from ~73 % to ~40 %, the plagioclase increased from ~27 % to ~60 %, and the K-feldspar did not crystallise. The maximum plagioclase An content increased from An₅₉ to An₆₇ (Figure 6.16). The effect of water cannot be ignored, since the anorthosite occurs in a subduction terrane environment. In this iteration, 0.5 wt% H₂O was added to each parental melt variety and run through similar pressure and temperature conditions to previous models (isobaric, 4 kbar). The addition of water generally encourages the formation of new hydrous phases in the late stages of crystallization and increases the An content. The amount of plagioclase decreased while the clinopyroxene proportions increased.

Addition of crustal contaminant and H₂O at 4kbar

The crustal composition used corresponded to a biotite-granite (36 % plagioclase, 25% K-feldspar, 30% quartz, 7% biotite, 2% muscovite) thought to be the most representative country rock into which the SNB was emplaced (Brown *et al.*, 2018). The contaminant was assimilated 20 times at a mass ratio of 1:2 to the liquid mass (parental melt) assuming that it happened in an open system under fractional crystallisation, resulting in isolation of early formed crystals . The same process was repeated with a change in H₂O from 0.5 to 0.7 wt% for every parental melt identified from the study by Schmidts and Jagoutz (2017) (i.e. tholeiitic basalt, basalt and andesite) (Figure 6.20). The modelling results show that a parental magma contaminated with crustal material will only crystallise plagioclase and clinopyroxene in the system. The change in the water content increases the An content of the plagioclase and causes a slight increase in the amount of clinopyroxenes, with the presence of crustal contaminants in the system, the effect of increased water level is negligible.

Melt 1 + crust + 0.5 wt% H₂O (Figure 6.20 a), the liquidus temperature is 1210.35 °C and the first phase to crystallise is plagioclase with An₈₂. The second phase is clinopyroxene at 1196.94 °C. The plagioclase An content decreases with decreasing temperature. The system crystallises plagioclase and clinopyroxene at a similar rate. By increasing the water content (Figure 6.20 b), the liquidus temperature

drops even further to 1197.78 °C, and the plagioclase and clinopyroxene crystallise simultaneously. The An content of the plagioclase slightly increases to An₈₃. The amount of clinopyroxene increases at a higher rate than the plagioclase and the An content drops to An₆₀, which is lower than the first simulation (Figure 6.21).

Melt 2 + crust + 0.5 wt% H₂O (Figure 6.20 c), the liquidus temperature is 1150.86 °C, where plagioclase and clinopyroxene crystallise simultaneously and the An content starts at An₆₃. Here the An content drops to An₄₇ (Figure 6.21) and the plagioclase crystals crystallise at a higher rate than the clinopyroxene. By increasing the water content (Figure 6.20 d), the first crystals to form are clinopyroxenes at 1152.67 °C, and the plagioclase comes later at 1139.67 °C with An₆₃. The last plagioclase crystals crystallise at 1071.35 °C with An₄₉.

Melt 3 + crust + 0.5 wt% H₂O (Figure 6.20 e), the liquidus temperature is 1163.94 °C and the first crystals to crystallise is clinopyroxene, and the second phase is plagioclase at 1151.47 °C with An₆₆. By increasing the water content (Figure 6.20 f), the liquidus temperature slightly increases to 1164.34 °C, and the first crystals are the clinopyroxenes. The addition of more water increases the amount of clinopyroxene in the system and causes plagioclase to crystallise later in the system at 1132.04 °C with An₆₄.

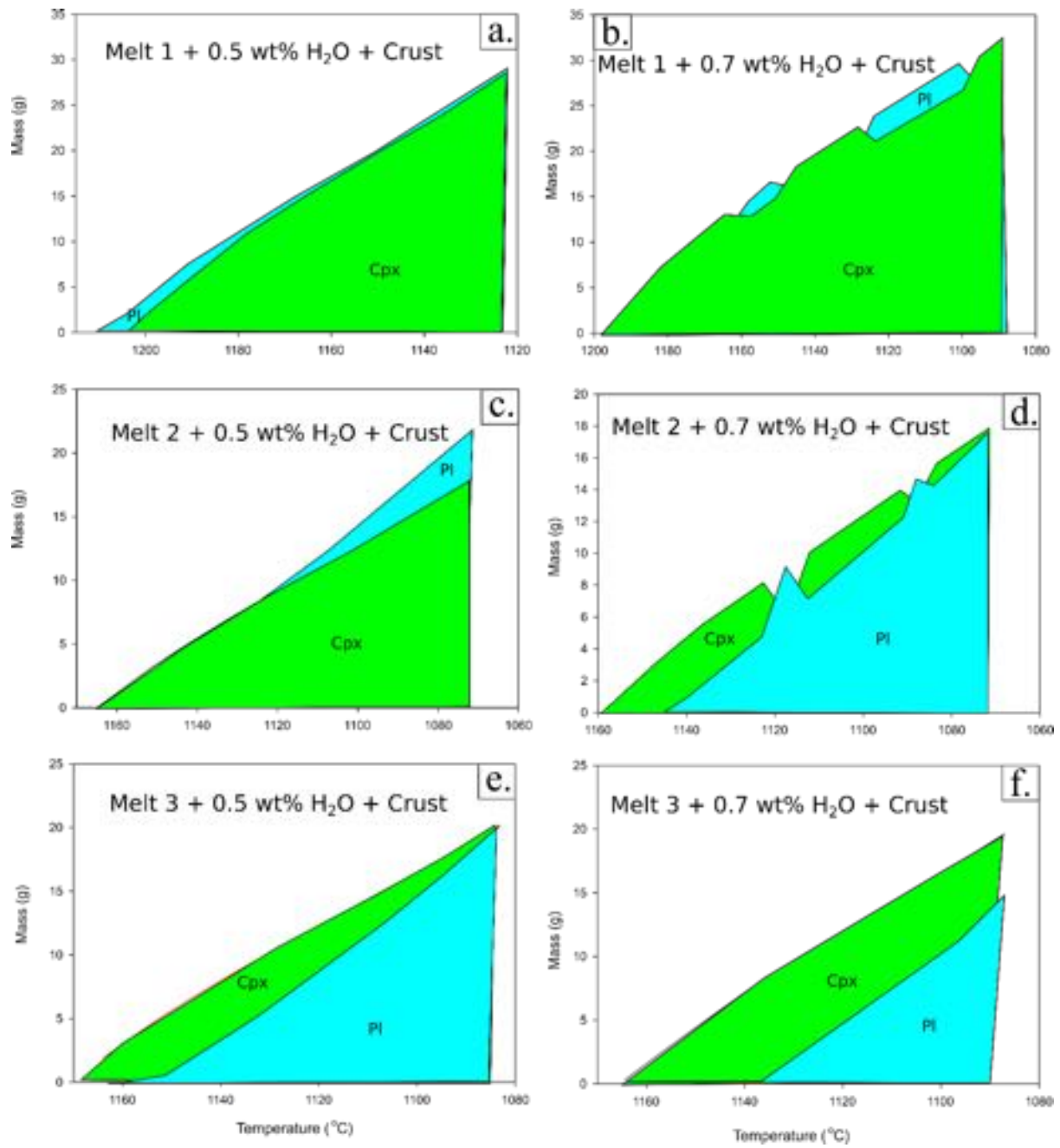


Figure 6.20: Mineral crystallisation in the system with the addition of felsic contaminants in the form of a solid and water. The diagrams display different parental compositions and changing the amount of water in the system from 5 to 7 wt%, and the amount of contaminants was kept constant at a ratio of 1:2 to the parental melt.

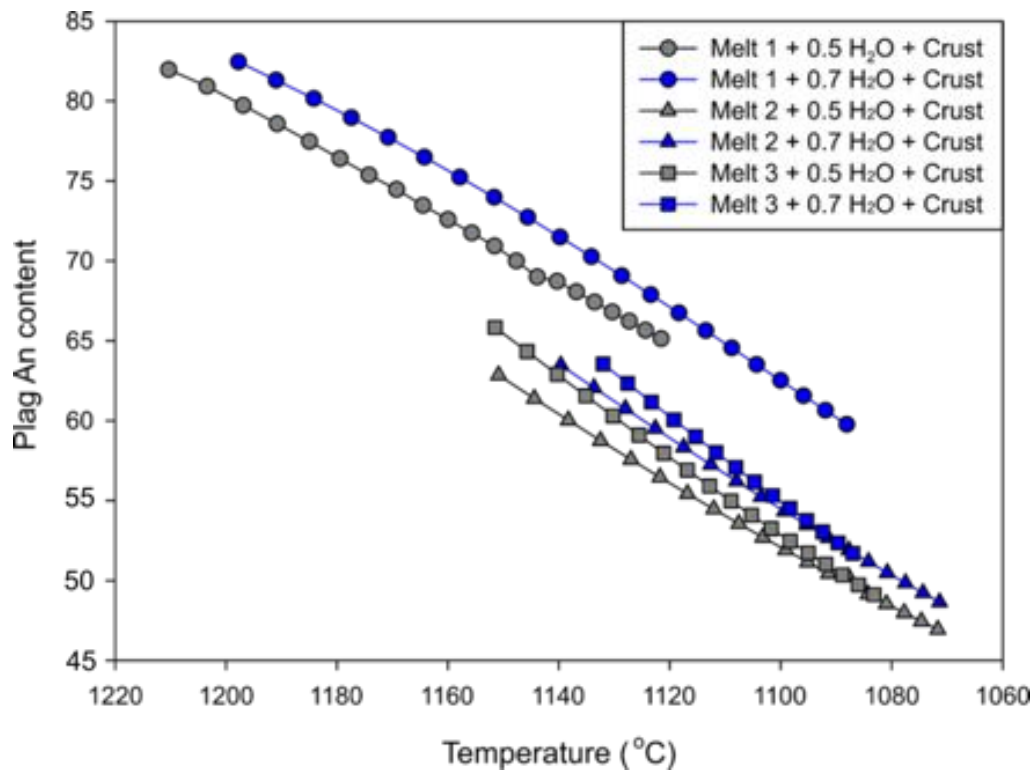


Figure 6.21: The change in plagioclase An content with the decrease in temperature. Generally, the An content decreases with a decrease in temperature. Melt 1 shows high An content compared to the other Melts (2 and 3). Melt 1 with higher water content (0.7 wt%) shows a wide range of An content compared to the 0.5 wt% H₂O. The dark grey lines represent primitive arc melts+ crust and 0.5 wt% H₂O crystallisation trend under an isobaric condition with a decrease in temperature, and the blue lines are the primitive arc melts+ crust + 0.7 wt% H₂O.

The effects of crustal contamination, H₂O, and change in pressure

The models were simulated at higher pressures starting at 25 kbar to understand the effect of pressure on the phase compositions and assemblages. During the modelling process, it was assumed that the magma crystallised in an open system, therefore crustal material was added into the evolving magma and early formed crystals were isolated from the evolving magma resulting in depletion of certain elements in the system. The crustal composition used correspond to a biotite-granite (36 % plagioclase, 25% K-feldspar, 30% quartz, 7% biotite, 2% muscovite) thought to be the most representative country rock into which the SNB was emplaced (Brown *et al.*, 2018). The biotite granite is more likely to be the general crustal composition of the region intruded by the Mesozoic granitic rocks (granodiorites, tonalite, quartz monzonite, quartz monzonite and trondjemite) (Kistler and Peterman, 1978). The contaminant was assimilated 20 times at a mass ratio of 1:2 to the liquid mass (parental melt). The main effect of higher temperatures and pressures are the increase in liquidus temperature, the order of phase saturation and higher plagioclase An content. At depth, the first phase to crystallize is clinopyroxene, followed by plagioclase at lower pressures. At lower temperatures and pressure, the production of the phases fluctuates as clinopyroxene decreases and plagioclase increases at the same time. The plagioclase An content decreases, as the pressure and temperature decreases.

Melt 1 (tholeiitic basalt) was simulated with crustal contamination and changing water content in the system (Figure 6.22). The melt only crystallises two phases which are clinopyroxene and plagioclase. For 0.4 wt% H₂O and crustal contamination (Figure 6.22 a, b), the first phase to form is the clinopyroxene at high temperatures and pressures (1390 °C and 20.52 kbar). The second phase is plagioclase at lower temperatures and pressures (1170 °C and 11.74 kbar). The melt crystallised ~58% clinopyroxene and ~42 % plagioclase. The plagioclase An content decreases with the decrease in temperature and pressure (Figure 6.23). The first plagioclase crystals have An₆₃ and the last crystals have An₄₅. For 0.6 wt% H₂O and crustal contamination (Figure 6.22 c, d), the first phase to crystallise is clinopyroxene at high temperatures and pressures (1368 °C and 19.33 kbar). There is an exponential increment of clinopyroxene minerals at 1148 °C, and then the amount fluctuates while decreasing after that. The second phase to crystallise is plagioclase at lower temperatures and pressures (1168 °C and 11.46 kbar). With 0.2 wt% more water, the melt crystallised ~57 % clinopyroxene and ~43 % plagioclase. The plagioclase An content decreases with the decrease in temperature and pressure (Figure 6.23). The first crystals of plagioclase have An₆₄ and the last ones to crystallise have An₄₇. For 0.8 wt% H₂O and crustal contamination (Figure 6.22 e, f), the first phase to crystallise is clinopyroxene at high temperatures and pressures (1366 °C and 19.10 kbar), followed by plagioclase at lower temperatures and pressures (1166 °C and 11.09 kbar). With high water level (0.8 wt%), the melt crystallised ~54% clinopyroxene and ~46 % plagioclase. The amount of plagioclase crystallising is higher than in the models with lower water content. The plagioclase An content decreases with the decrease in temperature

and pressure (Figure 6.23), and the first crystals of plagioclase have An_{65} and the last ones to crystallise have An_{44} .

The amount of plagioclase increases with the increasing water content, from ~42 to 46 %, while the amount of clinopyroxene decreases from ~58 to 54 %. The maximum plagioclase An content increases from An_{63} at low water levels to An_{65} at high water levels. Compared to the baseline experiments, the proportions are different from the baseline experiments, where there are high plagioclase proportions than the clinopyroxene. The baseline experiment of melt 1 has maximum plagioclase proportions of ~58 %, while the contaminated melt with water has maximum proportions of ~46 %. The clinopyroxene had a maximum proportion of ~42 %, while the contaminated melt with varying water levels has maximum proportions of ~58 %. The plagioclase An content of the baseline experiment had An_{68} , which is higher compared to the An content (An_{65}) of the contaminated melt with the different water level. The important effect the crustal contamination and the addition of water have is the encouragement of more clinopyroxene formation and a slight decrease in the plagioclase An content.

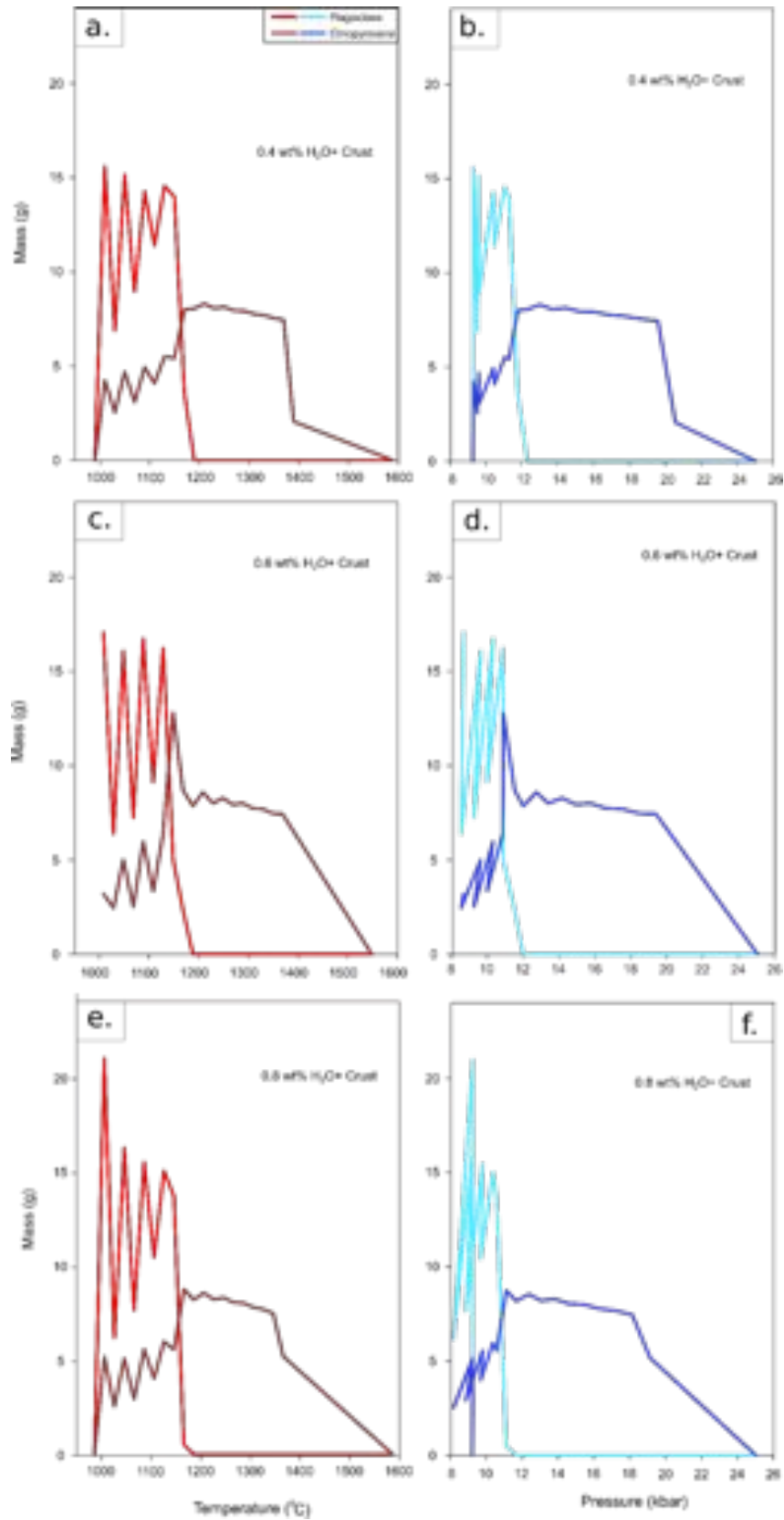


Figure 6.22: Models of melt 1: tholeiitic basalt with crustal contamination and changing water content in the system. Clinopyroxene forms at high temperatures and pressures, while the plagioclase forms at lower temperatures and pressure.

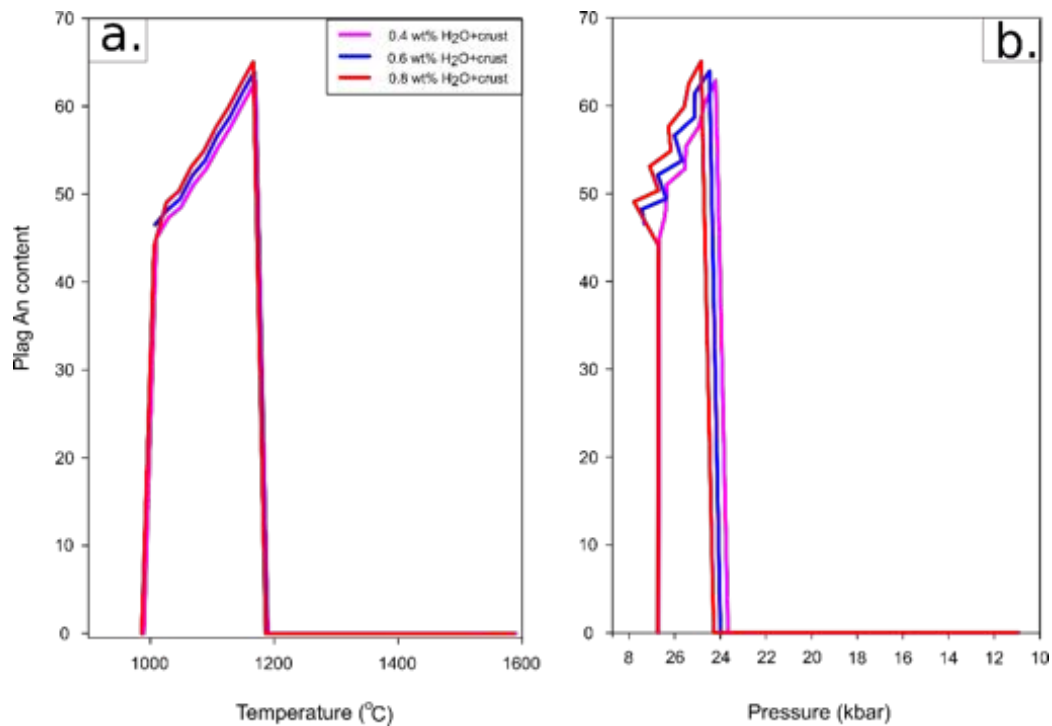


Figure 6.23: The change in plagioclase An content with changing temperature and pressure of melt 1: tholeiitic basalt. The increasing water content shows a slight increase in the plagioclase An content.

Melt 2 (basalt) was simulated with crustal contamination and changing water content in the system (Figure 6.24). Only two phases crystallise from the melt, with clinopyroxene crystallising at higher temperatures and pressures, followed by plagioclase at lower temperatures and pressures. For 0.4 wt% H₂O and crustal contamination (Figure 6.24 a, b), the first phase to form is the clinopyroxene at high temperatures and pressures (1379 °C and 23.08 kbar). The second phase is plagioclase at lower temperatures and pressures (1079 °C and 14.92 kbar). The melt crystallised ~79 % clinopyroxene and ~21 % plagioclase. The first crystals of plagioclase have An₂₇ and the last ones to crystallise have An₂₅. For 0.6 wt% H₂O and crustal contamination (Figure 6.24 c, d), the first phase to crystallise is clinopyroxene at high temperatures and pressures (1355 °C and 22.02 kbar). The amount of clinopyroxenes peaks at 1235 °C, 18.08 kbar. The second phase to crystallise is plagioclase at lower temperatures and pressures (1075 °C and 14.53 kbar). The melt at 0.6 wt% water, it crystallised ~69 % clinopyroxene and ~31 % plagioclase. The first crystals of plagioclase have An₂₇ and the last ones to crystallise have An₂₅. (Figure 6.25). For 0.8 wt% H₂O and crustal contamination (Figure 6.24 e, f), the first phase to crystallise is clinopyroxene at high temperatures and pressures (1351 °C and 21.75 kbar), followed by plagioclase at lower temperatures and pressures (1071 °C and 14.17 kbar). With high water level (0.8 wt%), the melt crystallised ~68 % clinopyroxene and ~32 % plagioclase. The first crystals of plagioclase have An₂₈ and the last ones to crystallise have An₂₆. (Figure 6.25).

The change in water level has little effect on the proportions and composition of the plagioclase, especially from 0.6 to 0.8 wt% water. From 0.4 to 0.6 wt% water, there is a huge increase in the amount of plagioclase (from ~21 to ~31 %), but little change occurs from 0.6 to 0.8 wt% water, with only 1% increment. The opposite effect occurs, as the amount of clinopyroxene decreases from ~79 to ~69 % (0.4 to 0.6 wt% water respectively), and little effect happens from 0.6 to 0.8 wt% water. The baseline experiment of melt 2 has maximum plagioclase proportions of ~54 %, while the contaminated melt with water has maximum proportions of ~32 %. The clinopyroxene had a maximum proportion of ~42 %, while the contaminated melt with varying water levels has maximum proportions of ~79 %. The maximum plagioclase An content of the baseline experiment had is An₂₉, which is slightly higher compared to the An content (An₂₈) of the contaminated melt with the different water level. The important effect the crustal contamination and the addition of water have is the encouragement of more clinopyroxene formation and a slight decrease in the plagioclase An content.

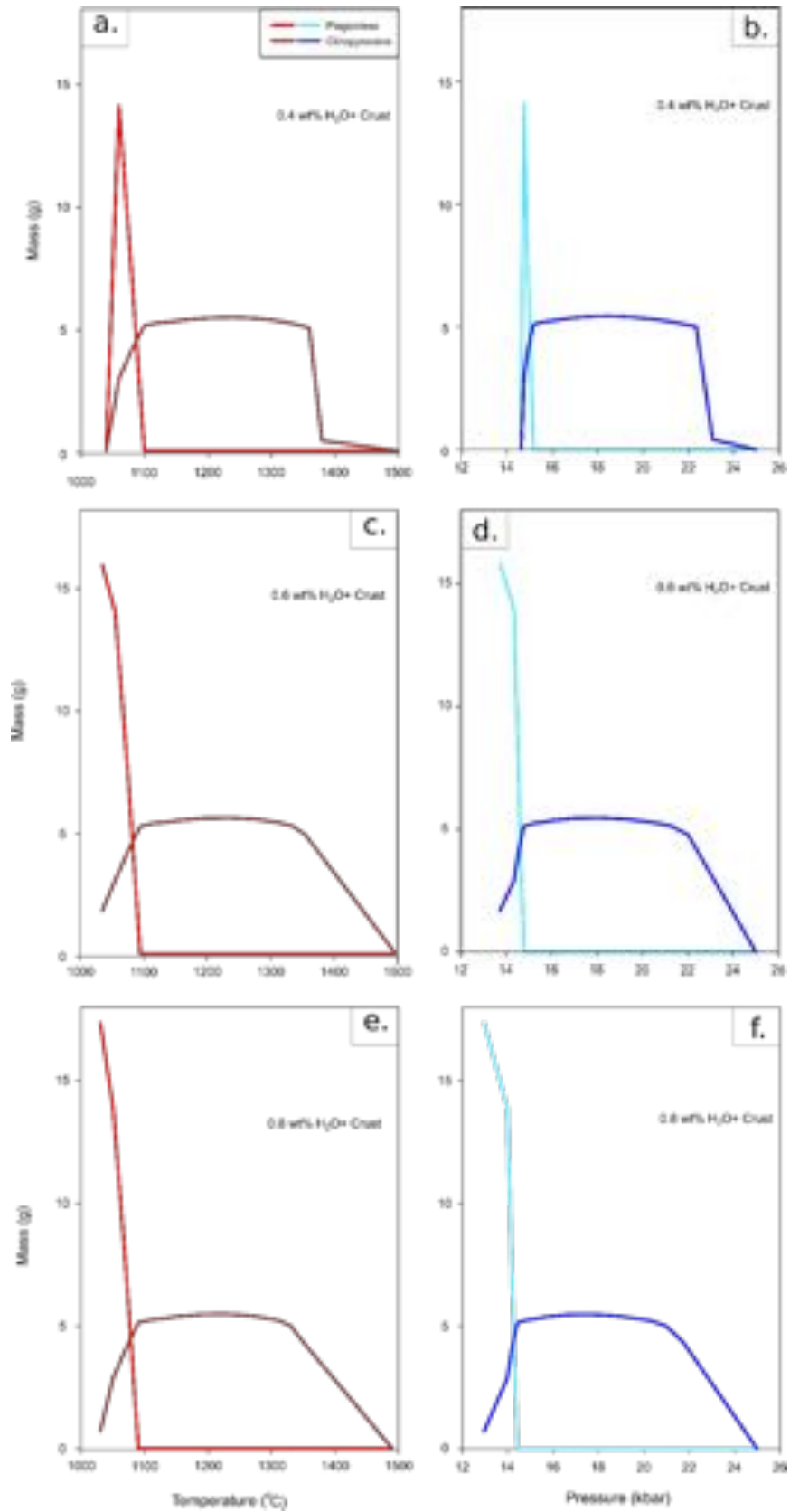


Figure 6.24: Models of melt 2: basalt with crustal contamination and changing water content in the system. Clinopyroxene forms at high temperatures and pressures, while the plagioclase forms at lower temperatures and pressure exponentially.

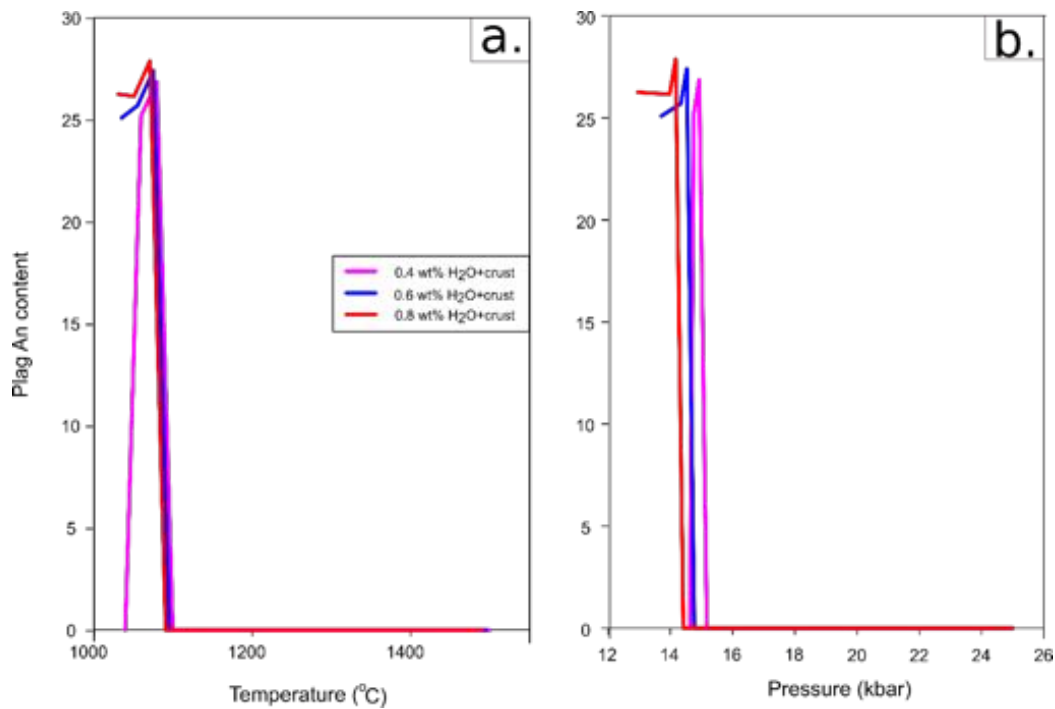


Figure 6.25: The change in plagioclase An content with changing temperature and pressure of melt 2: basalt. The increasing water content shows a slight increase in the plagioclase An content.

Melt 3 (andesite) was simulated with crustal contamination and changing water content in the system (Figure 6.26). Only two phases are crystallised from the melt, with clinopyroxene crystallising at higher temperatures and pressures, followed by plagioclase at lower temperatures and pressures. For 0.4 wt% H₂O and crustal contamination (Figure 6.26 a, b), the first phase to form is the clinopyroxene at high temperatures and pressures (1393 °C and 22.27 kbar). The second phase is plagioclase at lower temperatures and pressures (1112 °C and 13.93 kbar). The melt crystallised ~59 % clinopyroxene and ~41 % plagioclase. The first crystals of plagioclase have An₃₉ and the last ones to crystallise have An₃₁ (Figure 6.27). For 0.6 wt% H₂O and crustal contamination (Figure 6.26 c, d), the first phase to crystallise is clinopyroxene at high temperatures and pressures (1368 °C and 21.15 kbar). The amount of clinopyroxenes peaks at 1348 °C, 20.27 kbar. The second phase to crystallise is plagioclase at lower temperatures and pressures (1108 °C and 12.68 kbar). The melt at 0.6 wt% water, it crystallised ~62 % clinopyroxene and ~38 % plagioclase. The first crystals of plagioclase have An₄₃ and the last ones to crystallise have An₃₄. (Figure 6.27). For 0.8 wt% H₂O and crustal contamination (Figure 6.26 e, f), the first phase to crystallise is clinopyroxene at high temperatures and pressures (1365 °C and 20.75 kbar), followed by plagioclase at lower temperatures and pressures (1085 °C and 11.14 kbar). With high water level (0.8 wt%), the melt crystallised ~66 % clinopyroxene and ~34 % plagioclase. The first crystals of plagioclase have An₄₄ and the last ones to crystallise have An₂₈. (Figure 6.27).

The amount of clinopyroxene increases with the increasing water content, from ~59 to 66 %, while the amount of plagioclase decreases from ~41 to 34 %. The maximum plagioclase An content increases from An₃₉ at low water levels to An₄₄ at high water levels. The baseline experiment of melt 3 has maximum plagioclase proportions of ~53 %, while the contaminated melt with water has maximum proportions of ~41 %. The clinopyroxene had a maximum proportion of ~38 %, while the contaminated melt with varying water levels has maximum proportions of ~66 %. The plagioclase An content of the baseline experiment had An₃₆, which is lower compared to the An content (An₄₄) of the contaminated melt with the different water level. The crustal contamination and the addition of water cause the melt to crystallise two phases, while the baseline experiment had four phases. The important effect it has is the encouragement of more clinopyroxene formation and increase in the plagioclase An content with little proportions.

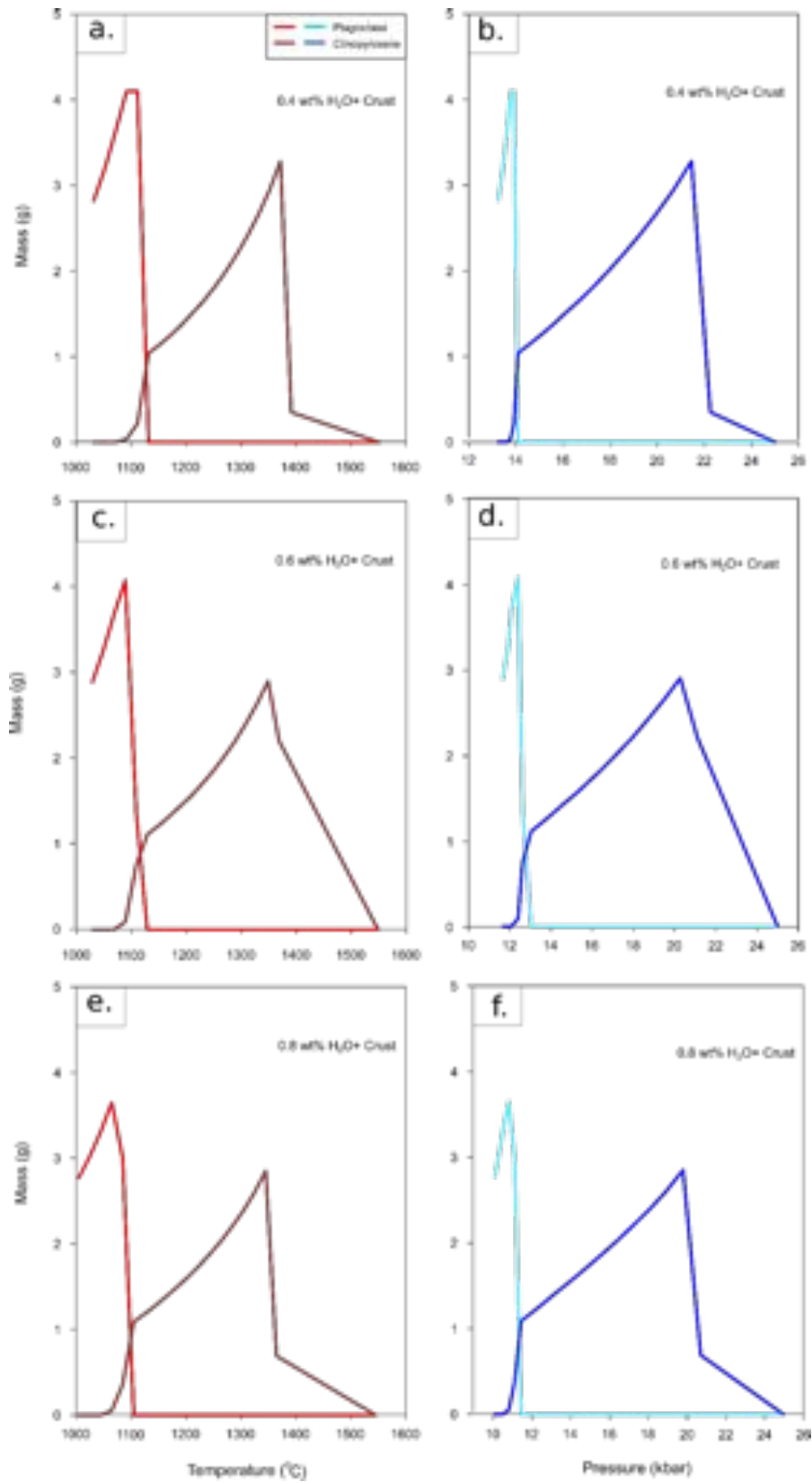


Figure 6.26: Models of melt 3: Andesite with crustal contamination and changing water content in the system. Clinopyroxene forms at high temperatures and pressures while the plagioclase forms at lower temperatures and pressure exponentially.

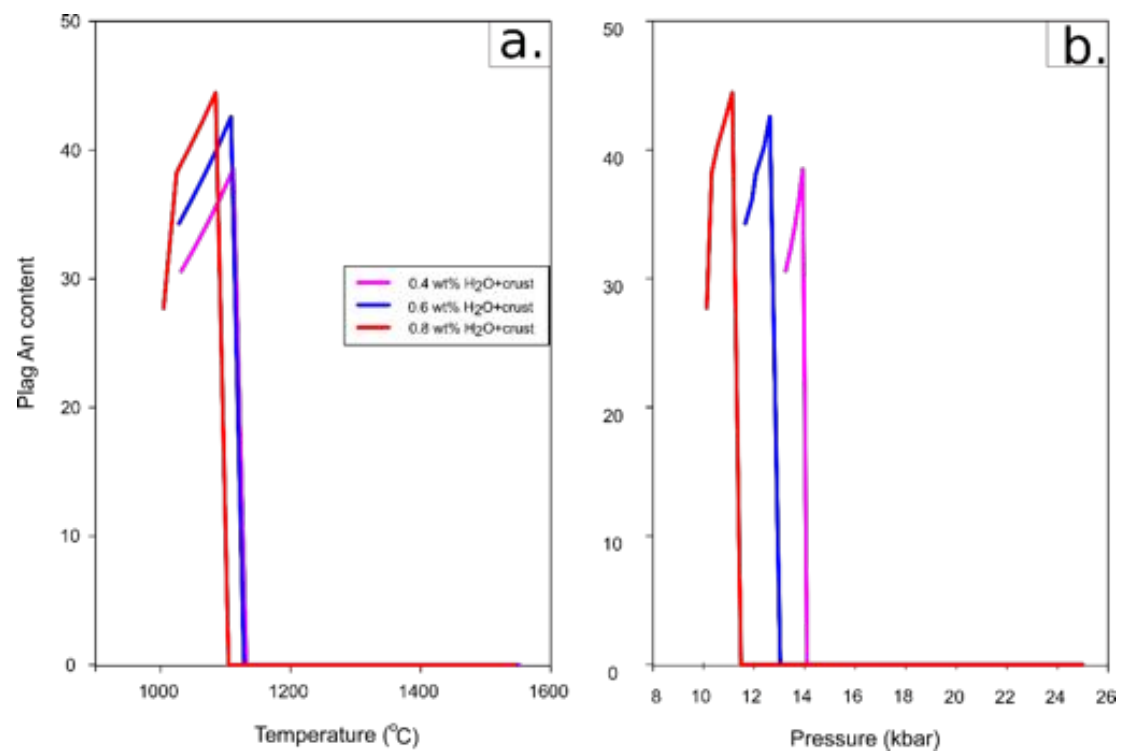


Figure 6.27: The change in plagioclase An content with changing temperature and pressure of melt 3: andesite.
The increasing water content shows a slight increase in the plagioclase An content.

7. Discussion

The temporal restriction and tectonic setting of Proterozoic massif-type anorthosites have been the topic of active debate for decades and many possible settings have been suggested, ranging from continental rifts, plumes and continental arcs (see Section 2.1). The continental arc setting hypothesis has gained traction as more geochemical and geochronological evidence becomes available, but the paucity of anorthosites with massif-type characteristics in Phanerozoic continental arcs has hindered the full acceptance of this hypothesis. The IRA is a small, isolated body of anorthosite which displays a similar plutonic appearance to massif-type anorthosites (albeit smaller), while its texture and geochemical nature is similar to cognate anorthosite in arcs, especially continental arcs. Many processes involved in the formation of massif-type anorthosite can be applied to the formation of IRA in continental arc setting. Other bodies of massif-type anorthosite have been described in Phanerozoic arc environments, but they are xenoliths or xenolithic terranes, which are Proterozoic in age. The two small bodies of anorthosite situated in the Western Sierras Pampeanas of Argentina (Ashwal and Bybee, 2017) and another in southern Peru near Ilo (Casquet *et al.*, 2010) are excellent examples but do not aid in achieving the aims of this discussion. Here we show that the IRA is coeval and consanguineous with the Sierra Nevada Batholith using U-Pb zircon and apatite geochronology, along with a complete suite of major and trace element geochemistry, as well as Sm-Nd and Lu-Hf isotopic data. Thereafter alphaMELTS modelling is used to understand how intrinsic variables such as fractionation pressure, water content, and crustal assimilation influence the formation of anorthosites and how this may explain the presence of the IRA in the SNB, and more importantly, the paucity of massif-type in Phanerozoic arc environments.

7.1 Age relationship between the IRA and SNB

Figure 7.1 shows representative age data for the Sierra Nevada Batholith in 3-dimensional space. In general, the ages of the SNB plutons decrease with latitude and longitude and they become younger towards the southeast. The Black Canyon Pluton is 96 ± 2 Ma (Carl, 2000; Mahan *et al.*, 2003), while the McDoogle Pluton is 94 ± 4 Ma (Mahan *et al.*, 2003; Stearns and Bartley, 2014). Two samples of the IRA (sample IRA 4-2 and IRA 18-6) yielded ages of 94.01 ± 0.05 and 95.29 ± 0.06 Ma, respectively (Figure 7.2). The ages of the IRA, Black Canyon Pluton and McDoogle Pluton fall within the age range of the SNB and precisely align with the expected ages of the southern SNB (Figure 7.1, 7.2). The U-Pb ages produced in this work show that the Indian Rock Anorthosite is coeval with the surrounding granitoids of the Sierra Nevada Batholith. The ages overlap with the 95 Ma Mesozoic high flux event (Ducea, 2001; Paterson and Ducea, 2015), which shows more evolved isotopic compositions (and will be discussed in details in Section 7.2). This disproves the hypothesis that the IRA may be a fragment of

an older, Proterozoic terrane, caught up in the SNB magmatic event. Consequently, the IRA is one of the few (or only) discrete bodies of anorthosite known coeval with a subduction-related, continental arc terrane.

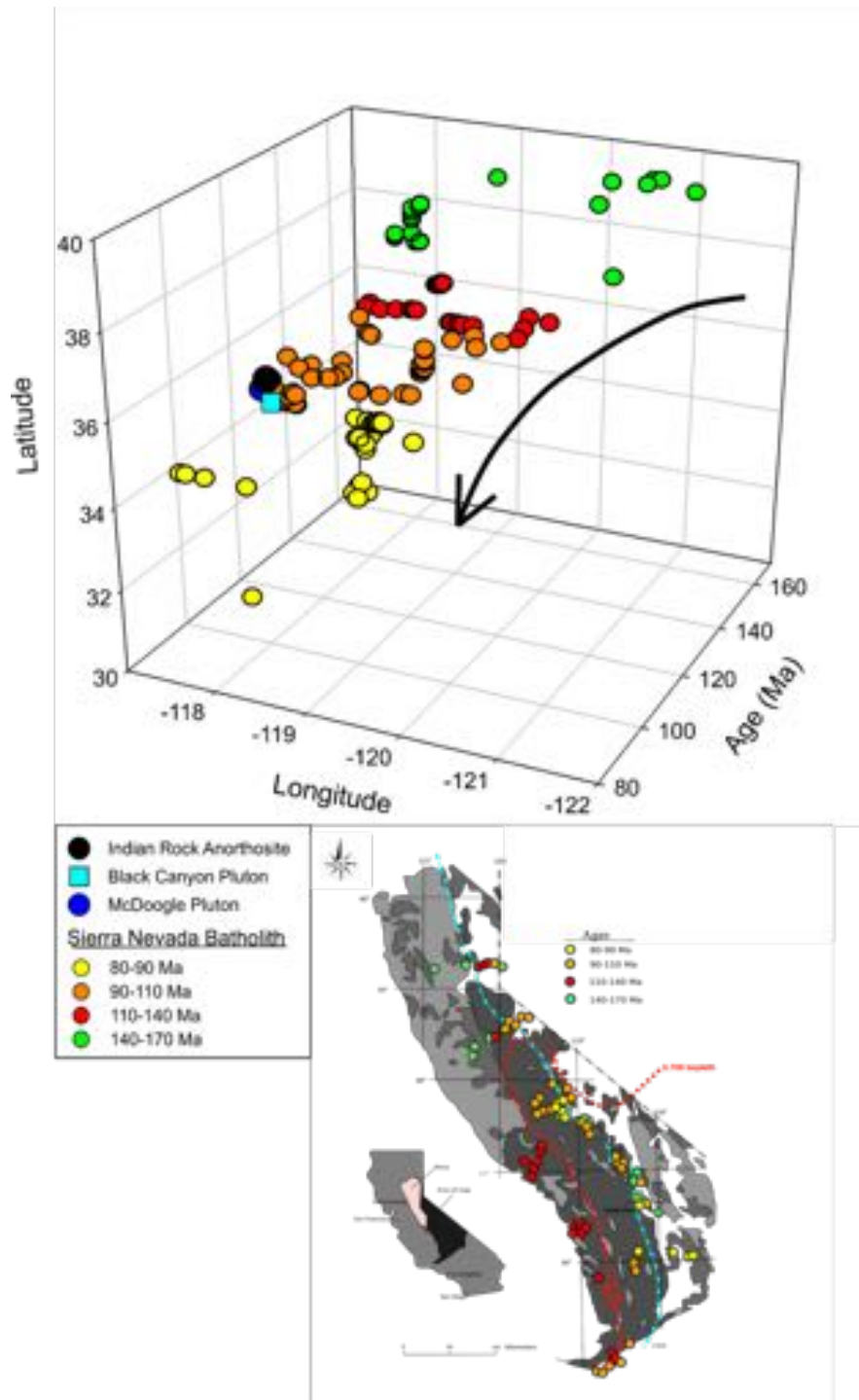


Figure 7.1: Spatial distribution of samples in the Sierra Nevada Batholith. This diagram shows the age distribution of the Sierra Nevada Batholith in a 3D space with the arrow showing the direction from older to younger rocks. The map of Sierra Nevada Batholith shows the age position of samples. The age and latitude-longitude data are from Kistler et al. (1986); Saleeby et al. (1987, 2007); Clemens-Knott (1992); Coleman et al. (1992); Olson (1992); Bradford (1995); Cousens (1996); Sisson et al. (1996); Truschel (1996); Ratajeski et al. (2001); Kylander-Clark et al. (2005); Glazner et al. (2008); Gray et al. (2008); Cecil et al. (2012); Nelson et al. (2013); and Chapman and Ducea (2019).

The apatite U-Pb ages of the IRA are 90.73 ± 0.19 and 107.17 ± 0.19 Ma for IRA 4-2 and IRA 18-6, respectively, and are intriguingly similar to the U-Pb zircon ages (Figure 7.2). Apatite has a closure temperature of 350-550 °C (Chamberlain and Bowring, 2001; Schoene and Bowring, 2007; Cochrane *et al.*, 2014; Schoene *et al.*, 2017) which aids in addressing thermochronologic processes during tectonothermal events (Kirkland *et al.*, 2018) and the late-stage crystallisation and subsolidus history of magmatic rock (Gawęda *et al.*, 2014). Zircons, on the other hand, have a U-Pb closure temperature of temperature > 900 °C (Schoene *et al.*, 2017). The zircon U-Pb and apatite U-Pb ages of sample IRA 4-2 show a difference of 3.27 Ma and 12 Ma for sample IRA 18-6. Sample IRA 4-2 has a U-Pb zircon age of 94.01 ± 0.05 Ma (> 900 °C) and the U-Pb apatite age of 90.73 ± 0.19 Ma (350-550 °C) which tells us it took approximately 3.27 Ma for the magmatic temperature to decrease from ~900 °C to ~550-350 °C, suggesting rapid cooling rates. The rapid cooling rates indicate that the plutons were emplaced in a weakly extensional deformation regime shown by the magmatic fabric of the plutons which are more prominent in the granodiorite of the McDoogie Pluton and the four westerly shear zones (Bench Canyon shear zone, Quartz Mountain shear zone, Kaiser Peak shear zone and Courtright/Wishon shear zone) which show steeply plunging stretching lineations. However, this is not the case for sample IRA 18-6, as the apatite U-Pb ages are older than the zircon U-Pb ages and do not agree with generally accepted closure temperatures of these minerals. More apatite geochronology is required to make a conclusive interpretation as later uplift and erosion history of this portion of the SNB.

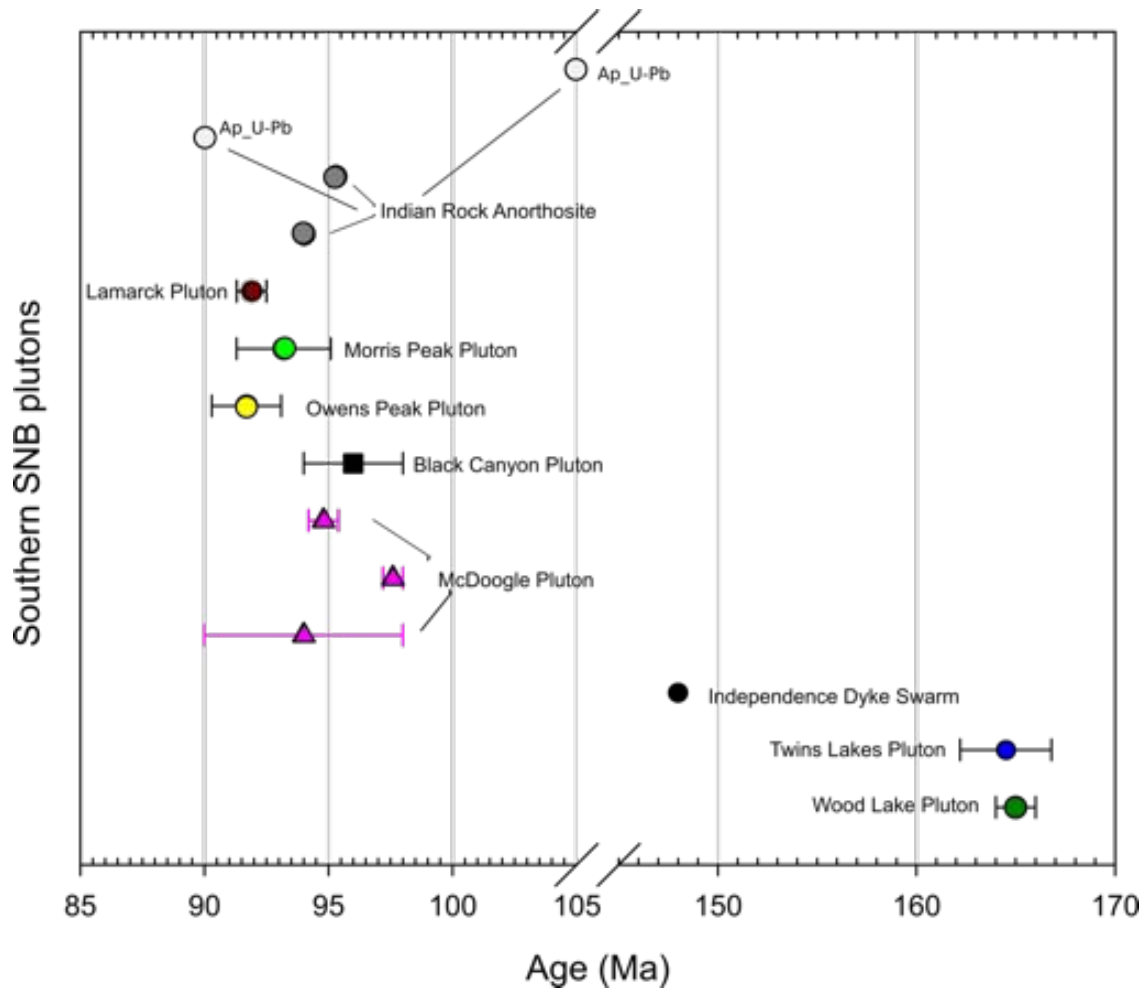


Figure 7.2: Timeline showing the ages of the Southern SNB and the zircon and apatite U-Pb ages of the IRA. IRA apatite U-Pb ages are shown as white dots, while grey dots show zircon U-Pb ages. Error bars are smaller than the symbols. The published U-Pb ages are from Mahan et al. (2003): McDoogle Pluton, Sawmill Lake Pluton, Carl (2000): Black Canyon Plutonic rock, Chen and Moore (1982) and Mahaffey (2016): Tinehama granodiorite, Wood Lake, Coleman et al. (2000): Lamarck granodiorite.

7.2 The Petrogenetic link between IRA and SNB

The zircon and apatite U-Pb data demonstrate that the IRA is coeval with the surrounding SNB lithologies and in this section; we test the petrogenetic link between the anorthosite and the surrounding granitoids to establish whether they are comagmatic. Moore (1963) has documented that the IRA and the granodiorite of the Black Canyon Pluton have a gradational contact and this observation is supported by the geochemical relationship between the rocks. Major element data demonstrates that the rocks share several compositional similarities. The Harker diagrams (Section 6.1.1) show fractionation trends between anorthosites of the IRA, BCP granodiorites and alaskite dykes from a more mafic source (basaltic) towards a highly evolved melt. A decreasing trend in CaO % and Al₂O₃ % with increasing SiO₂ % indicates fractional crystallisation by feldspars (e.g. plagioclase). The fractionation of the

plagioclase minerals will fractionate with CaO/Na₂O, therefore decreasing the elements in the melt resulting in the newly formed minerals having low CaO % and high Na₂O %. This can be observed using the plagioclase anorthite content in the analysed samples. The calculated anorthite content of the IRA ranges from An₇₉ – An₈₂, and the granodiorites of the BCP has An₃₁ while the McDoogie Pluton ranges from An₃₆ – An₄₄, and lastly the felsic dykes have An₁₈ – An₂₂. Elements like TiO₂ %, Ni (ppm), and Fe₂O₃ % show a positive trend with MgO %, which indicates fractional crystallisation controlled by ferromagnesian minerals (e.g. pyroxene). Comparison of the samples from this study and the broader Sierra Nevada Batholith as sampled from the *Navdat database* (<http://earthchem.org/>) (Figure 7.3) show similarities. For example, the alaskite dykes (this study) plots with the SNB alaskites. The granodiorites of the Black Canyon and McDoogie Pluton plot in-between the SNB gabbros and the SNB granodiorites. The Indian Rock Anorthosites plot with the SNB gabbros. The IRA and the granodiorites of the Black Canyon and McDoogie Pluton show primitive signatures in terms of their major elements; hence they plot with the gabbros (Figure 7.3). Mafic complexes in the central Sierra Nevada batholith record valuable information regarding the interaction of the mafic magmas and the generation of the continental crust (Nelson *et al.*, 2013). For example, the rocks in the Yosemite Valley Pluton show high ⁸⁷Sr/⁸⁶Sr (~0.7067), high δ¹⁸O values (6.6 – 7.3 ‰) and low ε_{Nd} (~ -3.4), and the homogeneity of ⁸⁷Sr/⁸⁶Sr in plagioclase and zircon data, shows the interaction between the two sources (Nelson *et al.*, 2013). More or less similar values can be observed with the granodiorites of the Black Canyon and McDoogie Pluton (Figure 7.3), with bulk rock ⁸⁷Sr/⁸⁶Sr (~ 0.70775 and ~0.70746), and ε_{Nd} (~-3.93 and ~-5.37) values, respectively. The plagioclase grains in the granodiorites have a calcic core and a sharp change to sodic rims, with calculated anorthite content ranging between An₃₁ and An₄₄ which form zonation seen in the petrographic studies, and this is commonly associated with mixing of mafic and felsic magma.

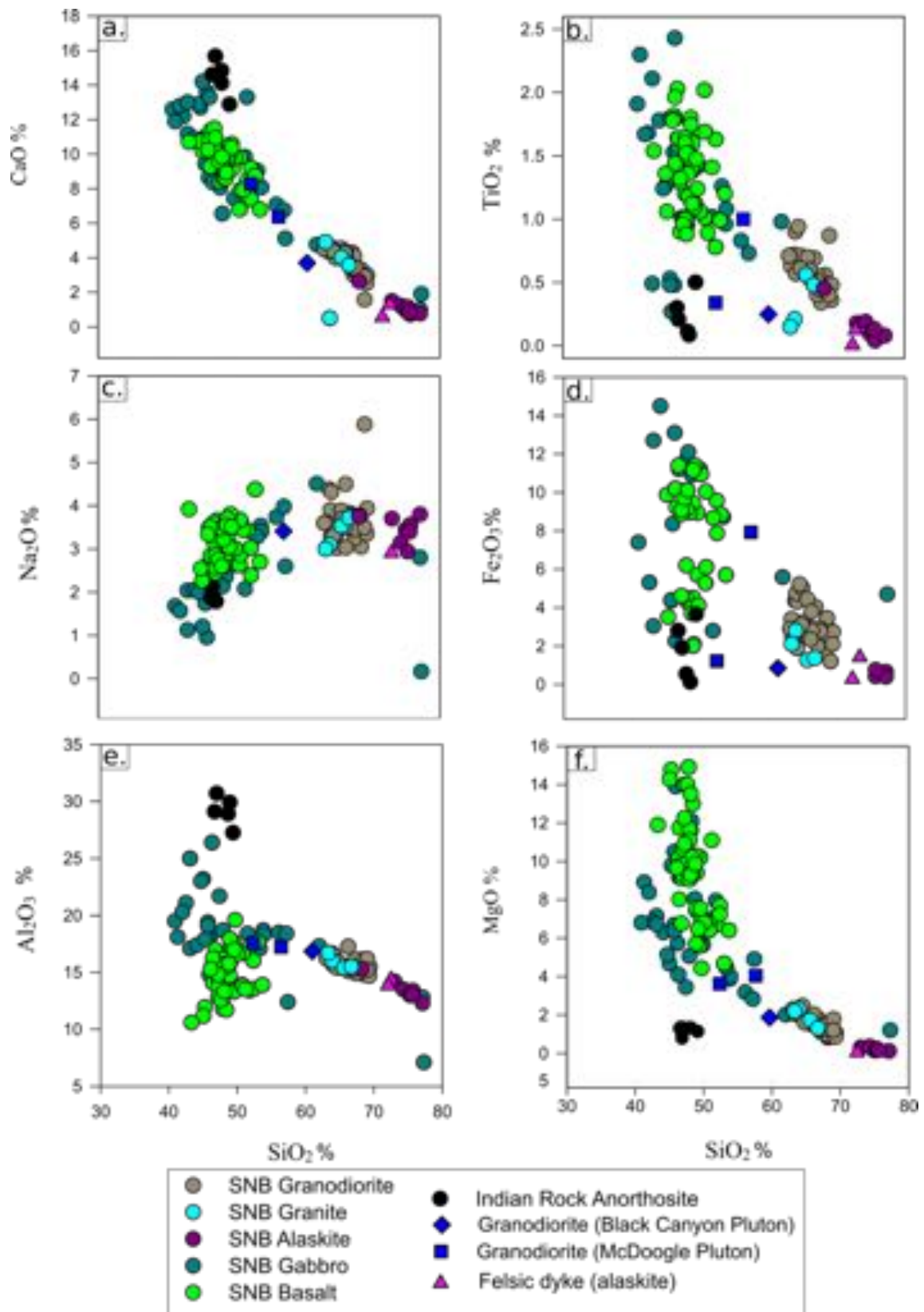


Figure 7.3: Harker diagrams of samples from this research compared with the Sierra Nevada Batholith samples from the Navdat (<http://earthchem.org/>) database.

The analysed samples (anorthosite, granodiorite, and felsic dykes) are enriched in incompatible REEs relative to chondrites, and this indicates partial melting of the source which leaves a restite that is enriched in HREEs (Hegner *et al.*, 2010). The chondrite-normalised REE patterns show that the IRA is more primitive than the granodiorites and the dykes (Figure 7.4). The IRA has a positive Eu anomaly compared to the granodiorites, which is caused by the fractionation of plagioclase grains leaving the melt Eu depleted, but this anomaly is very subtle ($[Eu/Eu^*]_N = \sim 1.12$). The IRA and SNB gabbros have similar Al_2O_3 % and CaO % concentrations, which are the main component in the formation of plagioclase minerals and they show crystallisation of plagioclase. The high MgO % signature in the SNB gabbros are due to high modal abundance of clinopyroxenes. During the formation of the anorthosite, plagioclase minerals which have high CaO/ Na_2O fractionate from the melt which increases the amount of Na_2O and decrease the amount of CaO in the melt. The granodiorites which forms from the melt will retrieve the melt signatures after the fractionation of the anorthosite-forming plagioclase. This can be observed in Figure 7.3, where the granodiorites show low CaO % and high Na_2O % relative to the anorthosites. The granodiorites of McDoogie and the Black Canyon Plutons have a more primitive signature compared to the SNB granodiorite. The initial Sr values of the rocks fall on the west side of the 0.706 isopleths, which are values that are >0.706 and they conform to the general initial Sr values of the SNB. The granitoids (granites, granodiorites, and monzonite) are depleted in Eu relative to other REE showing negative Eu anomalies and this is probably caused by retention of plagioclase in the source (Figure 7.4). The same characteristics have been identified in granitic rocks that are spatially associated with the massif-type anorthosites (Simmons and Hanson, 1978). The granitoids are typically depleted in HREEs and exhibit high U and Th relative to La and these exhibit signatures of a garnet-bearing lower crust (Hegner *et al.*, 2010). In a continental arc setting the granitoids are generated by the rising basaltic melt and heat from the mantle through the process of underplating (Bateman, 1992). REE patterns in the lower crust are highly fractionated, which means the Sm/Nd ratios will be low and the $^{148}Nd/^{144}Nd$ ratios will be lower than the chondritic mantle (Tarney and Windley, 1977; DePaolo, 1981). Therefore, rocks formed in the lower crust will retain the unradiogenic Nd isotope characteristics and low Sr isotope ratios of the lower crust, which is not the case with the samples from this study.

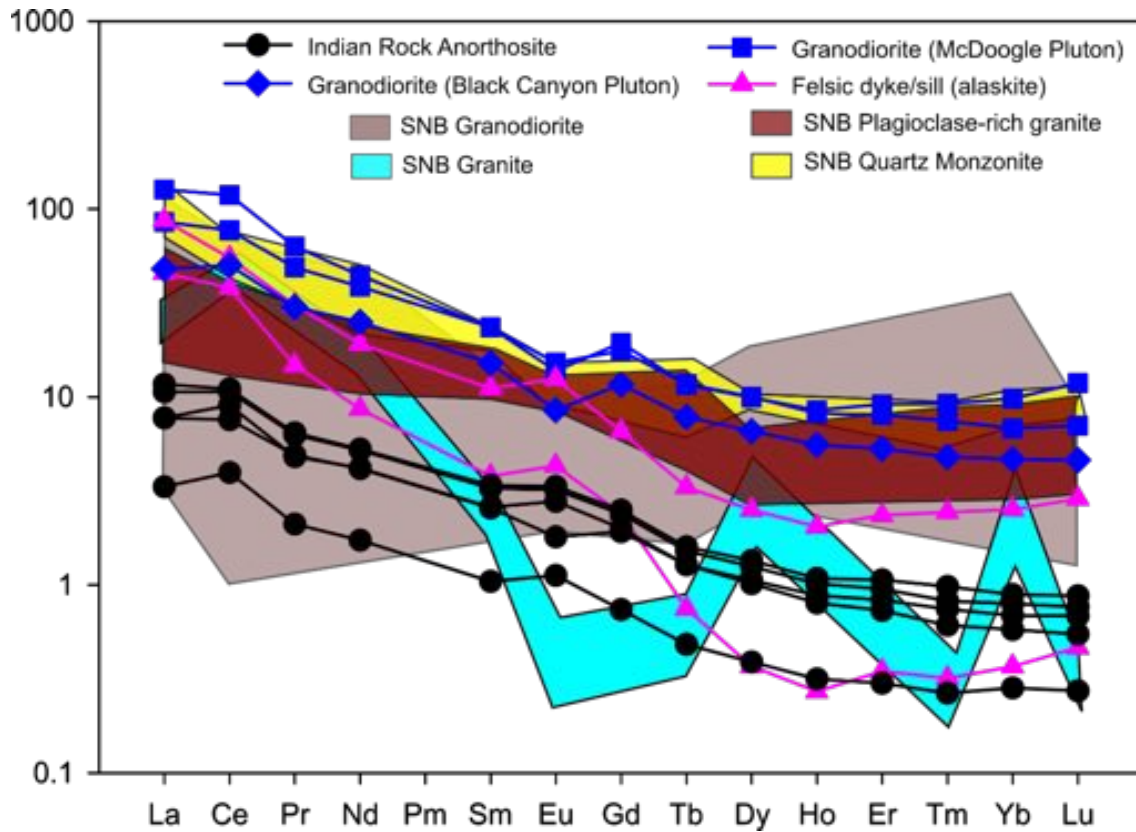


Figure 7.4: Chondrites normalised REE plots using chondrite values by Anders and Grevesse, (1989) of the samples of this study compared to Sierra Nevada Batholith samples extracted from the Navdat database.

The isotopic data from this study show fairly negative ϵ_{Nd} values and initial Sr values that fall in the range of the Mesozoic Sierra Nevada Batholith data extracted from DePaolo, (1981); Dodge and Bateman, (1988); Mukhopadhyay (1989); Mukhopadhyay and Manton (1994); Ducea and Saleeby (1998); and Ducea (2001) (Figure 7.5). The ϵ_{Nd} values of the granodiorite of the McDoogie Pluton (samples IRA 6-3) and the IRA cluster together showing a relationship between the two plutons, further showing an interaction between the pluton and potentially sharing a similar source or the magma parental of the anorthosites was contaminated by crustal components similar to the deep mafic crustally-derived granodiorites (Figure 7.5). The data shows the interaction between the young crust, the older crust and the upper mantle source. The low ϵ_{Nd} values (~ -4.41) and high $^{87}\text{Sr}/^{86}\text{Sr}$ (~ 0.7069) of bulk rocks indicate assimilation of the continental crust in a mantle-derived magma. The temperature and the water content are primary factors that affect the melting reactions, strain, melt segregation and accessory phase dissolution (which controls the Sr and Nd isotope systematics) (Zeng *et al.*, 2005). This explains the variable isotopic values in plutons like the McDoogie Pluton; also, the rate of melt-extraction (Zeng *et al.*, 2005) plays an essential role in determining the signatures.

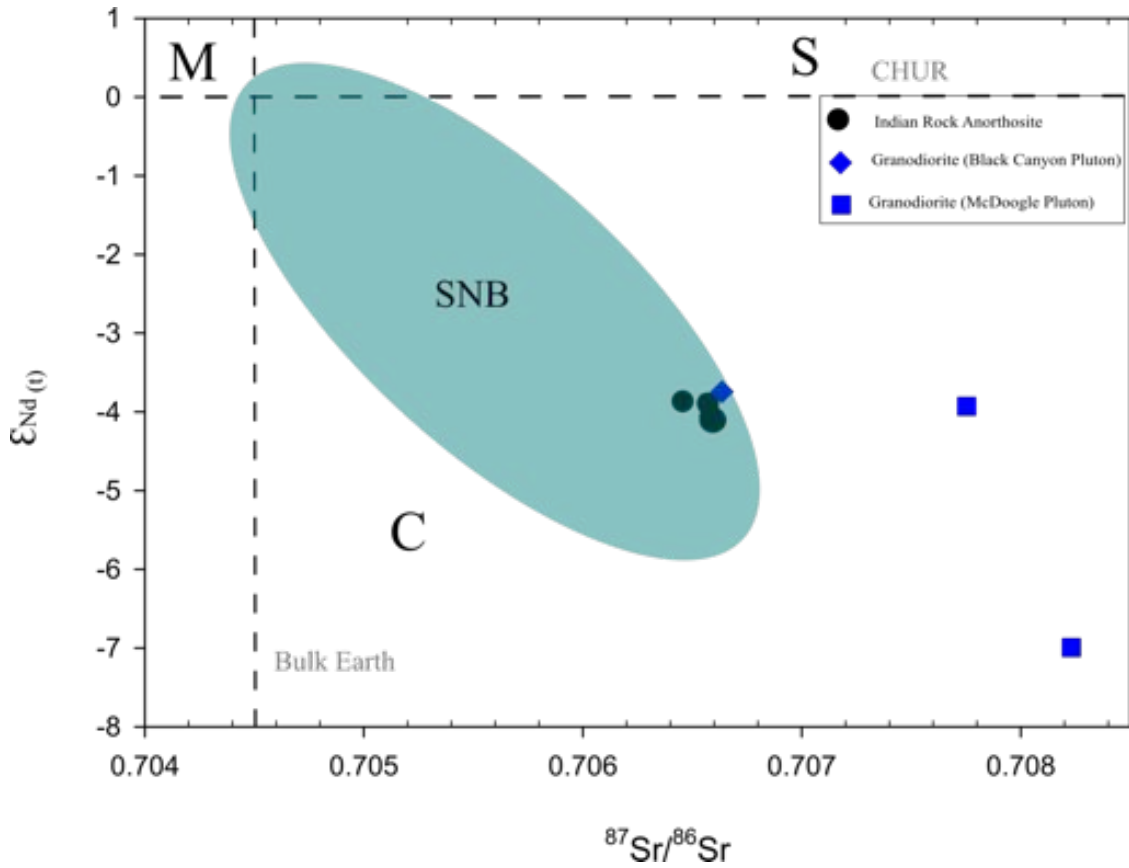


Figure 7.5: Samples from this research compared with Mesozoic isotopic data from the Sierra Nevada Batholith (green polygon). The data used for the green-oval was extracted from DePaolo, (1981); Dodge and Bateman, (1988); Mukhopadhyay (1989); Mukhopadhyay and Manton (1994); Ducea and Saleeby (1998); and Ducea (2001). The data includes compositions ranging from granodiorites, granites, quartz monzonite, and quartz diorite. M-young mantle, C-older lower crust, and S-sedimentary.

The isotopic composition of some plutons of the Sierra Nevada Batholith is plotted in Figure 7.6. The bulk rock initial $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd} values of this study are consistent with the signatures of the other rocks in the SNB. The IRA and the granodiorite of the Black Canyon Pluton are coincident with the isotopic signatures of the Lamarck Pluton, which occurs north-west of the IRA, while the McDoogle Pluton show variable signatures. The analysed samples (IRA, BCP, McDoogle Pluton and the felsic dykes) show negative Nb anomalies which suggest that the magma was derived either from depleted mantle or within subduction zone contaminated with old continent crust. The evolved isotopic ratio (relatively high initial $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd} values) implies that the rocks have experienced crustal contamination and similar values are observed in the rocks from the Yosemite Valley and xenoliths in the central Sierra Nevada Batholith (Beard and Glazner, 1995; Coleman *et al.*, 1995; Ducea and Saleeby, 1998; Ducea, 2001; Nelson *et al.*, 2013). The similarities in isotopic signatures suggest that the rocks may share a mantle source origin, which means they are comagmatic.

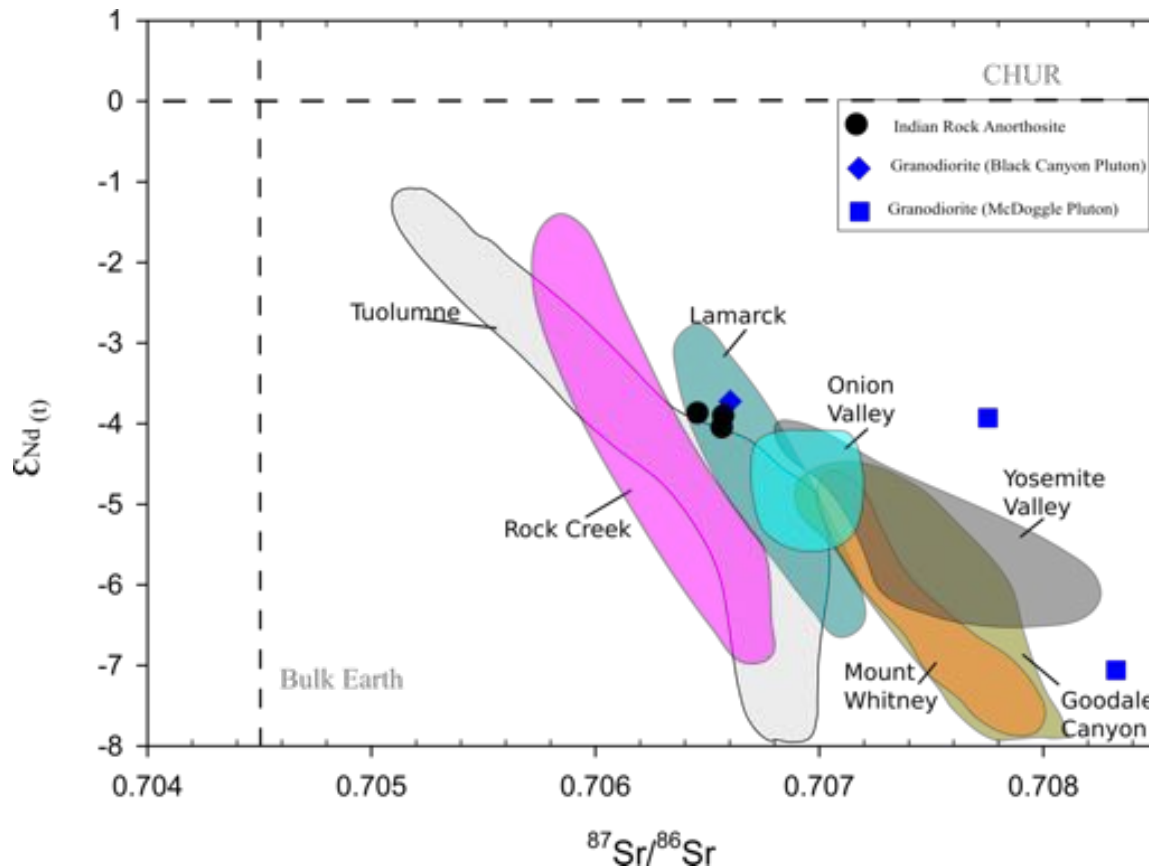


Figure 7.6: Plot of Indian Rock Anorthosite, the granodiorites of Black Canyon and McDoogle pluton compared to plutons of the southern Sierra Nevada Batholith. The data of the plutons were extracted from Nelson et al. (2013) and references therein.

7.3 Classifying the IRA

The age of the IRA discussed in Section 7.1 disproves the hypothesis that it is a Proterozoic-aged anorthosite. Section 7.2 demonstrates that the IRA and the surrounding SNB rocks are comagmatic or the magma parental of the anorthosites was contaminated by crustal components similar to the deep mafic crustally-derived granodiorites. Also, the section has shown that the McDoogle and Black Canyon Pluton are mantle-derived. The IRA has a gradational contact with the surrounding granodiorites of the BCP, which rules out the possibility of the anorthosite being xenolithic (Moore, 1963). In this section, the focus will be using texture and geochemical data to compare the IRA to cognate anorthosite inclusions, massif-type anorthosite and Archean anorthosite. Cognate anorthosite inclusions represent the accumulation of plagioclase from the same magma and the host, commonly found in arc terranes (Ashwal, 1993). Cognate anorthosite inclusions are found in island arcs and arc-root plutonic complexes (in continental arcs) and they mostly occur as both individual megacrysts or

coarse glomeroporphyritic aggregates ranging from >1 cm to hundreds of metres (Figure 7.7) (Morrison *et al.*, 1983; Ashwal, 2010).

In its field structure and plutonic nature, the IRA resembles a typical, albeit small, massif-type anorthosite. However, the high An content in the IRA (analysed: An₇₂ - An₈₉, calculated: An₇₉ - An₈₂) is more typical of Archaean anorthosites (An₆₁- An₉₄) and cognate anorthosite inclusions in arcs (An₆₃- An₁₀₀) (Kristmannsdóttir, 1971; Ashwal, 2010; Ashwal and Bybee, 2017). The adcumulus texture, mineral content and whole-rock geochemistry resembles are cognate anorthosite inclusions in continental arcs. The high An content has been previously linked to the water content in the parental melt (e.g., Stern, 1979; Arculus and Wills, 1980; Sisson and Grove, 1993). Cognate anorthosite inclusion in island arcs have compositions ranging from anorthosite to gabbroic anorthosite (80 - 99 % plagioclase) and the An content range between An₆₃ - An₁₀₀ (Ashwal, 1993). Textural varieties include “mottled” anorthosites containing pyroxene or olivine-rich areas (2 - 15 cm) and banded types with mafic layers. The texture, size, An content and the composition of the anorthosite inclusions in island arcs are similar to the IRA; however, the main difference is the geological setting, where the IRA occur in a continental arc. Cognate inclusions in continental arcs are described as gabbroic (noritic) anorthosite (77.5 - 90 % plagioclase) measuring to 1 km across. They have well developed magmatic structures which include layering ranging between a few mm and 1 m thick (Ashwal, 1993). In terms of texture, they are dominantly adcumulus with varying proportions of plagioclase (An₇₀ - An₁₀₀). The texture, An content, size, geological setting of these inclusions are similar to the IRA, but the main difference is the plagioclase modal abundance where the IRA is considered to be an anorthosite as it has over 90% plagioclase. The IRA shows most similarities to cognate inclusions in continental arcs, but it is slightly larger than the typical size of cognate continental arc inclusions and more typically anorthositic.

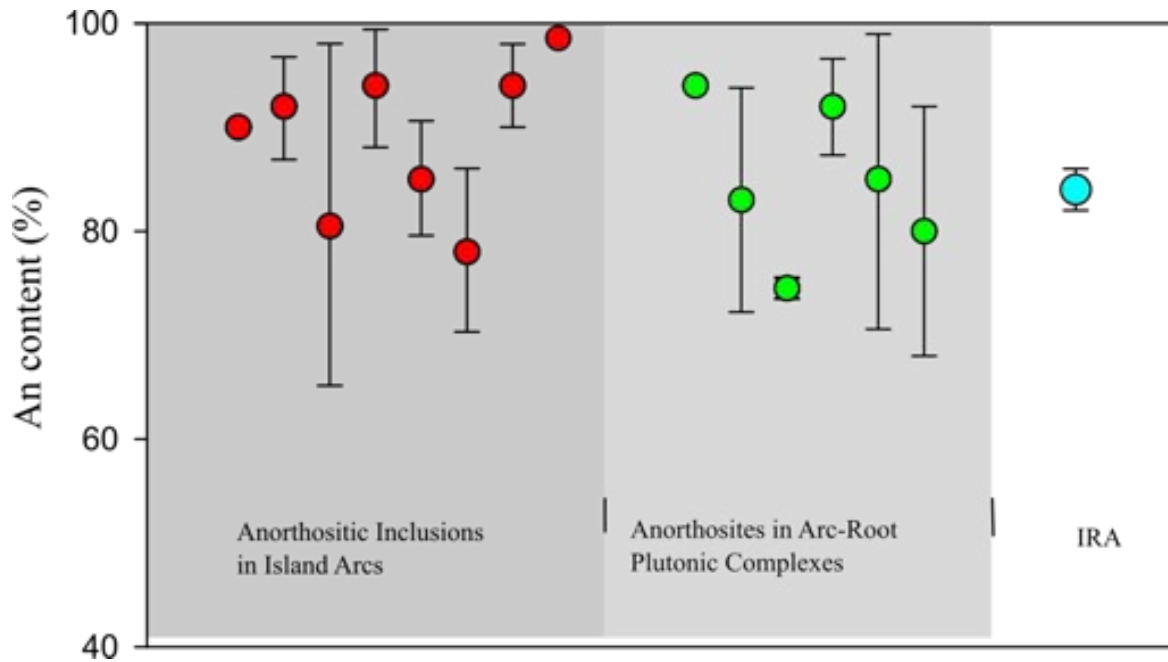


Figure 7.7: An content of cognate anorthosite inclusions and the IRA (data extracted from Ashwal (2010)). The grey areas show different types of cognate anorthosite inclusions in different settings. The red dots show anorthositic inclusions in island arcs, green dots are anorthosites in arc-root plutonic complexes and the sky-blue in the IRA.

Although the IRA shows similarities to cognate inclusions in continental arcs, it also has a typical massif-type anorthosite nature in terms of structures and morphology. Figure 7.8 shows the comparison of the analysed sample from this study and massif-type anorthosites. Geochemically, the IRA fall within the massif-type anorthosites which shows some similarities. The IRA has specific properties which are typical of massif-type anorthosites, i.e. plagioclase making up >90% of the rock, high CaO %, Al_2O_3 %, Na_2O %, Sr and LREE enrichment, positive Eu anomaly and low Fe-Ti oxides (Simmons and Hanson, 1978; Philpotts, 1981; Ashwal, 1993; Wiebe, 1993; Maji *et al.*, 2010; Ashwal and Bybee, 2017). The IRA is enriched in light REE, but it has a smaller positive Eu anomaly ($[\text{Eu}/\text{Eu}^*]_{\text{N}} = \sim 1.12$) compared to most massif-type anorthosites ($[\text{Eu}/\text{Eu}^*]_{\text{N}} = \sim 2.96$) (Figure 7.9). Small positive Eu anomalies have been observed in the Cascadian HAOT (High-alumina olivine tholeiite), which is the function of the fractionation process of a high-Al basalt (Bacon *et al.*, 1997). The magnitude of the Eu anomaly is strongly dependent on the oxidation potential in the magma system at the time of crystallisation (Weill and Drake, 1973). The oxygen fugacity in massif-type anorthosite magma may be sufficiently reduced to cause large Eu anomalies. The model presented by Weill and Drake (1973) showed Eu anomalies in plagioclase as a function of oxygen fugacity and temperature of the system. The plagioclase Eu anomaly decreases with increasing oxygen fugacity and increasing temperature (Hanson, 2003). Primary arc magmas can be produced by either melting the asthenospheric mantle

wedge arc volcanoes by decompression, volatile fluxing or both (Sisson and Bronto, 1998; Conder *et al.*, 2002; Chin *et al.*, 2012). This process adds water into the system, which oxidises the magma system and in turn affecting the plagioclase Eu affinity. A wet, more oxidised arc source for the IRA may result in a less pronounced Eu anomaly, as observed in Figure 7.9. Most igneous rocks in continental arcs like the SNB contains significant amount of volatiles (H_2O , CO_2), and the source is hydrous minerals carried into the Earth by the subducting slab (Sisson and Grove, 1993a, 1993b; Pearce and Peate, 1995; Grove *et al.*, 2012; Ducea *et al.*, 2015). The hydrous minerals release their bound H_2O into the overlying hotter mantle where melting take place and continues as buoyant hydrous mafic magmas ascend (Grove *et al.*, 2012). From this comparison study, the IRA can be thought of as a unique type of anorthosite having a plutonic nature, similar to massif-type anorthosite and adcumulus texture, mineral content similar to cognate inclusions in continental arcs.

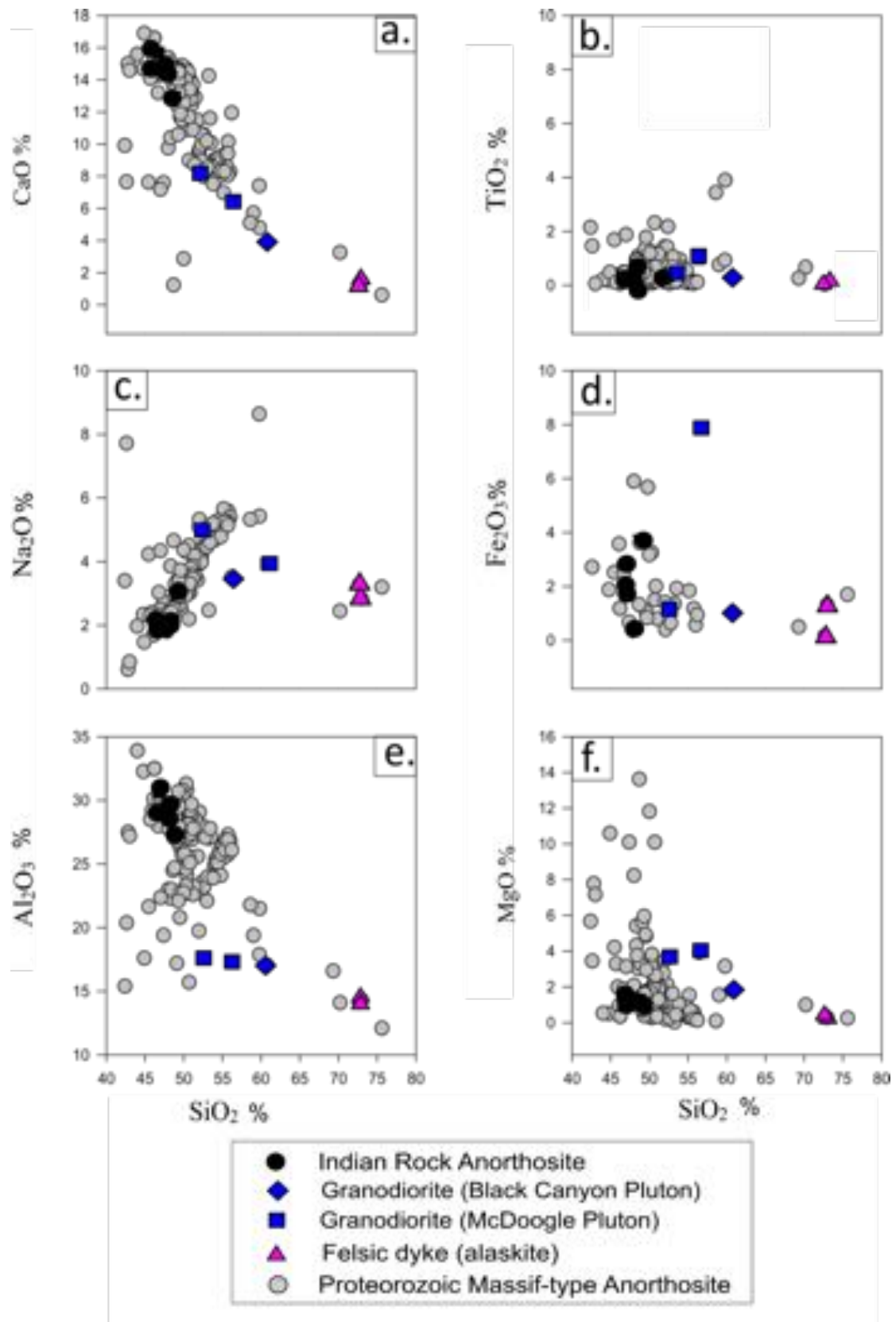


Figure 7.8: Samples from this study compared to the Proterozoic massif-type anorthosites. The data was extracted from Navdat database. The IRA falls at the centre of the Proterozoic massif-type anorthosites indicating similar major element compositions.

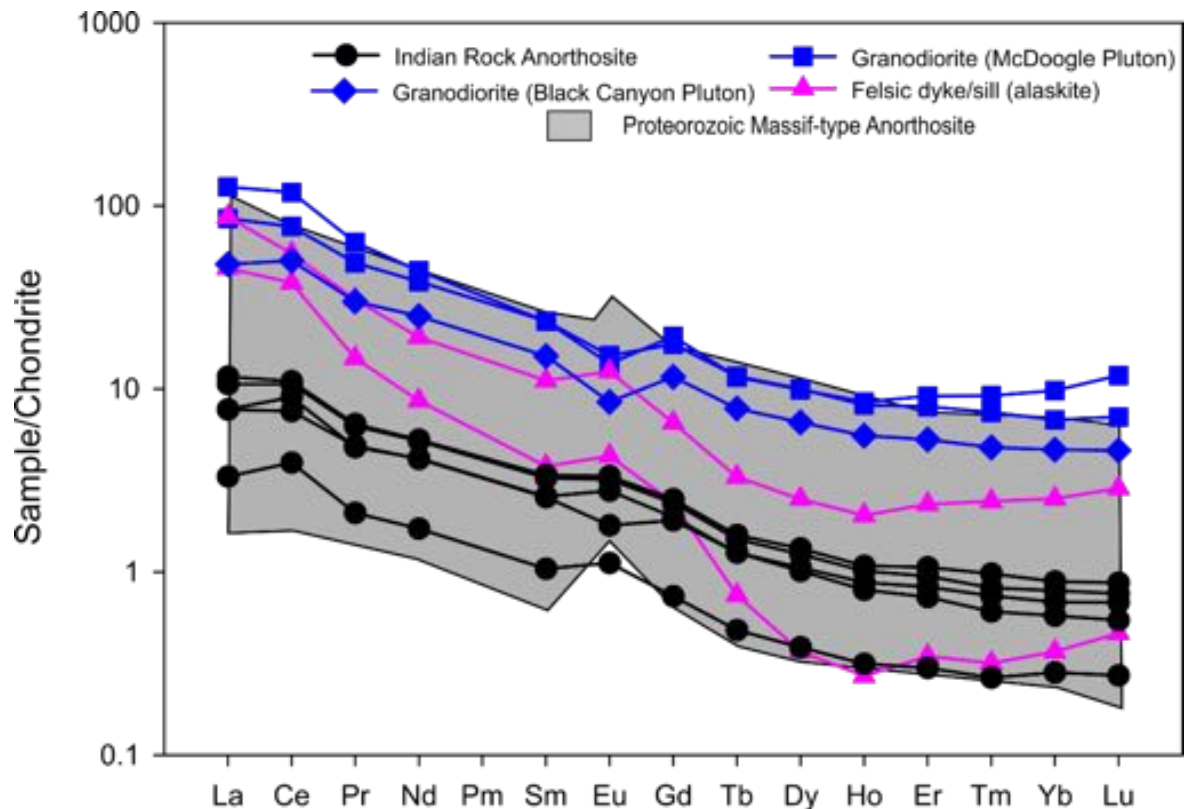


Figure 7.9: Chondrite-normalised REE using CI chondrites values by Anders and Grevesse, (1989). Samples from this study compared to the Proterozoic massif-type anorthosites. The data was extracted from Navdat database.

7.4 The Formation of the IRA

Previous sections established that the IRA is a unique type of anorthosite with similarities to both massif-type and cognate inclusion type anorthosite. The leading question is then what conditions are required to form a relatively small, isolated body of this type of anorthosite in a continental arc environment? A study done by Schmidts and Jagoutz (2017) has identified three possible parental magmas for the SNB which are tholeiitic basalt, basalt and andesitic melts. In Section 6.4, these different parental compositions received amounts of crustal assimilation under different conditions (i.e. pressure, temperature and water content) using alphaMELTS. The andesitic melt crystallises plagioclase and clinopyroxene with little K-feldspar, and the presence of water increases the amount of clinopyroxene (from ~27% to 60%). The basaltic melt crystallises plagioclase and clinopyroxenes, and the presence of water encourages new phases (hematite/ilmenite, apatite, melilite and perovskite) to appear at the end, which form the interstitial phases. The tholeiitic basalt melt under dry conditions crystallises plagioclase, pyroxene and other phases like K-feldspar, corundum and hematite. The addition of water in the melt encourages additional phases to form like biotite, hornblende, hematite/ilmenite, corundum, apatite, perovskite and pyrite. Polybaric experiments showed that

clinopyroxene forms at high temperatures and pressure (~1413.84 °C and ~23.26 kbars), while plagioclase forms at lower temperatures and pressure (~1187.17 °C and ~17.20 kbars). The crystallisation of the melts at high pressures (10-15 kbar) produces high Al-rich pyroxene at liquidus (Fram and Longhi, 1992; McLelland and Selleck, 2010). The low water content increases the amount of plagioclase, due to the plagioclase stability and Al depletion, and at low pressures, the amount of clinopyroxenes decreases (Leuthold *et al.*, 2015). Another important effect of pressure is the increased An content at lower pressure and high water level. At high-pressures (> 16 kbar) the plagioclase An content of all the melts range from An₃₆ to An₆₈ and at lower pressures (4 kbar) it ranges from An₅₉ to An₇₈. Tholeiitic basaltic melt has high An content compared to the basalt and andesite melt. This is due to the higher Al₂O₃ % and CaO % in the melt relative to the other melts (basalt and andesite). Crustal assimilation in the system increases the amount of plagioclase with high An content.

A hydrous melt parental melt that is capable of generating anorthosite has been proved not to be generally valid shown by the study done by Yoder (1969). Anorthosites are thought to have formed from a relatively dry melt because of their anhydrous mafic mineralogy (Selbekk *et al.*, 2000). The contact aureoles around anorthosite massif are also conspicuously dry, especially where they contain osumilite (Berg, 1977), which has a low tolerance for H₂O (Morse, 1982). In this case, there is a gradational contact between the IRA and the granodiorite (Moore, 1963). The presence of hydrous minerals like the hornblende indicates that the eastern part of the batholith has high levels of H₂O (Memeti *et al.*, 2010). It has been thought that anorthosite complexes form in arcs (Ashwal and Bybee, 2017), which means that the role of water in the formation of these rocks cannot be ignored if this tectonic setting is considered to be true. A water-rich parental melt of the anorthosite has a great effect on the phase equilibria controlling the amount and composition of the plagioclase grains (Sisson and Grove, 1993a; Ohtani, 2006). The presence of water can enhance melting and change of the phase relations of the mantle minerals (Sisson and Grove, 1993a; Chen *et al.*, 1998; Hirschmann, 2006; Ohtani, 2006; Grove *et al.*, 2012). Experimental work on hydrous basaltic compositions has shown that increased water content in the magma suppresses the amount of plagioclase and produces higher An content (Sisson and Grove, 1993a). A similar effect is observed in the alphaMELTS models, where the addition of water to the system has a general effect of lowering the amount of plagioclase produced and increasing the amount of clinopyroxenes produced. A schematic diagram of in Figure 7.10, albite-anorthite system shows the change of plagioclase composition with changing intensive variables like temperature, pressure, and water content. At constant pressure and low water content, decreasing temperature causes crystallisation which causes magma differentiation (black arrow in Figure 7.10). At the same temperature, the plagioclase An-content in the system increases with increasing water content (red) and decreasing pressure (green) (Bennet *et al.*, 2019). At high pressure, the plagioclase is more sodic than at lower pressures (Yoder, 1968; Ustunisik *et al.*, 2014), hence under decompression conditions from high to low pressures (yellow to red lines) causes resorption and crystallisation of more

calcic plagioclase. The opposite effect is observed when decompression occurs under hydrous conditions which results in degassing and crystallisation of more sodic plagioclase (Blundy and Cashman 2005). Whilst the addition of the water depresses the plagioclase loop causing the plagioclase composition to become more anorthitic (calcic) (green to red in Figure 7.10). The smaller amount of plagioclase produced tends to have high An content compared to the anhydrous melts. The high An content has been associated with water in the parental magma (Ashwal, 1993). Selbekk *et al.* (2000) noticed that the presence of water-bearing phases lowers the solidus temperature of the rock (also observed in Figure 7.10). With the presence of water in the melt, many new phases start to appear, and the plagioclase shows an increase in the An content having values close to the analysed samples. The effect of water on plagioclase grains has been noticed and noted by other petrologists including Simmons and Hanson (1978) and Selbekk *et al.* (2000). The water leads to more calcic and less Na₂O-rich melt compositions than the anhydrous melt (Selbekk *et al.*, 2000).

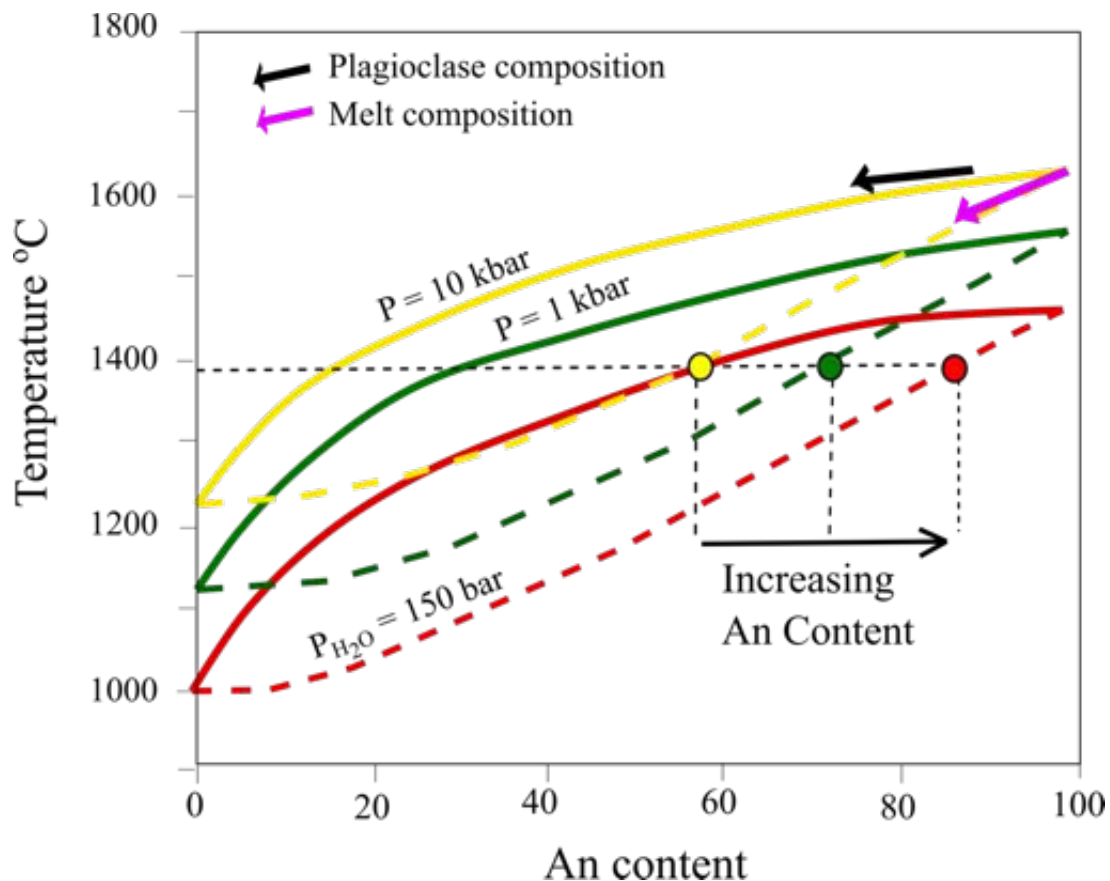


Figure 7.10: A schematic phase diagram of the albite-anorthite system showing the effect of changing variables (e.g. temperature, pressure and water content) on plagioclase composition (modified from Bennet *et al.*, (2019)).

A tholeiitic basalt melt under hydrous condition produces phases that were identified during petrological studies. The kind of melts with high Al (Al₂O₃ >17 %) show abundant plagioclase

accumulation (Whattam, 2018). There is significant evidence that favours high-Al melt as the parental magma composition and the high-Al in the melt would be in equilibrium with plagioclase cumulates (Emslie *et al.*, 1994). A partial melt of a tholeiitic source yield melts containing nearly 70% plagioclase (Simmons and Hanson, 1978). A tholeiitic melt or high Al melt has been favoured to be the parental melt of massif-type anorthosite by other researchers like Philpotts (1968); Simmons and Hanson (1978); Wiebe (1980); Morse (1982); Emslie (1985); Olson (1992); and Scoates and Mitchell (2000). Generation of the plagioclase rich magma is possible with the fractional crystallisation of pyroxene from a basaltic magma (Emslie, 1985). The observations from the alphaMELTS models show crystallisation of plagioclase when extracting pyroxenes from the melt. At high pressure, plagioclase rich melt can be derived by partial melting of a mafic parental source and extraction of pyroxenes from the melt (Simmons and Hanson, 1978). Massif-type anorthosites typically contain >90% plagioclase with low mafic minerals. Therefore, it can be said that the IRA formed from a hydrous high Al-melt in a high pressure, and high-temperature environment, where a small amount of plagioclase with high An content can accumulate.

Arc magmas are produced by melting of the downgoing oceanic crust, which has fairly constant $\epsilon_{Nd} \sim +10$ (Ducea, 2001). But in this case, the obtained values are negative which fits perfectly with the negative values of most California arc magmas (+5 and -12 with an average of ~ -6) (Ducea, 2001). A possible explanation for this is the second stage melting and/or fractionation and generation of the residue taking place in the complex transition between lower crust and upper mantle in the thickened continental crust (Hildreth and Moor bath, 1988). Figure 7.11 shows the change in ϵ_{Nd} values to a more evolved signature with increased distance from the margin. There are two possibilities illustrated in Figure 7.11. The first one (Figure 7.11 a) is the assimilation of older crustal material to the mantle source which makes the ϵ_{Nd} values to be negative, the second one (Figure 7.11 b) is the increased assimilation of the crustal material during crustal thickening, which is the more likely possibility. Crustal thickening in the area has been documented by researchers such as Ducea *et al.* (2015). The analysed IRA samples (this research) show more evolved signature and they occur at a back arc area which is further from the convergence margin meaning that it may have interacted with more evolved crustal material during crystallisation.

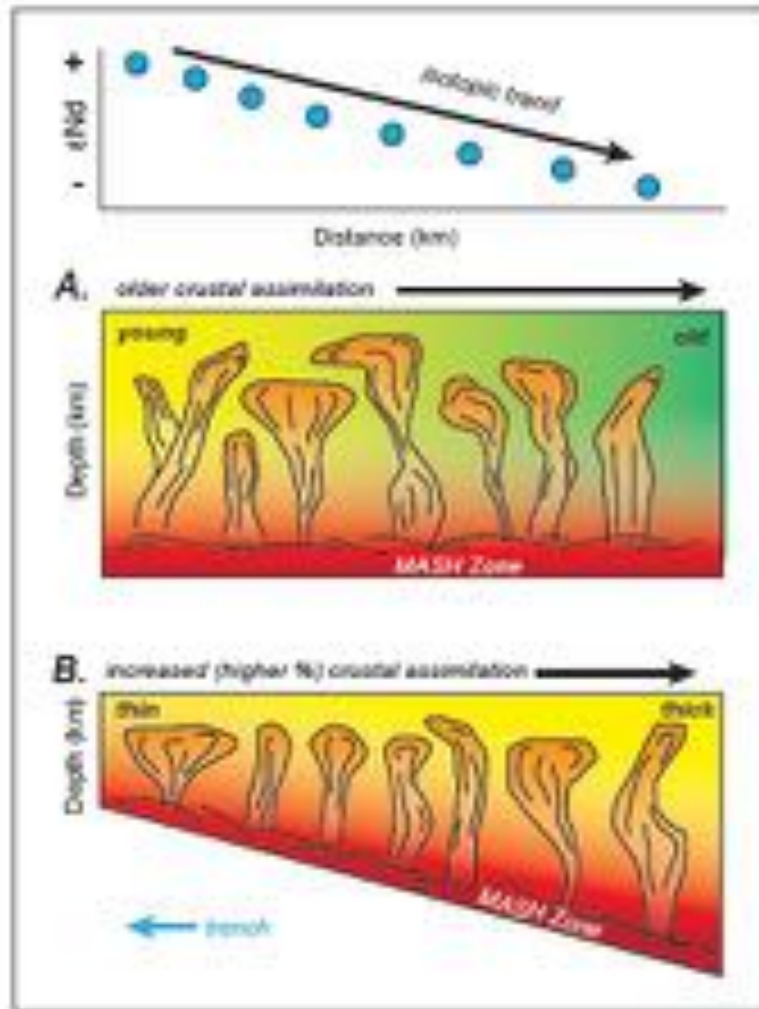


Figure 7.11: The change of ϵ_{Nd} values with distance from the trench or convergence margin (Chapman et al., 2017). a, Shows the assimilation of the older crustal material to the mantle source. b, Shows increased crustal assimilation to the mantle source during crustal thickening.

The models in Figure 7.12 show possible scenarios that may have occurred during the formation and emplacement of the IRA. The two magma sources (Depleted Mantle (DM) and Enriched Mantle I (EM I)) are compared to see which one is more likely to be the probable source for IRA. Figure 7.12 a shows two partial melt crystallisation trends from the DM (depleted mantle) field, which is contaminated by a younger and an older crust material with different radiogenic isotope compositions. The first path in Figure 7.12 a shows a trend of the DM, which is contaminated by a younger crust with ϵ_{Nd} values near CHUR to form the IRA and the surrounding granodiorites (McDoogie Pluton- IRA 6-3 and Black Canyon Pluton- IRA 18-2). The second model shows a path that is contaminated by an older crust to form sample IRA 18-1 (granodiorite of McDoogie Pluton). Figure 7.12 b shows a fractionation trend from an Enriched Mantle I (EM I) which shows two paths of evolution being contaminated by a younger and an older crust. The first path shows trends of the magma being contaminated by a younger crust to form the IRA (sample IRA 3-1, IRA 18-5 and IRA 18-7) and the granodiorites (sample IRA 6-3 and

IRA 18-2). The second path, which shows contamination by an older crust formed the granodiorite (sample IRA 18-1). An EM I melt (Figure 7.12 a) requires relatively less contamination to have the same degree of radiogenic enrichment compared to the DM depending on the concentration of Rb, Sr, Sm-Nd in the crust and mantle (Figure 7.12 b). However, the rocks show more mantle signature than a crustal signature (which is observed in Figure 7.3); therefore, this makes the DM source to be probable in this case. Contamination of the mantle-derived magma played a crucial role in forming the geochemical and isotopic signatures observed in the area. The negative Nb anomalies suggest the magma was derived from a depleted mantle or within a subduction zone contaminated with old continental crust (Mohan *et al.*, 2013; Polat *et al.*, 2009). In the field, many metamorphic screens were observed, and this shows some form of interaction of the crust and the magma (Memeti *et al.*, 2010). Contamination components are more likely to be broadly granitic in compositions and they are introduced into a mantle-derived anorthosite producing magma by either assimilation of deep crustal rocks or by blending with partial melt from the crust (Michot, 1964; Philpotts, 1968; Ashwal and Bybee, 2017). Different degrees of silicic contamination control the isotopic compositions of the rocks and this influences the mineral assemblages of the anorthosite (Ashwal and Bybee, 2017). The major element plots show that the granodiorites have a more primitive signature relative to the SNB granodiorites. Sample IRA 6-3 plot near the anorthosites and this indicate similar signatures which might be inherited from the deep crustal contamination.

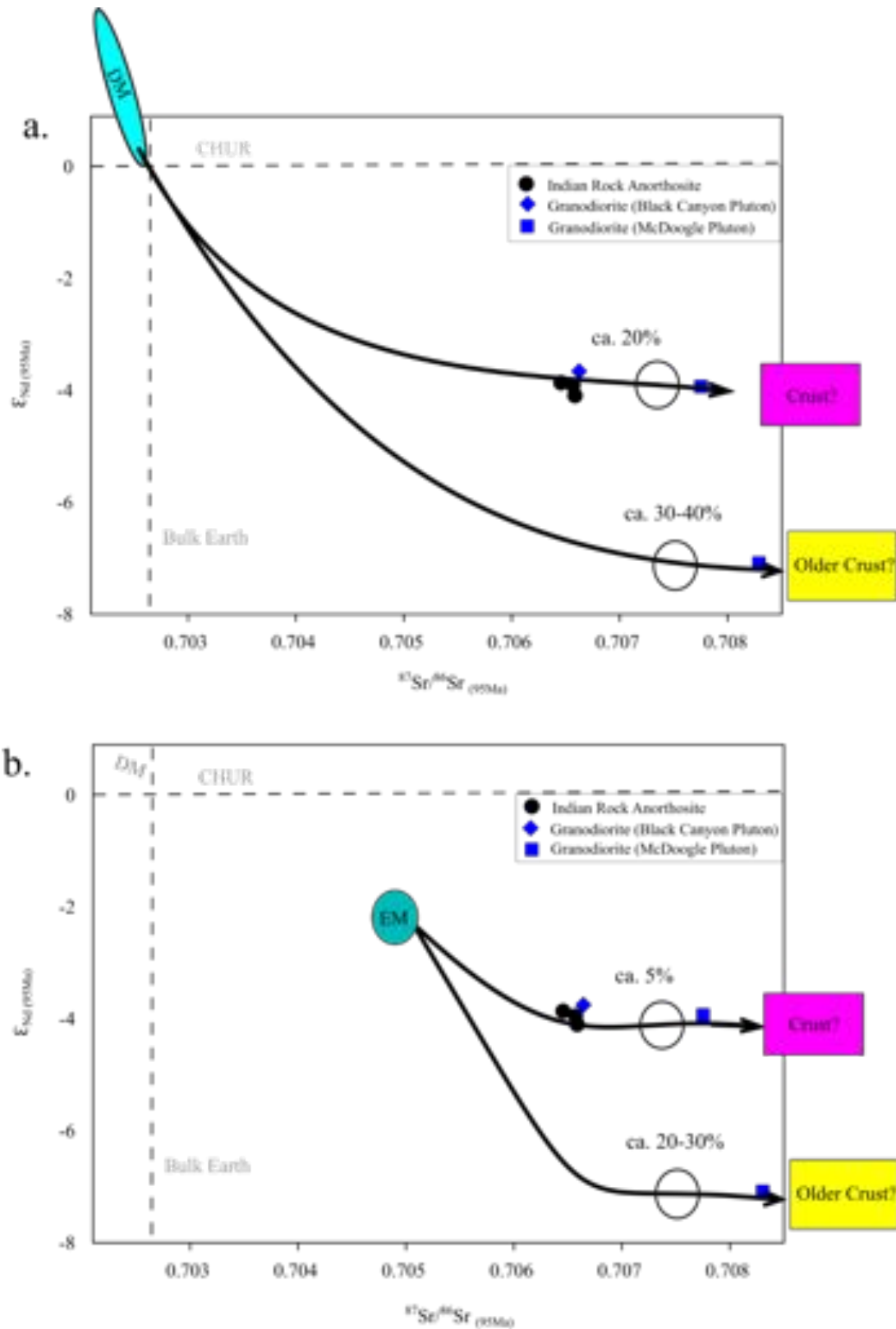


Figure 7.12: Fractionation trend of Depleted Mantle and the Enriched Mantle I. a, the DM fractionation trends with addition of the crustal material and this shows the contamination of the mantle source with the older and younger crustal material. b, The EM I source trend shows lesser crustal contamination compared to the DM to obtain the observed isotopic signatures in our samples.

The McDoogie Pluton shows more complex radiogenic isotope signatures, which were induced by contamination from the older and the younger crust. Bateman (1992), Barnes *et al.* (2004) and others have challenged the idea of plutons being emplaced by traditional mechanisms of diapirism, wall-rock assimilation and stoping. Saleeby *et al.* (2008) and Stearns and Bartley (2014) presented an emplacement model of plutons in the form of conduit feeders/stockwork of dykes. The Black Canyon Pluton was emplaced in a highly foliated wall rock which guided the dykes through diversely oriented fractures (Stearns and Bartley, 2014). This can be seen in the field by the magmatic fabric and the preferred orientation (NW-SE) of the mafic enclaves in McDoogie Pluton. The presence of mafic enclaves in many plutons of the central Sierra Nevada Batholith (e.g. Mount Givens Pluton) has been interpreted as the result of magma mingling (McNulty *et al.*, 2000). The enclaves reflect the input of mafic magma during the formation of the granodiorites (McNulty *et al.*, 2000). The isotopic compositions indicate magmatic mixing between different sources indicated by the Sr ratios. These show the complexities of an arc environment. The isotopic data shows contamination of the mantle-derived melt by crustal material. Melt derived from the mantle stalls and accumulate in the mid-crust, which results in fractionation crystallisation of early formed phases and assimilation of the wall-rocks (Tamura *et al.*, 2016). Subduction arc environments are known to be complex areas where a hydrated magma from the downgoing slab interacts with an older continental crust which is made up of crust of different ages (Whattam, 2018). The granodiorite of the McDoogie Pluton (sample IRA 18-1) shows contamination by a source that is more evolved which is an older crust in the region, while the granodiorite of the McDoogie Pluton (sample IRA 6-3) shows contamination by a less negative ϵ_{Nd} source, which is the younger crust. This shows that the McDoogie Pluton was emplaced in a diverse country-rock terrane reflected by the varying $\epsilon_{\text{Nd}} (95\text{Ma})$ vs $^{87}\text{Sr}/^{86}\text{Sr} (95\text{Ma})$ values. The process of multiple intrusions can explain the varying Sr and Nd values within a similar unit. Episodic construction of plutons results in many boundaries between intrusive units due to a wide variety of processes governed by the interaction with other chambers (Zák and Paterson, 2005). Kistler *et al.* (1986) concluded that the varying values are results of each intrusion unit evolving separately from the magma mixing of basaltic and granitic magma.

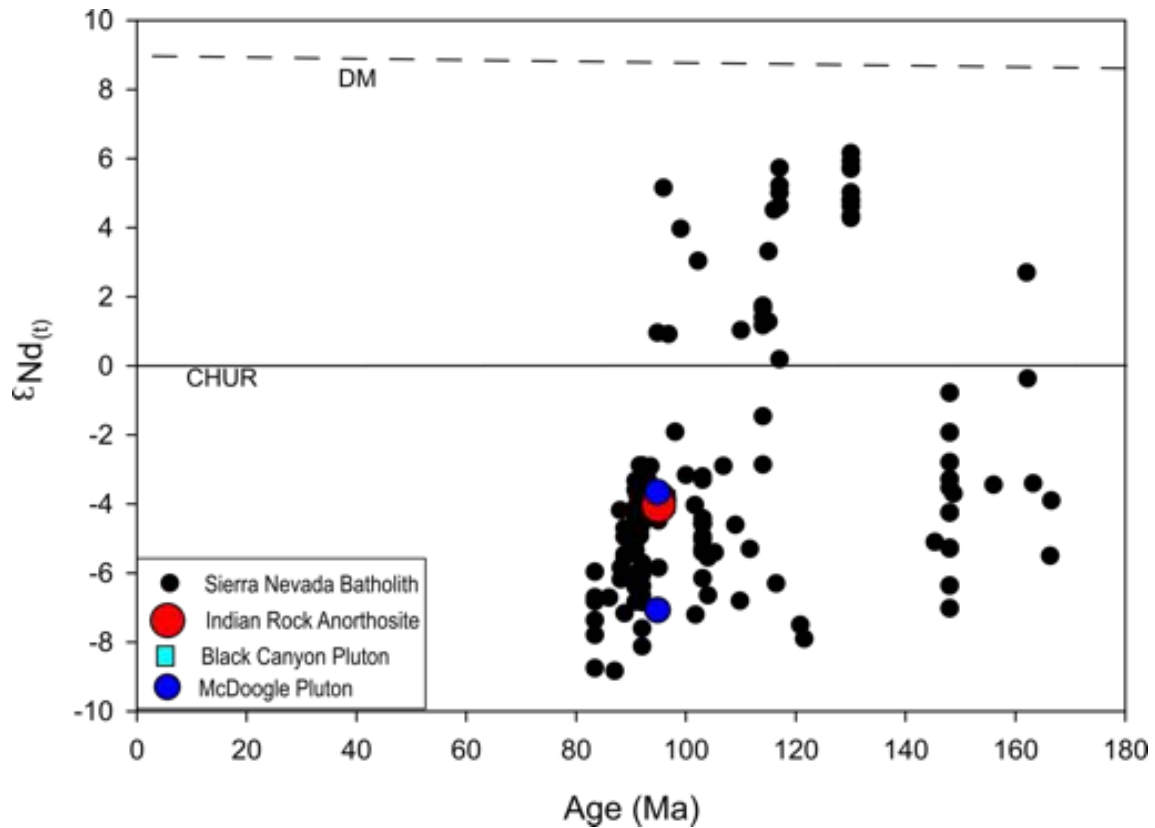


Figure 7.13: $\epsilon_{Nd(t)}$ vs crystallisation ages (Ma) plot. The crystallisation ages and $\epsilon_{Nd(t)}$ data are from Kistler *et al.* (1986); Saleeby *et al.* (1987, 2007); Clemens-Knott (1992); Coleman *et al.* (1992); Olson (1992); Bradford (1995); Cousens (1996); Sisson *et al.* (1996); Truschel (1996); Ratajeski *et al.* (2001); Kylander-Clark *et al.* (2005); Glazner *et al.* (2008); Gray *et al.* (2008); Cecil *et al.* (2012); Nelson *et al.* (2013); and Chapman and Ducea (2019).

The analysed samples display a more evolved zircon Lu/Hf values relative to the mantle values. Hafnium in the Earth is primarily sequestered in zircons that are abundant in continental crust and high-energy terrigenous sediments (Ingle *et al.*, 2003). Therefore, Hf isotopes are a useful indicator of continental contamination and sediment. The IRA samples show a vertical mixing array which comes down from DM to more evolved mantle signatures, referred to as isotopic pull-down. There are many possibilities which can contribute to the vertical mixing array such as assimilation during crustal thickening, under thrusting of isotopically evolved lower crust, inboard arc migration into evolved crust, or crustal mixing. In this case, the samples show vertical array from positive to negative epsilon Hf values indicating assimilation of the lower crustal material during crustal thickening. Convergent margins such as Sierra Nevada Batholith show complex isotopic signatures (Attia *et al.*, 2020) (Figure 7.14), which shows interactions of juvenile magma (with positive epsilon Hf values) in the system with old and young crustal material (with negative epsilon Hf values) leading to more evolved signatures in the rocks. Crustal thickening has similar effect of forming more evolved isotopic signatures by introducing more crustal or higher crustal material to the evolving magma in the system. Mantle-derived arc magmas rise in the overriding crust and when they cool, they differentiate into felsic liquids and

complementary mafic cumulates and restates (Chin *et al.*, 2012). Through alphaMELTS modelling, a large number of mafic components are formed in the system. Today there is not much of the mafic components in the batholith, and this may be due to the process of root removal proposed by Zandt *et al.* (2004). Root removal is the process where the batholith loses its ultramafic root (Zandt *et al.*, 2004). Many researchers have proposed a similar process in the massif-type anorthosite formation (Ashwal, 1993; Scoates, 2000; McLelland and Selleck, 2010). The fractionation of mafic minerals like olivine and pyroxene sink back into the mantle and the plagioclase crystals float in the increasingly dense residual magmas (Ashwal, 1993). Considering the thickness of the batholith, the lack of garnetiferous samples seems strange. The garnet-bearing samples are broken down by the crustal thinning of the eastern part of the SNB or by the heat from the magmatism (Wernicke *et al.*, 1996).

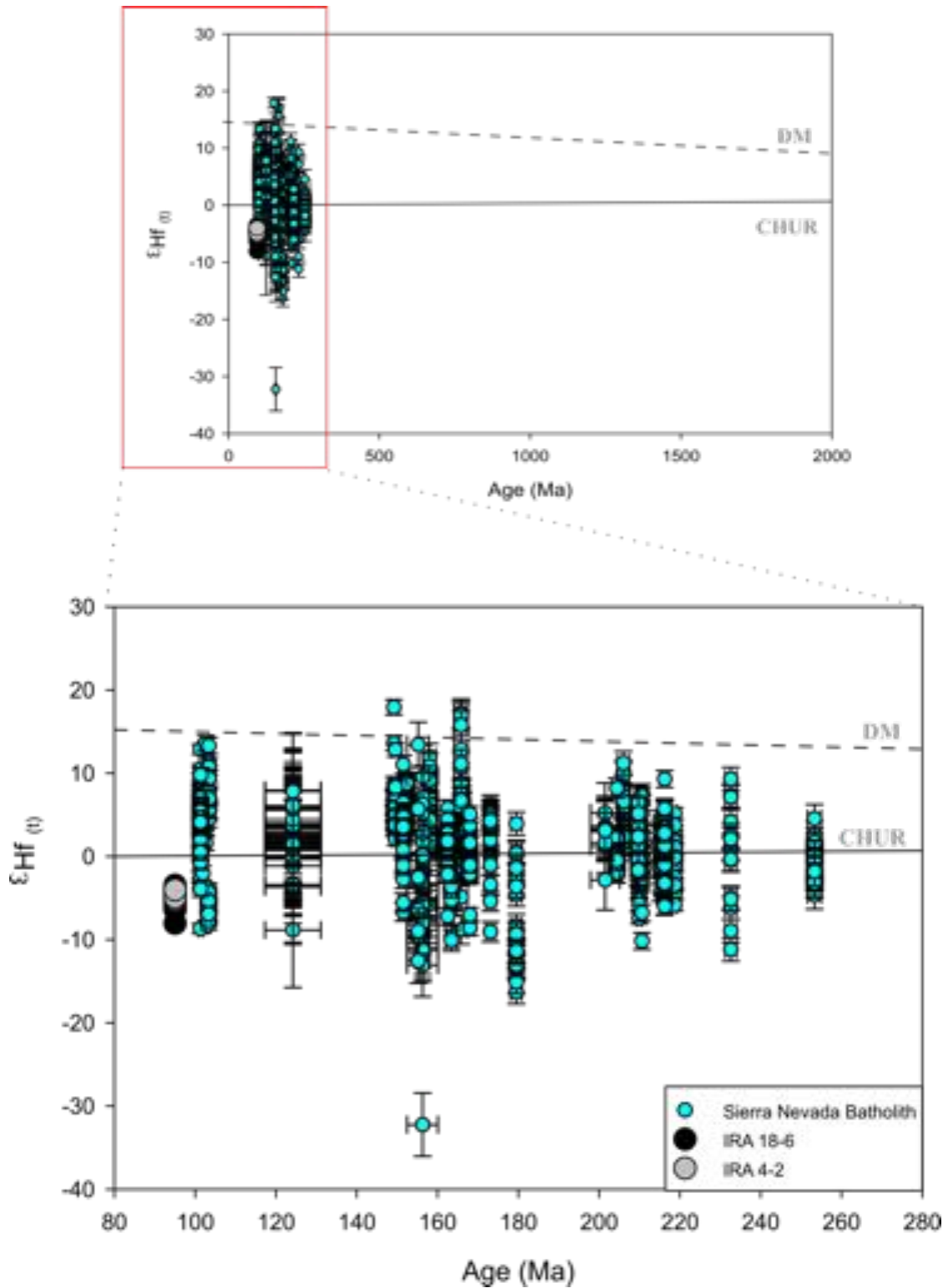


Figure 7.14: Zircon Lu-Hf model age of the eastern Sierra Nevada Batholith and the IRA. The eastern Sierra Nevada Batholith data was extracted from Attia et al. (2020). The graph of $\epsilon_{\text{Hf}(t)}$ vs crystallisation age (Ma). The evolution-depleted mantle (DM) is calculated back from present ratios of $^{176}\text{Lu}/^{177}\text{Hf} = 0.0524$ and $^{176}\text{Hf}/^{177}\text{Hf} = 0.28326$ (Workman and Hart, 2005). The second figure is the zoom out area of the red box.

7.5 Change in Subduction Zone Properties from Proterozoic to Phanerozoic Eon

This work has shown that the IRA is a unique type of anorthosite which has both the characteristics of massif-type anorthosite (plutonic nature) and cognate anorthosite from continental arc that formed in the new Phanerozoic arc environment. Can the IRA tell us about changes in the nature of subduction zones in the Phanerozoic compared to the Proterozoic? Previously many proxies have been used for subduction-related and intraplate settings, such as metamorphic conditions and magma compositions (Hawkesworth *et al.*, 2020). Plate tectonics started at some time between early Hadean and 700 Ma (Hawkesworth and Brown, 2018). According to the study by Lambert (1980), heat production was 3-6 times higher in the Archean compared to the present time. Brown (2007, 2014) has compiled variations of metamorphic conditions for the most part in the context of plate tectonics, which focused on the relatively high-grade rocks linked to the peak of metamorphism (Figure 7.15). The rocks are subdivided into a range of estimated apparent temperature and pressure values, which are granulites and ultrahigh-temperature (UHT) granulites, medium-temperature eclogite high-pressure (EHP) granulites, and high-pressure to ultrahigh-pressure (HP-UHP) metamorphic rocks. Alpine facies series constitute of blueschist to eclogite facies metamorphic rocks, which form during gravity-driven subduction for active compression at low thermal gradients of 5- 10 °C/km, and the Buchan facies series constitute of amphibolite to granulite facies metamorphic rocks form during buoyancy-driven rifting for an active extension at high thermal gradients of 30- 60 °C/km (Figure 7.16 a) (Zheng and Chen, 2017). Blueschists, which are HP-UHP, have been observed back to 700 Ma, they have been observed in the recent geological time as part of the bimodal suites, and rocks older than 700 Ma also appear to record bimodal associations which are displaced to higher T/P (Figure 7.16 b) (Hawkesworth *et al.*, 2020). Figure 7.16 b illustrates how the tectonic mechanism for the generation of different metamorphic facies series along plate boundaries have changed through Earth's history. Buchan facies series occurred from Archean to Cenozoic, whereas Alpine facies series mostly appeared since Neoproterozoic, and the Barrovian facies series has appeared since the Archean (Zheng and Zhao, 2020). The change in the facies series over time shows important spatial and temporal relationships between tectonic regimes along convergent plate boundaries.

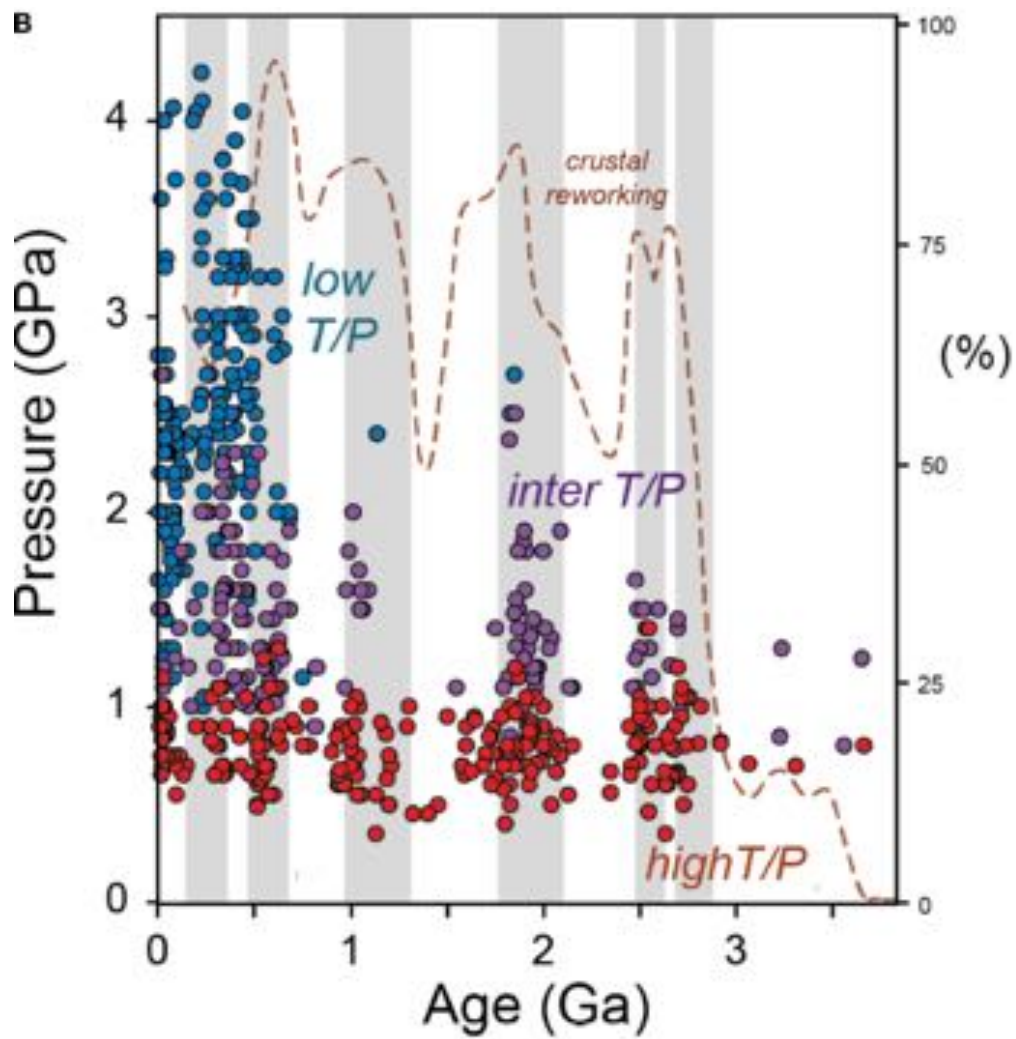


Figure 7.15: The graph showing maximum pressure in metamorphic terranes through the Earth's history, crustal reworking curve and the ages of supercontinents shown by grey areas (Hawkesworth et al., 2020). The red dots are high T/P, purple dots are inter T/P and the blue dots are low T/P.

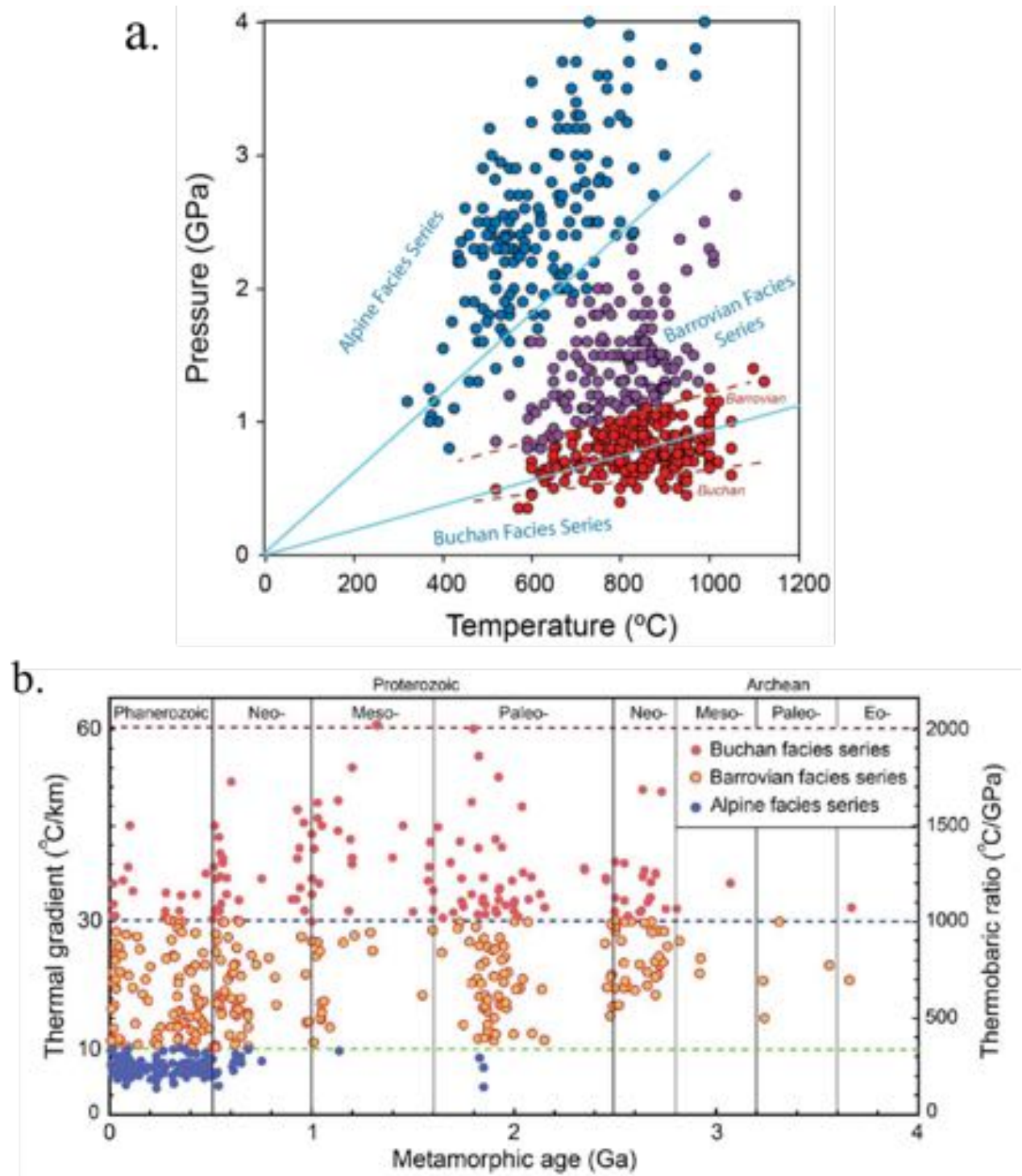


Figure 7.16: Variations in peak metamorphic conditions through time. a. Shows difference in temperature and pressure from the peak metamorphism estimates (Hawkesworth et al., 2020). b, Metamorphic age with thermal gradient grouped by metamorphic facies series (Zheng and Zhao, 2020). The facies series are from Zheng and Zhao (2020), peak metamorphism conditions estimates are from Brown (2007). The dashed lines show boundaries between the different facies series.

Present-day metamorphic belts are a distinctive feature of destructive plate margins that results in a bimodal distribution of apparent thermal gradients, which is attributed to metamorphism near low T/P (HP-UHP) and away from the high T/P subduction zones (Hawkesworth *et al.*, 2020). The modelling results from this research have shown us that the effect of high pressures increases the plagioclase An content during crystallisation. The IRA has high An content measured to be An₇₂ to An₈₉, higher than the typical Proterozoic massif-type anorthosite. At high pressures (>10 kbar), the system produces a small amount of plagioclase with high An content by partial melting of a mafic parental source (Simmons and Hanson, 1978), and due to instability of plagioclase at higher depths, which results in an isolated small scale body of massif-type anorthosite like the IRA. Also, the presence of water at depth increases the An content observed. The present-day (Phanerozoic) subduction zones are colder relative to the Archean and Proterozoic subduction zones, as illustrated in Figure 7.16 a. In a subduction environment, the temperature increase with increasing depth. At high temperature, both a dry melt and hydrous melt crystallise plagioclase with high An content (Figure 6.16). A hydrous melt shows higher plagioclase An content at high temperature which correlates to higher depth, and according to Figure 7.17, a colder subduction zone is capable of transporting water to higher depth compared to hotter subduction, which is crucial in the formation of the calcic rich plagioclase found in the IRA. The IRA provides evidence that the conditions in subduction zones may have changed, especially looking at the Proterozoic subduction zones relative to the Phanerozoic. The Proterozoic subduction zones were hotter, low pressure and drier when compared to the Phanerozoic subduction zones, which are colder, high pressure and wet. The Proterozoic massif-type anorthosites are large bodies, which have plagioclase with large Eu anomaly and lower An content. The large bodies of Proterozoic massif-type anorthosite formed at low pressure, reducing oxygen fugacity with low water content, which allowed for vast amounts of plagioclase with low An content and large Eu anomaly to accumulate. Low pressure environments allow for vast amounts of plagioclase crystallisation due to the plagioclase stability and Al depletion in the melt. Low water content systems form plagioclase with large Eu anomaly and low An content. Therefore, a high pressure, low water environment forms large amounts of plagioclase with large Eu anomalies and low An content. The IRA has a small Eu anomaly, which is a function of oxygen fugacity and temperature. The small, isolated body of anorthosite formed in a high pressure, relative high temperature, oxidising oxygen fugacity environment with high water content, which formed a small amount of plagioclase with high An content and a shallow Eu anomaly. Therefore, the IRA shows that Phanerozoic subduction zones are high pressure, high temperature and capable of transporting water to deeper levels due to the colder subducting slab, and the oxygen fugacity is sufficiently oxidised.

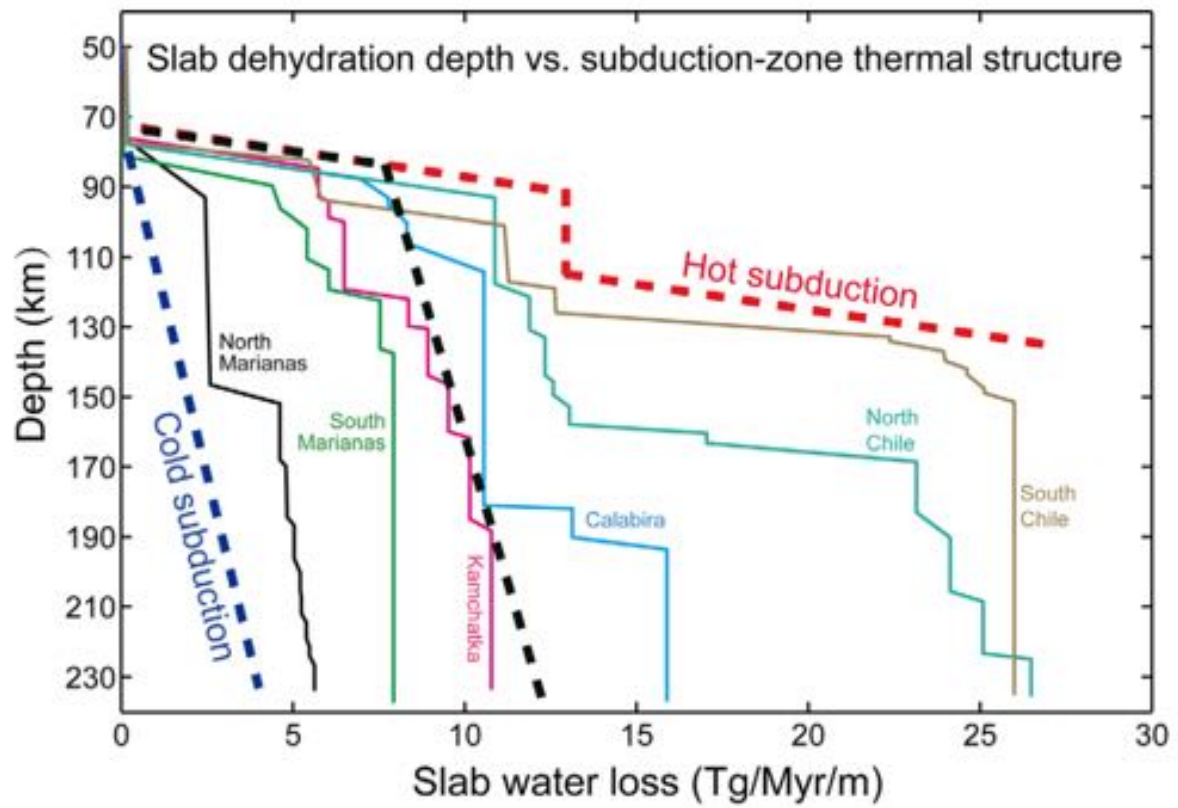


Figure 7.17: Different rates of water loss in oceanic subduction zones as a function of depth. Cold subduction zones lose little water even at 230 km depth (blue-dashed line), while the hot subductions lose water at a faster rate (red-dashed line) (Zheng et al., 2016).

8. Conclusions, Recommendations, and Future work

Massif-type anorthosites are intrusive rocks that are restricted to the Proterozoic Eon. The IRA forms one of the unusual cases where we see one of the rare or never been documented anorthosite body having characteristics of both the massif-type anorthosite and the cognate anorthosite in Phanerozoic continental arc. This rock was documented and partially described by Moore (1963), but due to the lack of information, it was inconclusive to state whether it is a fragment of an older Proterozoic crust or a cognate anorthosite inclusions in continental arcs.

1. The zircon U-Pb data of sample IRA 4-2 and IRA 18-6 from the IRA present ages of 94.01 ± 0.05 and 95.29 ± 0.06 Ma respectively. The zircon U-Pb ages confirm that the IRA is coeval with the SNB. The major and trace element data, accompanied by the radiogenic isotopes show fractionation trends, which links the McDoogle Pluton, IRA and the Black Canyon Pluton. The similarities in isotopic geochemistry ($^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd} values) with its immediate pluton confirm that the IRA is coeval with the surrounding granodiorites of the Black Canyon and McDoogle Pluton and the SNB in general.
2. The compositional trends (major elements) in the plutons are consistent with the role of fractional crystallisation; especially seeing compatible elements such as TiO_2 %, Al_2O_3 %, Fe_2O_3 %, MgO %, CaO %, and Sr generally decrease as SiO_2 % increases. The low content in MgO %, Ni (ppm), Cr (ppm), and Co (ppm) indicate plagioclase crystallisation and accumulation after removal of olivine and clinopyroxene, while the negative Nb anomalies suggest that the magma was derived either from depleted mantle or within subduction zone contaminated with old continent crust.
3. The zoned plagioclase with a high calcic composition ($\sim\text{An}_{84}$) is due to water-rich mafic parental magma mixing with felsic crustal material. The low ϵ_{Nd} values (~-4.41) and high $^{87}\text{Sr}/^{86}\text{Sr}$ (~ 0.7069) of bulk rocks indicate assimilation of the continental crust in a mantle-derived magma. The minor variation in isotopic compositional signatures is due to different interactions of the mafic melt with old and young crustal rocks.
4. The addition of water to the system changes the phases and the sequence of crystallisation and most noticeably, it suppresses the amount of plagioclase in the system and increases the plagioclase An content.
5. A probable parental magma of the IRA is a hydrous high-Al tholeiitic basalt in a high pressure, and high temperature environment, where a small amount of plagioclase with high An content can accumulate. The IRA closely relates to cognate anorthosite that occurs in continental arcs, but it also shows some characteristics of a typical massif-type anorthosite. It can be said that it

is unique in a sense that it can be identified by high An content and shallow positive Eu anomaly.

6. The change in the condition of subduction zones from Proterozoic to Phanerozoic Eon is marked by an increase in pressure, water content, and temperatures. The IRA can be used as a proxy for understanding the effects of high pressures and increased water content in subduction zones on the formation of massif-type anorthosites. The IRA shows that the new Phanerozoic subduction zones are wet, high pressure and relatively high temperatures compared to the Proterozoic subduction zones.
7. The young Phanerozoic subduction zones are hydrated, and they occur at high pressure and high temperature, under oxidising conditions. These conditions reduce the production of plagioclase in the differentiating magma derived from the mantle wedge and increases the An content of the plagioclase. This explains why so few discrete bodies of anorthosite are found in Phanerozoic arcs and where they are found as discrete bodies (e.g. the Indian Rock Anorthosite) or as cognate inclusions in arc volcanics, the An content ranges from 70 – 100 %. These results suggest that fundamental changes in continental arc dynamics between the Proterozoic and Phanerozoic may be responsible for the temporal restriction of Proterozoic anorthosites.

Recommendation and future work

To further this study, more field observations need to be documented to understand better the interactions or relationship of this body with the host granodiorites and the textural change within the body for a better understanding on the emplacement of the IRA in the Sierra Nevada. More samples need to be analysed for a good representation of the larger southeastern Sierra Nevada Batholith overview. *In-situ* samples need to be analysed to understand the spatial variation within the anorthosite body. Zircon *in-situ* oxygen isotopes will better explain the magmatic reservoir and the magmatic signature of the parental magma.

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Appendices

Appendix A: Petrology

Table A1: Field and Petrological description of the samples.

Sample ID	Lithology Field name	Field description	Motivation for sampling	Lithology Petrographic name	TS description
IRA08/05	Anorthosite	Boulder of anorthosite with heterogeneous texture. Patches of cpx occur as oikocrysts with plagioclase inclusions.	Thin section and geochronology	Anorthosite	Zoned plagioclase with cpx occurring in the fractures, The cpx grains is altered to chlorite and amphibole (hornblende with simple twinning). The cpx in the fractures of plag grains is not altered compared to the cumulates which are altered.
IRA06/03	Tonalite	mafic tonalite with biotite and minor k-spar. The grain size are between 3-5 mm and very little qtz and k-spar grains. (in places it almost looks like gabbro)	Thin section and geochronology	Granodiorite	Zoned plagioclase with an euhedral shape. Granophyric texture with qtz inclusions. The biotite has altered to chlorite (chloritization). Cpx is altered to hornblende. The biotite has been altered to chlorite.

IRA07/04	Granite	white and medium grained granite, possibly alaskite cross cutting the rock sub-horizontal.	Thin section	Alaskite	Interlocking texture, zoned plagioclase, granophyric texture (quartz inclusion), hornblende having inclusion of quartz, k-spar(microcline) having simple twinning. The biotite altering to chlorite. The cpx occur as inclusions in k-spar grains.
IRA04/02	Anorthosite	Pink, purple anorthosite with cpx, magnetite and sulphides.	Thin section and geochronology	Anorthosite	The plagioclase grains are zoned with the inclusions of cpx in cracks/fractures found on the plagioclase grains, randomly orientated. The cpx grains are more concentrated in the cores of the zoned plagioclase, the rims have less cpx. The cpx grains have altered to chlorite and a hornblende, the cpx can be observed in the chlorite (remnant/relic texture). The Microcline occurs between plagioclase grains.
IRA03/01		Greyish-green rock containing plagioclase, cpx and chlorite		Anorthosite	Zoned plagioclase with cpx occurring in the cracks of the plagioclase. The k-spar has simple twinning. The cpx is altered to hornblende and chlorite.
		Enclaves of mafic material. Maybe mica fish in various degrees. Close to the			

		kmg and km unit. Looks like enclaves of the km unit.			
IRA18/01	Mc Doggle Pluton (granodiorite)	Grey with dark (hornblende) minerals occurring as oikicrysts with quartz (granodiorite) . The grain size between 0.2-0.5 cm. The rock contains mafic enclaves (L=8-30 cm, W=2-8cm) enlogated in a NW-SE trend. The rock is foliated (NW-SE) with the mafic enclaves are more deformed than the host, which shows more resistance to the force.	In contact with the Km unit (mafic pluton) hosting the anorthosite	Granodiorite	Zoned plagioclase with an euhedral shape. Granophyric texture with qtz inclusions. The biotite has altered the chlorite. Cpx is altered to an amphibole (hornblende). The biotite has been altered to chlorite.

IRA18/02	Mc Gann pluton (granodiorite)	Dark coloured medium grained (0.2-0.5 cm) rock with biotite, hornblende and cpx.	In contact with the Km unit (mafic pluton) hosting the anorthosite . Corresponds to IRA17/6	Granodiorite	The biotite is chloritized and occurs as cumulates and interstitial minerals. The k-spar (simple twinning) (0.3-0.5 mm) is elongated and contains inclusions of qtz. Chlrite oikocrysts have inclusions of irregular shaped qtz grains. The hornblende grains are cumulates and occur with oxides.
IRA18/03	Felsic dyke (Alaskite)	The felsic dyke has grain size of 0.1-0.2 cm and the granodiorite is more coarse grained near the contact with the dyke. The granodiorite is from the Km unit (mafic pluton).	The dyke cutting the Km (mafic pluton) which is in contact with the anorthosite	Alaskite	Interlocking texture, zoned plagioclase, granophyric texture (quartz inclusion), hornblende having inclusion of quartz, microcline having simple twinning. The biotite altering to chlorite. The cpx occur as inclusions in k-spar grains.

IRA18/4a	Anorthosite with the mafic interstitial vein	The anorthosite is grey and has a grain size of 0.2-0.4 cm. The mafics (pyroxene) has a poikilitic texture with plag occurring as inclusions. The mafics occur as intercumulus and cumulus showing two stage crystallization. On the weathered surface the mafics are altered (weathered) and on the fresh surface they have a green colour. The pyroxene (oikocrysts) are 2 cm with inclusions of plag.	TS of the mafic vein and texture	Anorthosite	The plagioclase grains are zoned and occur as cumulates with annealing texture and inclusion in the cpx oikocrysts. The cpx occurs in the cracks of the plagioclase grains. The oxides occur as interstitial, between the plagioclase grains.
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IRA18/4b	Mafic interstitial vein in Anorthosite	Anorthosite with a mafic vein of cpx	Thin section of the mafic vein and texture	Anorthosite	The cpx vein has inclusions of euhedral plagioclase
IRA18/5	Anorthosite	Large crystals of plagioclase grains with 'blobs' of cpx	For geochronology	Anorthosite	The mafic (cpx) occurs as interstitial (oikocrysts) with inclusions of euhedral plagioclase grains. The cpx has been chloritized. An opx grain occurring as an inclusion in the chlorite? The oxides occur as interstitial. K-spar with simple twinning occurring in the chlorite.
IRA18/6	Anorthosite	Grey anorthosite with "mottled" texture made up of cps	For geochronology	Anorthosite	The plagioclase grains are zoned and occurs as cumulates with annealing texture and inclusion in the cpx oikocrysts. The cpx occurs in the cracks of the plagioclase grains. The oxides occur as interstitial, between the plagioclase grains.
IRA18/7	Anorthosite	Grey anorthosite with cpx, magnetite and sulphides.	For geochronology	Anorthosite	Microcline occurs as cumulate mineral with plagioclase grains. the cpx occurs as an interstitial mineral cutting the plagioclase grains. Chlorite occurring between plag having a more cumulate texture and a relic texture of cpx. The hornblende seems to be formed from cpx. The oxides occur with the cpx and chlorite. Qtz grains occur with the plagioclase.

Table A.2: Point count data and mineral modal abundance of Indian Rock Anorthosite.

Modal proportion calculation																			
	Rock name	Anorthosite	%		Anorthosite	%		Anorthosite	%		Anorthosite	%		McDoogle Pluton (granodiorite)	%		McDoogle Pluton (granodiorite)		
	Sample ID	IRA 3-1			IRA 4-2			IRA 4-2B			IRA 8-5			IRA 18-5			IRA 18-7	%	
Minerals	Plagioclase	204	85.00	91	189	78.75	86	205	85	90	194	81	96	190	79	97	181	77	90
	Clinopyroxene	20	8.33	9	31	12.92	14	23	10	10	8	3	4	6	3	3	30	13	10
	Quartz	0	0.00		0	0.00		0	0		0	0		0	0		1	0	
	K-feldspar	4	1.67		5	2.08		0	0		3	1.25		0	0		5	2	
	Hornblende	0	0.00		0	0.00		0	0		0	0		3	1		3	1	
	Chlorite	0	0.00		9	3.75		3	1.25		0	0		4	2		4	2	
	Opaque	3	1.25		6	2.50		9	3.75		2	1		1	0		2	1	
	Blank spaces	9	3.75		0	0.00		0	0		33	14		36	15		9	4	
	Total	240	100	100	240	100	100	240	100	100	240	100	100	240	100	100	235	100	100

Table A2: Point count data and mineral modal abundance of granodiorites of the Black Canyon Pluton and McDoogle Pluton.

		Black Canyon Pluton (granodiorite)		Felsic Dyke		Felsic Dyke	
	Sample ID	IRA 6-3	%	IRA 18-1	%	IRA 18-2	%
Minerals	Plagioclase	103	43	132	55	120	50
	Clinopyroxene	36	15	29	12	24	10
	Quartz	84	35	48	20	48	20
	K-feldspar	12	5	24	10	24	10

	Hornblende	2	1	2	1	14	6
	Biotite	2	1	5	2	10	4
	Total	240	100	240	100	240	100

Table A3: Point count data and mineral modal abundance of alaskite dykes

	Sample ID	IRA 7-4	%	IRA 18-3	%
Minerals	Plagioclase	24	10	36	15
	Clinopyroxene	10	4	12	5
	Quartz	144	60	168	70
	K-feldspar	2	1	5	2
	Hornblende	2	1	5	2
	Biotite	5	2	7	3
	Opaque	5	2	3	2
	Blank spaces	48	20	5	3
	Total	240	100	241	100

Appendix B: Major and Trace Element data

Trace Elements Systematics methodology

Whole-rock trace element data were obtained using a Thermo-Fisher iCAP RQ ICPMS instrument. To obtain trace elements, 50 mg of every sample was dissolved using microwave digestion in 6 mL of 2:1 HF: HNO₃ for 55 minutes at 190 °C. After digestion, the solutions were rinsed with 10% nitric acid and then dried down at 70 °C on a hot plate. 2 mL of nitric acid was added and dried at 60 °C for 24 hours, before increasing the temperature 70 °C before removing them from the hotplate. Finally, 300 µL of nitric acid was added to the solutions, and they were capped for dilution with an internal standard that contains Rh,

Re, In, and Bi in a 5% HNO₃ solution. The solutions were analysed using the Thermo-Fisher iCAP RQ ICPMS at EarthLab, University of the Witwatersrand. Table B1 shows the standard reference values obtained during the ICPMS analyses.

Table B1: Standards used in ICPMS trace element analyses BHVO2 and BCR2.

Elements	BHVO2	BCR2	BHVO2	BCR2
Ce	38.8	55.9	50.7525	36.817
Co	43.8	37.1	36.782	45.7795
Cr	283	15.1	16.345	288.2405
Cs	0	0.93	1.1175	0.116
Eu	2.01	1.93	2.006	1.911
Hf	4.2	4.8	4.8525	4.2285
La	14.8	24.5	24.8265	15.4035
Lu	0.275	0.494	0.4745	0.2575
Nd	24.6	28.9	27.646	23.459
Rb	0	48.5	48.3495	10.223
Sc	31.2	33	32.588	30.462
Sm	6	6.6	6.417	5.698
Ta	1.04	0.7	0.7315	1.0075
Tb	0.88	1.05	0.974	0.84
Th	1.06	5.9	5.6385	1.152
Yb	1.89	3.3	3.3065	1.958

Major Elements Systematics methodology

The whole-rock major element data were obtained using the PANalytical PW2404 spectrometer XRF machine. For analyses, the losses of ignition of samples (LOI) were determined by igniting 1 g of the sample powder in a furnace for 40 minutes at 1000 °C. Then the samples were mixed with lithium tetraborate flux and heated in a furnace before producing fusion beads (Johnson *et al.*, (1999) for more details). Table B2 shows the standard reference values obtained during the XRF analyses.

Table B2: Standards used in XRF major element analyses.

	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	TiO ₂	P ₂ O ₅	Cr ₂ O ₃	NiO	TOTAL	LOI
NIM-P	51.49	4.28	1.28	10.36	0.23	25.42	2.64	0.35	0.09	0.20	0.02	3.55	0.07	99.97	-0.23
NIM-D	38.76	0.32	1.72	13.92	0.22	44.06	0.31	0.04	0.00	0.03	0.01	0.41	0.27	100.07	-0.51
W2	53.15	15.60	1.09	8.86	0.16	6.44	11.00	2.24	0.63	1.08	0.13	0.02	0.01	100.42	0.19
NIM-S	64.04	17.28	0.14	1.13	0.01	0.46	0.66	0.38	15.42	0.05	0.12	0.01	0.00	99.69	0.42
GSP2	67.38	15.02	0.50	4.02	0.05	0.97	2.12	2.81	5.47	0.67	0.29	0.01	0.00	99.31	0.87
BCR2	55.02	13.71	1.40	11.35	0.19	3.64	7.19	3.23	1.80	2.29	0.36	0.02	0.00	100.21	0.10
NIM-N	53.48	16.76	0.90	7.33	0.17	7.55	11.57	2.46	0.24	0.20	0.02	0.01	0.02	100.72	0.04
NIM-G	76.56	12.23	0.20	1.61	0.01	0.05	0.79	3.32	5.07	0.10	0.01	0.01	0.00	99.97	0.73
PCC1	44.55	0.73	0.87	7.02	0.12	45.99	0.60	0.06	0.00	0.01	0.01	0.41	0.33	100.71	5.09
BHVO2	50.67	13.74	1.25	10.12	0.17	7.32	11.46	2.22	0.52	2.76	0.27	0.06	0.02	100.58	-0.48
AGV2	60.40	17.15	0.68	5.52	0.10	1.82	5.27	4.25	2.92	1.05	0.48	0.01	0.00	99.67	1.80
G2	69.80	15.39	0.27	2.17	0.03	0.76	1.95	4.11	4.52	0.49	0.14	0.01	0.00	99.63	0.64
DTS1	40.61	0.27	0.87	7.03	0.12	49.90	0.17	0.07	-0.01	0.02	0.01	0.58	0.31	99.95	-0.02

Appendix C: EPMA data

Table C1: EPMA data of plagioclase from the Indian Rock Anorthosite.

		Comment	Na ₂ O	MgO	Al ₂ O ₃	SiO ₂	K ₂ O	CaO	Cr ₂ O ₃	MnO	FeO	BaO	TiO ₂	Total	An content
Anorthosite	IRA18-4a	core	1.22	0.00	34.43	45.72	0.02	18.20	0.02	0.01	0.35	0.03	0.02	100.01	89
		core	1.34	0.00	34.32	45.63	0.01	18.08	0.01	0.00	0.38	0.00	0.04	99.81	88
		core	1.65	0.00	34.03	47.57	0.04	17.81	0.01	0.02	0.40	0.00	0.01	101.52	85
		core	1.76	0.00	34.07	46.64	0.02	17.82	0.02	0.01	0.28	0.00	0.00	100.61	85
		core	2.03	0.00	34.19	47.56	0.01	17.25	0.00	0.00	0.18	0.00	0.00	101.22	82
		core	2.00	0.00	34.00	47.89	0.02	17.41	0.00	0.01	0.19	0.00	0.00	101.52	83
		core	1.67	0.00	34.36	46.94	0.01	17.60	0.02	0.00	0.23	0.01	0.00	100.84	85
		core	1.76	0.00	32.09	47.92	0.05	17.69	0.00	0.02	0.33	0.02	0.02	99.90	84
	IRA18-5	intermediate	1.91	0.00	32.01	48.33	0.03	17.26	0.00	0.00	0.42	0.04	0.00	99.99	83
		intermediate	1.83	0.00	32.12	47.76	0.04	17.55	0.00	0.02	0.38	0.08	0.02	99.79	84
		intermediate	1.80	0.00	32.03	47.47	0.03	17.56	0.00	0.00	0.37	0.02	0.01	99.29	84
		intermediate	1.76	0.00	32.42	47.45	0.02	17.66	0.00	0.00	0.40	0.02	0.02	99.75	85
		intermediate	1.91	0.00	31.91	48.17	0.03	17.49	0.00	0.01	0.32	0.00	0.05	99.89	83
		intermediate	1.86	0.00	32.10	47.90	0.04	17.43	0.01	0.00	0.30	0.00	0.03	99.68	84
		core	1.89	0.00	31.97	48.34	0.02	17.55	0.03	0.00	0.30	0.00	0.03	100.14	84
		core	1.98	0.00	31.37	48.10	0.03	17.14	0.01	0.00	0.36	0.04	0.07	99.09	83
		core	2.01	0.00	31.50	47.81	0.02	17.29	0.01	0.01	0.37	0.00	0.03	99.05	82
		intermediate	1.31	0.00	31.95	46.67	0.02	18.34	0.00	0.03	0.39	0.01	0.00	98.73	88
		intermediate	1.56	0.00	31.86	47.19	0.01	18.01	0.00	0.01	0.35	0.00	0.06	99.04	86
		intermediate	1.64	0.00	31.75	47.49	0.03	17.73	0.00	0.00	0.33	0.02	0.05	99.05	85
		intermediate	1.79	0.00	31.81	47.96	0.02	17.59	0.00	0.00	0.30	0.02	0.03	99.52	84
		intermediate	1.58	0.00	32.08	47.22	0.02	18.15	0.00	0.00	0.36	0.01	0.05	99.48	86
		intermediate	1.80	0.00	31.96	48.12	0.03	17.51	0.02	0.01	0.40	0.00	0.01	99.88	84
		intermediate	1.36	0.00	32.50	47.14	0.04	18.29	0.00	0.00	0.36	0.01	0.00	99.71	88
		intermediate	1.45	0.00	32.49	47.84	0.07	18.13	0.02	0.00	0.42	0.01	0.03	100.46	87
		intermediate	1.45	0.00	32.26	47.27	0.03	18.14	0.01	0.01	0.40	0.02	0.01	99.58	87
		rim	2.04	0.00	30.99	47.83	0.05	17.12	0.00	0.02	0.34	0.00	0.00	98.38	82
		rim	2.16	0.00	30.91	48.01	0.02	17.28	0.02	0.04	0.28	0.03	0.02	98.79	81
		rim	2.31	0.00	31.52	49.55	0.04	16.81	0.00	0.00	0.34	0.00	0.06	100.64	80
		rim	1.89	0.00	31.11	47.43	0.02	17.53	0.00	0.00	0.36	0.05	0.00	98.40	84
		rim	1.88	0.00	31.71	47.36	0.03	17.42	0.00	0.00	0.29	0.00	0.03	98.71	83
		rim	2.01	0.00	31.05	48.19	0.03	17.28	0.00	0.00	0.36	0.04	0.02	98.99	82
		inclusion in Cpx	2.31	0.00	31.52	49.55	0.04	16.81	0.00	0.00	0.34	0.00	0.06	100.64	80
		inclusion in Cpx	1.91	0.00	32.01	48.33	0.03	17.26	0.00	0.00	0.42	0.04	0.00	99.99	83

	inclusion in Cpx	1.83	0.00	32.12	47.76	0.04	17.55	0.00	0.02	0.38	0.08	0.02	99.79	84
	inclusion in Cpx	1.80	0.00	32.03	47.47	0.03	17.56	0.00	0.00	0.37	0.02	0.01	99.29	84
	inclusion in Cpx	1.76	0.00	32.42	47.45	0.02	17.66	0.00	0.00	0.40	0.02	0.02	99.75	85
	inclusion in Cpx	1.91	0.00	31.91	48.17	0.03	17.49	0.00	0.01	0.32	0.00	0.05	99.89	83
	inclusion in Cpx	1.86	0.00	32.10	47.90	0.04	17.43	0.01	0.00	0.30	0.00	0.03	99.68	84
	inclusion in Cpx	1.89	0.00	31.97	48.34	0.02	17.55	0.03	0.00	0.30	0.00	0.03	100.14	84
	inclusion in Cpx	1.48	0.00	36.07	43.24	0.05	18.20	0.01	0.01	0.64	0.02	0.03	99.72	87
	inclusion in Cpx	1.60	0.00	37.25	43.09	0.04	17.90	0.00	0.01	0.77	0.03	0.00	100.69	86
	inclusion in Cpx	1.40	0.00	37.76	42.79	0.04	18.44	0.00	0.03	0.77	0.00	0.04	101.27	88
	inclusion in Cpx	1.12	0.00	33.09	53.11	0.13	12.62	0.00	0.04	0.42	0.00	0.02	100.54	85
	inclusion in Cpx	1.13	0.00	37.43	48.27	0.04	13.27	0.00	0.01	0.60	0.01	0.01	100.77	86
	rim	1.67	0.00	38.53	47.42	0.04	13.19	0.00	0.04	0.60	0.00	0.04	101.53	81
	rim	1.96	0.00	33.50	48.54	0.24	15.26	0.00	0.01	0.57	0.00	0.00	100.09	80
	rim	2.39	0.00	30.67	50.32	0.06	16.86	0.01	0.02	0.73	0.08	0.00	101.13	79
	inclusion in Cpx	1.15	0.00	34.25	47.74	0.02	16.55	0.00	0.00	0.60	0.10	0.00	100.42	88
	inclusion in Cpx	1.27	0.00	36.33	45.62	0.03	17.44	0.00	0.03	0.70	0.00	0.02	101.45	88
	inclusion in Cpx	1.63	0.00	28.25	51.67	0.05	16.67	0.00	0.00	0.61	0.00	0.04	98.91	85
	inclusion in Cpx	1.48	0.01	34.59	44.95	0.02	16.25	0.00	0.04	0.65	0.07	0.01	98.06	86
	rim	2.39	0.01	33.96	50.39	0.04	14.50	0.01	0.00	0.02	0.38	0.07	101.78	76
	rim	1.50	0.00	35.00	48.30	0.04	15.48	0.01	0.00	0.02	0.38	0.00	100.72	84
	rim	1.73	0.00	34.80	48.62	0.02	15.23	0.02	0.00	0.00	0.38	0.07	100.87	82
	rim	1.76	0.00	34.75	48.65	0.02	15.25	0.01	0.01	0.01	0.32	0.08	100.87	82
	rim	1.97	0.00	34.19	48.10	0.02	14.47	0.01	0.03	0.00	0.34	0.00	99.14	80
	rim	2.07	0.00	33.94	47.97	0.01	14.17	0.01	0.01	0.00	0.33	0.00	98.52	79
	rim	1.47	0.00	34.79	47.03	0.03	15.12	0.03	0.00	0.00	0.40	0.01	98.87	84
	rim	1.56	0.00	34.74	47.72	0.04	15.15	0.04	0.00	0.00	0.45	0.05	99.75	83
	rim	2.79	0.00	33.01	49.28	0.04	13.27	0.02	0.00	0.01	0.36	0.01	98.79	72
	rim	1.36	0.00	34.95	46.03	0.04	15.45	0.02	0.00	0.01	0.49	0.00	98.36	85
	rim	1.41	0.00	34.94	46.37	0.02	15.28	0.01	0.00	0.00	0.38	0.00	98.41	85
	rim	2.47	0.01	33.29	48.77	0.04	13.85	0.03	0.00	0.00	0.38	0.00	98.84	75
	rim	1.31	0.00	34.91	46.96	0.02	15.30	0.02	0.01	0.00	0.40	0.00	98.95	86
JRA18-7	intermediate	1.68	0.00	34.64	47.46	0.03	14.88	0.01	0.00	0.02	0.41	0.00	99.13	82
	intermediate	1.65	0.00	34.56	47.55	0.06	15.08	0.04	0.01	0.04	0.38	0.01	99.37	83
	intermediate	1.36	0.00	34.86	46.98	0.02	15.49	0.02	0.00	0.00	0.40	0.00	99.13	85
	inclusion in Cpx	1.87	0.00	34.65	48.71	0.02	15.06	0.01	0.02	0.00	0.32	0.00	100.66	81
	inclusion in Cpx	2.18	0.04	33.80	49.37	0.06	14.37	0.00	0.00	0.00	0.34	0.02	100.17	78
	inclusion in Cpx	2.44	0.02	33.83	50.19	0.03	14.16	0.00	0.00	0.00	0.30	0.13	101.11	76
	inclusion in Cpx	2.13	0.02	34.08	49.20	0.04	14.62	0.00	0.02	0.00	0.25	0.08	100.45	79

	inclusion in Cpx	1.51	0.00	35.08	48.38	0.15	15.76	0.00	0.00	0.01	0.31	0.00	101.20	84
	inclusion in Cpx	2.35	0.00	33.57	50.29	0.05	14.34	0.01	0.02	0.00	0.40	0.00	101.03	76
	inclusion in Cpx	1.22	0.00	35.19	47.89	0.05	16.01	0.03	0.01	0.02	0.44	0.00	100.86	87
	inclusion in Cpx	1.59	0.00	34.82	48.42	0.03	15.54	0.04	0.00	0.00	0.45	0.04	100.92	83
	inclusion in Cpx	1.99	0.00	33.55	49.12	0.06	14.81	0.02	0.00	0.00	0.36	0.03	99.93	80
	inclusion in Cpx	1.47	0.00	34.95	48.29	0.03	15.81	0.06	0.00	0.00	0.44	0.00	101.06	85
	inclusion in Cpx	2.07	0.01	34.33	49.69	0.03	14.86	0.02	0.00	0.00	0.34	0.04	101.38	79
	inclusion in Cpx	1.74	0.00	34.91	48.74	0.03	15.41	0.00	0.00	0.01	0.27	0.07	101.17	82
	intermediate	1.47	0.02	34.64	47.13	0.03	15.05	0.00	0.00	0.00	0.34	0.05	98.73	84
	intermediate	2.01	0.00	34.30	47.99	0.03	14.29	0.00	0.00	0.00	0.27	0.04	98.93	79
	intermediate	2.81	0.00	32.08	48.68	0.07	16.00	0.01	0.00	0.37	0.05	0.04	100.12	75
	intermediate	2.44	0.00	32.39	48.67	0.13	15.30	0.00	0.03	0.43	0.00	0.00	99.40	77
	intermediate	1.39	0.00	33.43	45.17	0.05	17.90	0.00	0.01	0.46	0.02	0.02	98.45	87
	intermediate	1.23	0.00	34.08	44.89	0.07	18.19	0.00	0.00	0.46	0.07	0.01	99.00	89
	intermediate	1.20	0.00	33.93	44.98	0.02	18.38	0.00	0.02	0.44	0.00	0.03	99.01	89
	intermediate	1.40	0.00	33.63	45.22	0.04	18.03	0.01	0.01	0.46	0.05	0.03	98.88	87
	intermediate	1.37	0.00	33.84	45.14	0.02	17.92	0.00	0.00	0.36	0.04	0.00	98.69	88
	intermediate	1.47	0.00	33.75	45.47	0.04	18.01	0.00	0.02	0.36	0.00	0.01	99.13	87
	intermediate	2.81	0.00	31.88	47.81	0.05	15.84	0.00	0.01	0.32	0.00	0.01	98.71	76
	intermediate	1.74	0.00	33.19	45.33	0.03	17.42	0.01	0.00	0.35	0.02	0.03	98.13	84
	intermediate	2.43	0.00	32.22	46.50	0.01	16.64	0.01	0.00	0.41	0.08	0.04	98.34	79
	intermediate	1.35	0.00	33.53	44.60	0.04	18.02	0.02	0.00	0.43	0.01	0.03	98.02	88
	intermediate	1.39	0.00	33.79	44.86	0.03	18.18	0.00	0.02	0.37	0.02	0.04	98.69	88
	intermediate	1.33	0.00	33.26	44.83	0.03	18.16	0.02	0.00	0.44	0.01	0.00	98.08	88
	intermediate	1.40	0.00	33.16	45.45	0.02	18.04	0.04	0.00	0.41	0.00	0.07	98.59	88
	intermediate	1.34	0.00	33.43	44.77	0.05	18.18	0.01	0.04	0.43	0.01	0.00	98.26	88
	intermediate	2.79	0.00	31.88	47.73	0.03	15.61	0.00	0.01	0.18	0.07	0.03	98.34	75
	intermediate	1.51	0.00	33.81	45.25	0.03	18.08	0.00	0.00	0.40	0.01	0.01	99.11	87
	intermediate	1.12	0.00	36.22	45.48	0.05	17.16	0.01	0.01	0.33	0.00	0.02	100.40	89
IRA3-I	core	1.21	0.00	34.59	46.39	0.00	16.23	0.00	0.01	0.57	0.00	0.03	99.05	88
	core	1.52	0.00	38.39	40.14	0.02	18.28	0.00	0.01	0.51	0.00	0.02	98.88	87

	core	1.51	0.00	35.01	43.56	0.04	19.19	0.00	0.01	0.85	0.00	0.03	100.20	87
	core	1.48	0.00	34.19	44.43	0.02	18.63	0.00	0.02	0.59	0.03	0.05	99.44	87
	core	1.55	0.00	33.61	44.13	0.01	18.58	0.00	0.00	0.57	0.00	0.00	98.44	87
	intermediate	2.62	0.00	28.49	52.47	0.05	16.64	0.00	0.01	0.78	0.00	0.00	101.06	78
	intermediate	1.83	0.00	33.90	49.96	0.06	13.91	0.00	0.01	0.53	0.00	0.04	100.23	80
	intermediate	2.34	0.00	31.33	46.61	0.51	17.85	0.00	0.00	0.57	0.00	0.04	99.26	79
	intermediate	1.63	0.00	34.00	51.03	0.05	13.41	0.00	0.00	0.13	0.00	0.02	100.26	82
	core	1.19	0.00	34.50	50.06	0.04	15.05	0.00	0.00	0.74	0.01	0.01	101.59	87
	core	1.20	0.00	34.89	49.92	0.04	14.05	0.00	0.00	0.81	0.00	0.00	100.91	86
	intermediate	0.95	0.00	23.88	57.25	1.52	12.66	0.00	0.02	4.52	0.07	0.04	100.91	78
	core	1.60	0.00	30.28	50.17	0.02	18.25	0.00	0.01	0.86	0.00	0.06	101.26	86
	intermediate	1.50	0.00	32.68	53.55	0.03	12.54	0.00	0.01	0.68	0.00	0.04	101.02	82
	intermediate	1.47	0.00	32.74	54.00	0.03	12.42	0.00	0.03	0.78	0.00	0.01	101.48	82
	intermediate	2.67	0.00	33.67	48.44	0.12	14.85	0.00	0.00	0.76	0.00	0.07	100.58	75
	intermediate	2.38	0.00	34.44	46.70	0.07	16.27	0.00	0.00	0.53	0.00	0.04	100.43	79

Average (An)	83
n	118

Appendix D: Rb-Sr and Sm-N

Table D1: Age and ϵ_{Nd} data of the Sierra Nevada Batholith with location.

Sample	Age	LATITUDE	LONGITUDE	Ep_Nd(t)	Source
0102-11C	83.4	36.2297	-118.3567	-6.82	Kylander-Clark_et al_2005
0102-13M	83.4	36.7212	-118.3519	-5.96	Kylander-Clark_et al_2005
020517-1C	83.4	36.1822	-117.6326	-7.36	Kylander-Clark_et al_2005
020517-2C	83.4	36.1682	-117.6958	-8.75	Kylander-Clark_et al_2005
020518-2C	83.4	36.2181	-117.902	-7.79	Kylander-Clark_et al_2005
020518-2M	83.4	36.2181	-117.9018	-6.7	Kylander-Clark_et al_2005
Z-63	86	37.8683	-119.4383	-6.71	Kistler_et al_1986
Z-18	87	37.8733	-119.3683	-8.83	Kistler_et al_1986
Z-53	88	37.855	-119.485	-4.18	Kistler_et al_1986
HD02-110	88.1	37.7519	-119.3877	-5.84	Gray_2003
HD02-94	88.1	37.7956	-119.4252	-6.18	Gray_2003
HD01-77A	88.8	37.8411	-119.4481	-7.16	Gray_2003
HD01-80	88.8	37.8388	-119.4509	-5.55	Gray_2003
HD01-84	88.8	37.7427	-119.4039	-4.7	Gray_2003
HD02-109	88.8	37.7049	-119.356	-5.46	Gray_2003
HD02-111	88.8	37.7451	-119.3859	-4.95	Gray_2003
Z-58	90	37.8517	-119.4733	-5.22	Kistler_et al_1986
G22	90.2	38.776	-119.8968	-5.4	Cecil_et al_2012
HD01-10	90.8	37.7452	-119.4659	-6	Gray_2003
HD01-21	90.8	37.8056	-119.47	-6.83	Gray_2003
HD01-30	90.8	37.8211	-119.4692	-5.09	Gray_2003
HD01-39	90.8	37.8116	-119.4855	-4.73	Gray_2003
HD01-40	90.8	37.7354	-119.5147	-4.38	Gray_2003
HD01-48	90.8	37.7486	-119.5713	-3.33	Gray_2003
HD01-6	90.8	37.7612	-119.5394	-4.09	Gray_2003
HD02-107	90.8	37.881	-119.4142	-6.4	Gray_2003
HD02-112	90.8	38.0514	-119.5157	-5.34	Gray_2003
HD02-119	90.8	37.7215	-119.5689	-3.59	Gray_2003
HD02-97	90.8	37.9728	-119.2963	-4.78	Gray_2003
HD01-43	91.1	37.746	-119.5324	-3.74	Gray_2003
15A	91.6	37.2336	-118.6589	-6.53	Coleman_et al_1992
20A	91.6	37.1719	-118.5642	-4.6	Coleman_et al_1992
6A	91.6	37.2103	-118.6442	-3.81	Coleman_et al_1992
76A	91.6	37.1478	-118.5808	-4.32	Coleman_et al_1992
80C	91.6	37.1428	-118.5833	-4.7	Coleman_et al_1992
80fG	91.6	37.1461	-118.5833	-4.18	Coleman_et al_1992

80F-L	91.6	37.1461	-118.5833	-3.84	Coleman_et al_1992
83C	91.6	37.1439	-118.5886	-4.32	Coleman_et al_1992
87A	91.6	37.2306	-118.6519	-4.31	Coleman_et al_1992
SAB-1b	91.6	37.1867	-118.6208	-4.73	Coleman_et al_1992
UB-1a	91.6	36.0686	-118.4069	-2.88	Coleman_et al_1992
UB-1b	91.6	36.1039	-118.4214	-4.92	Coleman_et al_1992
AD15-93	92	36.9561	-118.3433	-6.38	Bradford_1995
AD18-93	92	36.9678	-118.3422	-8.12	Bradford_1995
AD1a-93	92	36.9581	-118.3417	-5.91	Bradford_1995
AD21-93	92	36.9619	-118.3456	-5.7	Bradford_1995
AD22-93	92	36.965	-118.35	-6.04	Bradford_1995
AG2-93	92	36.9547	-118.345	-6.85	Bradford_1995
AQMI-93	92	36.9464	-118.3486	-6.65	Bradford_1995
BS1-93	92	36.9475	-118.3542	-2.88	Bradford_1995
LS4-93	92	36.9581	-118.3417	-7.6	Bradford_1995
TQM2-93	92	36.9836	-118.3281	-5.95	Bradford_1995
86S40a	92.1	36.7767	-118.36	-4.54	Sisson_et al_1996
86S45	92.1	36.7833	-118.3833	-4.4	Sisson_et al_1996
86S49	92.1	36.785	-118.3683	-4.41	Sisson_et al_1996
HD01-55	93	37.8625	-119.2674	-3.35	Gray_2003
HD01-71	93	37.8469	-119.4878	-3.28	Gray_2003
HD01-35	93.5	37.9074	-119.2653	-2.91	Gray_2003
5.00	94.8	34.9857	-118.7291	0.96	Saleeby_et al_2007
K-37-67	95	37.83	-119.7417	-5.85	Kistler_et al_1986
K-5-64	95	37.8767	-119.6717	-4.48	Kistler_et al_1986
5	95.9	34.9266	-118.8651	5.15	Saleeby_et al_2007
1	96.8	34.9043	-118.9178	0.92	Saleeby_et al_2007
TC-42	98	35.2283	-118.4961	-1.91	Saleeby_et al_1987
2	99	34.9256	-118.9279	3.97	Saleeby_et al_2007
CM-9	100	35.0408	-118.5683	-3.16	Saleeby_et al_1987
IP0203	101.6	36.7603	-118.2618	-4.03	Kylander-Clark_et al_2005
G20	101.7	38.7048	-120.0899	-7.20	Cecil_et al_2012
6b	102.2	34.9204	-118.8269	3.04	Saleeby_et al_2007
Rs-3	103	37.7284	-119.63	-3.20	Nelson_et al_2013
RS-6	103	37.7284	-119.63	-3.30	Nelson_et al_2013
YOS-1	103	37.7258	-119.6244	-4.55	Ratajeski_et al_2001
YOS-1	103	37.7258	-119.6244	-4.55	Ratajeski_et al_2001
YOS-103a	103	37.7284	-119.63	-6.15	Ratajeski_et al_2001
YOS-103a	103	37.7284	-119.63	-6.15	Ratajeski_et al_2001
YOS-103b	103	37.7284	-119.63	-5.03	Ratajeski_et al_2001
YOS-103b	103	37.7284	-119.63	-5.03	Ratajeski_et al_2001
YOS-104	103	37.7284	-119.63	-4.94	Ratajeski_et al_2001
YOS-104	103	37.7284	-119.63	-4.94	Ratajeski_et al_2001
YOS-105b	103	37.7248	-119.6675	-4.42	Ratajeski_et al_2001
YOS-105b	103	37.7248	-119.6675	-4.42	Ratajeski_et al_2001

YOS-18a	103	37.7283	-119.6274	-4.57	Ratajeski_et al_2001
YOS-18a	103	37.7283	-119.6274	-4.57	Ratajeski_et al_2001
YOS-18c	103	37.7283	-119.6274	-5.30	Ratajeski_et al_2001
YOS-18c	103	37.7283	-119.6274	-5.30	Ratajeski_et al_2001
YOS-55a	103	37.723	-119.6663	-5.39	Ratajeski_et al_2001
YOS-55a	103	37.723	-119.6663	-5.39	Ratajeski_et al_2001
YOS-61	103	37.7127	-119.6663	-4.58	Ratajeski_et al_2001
YOS-61	103	37.7127	-119.6663	-4.58	Ratajeski_et al_2001
DPKg	104	37.62	-119.09	-6.64	Cousens_1996
RCKg	104	37.5933	-119.06	-5.55	Cousens_1996
G19	105.2	38.6457	-120.1343	-5.4	Cecil_et al_2012
G23	106.8	38.7648	-119.8484	-2.9	Cecil_et al_2012
G06	109	39.3334	-120.2906	-4.6	Cecil_et al_2012
G21	109.8	38.6945	-119.9956	-6.8	Cecil_et al_2012
PC-35	110	34.9194	-118.7961	1.03	Saleeby_et al_1987
G10	111.6	39.3276	-120.3901	-5.3	Cecil_et al_2012
FG	114	37.0505	-119.7738	1.39	Truschel_1996
FG	114	37.1472	-119.7318	1.74	Truschel_1996
FG	114	37.3187	-119.6612	-1.46	Truschel_1996
FG	114	37.3647	-119.633	-2.86	Truschel_1996
FG	114	37.2062	-119.8767	1.66	Truschel_1996
FG	114	37.1918	-119.7223	1.17	Truschel_1996
WKB	115	35.9489	-118.8806	3.31	Clemens-Knott_1992
CM-630	115	35.125	-118.7228	1.28	Saleeby_et al_1987
WKB	116	36.4783	-119.0194	4.51	Clemens-Knott_1992
G09	116.4	39.3436	-120.3385	-6.3	Cecil_et al_2012
WKB	117	36.4524	-119.1518	5.01	Clemens-Knott_1992
WKB	117	36.4393	-119.1469	5.73	Clemens-Knott_1992
WKB	117	36.4283	-119.2034	5.22	Clemens-Knott_1992
WKB	117	36.4518	-119.1879	4.62	Clemens-Knott_1992
WR-643	117	35.0244	-118.7067	0.19	Saleeby_et al_1987
G11	120.8	39.3123	-120.4946	-7.5	Cecil_et al_2012
G18	121.5	38.561	-120.2663	-7.9	Cecil_et al_2012
WKB	130	36.4977	-119.1066	5.7	Clemens-Knott_1992
WKB	130	36.5244	-119.0998	5.93	Clemens-Knott_1992
WKB	130	36.5108	-119.0908	5.74	Clemens-Knott_1992
WKB	130	36.5438	-119.0605	4.8	Clemens-Knott_1992
WKB	130	36.4575	-119.077	4.28	Clemens-Knott_1992
WKB	130	36.449	-119.1084	4.8	Clemens-Knott_1992
WKB	130	36.479	-119.1419	4.34	Clemens-Knott_1992
WKB	130	36.4926	-119.1462	5.02	Clemens-Knott_1992
WKB	130	36.5023	-119.1446	4.62	Clemens-Knott_1992
WKB	130	36.4921	-119.1264	6.16	Clemens-Knott_1992
G04	145.3	32.3247	-120.5981	-5.1	Cecil_et al_2012
BP918B	148	37.0663	-118.3536	-2.79	Glazner_et al_2008
BP918B	148	37.0663	-118.3536	-2.79	Glazner_et al_2008

BP918C	148	37.0663	-118.3536	-1.93	Glazner_et al_2008
BP918C	148	37.0663	-118.3536	-1.93	Glazner_et al_2008
ID93-37	148	36.9796	-118.4013	-4.25	Glazner_et al_2008
ID93-37	148	36.9796	-118.4013	-4.25	Glazner_et al_2008
ID93-39	148	36.9797	-118.4039	-3.28	Glazner_et al_2008
ID93-39	148	36.9797	-118.4039	-3.28	Glazner_et al_2008
ID95-10	148	36.6488	-118.1175	-5.29	Glazner_et al_2008
ID95-10	148	36.6488	-118.1175	-5.29	Glazner_et al_2008
ID95-11	148	36.6478	-118.1182	-0.78	Glazner_et al_2008
ID95-11	148	36.6478	-118.1182	-0.78	Glazner_et al_2008
IN919D	148	36.7843	-118.3113	-7.02	Glazner_et al_2008
IN919D	148	36.7843	-118.3113	-7.02	Glazner_et al_2008
IN919E	148	36.7843	-118.3113	-6.36	Glazner_et al_2008
IN919E	148	36.7843	-118.3113	-6.36	Glazner_et al_2008
IN919F	148	36.7843	-118.3113	-5.27	Glazner_et al_2008
IN919F	148	36.7843	-118.3113	-5.27	Glazner_et al_2008
TP94-34	148	36.9796	-118.4004	-3.53	Glazner_et al_2008
TP94-34	148	36.9796	-118.4004	-3.53	Glazner_et al_2008
TP-9A	148	36.9873	-118.4143	-4.25	Glazner_et al_2008
TP-9A	148	36.9873	-118.4143	-4.25	Glazner_et al_2008
G17	148.7	38.5463	-120.352	-3.7	Cecil_et al_2012
Mzmv	156	37.665	-119.0733	-3.44	Cousens_1996
G02	162	39.3307	-120.1947	2.7	Cecil_et al_2012
G01	162.2	39.2978	-121.0882	-0.37	Cecil_et al_2012
G16	163.2	38.4356	-120.5509	-3.4	Cecil_et al_2012
G14	166.3	38.3721	-120.5542	-5.5	Cecil_et al_2012
G03	166.5	39.3219	-120.633	-3.9	Cecil_et al_2012

Appendix E: Zircon and Apatite

Zircon and Apatite methodology

For zircon and apatite separation, the samples were coarsely crushed in a jaw crusher. This fragmentary material was coarsely milled for 10 seconds and passed through a set of 1700 µm and 500 µm mesh sieves. The remaining material was re-milled for a few seconds and passed through the sieves again. This was done to avoid milling the samples into a fine powder. Sample (IRA18-6) was selected for crushing using the Selfrag instrument to maximize the recovery of zircon. For this preparation procedure, the rock was cut into 2 x 4 x 1 cm dimensions and crushed/liberated using the Selfrag at Geology Department, University of Pretoria. The machine measured 383 discharges, 15 mm gap, at a frequency of 5 Hz, 150 kV, with 6.851×10^4 J. Both the milled and Selfrag samples were density separated into heavy and light minerals using a Wilfley table. The heavy mineral fraction was further separated using a handheld magnet to remove magnetic material such as magnetite. The Frantz magnetic separator was used at four different currents of 0.2A, 0.5A, 1A and 1.5A. The final, highest current, non-magnetic separate was passed through heavy liquids (Bromoform with a density of 2.89 g/cm³ and Diiodomethane with a density of 3.32 g/cm³) to separate the densest non-magnetic minerals, followed by hand picking for zircon and apatite. The zircon and apatite grains were mounted onto an epoxy resin, which was later polished to expose the internal textures of the grains. The LA-ICPMS (Laser Ablation-ICPMS) analyses were performed at the School of Geosciences, Earth Lab, University of the Witwatersrand. Table E1 and E2 shows zircon and apatite standard values obtained during the LAICPMS.

Table E1: Data of zircon standards

Zi re on St an da rd	Fin al2 07_ 235	Final 207_2 35_In t2SE	Fin al2 06_ 238	Final 206_2 38_In t2SE	ErrorCo rrelation _6_38vs 7_35	Fin al2 07_ 206	Final 207_2 06_In t2SE	ErrorCo rrelation n_38_6 vs7_6	Fin al2 08_ 232	Final 208_2 32_In t2SE	Fin al2 06_ 208	Final 206_2 08_In t2SE	Final Age 207_ 235	FinalA ge207_ 235_In t2SE	Final Age 206_ 238	FinalA ge206_ 238_In t2SE	Final Age 208_ 232	FinalA ge208_ 232_In t2SE	Final Age 207_ 206	FinalA ge207_ 206_In t2SE	App rox_ U_P U_P PM	Appro x_U_P PM_In t2SE	App rox_ Th_ Th_ PPM	Appro x_Th_ PPM_I nt2SE	App rox_ Pb_ PPM	Appro x_Pb_ PPM_I nt2SE	Final _U_ Th_ Ratio	Final_ U_Th_ Ratio_I nt2SE	Fin al2 38_ 206	Final 238_2 06_In t2SE
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B B	0.7 282	0.009 9	0.0 889 0	0.000 45	0.12421	0.0 597 7	0.000 83	0.23538	0.0 284 1	0.000 64	18. 15	0.40	554. 0	5.8	549. 0	2.7	566	13	569	30	117. 2	2.7	18.0 7	0.39	44.7	1.1	6.49 1	0.035	11. 248 59	0.056 93889
B B	0.7 63	0.014	0.0 878 8	0.000 43	0.235	0.0 634	0.001 2	0.02246	0.0 324	0.001 0	14. 94	0.53	573. 5	8.2	542. 9	2.6	644	20	686	39	105. 1	2.2	17.5 4	0.29	49.8	1.3	5.95 8	0.050	11. 379 15	0.055 67861
B B	0.7 26	0.012	0.0 855 1	0.000 45	0.11341	0.0 617	0.001 1	0.1732	0.0 267 3	0.000 62	15. 10	0.42	552. 3	6.9	528. 9	2.6	533	12	631	36	100. 9	2.2	18.9 2	0.24	44.5	1.1	5.29 2	0.080	11. 694 54	0.061 54301
B B	0.7 24	0.012	0.0 856 8	0.000 49	- 0.04559 8	0.0 613	0.001 1	0.34472	0.0 266 2	0.000 67	13. 81	0.40	551. 6	7.3	529. 9	2.9	531	13	615	39	87.9	2.0	18.1 2	0.22	42.5	1.1	4.81 6	0.081	11. 671 34	0.066 74783
B B	0.7 21	0.013	0.0 850 4	0.000 53	0.19429	0.0 618	0.001 1	0.2044	0.0 254 7	0.000 64	13. 22	0.36	549. 6	7.5	526. 0	3.1	508	13	631	40	84.4	1.8	19.0 8	0.25	42.5	1.0	4.42 1	0.075	11. 759 17	0.073 28741
Pl es	0.3 905	0.004 9	0.0 532 3	0.000 23	0.12639	0.0 532 8	0.000 69	0.20247	0.0 166 7	0.000 35	19. 77	0.44	334. 4	3.6	334. 3	1.4	334. 0	7.0	326	28	250. 1	5.5	36.1 5	0.90	52.8	1.5	6.95 4	0.036	18. 786 4	0.081 17362
Pl es	0.4 355	0.005 6	0.0 530 9	0.000 26	0.18835	0.0 596 0	0.000 78	0.20399	0.0 221 5	0.000 39	12. 63	0.24	366. 4	3.9	333. 4	1.6	442. 7	7.7	569	28	252. 2	7.4	43.1	1.7	82.3	2.6	6.05 2	0.064	18. 835 94	0.092 24608
Pl es	0.4 261	0.005 7	0.0 529 7	0.000 25	0.19275	0.0 582 1	0.000 77	0.13633	0.0 211 4	0.000 43	14. 24	0.39	359. 6	4.0	332. 7	1.5	422. 7	8.5	519	29	247. 2	7.3	39.9	1.6	75.0	3.2	6.38 8	0.065	18. 878 61	0.089 10048
Pl es	0.3 833	0.004 9	0.0 528 6	0.000 23	0.09552 9	0.0 526 9	0.000 69	0.2383	0.0 162 2	0.000 33	16. 55	0.35	328. 9	3.6	332. 0	1.4	325. 2	6.6	302	29	268. 1	5.5	47.0	1.1	67.2	1.7	5.71 4	0.028	18. 917 9	0.082 31396
Pl es	0.4 527	0.005 4	0.0 534 3	0.000 25	0.17629	0.0 615 1	0.000 74	0.21416	0.0 263 0	0.000 45	13. 49	0.26	378. 8	3.7	335. 5	1.5	524. 5	8.8	640	26	255. 6	5.4	34.1 2	0.88	79.3	2.1	7.56 2	0.044	18. 716 08	0.087 57289
T e m o r a	0.5 13	0.010	0.0 676 9	0.000 43	0.05015 5	0.0 550	0.001 2	0.24359	0.0 210 3	0.000 42	6.5 3	0.14	418. 6	7.1	422. 1	2.6	420. 6	8.3	373	44	73.1	1.6	32.5 2	0.79	61.5	1.7	2.26 4	0.012	14. 773 23	0.093 84679
T e m o r a	0.5 259	0.008 2	0.0 691 3	0.000 36	0.03102 6	0.0 559 8	0.000 93	0.27924	0.0 211 7	0.000 33	6.0 16	0.094	428. 2	5.5	430. 9	2.2	423. 4	6.5	422	36	105. 6	2.2	49.4 2	0.97	91.0	1.9	2.12 4	0.010	14. 465 5	0.075 33025
T e m	0.5 130	0.008 5	0.0 669 8	0.000 38	0.1515	0.0 555 8	0.000 95	0.1237	0.0 203 9	0.000 31	4.1 56	0.066	419. 5	5.7	417. 9	2.3	408. 5	6.1	411	37	99.0	2.1	67.4	1.5	122. 0	2.6	1.46 58	0.0061	14. 929 83	0.084 70193

[illegible]

G J	0.8 120	0.007 2	0.0 978 4	0.000 35	0.16136	0.0 599 9	0.000 54	0.22832	0.0 303 2	0.000 89	89. 1	3.0	602. 8	4.0	601. 7	2.0	608	18	592	20	291. 1	6.1	9.63	0.21	25.8 4	0.88	30.1 6	0.18	10. 220 77	0.036 56244
G J	0.8 215	0.006 8	0.0 985 0	0.000 38	0.10903	0.0 604 6	0.000 53	0.31892	0.0 316 0	0.000 96	86. 7	2.8	608. 6	3.8	605. 6	2.2	628	19	615	19	283. 9	5.4	9.40	0.19	25.7 6	0.85	30.2 1	0.17	10. 152 28	0.039 16617
G J	0.8 120	0.006 2	0.0 972 2	0.000 35	0.09983 7	0.0 605 6	0.000 49	0.34967	0.0 304 7	0.000 93	88. 2	2.8	603. 0	3.5	598. 1	2.1	607	18	617	18	289. 5	6.1	9.60	0.21	25.6 8	0.88	30.1 4	0.18	10. 285 95	0.037 03026
G J	0.8 167	0.006 7	0.0 978 8	0.000 36	0.25487	0.0 604 2	0.000 48	0.19976	0.0 309 7	0.000 92	87. 0	2.6	605. 6	3.7	601. 9	2.1	617	18	612	18	285. 8	5.7	9.48	0.20	25.6 2	0.78	30.1 4	0.20	10. 216 59	0.037 57635
G J	0.8 143	0.006 9	0.0 974 6	0.000 37	0.18702	0.0 606 4	0.000 53	0.25075	0.0 307 0	0.000 96	89. 4	3.0	604. 4	3.9	599. 5	2.2	610	19	618	19	287. 2	5.8	9.51	0.20	25.7 6	0.86	30.2 3	0.20	10. 260 62	0.038 95372
91 50 0	1.9 45	0.031	0.1 815	0.001 1	0.18256	0.0 769	0.001 2	0.20933	0.0 539	0.001 0	7.5 0	0.14	1093	11	1074 .7	6.0	1060	19	1096	32	33.8 3	0.64	13.3 4	0.26	63.2	1.4	2.52 4	0.015	5.5 096 42	0.033 39177
91 50 0	1.8 46	0.033	0.1 791	0.001 2	0.13595	0.0 755	0.001 4	0.22679	0.0 537	0.001 5	10. 80	0.32	1059	12	1061 .9	6.5	1056	29	1058	37	26.6 5	0.52	7.56	0.15	35.0	1.1	3.52 7	0.024	5.5 834 73	0.037 4102
91 50 0	1.8 80	0.034	0.1 784	0.001 3	0.1177	0.0 765	0.001 4	0.26702	0.0 535	0.001 4	10. 51	0.28	1069	12	1058 .1	6.9	1051	26	1083	37	26.0 7	0.54	7.37	0.16	34.1 9	0.93	3.53 6	0.027	5.6 053 81	0.040 84639
91 50 0	1.8 49	0.034	0.1 806	0.001 1	0.06382 3	0.0 747	0.001 4	0.26573	0.0 542	0.001 5	10. 41	0.29	1059	12	1070 .3	6.2	1065	29	1023	39	26.0 4	0.51	7.45	0.14	35.6	1.0	3.49 2	0.024	5.5 370 99	0.033 72541
91 50 0	1.9 23	0.037	0.1 818	0.001 2	0.16277	0.0 762	0.001 4	0.19116	0.0 541	0.001 6	10. 74	0.31	1082	13	1076 .3	6.5	1062	30	1067	39	25.2 1	0.45	7.18	0.13	34.3 9	0.98	3.50 2	0.025	5.5 005 5	0.036 30726

Table E2: Data of apatitestandards

Apatite STD	Final20 7_235	Final 207_ 235_ Int2S E	Final207_23 5_Prop2SE	Final206 _238	Final206 _238_Int 2SE	Final206_23 8_Prop2SE	ErrorCorrel ation_6_38v s7_35	Final238_206_P rop2SE	Final207_20 6	Final207_ 206_Int2 SE	Final207_ 206_Prop 2SE	ErrorCorr elation_3 8_6vs7_6	FinalAge 206_238	Final Age2 06_2 38_I nt2S E	FinalA ge206_ 238_Pr op2SE	Final Age2 07_2 35	FinalA ge207_ 235_Int 2SE	FinalAge 207_235_ Prop2SE
MAD	0.610	0.027	0.028	0.0774	0.0011	0.0052	0.32315	0.8680034	0.0573	0.0025	0.0025	0.038496	480.5	6.6	31	483	17	17

MAD	0.608	0.020	0.021	0.0759	0.0010	0.0051	0.2812	0.8852922	0.0573	0.0017	0.0017	0.13588	471.7	6.1	31	482	13	14
MAD	0.591	0.016	0.017	0.0750	0.0011	0.0051	0.30225	0.9066667	0.0570	0.0016	0.0016	0.23183	466.2	6.4	30	474	11	12
MAD	0.586	0.022	0.023	0.07371	0.00095	0.0050	0.25135	0.9202741	0.0572	0.0020	0.0020	0.18392	458.5	5.7	30	468	14	15
MAD	0.607	0.020	0.021	0.0766	0.0013	0.0052	-0.040395	0.8862287	0.0574	0.0021	0.0021	0.47688	476.0	7.9	31	481	12	13
MAD	0.613	0.029	0.029	0.0775	0.0011	0.0053	0.27644	0.8824142	0.0573	0.0025	0.0025	-0.096351	481.1	6.7	31	484	18	18
MAD	0.594	0.027	0.028	0.0758	0.0011	0.0051	0.24696	0.8876296	0.0566	0.0027	0.0027	0.058547	470.9	6.8	31	472	16	17
MAD	0.64	0.17	0.17	0.0807	0.0015	0.0056	0.17482	0.8598862	0.056	0.015	0.015	-0.12568	500.1	9.1	33	477	94	94
MAD	0.600	0.098	0.099	0.0743	0.0012	0.0051	0.51782	0.923831	0.0578	0.0094	0.0094	-0.44149	461.9	7.2	30	460	61	61
MAD	0.619	0.021	0.022	0.0773	0.0012	0.0053	0.0714	0.8869862	0.0574	0.0019	0.0019	0.41947	480.0	7.3	32	489	13	14
MAD	0.58	0.17	0.17	0.0757	0.0018	0.0053	0.64644	0.9248773	0.054	0.015	0.015	-0.58034	470	11	32	431	96	96
MAD	0.651	0.098	0.099	0.0822	0.0011	0.0056	0.67387	0.8287898	0.0569	0.0083	0.0083	-0.57617	509.5	6.8	33	499	57	57
MAD	0.599	0.025	0.026	0.0756	0.0012	0.0051	0.41714	0.8923322	0.0579	0.0027	0.0027	-0.014395	469.6	7.3	31	475	16	16
MAD	0.583	0.027	0.028	0.0745	0.0010	0.0050	-0.039446	0.9008603	0.0566	0.0029	0.0029	0.33157	463.4	6.3	30	465	17	17
MAD	0.608	0.021	0.022	0.0770	0.0011	0.0052	0.29892	0.877045	0.0571	0.0017	0.0017	-0.067668	478.2	6.6	31	481	13	14
MAD	0.57	0.12	0.12	0.0729	0.0015	0.0051	0.55596	0.959655	0.057	0.011	0.011	-0.44614	453.5	9.1	30	440	70	70
MAD	0.576	0.092	0.092	0.0767	0.0015	0.0053	0.53461	0.9009177	0.0536	0.0079	0.0079	-0.43842	476.3	9.2	32	460	62	62
MAD	0.599	0.021	0.022	0.0756	0.0010	0.0051	0.32915	0.8923322	0.0575	0.0021	0.0021	0.1209	469.6	6.0	31	476	13	14
DUR	0.224	0.016	0.016	0.00670 9	0.00009 9	0.00046	0.28649	10.2198	0.241	0.016	0.016	0.002400 6	43.11	0.63	2.9	204	13	13
DUR	0.232	0.026	0.027	0.00670	0.00025	0.00051	-0.034981	11.3611	0.249	0.028	0.028	0.23581	43.1	1.6	3.3	210	21	21
DUR	0.207	0.012	0.012	0.00650	0.00016	0.00046	0.025356	10.88757	0.232	0.015	0.015	0.39029	41.7	1.1	2.9	190	10	10
DUR	0.218	0.014	0.014	0.00646	0.00021	0.00048	0.35728	11.50208	0.246	0.016	0.016	-0.10783	41.5	1.3	3.0	200	12	12
DUR	0.206	0.013	0.013	0.00653	0.00017	0.00046	0.37503	10.78776	0.225	0.013	0.013	0.097875	41.9	1.1	3.0	190	11	11
DUR	0.212	0.016	0.016	0.00673	0.00024	0.00050	0.49964	11.03926	0.226	0.016	0.016	0.16084	43.3	1.5	3.2	195	13	13
DUR	0.194	0.012	0.012	0.00665	0.00016	0.00047	0.18719	10.62807	0.211	0.013	0.013	0.20352	42.7	1.0	3.0	179.8	9.9	10
DUR	0.231	0.031	0.031	0.00692	0.00033	0.00056	0.37979	11.69434	0.223	0.023	0.023	0.12851	44.4	2.1	3.6	208	24	24
DUR	0.194	0.011	0.011	0.00627	0.00016	0.00045	0.19247	11.44662	0.221	0.013	0.013	0.20949	40.3	1.1	2.9	179.3	9.0	9.2
DUR	0.187	0.011	0.012	0.00643	0.00016	0.00046	0.55149	11.12592	0.208	0.011	0.011	-0.15029	41.3	1.0	2.9	173.9	9.6	9.8
DUR	0.176	0.010	0.010	0.00634	0.00022	0.00047	0.00088415	11.69282	0.200	0.013	0.013	0.11412	40.8	1.4	3.0	164.3	8.8	9.0
DUR	0.187	0.013	0.013	0.00630	0.00022	0.00047	0.32755	11.84177	0.214	0.015	0.015	0.14294	40.5	1.4	3.0	174	11	11
DUR	0.174	0.011	0.011	0.00631	0.00020	0.00046	0.37006	11.55312	0.199	0.012	0.012	0.15403	40.5	1.3	3.0	162.6	9.6	9.8

DUR	0.232	0.014	0.014	0.00680	0.00018	0.00048	0.054443	10.38062	0.244	0.015	0.015	0.39104	43.7	1.1	3.1	212	12	12
DUR	0.1729	0.008 7	0.0089	0.00634	0.00020	0.00046	0.41955	11.44404	0.196	0.011	0.011	0.16221	40.7	1.3	3.0	161.7	7.5	7.7
DUR	0.207	0.011	0.011	0.00646	0.00017	0.00046	0.28887	11.02282	0.234	0.013	0.013	0.23079	41.5	1.1	2.9	190.5	9.0	9.3
DUR	0.189	0.018	0.018	0.00640	0.00028	0.00051	0.57833	12.45117	0.215	0.019	0.019	-0.32361	41.1	1.8	3.2	174	15	15
DUR	0.201	0.011	0.011	0.00626	0.00017	0.00045	0.34423	11.48322	0.233	0.012	0.012	0.24934	40.2	1.1	2.9	185.4	9.0	9.2
McC	2.917	0.050	0.060	0.1048	0.0016	0.0071	0.48949	0.6464513	0.2013	0.0036	0.0036	0.13309	642.4	9.5	42	1386	13	16
McC	3.916	0.072	0.085	0.1109	0.0015	0.0075	0.39563	0.6098151	0.2550	0.0038	0.0038	0.43328	678.2	8.7	43	1616	15	18
McC	3.865	0.078	0.090	0.1126	0.0021	0.0077	0.49941	0.6073149	0.2477	0.0045	0.0045	0.41117	688	12	45	1605	16	19
McC	3.73	0.11	0.12	0.1082	0.0018	0.0074	0.73132	0.6320875	0.2462	0.0046	0.0046	0.011309	662	11	43	1575	24	26
McC	2.725	0.077	0.083	0.0989	0.0017	0.0068	0.72413	0.6952105	0.1965	0.0033	0.0033	0.05111	608	10	40	1339	23	25
McC	4.173	0.073	0.087	0.1154	0.0017	0.0078	0.41798	0.5857105	0.2607	0.0050	0.0050	0.27017	704.2	9.6	45	1668	14	17
McC	2.888	0.059	0.067	0.1015	0.0015	0.0069	0.564	0.6697566	0.2053	0.0033	0.0033	0.26161	623.2	9.0	40	1378	15	18
McC	3.489	0.067	0.078	0.1122	0.0017	0.0076	0.21741	0.6037093	0.2244	0.0038	0.0038	0.58054	685	10	44	1524	15	17
McC	3.011	0.073	0.080	0.1041	0.0012	0.0070	0.49946	0.6459465	0.2107	0.0038	0.0038	0.34386	638.6	7.2	41	1409	18	20
McC	2.956	0.040	0.053	0.1012	0.0013	0.0068	0.38811	0.6639691	0.2121	0.0027	0.0027	0.61726	621.2	7.7	40	1396	10	13
McC	3.399	0.091	0.099	0.1079	0.0015	0.0073	0.1338	0.6270179	0.2285	0.0048	0.0048	0.58862	660.4	8.9	42	1502	21	23