

The nature and origin of gold mineralization at Damang mine, Ghana

by

ALISTAIR J.R. WHITE

Thesis submitted to the University of Oxford

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Supervised by Dr D.J. Waters, Prof. L.J. Robb and Mr E.J. Baltis

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Abstract

The Palaeoproterozoic Birimian terrane of the West African Man Shield is a gold province of global importance. Gold mineralization at the Damang deposit, however, is unique amongst known gold deposits in Ghana. It comprises a stratigraphically controlled auriferous quartz-pebble conglomerate, which is overprinted by later orogenic gold contained in a sub-horizontal fault-fracture quartz vein array. Furthermore, this gold mineralization is hosted in sediments of the Tarkwaian System, rather than volcanics of the underlying Birimian Supergroup. This thesis integrates petrological, geochemical, thermodynamic and geochronological data to investigate the Damang deposit in the context of the regional tectonic and metallogenic framework.

Staurolite-grade, amphibolite facies peak metamorphism at Damang occurred at around 595°C and 5.5 kbar at 2005±26 Ma. This was immediately followed by a short period (~5 Ma) of rapid exhumation, at a rate of approximately 2.6 mm/yr. This decompression led to the formation of the fault-fracture mesh, which was enhanced by the silicification of host rocks during an earlier ‘Pink Haematite’ hydrothermal event that was associated with the intrusion of voluminous dolerite bodies.

Orogenic gold mineralization occurred at the end of this exhumation period at conditions of 400–450°C and 1–2 kbar. These conditions are in broad agreement with other Ghanaian deposits. A gold-bearing, CO₂-rich (X(CO₂) ~ 0.7), aqueous fluid generated through metamorphic devolatilization reactions close to peak metamorphism and during the subsequent exhumation fed the fault-fracture mesh. This fluid imparted a potassic–carbonation–sulphidation alteration overprint similar to that seen at orogenic gold deposits globally. This alteration signature is characterized and identified with portable infrared spectroscopic techniques. Exhumation rates following gold mineralization were less than 0.01 mm/yr, while ⁴⁰Ar/³⁹Ar dating of biotites give ages ranging between 1978.8±6.2 Ma and 1898±11 Ma, indicating extremely prolonged (~80 Ma) cooling through the interval 300–250°C.

Ultimately, this thesis demonstrates that the Damang deposit is part of the same regional metallogenic event that formed the other Ghanaian deposits. This implies that the Tarkwaian System elsewhere in Ghana may be prospective for orogenic gold mineralization.

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Extended Abstract

The Palaeoproterozoic Birimian terrane of the West African Man Shield is a gold province of global importance. It contains two distinct styles of gold mineralization: orogenic gold deposits hosted by regional-scale, sub-vertical shear zones in volcano-sedimentary greenstone belts of the Birimian Supergroup, and quartz-pebble conglomerate, palaeoplacer-style deposits hosted in Tarkwaian System sediments that overly the Birimian belts. Gold mineralization at the Damang deposit, however, is unique amongst known gold deposits in Ghana. It comprises a stratigraphically controlled auriferous quartz-pebble conglomerate, which is overprinted by later orogenic gold contained in a sub-horizontal fault-fracture quartz vein array. Importantly, this orogenic gold mineralization is hosted in a starkly different structural setting to that typically found in the Man Shield and is also held within sediments of the Tarkwaian System, rather than volcanics of the underlying Birimian Supergroup. This thesis integrates petrological, geochemical, thermodynamic and geochronological data to investigate the Damang deposit in the context of the regional metallogenic framework of Ghana.

The Damang deposit has a history comprising various distinct structural, igneous, metamorphic and hydrothermal events. The oldest rock units at Damang are volcanoclastics of the Birimian Supergroup, which are intruded by the so-called Diorite Porphyry. U-Pb isotope zircon analysis constrains the formation of these units at 2178 ± 9.3 Ma and 2164.6 ± 8.0 Ma respectively. The first stage of gold mineralization at Damang occurred during sedimentation of the Tarkwaian System, on top of the progressively deforming Birimian Supergroup, with palaeoplacer-style gold deposited in the Banket Series conglomerates. Orogenic gold mineralization is the second stage of gold mineralization at Damang. This is the latest major

identifiable geologic event, which overprints earlier events and was promoted by many of the preceding structural and hydrothermal episodes.

A first order control on orogenic gold mineralization is provided by regional northeast – southwest-trending folding (Damang Anticline) and faulting (Damang Fault) formed during northwest – southeast compression (TD₁). The timing of movement on the Damang Fault is constrained between growth of biotite and garnet in the Tarkwa Phyllite during regional prograde metamorphism.

East-northeast – west-southwest-trending reverse faulting (TD₂) generated large-scale fluid pathways at intersections with TD₁ structures. TD₂ faulting was broadly coeval with the intrusion of a suite of calc-alkaline – tholeiitic gabbros – diorites and a pervasive oxidizing Cal+Chl+Hem+Ab+Qtz so-called ‘Pink Haematite’ alteration. Extensive silicification associated with this Pink Haematite event altered host rock rheology such that brittle fracturing was promoted during the later orogenic gold mineralization episode. Gold precipitation during this later event, however, is apparently hindered by the oxidation of magnetite to haematite.

Staurolite-grade, amphibolite facies peak metamorphism occurred at around 595°C and 5.5 kbar, as demonstrated by thermodynamic modelling and construction of pressure-temperature pseudosections with the computer program THERMOCALC. U-Th-total Pb ages of metamorphic monazites in the Tarkwa Phyllite constrain peak metamorphism at 2005 ± 26 Ma. Peak metamorphism at Damang is therefore significantly higher-grade and younger in age than is typically associated with the Tarkwaian System across Ghana (greenschist facies peak at 2130–2080 Ma). This includes at the neighbouring Tarkwa mine where amphibolite facies minerals are not apparent. The metamorphic history of Damang is, however, consistent with a regional-scale prograde metamorphic cycle in which deeper, higher-grade rocks reach their metamorphic peak at a later time than lower-grade domains.

Orogenic gold mineralization, hosted in a sub-horizontal, fault-fracture quartz vein array, formed during further regional east-southeast – west-northwest compression (TD₃). Gold is contained within quartz veins and associated localized, rock-buffered potassic–carbonation–

sulphidation alteration selvages comprising an $\text{Ank}+\text{Ms}+\text{Py}\pm\text{Bt}\pm\text{Po}$ assemblage in sedimentary lithologies, and $\text{Ank}+\text{Bt}+\text{Po}\pm\text{Tur}\pm\text{Py}$ coupled with the destruction of hornblende in dolerite intrusives. Gold precipitation occurred through the sulphidation of magnetite, the abundance of which controls the average gold grade in each lithology. This alteration style is comparable to that seen at other orogenic gold deposits elsewhere in Ghana, and indeed globally, implying a link between mineralization at Damang and the rest of Ghana.

Mineralogical changes produced during gold-associated hydrothermal alteration impart a distinctive signature that is characterized with portable infrared spectroscopic techniques. Results show that mineralized sedimentary lithologies have an increased Ferrous-iron Response and $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio and a decreased Water/OH Absorption Depth. These relate to the sulphidation of magnetite, growth of hydrothermal muscovite and overall mineralogical changes respectively. By contrast, mineralized dolerite intrusives display a decrease in both the Ferrous-iron Response and $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio, which is related to the formation of sulphide phases and the conversion of hornblende to biotite respectively. Variations in Ferrous-iron Response are detectable into the weathered profile. This spectral signature is consistent across the Damang region, making this technique a viable and useful exploration tool for use both at Damang and in similar terrains.

Orogenic gold mineralization at Damang occurred under conditions of 400–450°C and 1–2 kbar with a fluid $X(\text{CO}_2)$ of around 0.7, as constrained through pressure-temperature and temperature- $X(\text{CO}_2)$ pseudosections constructed with THERMOCALC for gold-associated alteration assemblages. These conditions are in broad agreement with those determined for many Ghanaian orogenic gold deposits. $^{40}\text{Ar}/^{39}\text{Ar}$ dating of both hydrothermal and metamorphic biotite constrains the timing of post-mineralization cooling, through approximately $300\pm 50^\circ\text{C}$, to within the range $1978.8\pm 6.2\text{ Ma} - 1898\pm 11\text{ Ma}$. This range is interpreted to relate to variations in Ar retention during cooling as a function of grain size. Numerical modelling of Ar diffusion through biotite using the computer program DIFFARG implies a cooling rate below 270°C of around $0.15^\circ\text{C}/\text{Ma}$. This age range also places a minimum age constraint on the timing of orogenic gold mineralization, which is younger than typ-

ically quoted from elsewhere in Ghana (2060–2090 Ma) but is consistent with Damang’s metamorphic history.

Immediately following peak metamorphism the Damang region underwent rapid exhumation at a rate of around 2.6 mm/yr for ~ 5 Ma, as determined through thermal modelling using the computer program THERMODYL. This exhumation is related to regional TD₃ compression and formation of the fault-fracture array. Generation of a mineralizing fluid through chlorite dehydration, and similar reactions, immediately prior to peak metamorphism was supplemented by fluid generated during rapid decompression and fed the fault-fracture array, leading to quartz vein precipitation and gold-associated alteration. Following the gold mineralization event, regional exhumation slowed to less than 0.01 mm/yr, likely nearer 0 mm/yr.

These results indicate that while the Damang deposit is currently unique, its structural, metamorphic and hydrothermal history is not necessarily so. Therefore, the Tarkwaian System across the Man Shield is prospective for orogenic gold mineralization, thereby expanding the terrane’s potential gold endowment. Furthermore, the post-peak metamorphic timing of orogenic gold mineralization at Damang, with fluid generation related to regional exhumation, is inconsistent with a metamorphic model of deposit formation in which mineralization occurs in association with prograde devolatilization.

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*The Damang exploration and mineral resource team... and me.
July 2009*

For Sam

For everything



November 2008



June 2009



July 2010



April 2011



*“All that is gold does not glitter,
Not all who wander are lost;”*

J.R.R. Tolkien
The Lord of the Rings

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List of Abbreviations

Mineral abbreviations used throughout this thesis follow generally those of ([Kretz, 1983](#)) with those used regularly listed here:

Act	actinolite	Grt	garnet	Ab	albite
Ank	ankerite	Hbl	hornblende	An	anorthite
Bt	biotite	Hem	haematite	Po	pyrrhotite
Cal	calcite	Ilm	ilmenite	Py	pyrite
Chl	chlorite	Mag	magnetite	Qtz	quartz
Cld	chloritoid	Mnz	monazite	St	staurolite
Cum	cummingtonite	Ms	muscovite	Tur	tourmaline
Dol	dolomite	Pg	paragonite	Zrn	zircon
Ep	epidote	Pl	plagioclase		

Chapter 1

Introduction



The Damang gold pour.

Ore deposits are occurrences of minerals and/or metals that can be extracted from the Earth at a profit. Ore deposits, therefore, are anomalies within the crust and require the action of one or more mechanisms to raise element concentrations significantly above average crustal levels. As such, they provide insights into geological processes that are not possible in other settings. Such understanding is often enhanced by extensive drilling and sampling, providing a multitude of three-dimensional data that is rarely available elsewhere. In particular, ore deposits formed from the action of hydrothermal fluids allow for the study of the generation, characteristics and movement of fluids by investigating their interaction with crustal rocks. With this as an overriding motivation, this thesis investigates an unusual hydrothermal gold deposit, Damang, in southwest Ghana.

1.1 Gold Mineralization in Space and Time

Gold, as with many other mineral deposits, is known to be unevenly, but not randomly, distributed across the globe (Fig. 1.1). Gold occurrences show a strong correlation with particular geological tectonic settings and geological epochs, which are themselves strongly interconnected (Sawkins, 1990; Barley & Groves, 1992; Groves *et al.*, 1998, 2005; Robb, 2005; Groves & Bierlein, 2007).

1.1.1 The Tectonic Settings of Gold Mineralization

Gold deposits in the Earth's crust form in a wide range of geological environs; primarily occurring in convergent margin settings but also in extensional regimes and in association with anorogenic magmatism (Fig. 1.2). Commonly, gold occurs in conjunction with other precious or base metals, often as a by-product, while in other cases gold may form monometallic deposits. This distinction has important implications for the mineralization process. One of the most recent and comprehensive reviews of mineral deposits in relation to tectonic setting is that of Groves & Bierlein (2007).

The two most important gold-bearing deposits formed in broadly extensional settings are

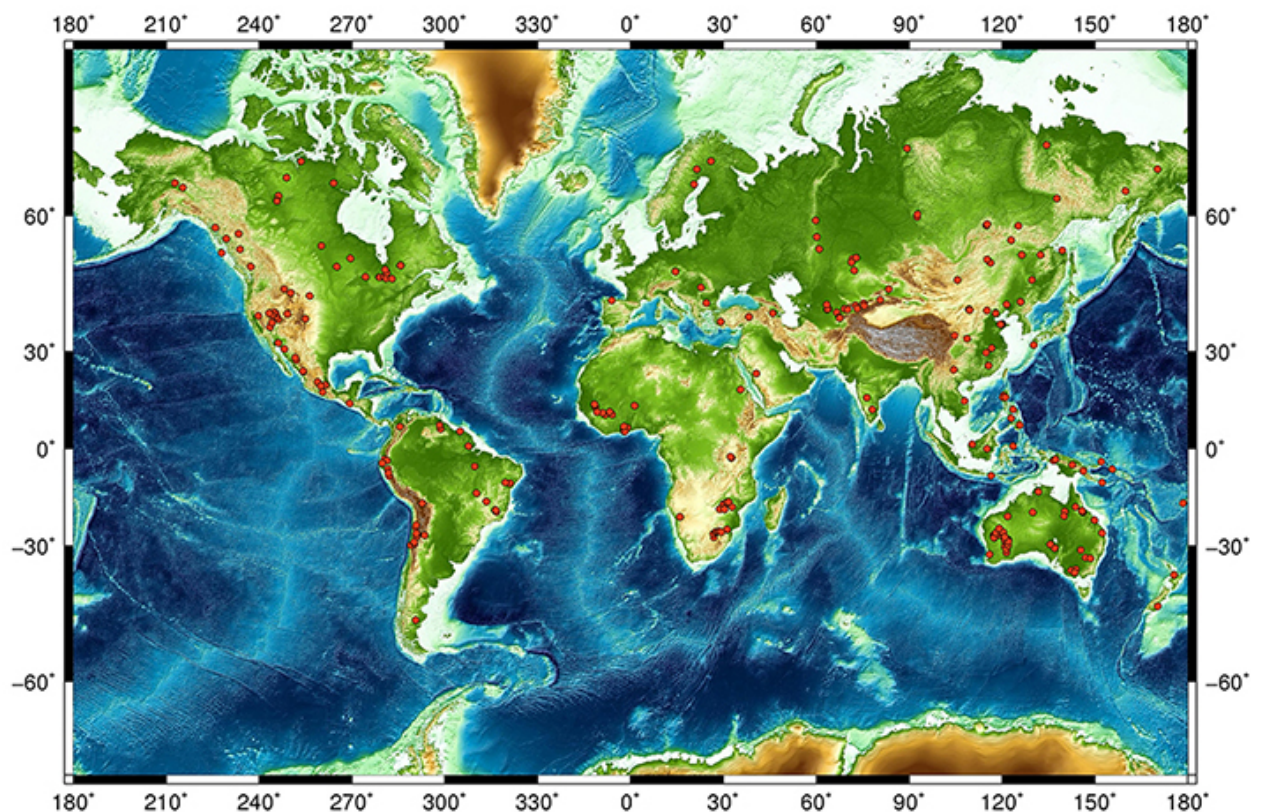


Figure 1.1: Map of global gold deposits containing resources of over 3 Moz of gold and where gold is the major product. Important gold provinces are clearly visible, including the Witwatersrand Basin and Barberton greenstone belt in Southern Africa, the Yilgarn craton in Western Australia, the Lachlan fold belt in southeast Australia, the Tien Shan mountain belt in Kazakhstan, the Birimian terrane of West Africa, porphyry gold deposits of Chile, the Abitibi greenstone belt in eastern Canada and the Carlin-type deposits of Nevada. Data from [Gosselin & Dubé \(2005\)](#).

1) the epigenetic Carlin-type deposits, named for the prototype in Nevada, USA, and 2) syngenetic volcanogenic massive sulphide (VMS) deposits. Carlin-type deposits are hosted in mildly extended continental crust although the exact process behind their formation is controversial, since they appear to be associated with both orogenic and intrusion-related gold deposits ([Cline *et al.*, 2005](#)). Gold is often a minor component of VMS deposits formed at oceanic spreading ridges, particularly in back-arc settings ([Groves *et al.*, 1998](#)).

Gold deposits formed in compressional settings are economically more significant than those in extensional settings, due to their greater abundance. Different deposit styles form at all stages of an orogenic cycle ([Groves & Bierlein, 2007](#); [Bierlein *et al.*, 2009](#)). Porphyry Cu-Au deposits, and commonly associated skarns, such as those in the Chilean Andes, form in oceanic island arc and continental margin volcanic arc settings. These terranes also host epithermal $\text{Au} \pm \text{Ag} \pm \text{Cu}$ mineralization at shallow crustal levels, which may be genetically

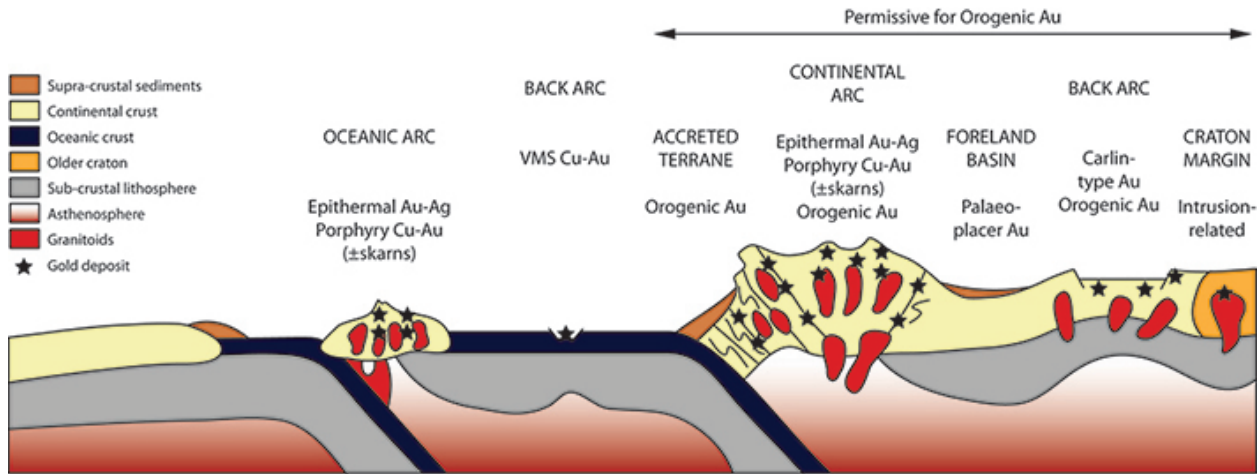


Figure 1.2: Schematic diagram illustrating the range of tectonic settings in which gold deposits form. While the term 'accreted terrane' is here shown in association with a fore arc setting, all continental collision zones inboard of the arc are permissive for orogenic gold deposits. Modified from Groves *et al.* (1998).

linked with the deeper porphyry systems (Hedenquist *et al.*, 2000; Taylor, 2007). Late in the main orogenic stage, so-called orogenic gold deposits, such as those in the Yilgarn craton of Western Australia, the Superior craton of eastern Canada, or Damang mine itself, predominate in metamorphic terranes. These types of deposit are discussed in more detail in section 1.2.1. Finally, during the late- to post-orogenic stage, Au-bearing palaeoplacer type deposits may form, such as the Tarkwaian deposits of Ghana or the 'super-giant' Witwatersrand basin palaeoplacers. This late-orogenic stage may also involve extensional rifting, resulting in formation of the Carlin-type deposits mentioned above.

Gold also occurs in a number of anorogenic, or otherwise uncertain, tectonic settings, such as the enigmatic and rather poorly-defined iron-oxide-copper-gold (IOCG) deposits and other reduced-intrusion related gold deposits (Hitzman *et al.*, 1992; Hart, 2007).

1.1.2 Gold and Geologic Time

The distribution of mineral deposits through time is an altogether more complex problem and has been the subject of a number of in depth studies and reviews (Barley & Groves, 1992; Kerrich & Cassidy, 1994; Goldfarb *et al.*, 2001). The temporal distribution of global mineral deposits is thought to relate to a number of different factors. Some of these factors are the development of plate tectonic styles such as the transition from plume-related tectonics

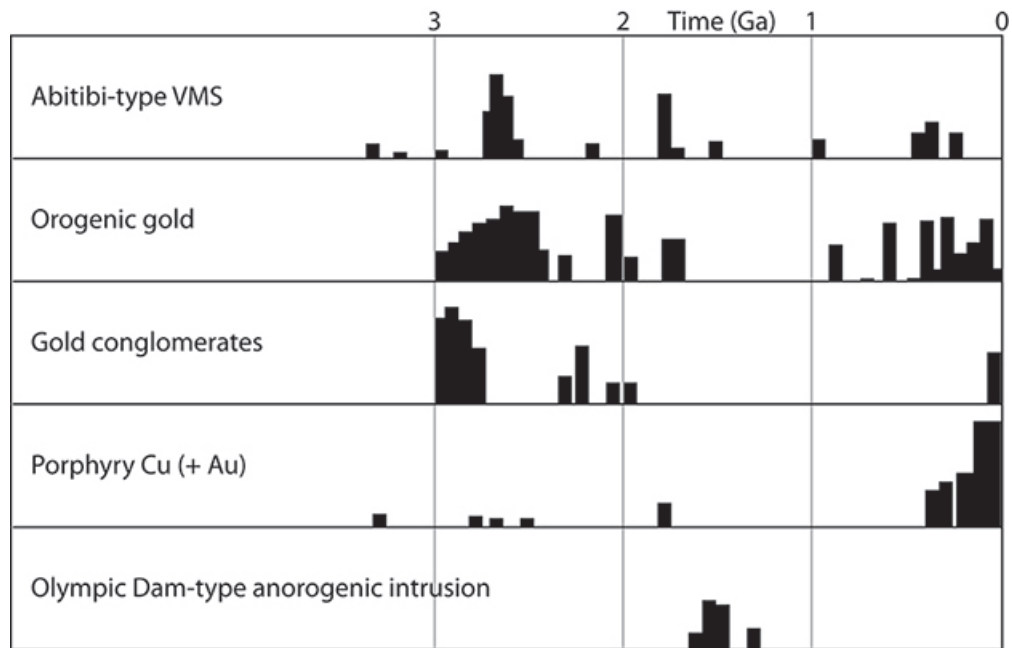


Figure 1.3: The temporal distribution of a range of orogenic and anorogenic gold deposit types. Each bar represents a 50 Myr interval, with the height proportional to the estimated amount of ore formed during that interval. Modified from [Barley & Groves \(1992\)](#) and [Groves & Bierlein \(2007\)](#).

in the Archaean to modern-day plate tectonics from the late Archaean – early Proterozoic onwards, the rate of growth of continental crust, supercontinent cycles, secular changes in global heat production, evolution of the atmosphere and hydrosphere - particularly during the oxygenation event - and preservational bias for different deposit types ([Barley & Groves, 1992](#); [Groves *et al.*, 2005](#); [Robb, 2005](#); [Kesler & Wilkinson, 2006](#); [Groves & Goldfarb, 2007](#)).

Therefore, a better understanding of the distribution and formation of ore deposits provides insights into 4-dimensional global tectonic processes. Likewise, the converse is true and improved knowledge of the geological processes that give rise to and operate during tectonic events aids in the exploration for future ore deposits.

Gold deposits of all kinds have formed episodically throughout the past 3000 Ma of Earth history. However, individual types of gold deposit show an even more heterogeneous distribution (Fig. 1.3, [Barley & Groves, 1992](#)). Deposits preserved in the Archaean and Palaeoproterozoic are dominated by orogenic gold deposits in granite-greenstone terranes, such as the Yilgarn craton of Western Australia, the Superior Province of Canada and the West African craton, Abitibi-type VMS deposits and the Witwatersrand palaeoplacers. These are all related to major shield forming and cratonization events. Mesoproterozoic gold is

primarily hosted within anorogenic igneous intrusions and IOCG deposits, such as Olympic Dam in Australia. This likely represents a prolonged period of essentially stable continents with no major orogenic events ([Barley & Groves, 1992](#)). The Neoproterozoic – Phanerozoic has seen another period of orogenic gold formation with numerous metallogenic epochs, such as the Lachlan fold belt of eastern Australia, the Tien Shan mountain range of Central Asia, regions of eastern China and Siberia and the Alaska–Yukon mountains, along with extensive porphyry Cu-Au and epithermal Au deposit formation in the circum-Pacific regions. This recent dominance of porphyry and epithermal deposits is likely related preservational bias due to exhumation and erosion of these relatively shallow forming deposit types from Mesozoic and older terranes ([Kesler & Wilkinson, 2006](#)).

1.2 Orogenic Gold Deposits

The term ‘orogenic gold deposit’ is applied here to the kind of epigenetic gold deposits that are hosted in broadly deformed and metamorphosed terranes (see reviews by [Goldfarb *et al.*, 2001, 2005](#); [Dubé & Gosselin, 2007](#)). There are geologic sub-divisions within this group. For example, Archaean deposits are primarily hosted in volcanic greenstone host rocks while ‘slate-’ or ‘turbidite-’hosted deposits dominate in the Palaeoproterozoic and Phanerozoic. Many of the fundamental characteristics of these deposits however, are the same.

The vast majority of orogenic gold deposits are spatially associated with first-order, deep-crustal shear zones cutting thick volcano-sedimentary successions, with individual lodes located along second- and third-order shears ([Goldfarb *et al.*, 2001, 2005](#); [Dubé & Gosselin, 2007](#)). There is also often a close spatial association with granitoid intrusions and lamprophyre dykes ([Rock & Groves, 1988](#)). Wall rock sulphidation (arsenopyrite and pyrite) is abundant, as is potassic alteration (sericite and biotite) and carbonation. Gold is commonly accompanied by Ag, As, W, Bi, Te and Sb while base metals are relatively rare. Vein-forming fluids are typically low salinity, CO₂-rich, aqueous fluids ([Mikucki & Ridley, 1993](#); [Ridley & Diamond, 2000](#)). Formation occurs over a wide range of temperatures with a distinct cluster between 250–400°C. Mineralization styles vary as a function of metamorphic grade, from

low temperature brittle stockworks and breccias with sericite-chlorite alteration, through to higher temperature crack-seal veining and disseminated mineralization with biotite alteration. This range of local-scale morphologies has led to the development of a number of competing genetic models to explain their formation.

1.2.1 Genetic Models of Orogenic Gold Deposits

The basic features of orogenic gold deposits, as described in the preceding section, are well characterised. There remains, however, a debate regarding the mode of formation of these deposits, particularly with regards to 1) the origin of the mineralizing fluid and 2) the relative timings of mineralization and regional metamorphism (see discussions in [Kerrick, 1993](#); [Kerrick & Cassidy, 1994](#); [Groves *et al.*, 2003](#)). At present, two opposing models predominate; the crustal continuum model and the metamorphic model.

The crustal continuum model (Fig. 1.4) states that deposits form over a wide range of pressure-temperature conditions as fluids exploit a deep-structural architecture ([Groves, 1993](#); [Groves *et al.*, 1998](#); [McCuaig *et al.*, 1993](#); [McCuaig & Kerrich, 1998](#); [Goldfarb *et al.*, 2005](#)). This model has typically been favoured in recent years since all orogenic gold deposits, irrespective of metamorphic grade, exhibit the same metal budget and alteration systematics, thereby forming a continuum. In general, mineralization occurs late in the orogenic cycle and typically post-dates regional metamorphism in greenschist facies terranes ([Groves, 1993](#)). Yet, in order for the proposed continuum to be valid, higher-grade terranes must undergo mineralization at conditions similar to those for peak metamorphism rather than becoming overprinted at much lower temperatures. This hypothesis has been disputed however, particularly with regards to the Griffins Find, granulite-facies deposit in Western Australia. [Tomkins & Grundy \(2009\)](#) show that mineralization at Griffins Find occurred after peak metamorphism at much lower temperatures (likely less than 650°C), and indeed suggest the same for all high-grade deposits. Furthermore, [Phillips & Powell \(2009\)](#) point out that the presence of an aqueous fluid, of any source, which is required for gold transport and precipitation, is incompatible with granulite facies conditions involving partial melting,

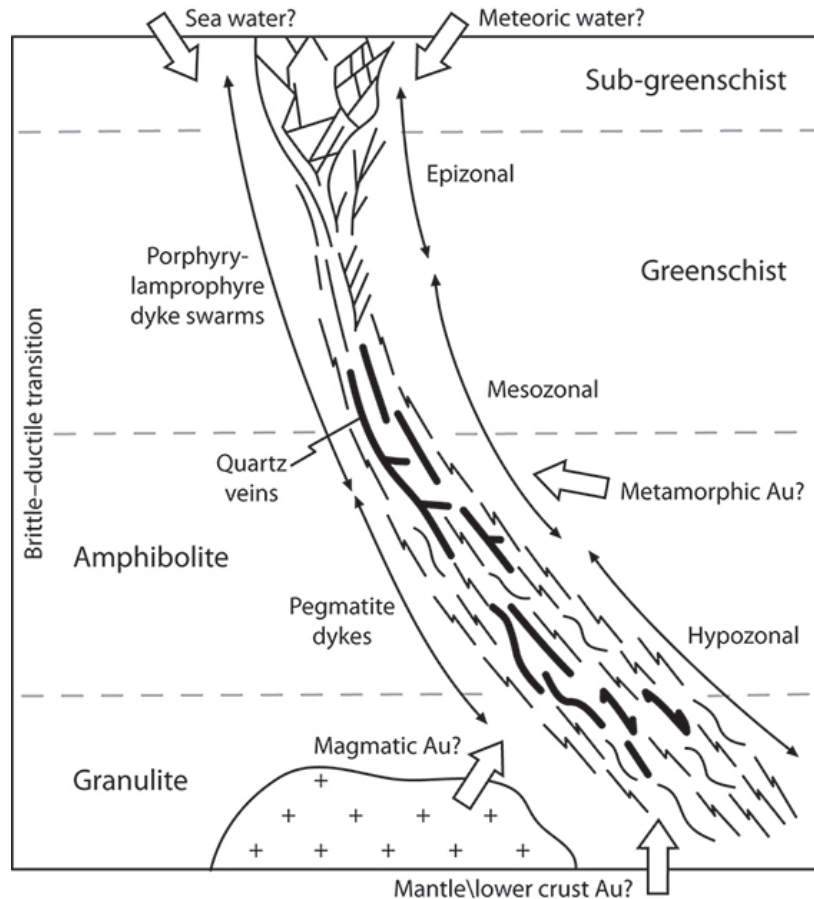


Figure 1.4: The crustal continuum model for the formation of orogenic gold deposits, with deposits forming at a range of crustal depths and temperatures. Modified from Groves (1993) and Goldfarb *et al.* (2001).

since any fluid would be consumed to produce further melt. Therefore, mineralization cannot occur at such high temperature conditions. The continuum model does allow for a range of different fluid sources (Fig. 1.4) including magmatic, metamorphic and mantle sources as well as the involvement of meteoric waters at shallower depths.

By contrast, the metamorphic model suggests that the mineralizing fluid is generated, in all cases, through metamorphic devolatilization of hydrated and carbonated mafic rocks close the greenschist–amphibolite facies transition during prograde metamorphism (Groves *et al.*, 1987; Phillips & Powell, 1993; Ridley & Diamond, 2000; Phillips & Powell, 2009, 2010). This is supported by Pitcairn *et al.* (2006), who showed that metamorphic devolatilization at the greenschist–amphibolite facies transition can leach vast quantities of gold from surrounding rocks. The metamorphic model explains occurrences of deposits in higher-grade terranes through continued prograde metamorphism following mineralization. There are however, problems with this model as well. Sediment-hosted orogenic gold deposits in non-Archaeon terranes may not contain sufficient mafic material to produce the proposed fluid during pro-

grade metamorphism. Although, carbonaceous sediments, which may be present, could be a significant source of both gold and mineralizing fluids ([Large *et al.*, 2011](#)). The metamorphic model also cannot explain the apparent post-peak metamorphic age for the Griffins Find deposit ([Tomkins & Grundy, 2009](#)), since mineralization occurs temporally and spatially with the devolatilization. However, as is discussed by [Pettke *et al.* \(1999\)](#); [Stüwe *et al.* \(1993\)](#) and [Stüwe \(1998\)](#), devolatilization of prograding rocks, as per the metamorphic model, may produce fluid that rises in to overlying rocks that have already passed their metamorphic peak, thereby giving the impression of post-peak mineralization.

1.3 Gold Economics

Gold deposits are of economic importance since gold has the unique role of being both an industrial commodity and a store of value. As such, it has been relatively resilient to the financial crisis that affected global economies during 2008–2011. A detailed review of

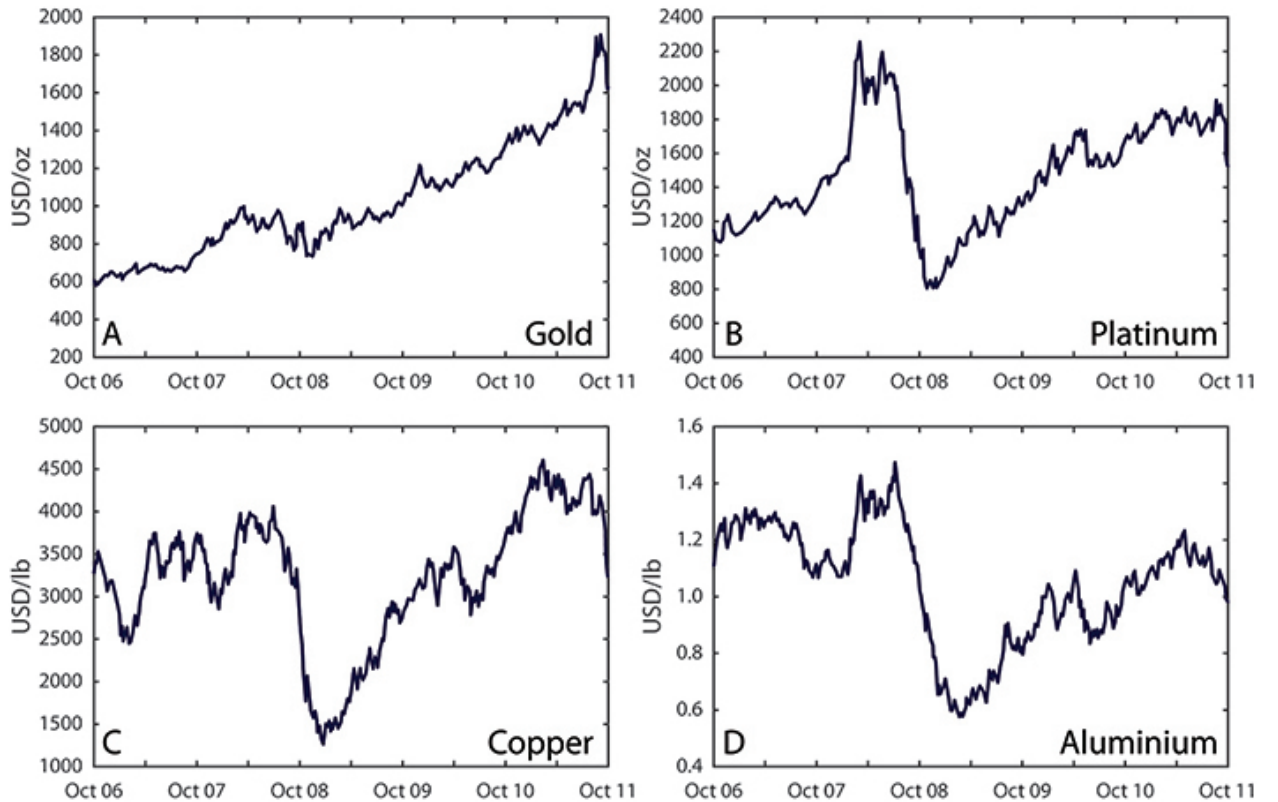


Figure 1.5: Commodity prices for gold (A), platinum (B), copper (C) and aluminium (D) over the past 5 years. While platinum, copper and aluminium experienced a large drop in price during 2008, gold was much more resilient due to continued demand. Data based on commodity price at New York close and available from [Kitco \(2011\)](#).

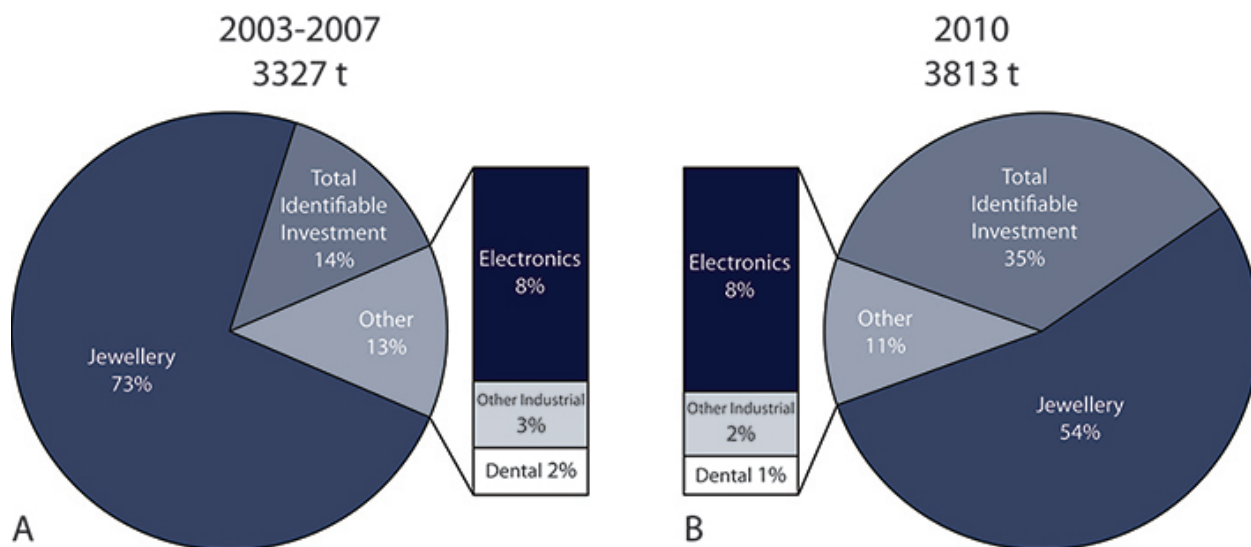


Figure 1.6: Demand for gold for the periods 2003–2007 (A) (5-year average), prior to the global financial crisis, and 2010 (B). Note the dramatic decrease in demand for jewellery and corresponding increase in demand for investment gold in 2010. Data from [Council \(2011\)](#)

the economics of gold is, of course, well beyond the scope of this thesis. However, a brief summary is presented here that illustrates some important characteristics and emphasises the need for the continued study of, and exploration for, gold deposits to meet the demands of modern society.

1.3.1 Gold Price

The price of gold, as with other commodities, is a delicate balance between supply and demand, which are themselves a complex function of financial, commercial and social factors ([Crowson, 2011](#)). Since 2002 the price of gold has continued to rise steadily; the highest recorded gold price of around 1900 USD/oz (based on New York close) occurred in mid-2011 (Fig. 1.5A; [Kitco, 2011](#)). In contrast, other precious (platinum) and industrial (copper and aluminium) metals experienced a major drop in their commodity prices during 2008. Of these, only copper has recovered close to its pre-2008 price (Fig. 1.5B–D). Despite a slight price drop during 2008, gold withstood the global financial crisis, indicating the demand that existed, and still exists, for gold. This high demand, coupled with such high prices, are driving exploration efforts across the world's gold provinces.

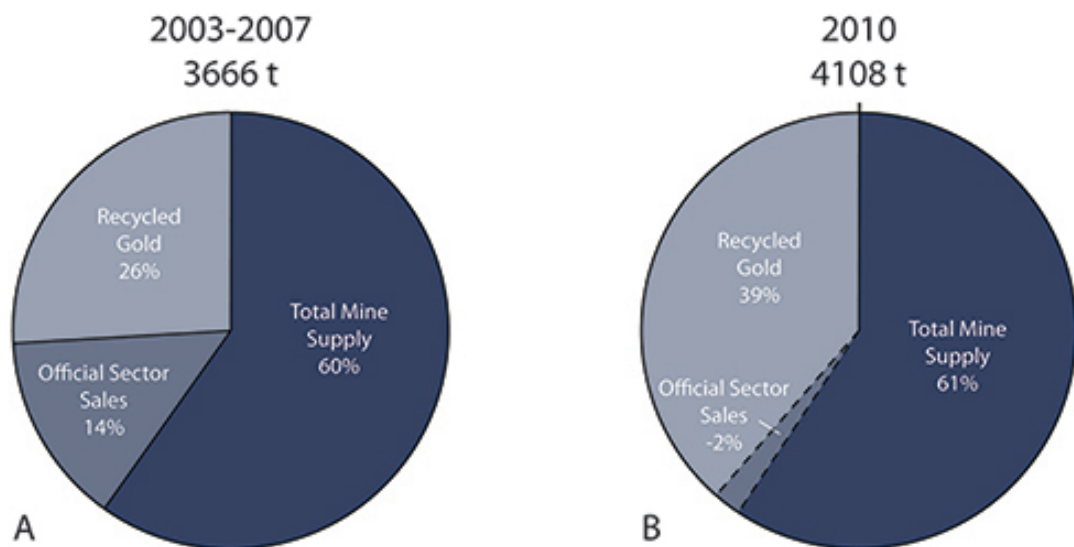


Figure 1.7: Supply of gold for the periods 2003–2007 (A) (5-year average), prior to the global financial crisis, and 2010 (B). Note the dramatic increase in the proportion of recycled gold in 2010. Data from [Council \(2011\)](#)

1.3.2 Gold Demand

Demand for gold can be divided into three broad categories: jewellery, industrial and investment. Industrial use is primarily in electronics due to gold’s high conductivity and relatively inert state that is resistant to corrosion, but also includes gold alloys that are used extensively in dentistry. Investment uses of gold include hoarding a store of bullion coins or bars as a hedge against inflation, either personally or by national governments.

Figure 1.6 shows a break down of gold demand averaged for the 5 years between 2003–2007, prior to the global financial crisis, along with demand in 2010. The most noticeable difference is a decrease in demand for jewellery from almost 75% to just over 50%. This has been balanced by an increase in demand for investment gold. Interestingly, total annual demand has increased slightly from 3327 to 3813 tonnes.

1.3.3 Gold Supply

While all gold is ultimately sourced from mine production, annual supplies are from three major sources; new mine production, ‘official sector sales’ (the sale of gold produced at an earlier time into the market), and recycled gold from pre-existing sources (e.g. electrical

components). Figure 1.7 shows a break down of gold supply averaged for the 5 years between 2003–2007 and for 2010. As with gold demand, total annual supply has increased slightly from 3666 to 4108 tonnes, although there has been a major shift in the balance of sources. Mine supply has remained similar, while recycled gold has drastically expanded, replacing official sector sales, which show a negative value for 2010. This indicates that gold has been purchased from the market and stored.

1.3.4 Gold Discovery Trends

The preceding sections have shown that supply and demand for gold has been growing in the recent past. However, future projections are more worrying. McKeith *et al.* (2010) investigated the rate and quality of gold discoveries, and the effectiveness of the exploration industry since 1950. They highlight a number of discovery trends since the gold discovery boom of the late 1970s and 1980s:

1. Total ounces of newly discovered gold, per year, have been steadily decreasing. In particular, there has been a marked decrease in the number of deposits discovered with resources greater than 1 Moz. Consequently, the bulk of mined gold is produced from deposits that were discovered many years ago and are not being replaced.
2. Financial and political risk associated with new discoveries is increasing as the traditional gold provinces are becoming depleted.
3. Exploration effectiveness, as measured by return on investment, has declined. During the 1960s, one US dollar spent on exploration would, on average for the decade, return 50 US dollars of value in the ground. From 2000–2009, this average had dropped to just 6 US dollars, despite the real gold price more than doubling between these two decades. This is related not just to lack of discovery, but also rising costs of exploration.

Despite these trends, recent studies suggest that only a fraction of possible gold deposits have been discovered (Frimmel, 2008; Carranza *et al.*, 2009; Kesler & Wilkinson, 2010; Wilkinson & Kesler, 2010). Ultimately, this highlights the need for continued gold exploration, backed

by sound research, first to improve the rate of discovery, and second, to reduce exploration costs by developing cheaper and more efficient exploration methodologies.

1.4 History of Damang Gold Mine

Industrial gold mining in the Damang region dates back to the late 19th century although local artisanal workings undoubtedly occurred before this time. At the southern end of the current Damang mining lease, the Abosso palaeoplacer mine was in production from 1882 until 1956. An underground operation, Abosso mine expanded northwards to incorporate the Adjah Bippo and Cinnamon Bippo underground mines in 1920. In 1989, after a 33 year hiatus, activity in the area restarted when Ranger Exploration began investigating the possibility of re-treating the old Abosso mine tailings. Early exploration pitting, trenching and later drilling, initially targeting palaeoplacer mineralization, coupled with the help of observing local artisanal workers, discovered epigenetic gold associated with quartz veins defining a feasible mining operation over a strike length of 3 km. Mining and gold production from the main Damang open pit began in November 1997. Gold Fields' first involvement at Damang was in 2001 when their wholly owned subsidiary, Abosso Goldfields Ltd., purchased part of Ranger's 90% interest in the mine. Extensive drilling and a feasibility study in 2004 supported the expansion of the main pit through the Damang Pit Cutback, which commenced in 2005 along with mining at other prospects in the area, such as Amoanda and Tomento. Over the following years, additional hydrothermal and palaeoplacer prospects were developed between Abosso and Damang (Rex in particular) as the Damang Pit Cutback continued. Continued exploration and resource estimation at present indicate that Damang's current life of mine is upwards of 9 years.

1.5 Previous Studies

Damang mine has been the focus of a small number of previous studies. Most notable are the Masters thesis of [Brabham \(1998\)](#), the PhD thesis of [Pigois \(2003\)](#) and the paper of

Tunks *et al.* (2004). Brabham (1998), in addition to describing many general aspects of the deposit, constructed a stratigraphic framework, with lateral correlation of units between the east and west limbs of the Damang anticline as well as along strike between Damang and Tarkwa. Pigois (2003) focussed more strongly on structural aspects of the deposit as well as carrying out a comprehensive detrital zircon age study of the Tarkwaian stratigraphy. Other geochronological methods, including U-Pb from hydrothermal xenotime and $^{40}\text{Ar}/^{39}\text{Ar}$ dating of biotite, were completed but less extensively. The paper of Tunks *et al.* (2004) is the result of a detailed mapping and structural analysis study, which defines the regional structural framework that is currently the basis of much exploration at Damang.

1.6 Aims and Overview of Thesis

This study is focussed on a more complete petrological, geochemical and geochronological investigation of the Damang deposit that will complement the previous studies to provide a more complete picture of the deposit as a whole and its place within the regional framework. There is little to be gained from reproducing previous work such as the detailed structural work of Pigois (2003) and Tunks *et al.* (2004) or the detrital zircon study of Pigois *et al.* (2003). This thesis therefore, focusses on other aspects of the deposit with five major aims:

1. Fully characterize the petrology and geochemistry of the Damang host rocks and successive hydrothermal alteration events,
2. Determine pressure-temperature-fluid composition conditions of regional metamorphism and orogenic gold mineralization,
3. Constrain the timing of the the different tectonic, magmatic, metamorphic and hydrothermal events through a comprehensive geochronological study, utilizing different methods to those of Pigois *et al.* (2003),
4. Integrate these results into a coherent geological history and mineralization model for the Damang deposit, in the context of the regional tectonic and metallogenic framework,

5. Assess and develop the application of portable infrared spectroscopy as an exploration tool at Damang, such that it could be implemented in future exploration programs at Damang, regionally, and in other tropical terrains globally.

This thesis is organized into a series of broadly self-contained chapters that are then integrated and discussed together at the end. Chapter 2 describes the regional geology of Ghana in the context of the West African craton, along with an overview of the general geology of the Damang deposit. Chapter 3 addresses the petrology and geochemistry of the deposit, considering primary lithologies and the effects of regional metamorphism and hydrothermal activity, including mineralization. Chapter 4 discusses the thermodynamic modelling that was undertaken to extract pressure-temperature-fluid composition information from the rocks at Damang and build up a full metamorphic and metasomatic history. Chapter 5 presents the results of the geochronological study, addressing each of the different techniques in turn. Chapter 6 describes the application of portable infrared spectroscopy and its potential use as an exploration tool at Damang and in similar terrains. Finally, chapter 7 presents the final discussion and conclusions, drawing on each of the preceding chapters to develop a mineralization model for Damang and some key points that may aid future exploration.

1.7 Supporting Publications

The majority of the work presented in chapter 6 has been published in a special publication of the Society of Economic Geologists (White *et al.*, 2010). At the time of submission, three further papers, focussing on the contents of chapters 3 and 4, chapter 5, and chapter 7 respectively are being prepared for publication.

Chapter 2

Tectonic, Geologic and Metallogenic Framework of Ghana and the Damang Region



Lima pit and the surrounding sub-tropical rainforest.

An overall aim of this thesis is to relate the Damang deposit to the regional geologic framework of Ghana and West Africa. This chapter, therefore, addresses the geology, structure, metamorphism, geochronology and gold mineralization of Ghana and West Africa. This provides the framework against which the Damang deposit is compared. Following this, a discussion of the geology of the Damang region itself is presented, focussing on its structural history and general aspects of its gold mineralization.

2.1 The West African Craton

The West African Craton is an area of approximately 4,500,000 km², extending from Mauritania in the northwest to Ghana in the southeast (Fig. 2.1). It comprises two large Archaean – Palaeoproterozoic shield areas, the northern Reguibat Shield of Mauritania and the southern Man Shield, which stretches between Senegal and Ghana (Fig. 2.2) (Ennih & Liégeois, 2008). The central portion of the craton is overlain by Mesoproterozoic – Palaeozoic sedimentary successions and the craton margins are bounded to the east and west by Neoproterozoic (Pan-African) fold belts. Accretion, deformation, metamorphism and associated gold mineralization of the Man Shield occurred during a major period of crustal growth through a series of orogenic events. In Ghana, this is represented by the 2130 – 1980 Ma Eburnean Orogeny, which was a major period of crustal growth (Abouchami *et al.*, 1990; Boher *et al.*, 1992; Eisenlohr & Hirdes, 1992; Feybesse & Milési, 1994; Allibone *et al.*, 2002; Hirdes & Davis, 2002; Feybesse *et al.*, 2006; Harcouët *et al.*, 2007), while in Burkina Faso it includes the earlier 2170–2130 Tangaean Event (Hein, 2010).

2.2 The Palaeoproterozoic of Ghana

Ghana lies on the southern margin of the Man Shield (Fig. 2.2) (Ennih & Liégeois, 2008). The western portion of Ghana contains the Palaeoproterozoic Birimian terrane, which comprises a number sub-parallel, northeast – southwest-trending, several hundred kilometre long,

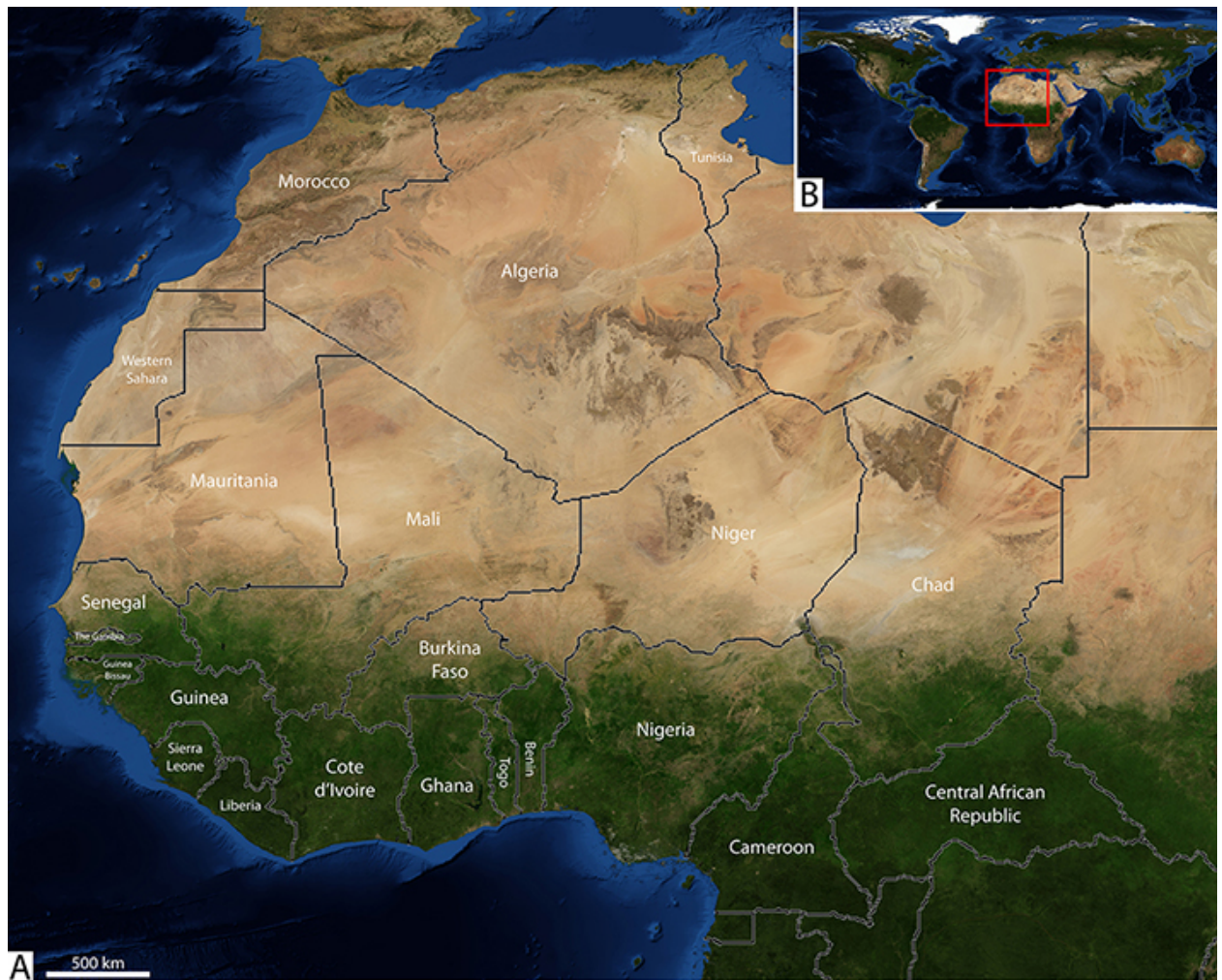


Figure 2.1: NASA Blue Marble (Stöckli *et al.*, 2005) MODIS satellite image of West Africa with political borders.

volcano-sedimentary greenstone belts that are intruded by Belt-type (formerly termed Dixcove type) tonalitic to granitic plutons (Hirdes *et al.*, 1992; Taylor *et al.*, 1992; Davis *et al.*, 1994; Oberthür *et al.*, 1998). The intervening basins contain large quantities of Birimian clastic sediments and are themselves intruded by the voluminous Basin-type (formerly termed Cape Coast type) granitoids (Hirdes *et al.*, 1992; Taylor *et al.*, 1992; Davis *et al.*, 1994). Geophysical data suggests that the Birimian extends northeastwards beneath the Palaeozoic Volta Basin (Hastings, 1982). The volcanic greenstone belts are occasionally overlain by clastic sediments of the younger Tarkwaian System (Davis *et al.*, 1994; Pigois *et al.*, 2003). Each of these lithological units, will now be discussed in turn.

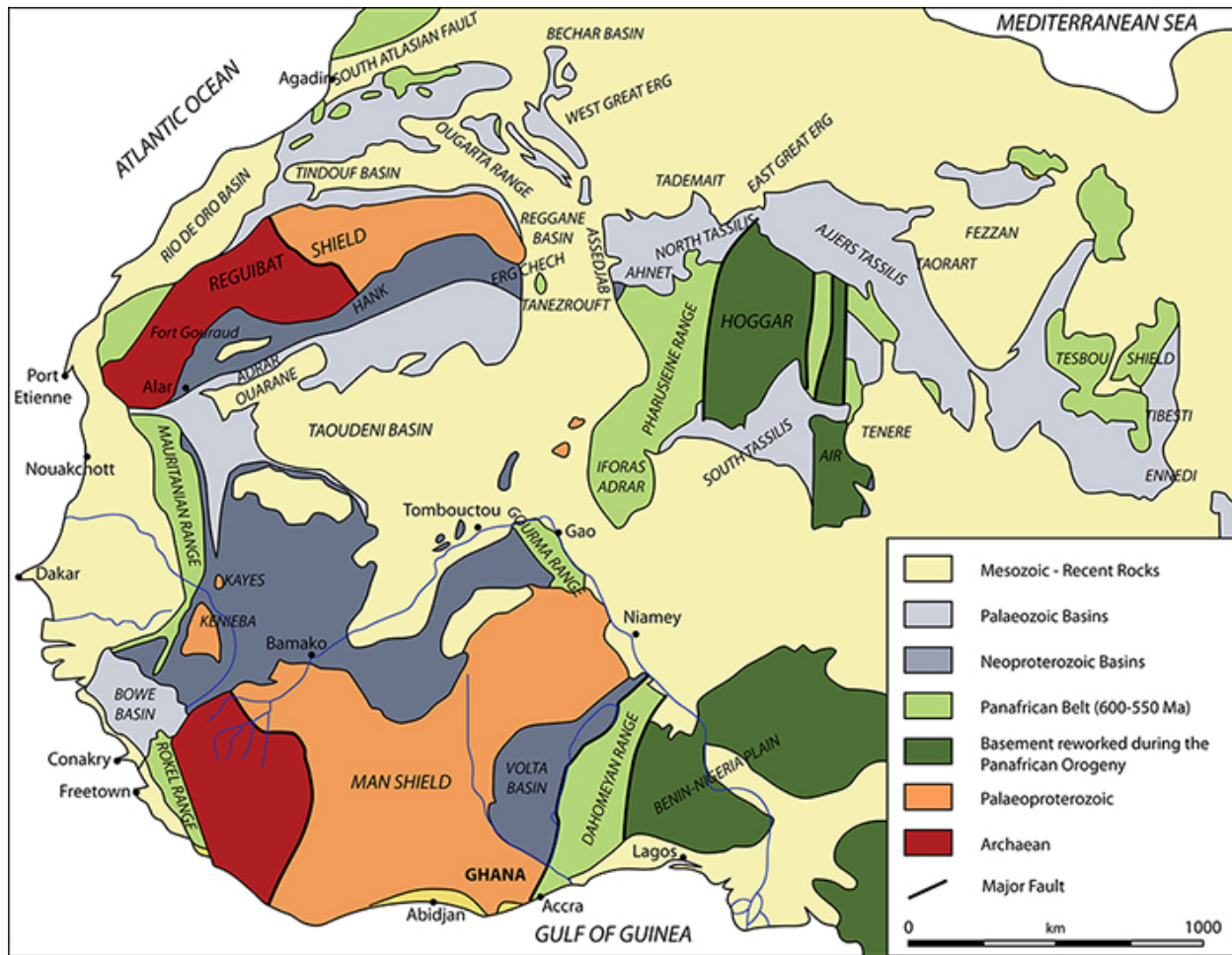


Figure 2.2: Simplified tectonic map of West Africa, modified after [Feybesse et al. \(2006\)](#). Economic gold mineralization is dominantly hosted within the Palaeoproterozoic ‘Birimian terrane’ of the Man Shield.

2.2.1 Birimian Supergroup

The Birimian Supergroup of Ghana, named for the Birim river in eastern Ghana, comprises belts of extrusive volcanics and associated volcanoclastic sediments, which are separated by clastic sedimentary basins. There has been strong debate for many years over the exact nature and stratigraphy of the Birimian, with ideas divided between anglophone and francophone workers. This has been exacerbated by a lack of both geochronological data and outcrop, precluding detailed field observations ([Hastings, 1982](#)). Early anglophone workers termed the sedimentary and volcanic parts the Lower and Upper Birimian respectively, implying that the sedimentary basins are older and the volcanics were erupted and deposited on top ([Junner, 1935, 1940](#)). In contrast, the stratigraphy accepted by the francophone workers generally used the opposite stratigraphy placing the volcanics prior to the sediments

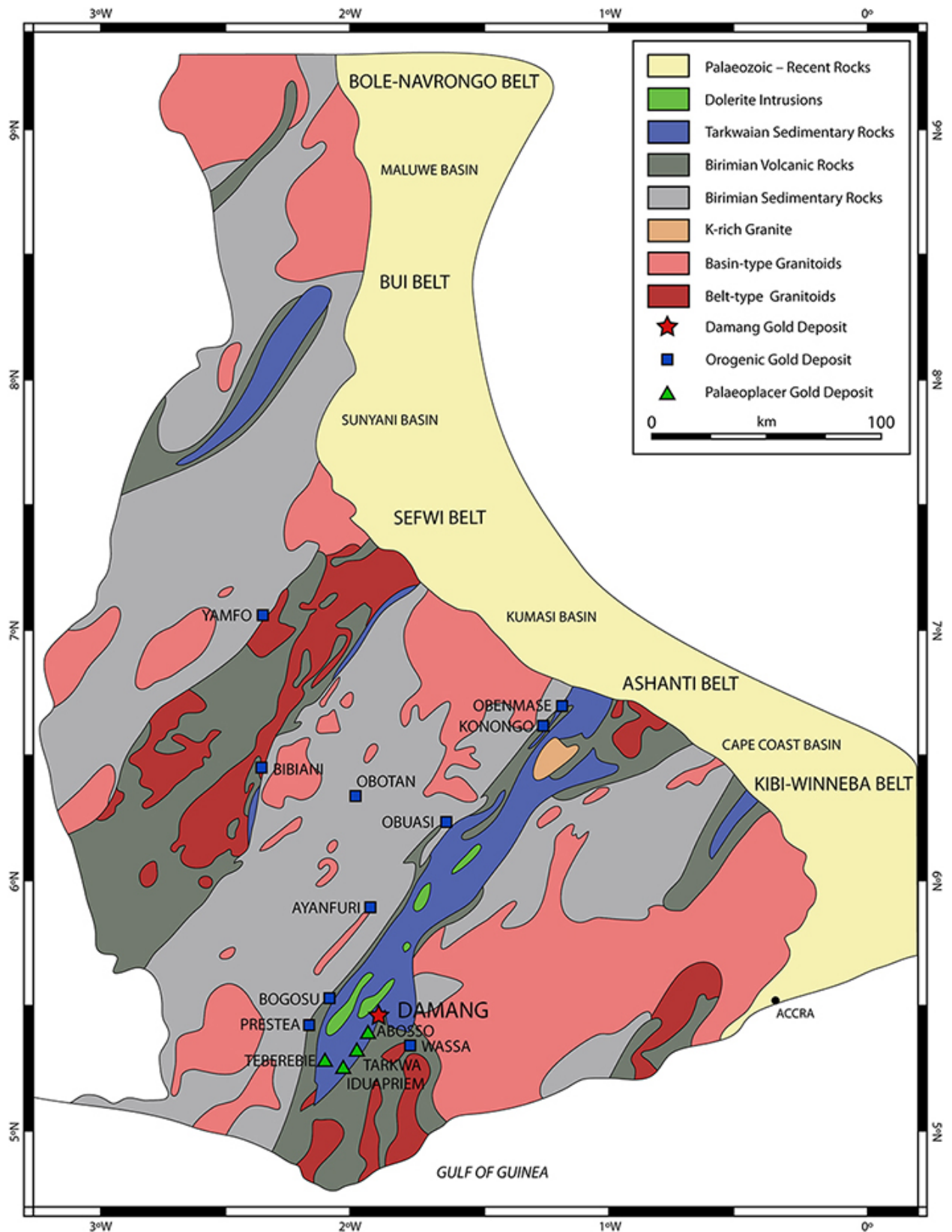


Figure 2.3: Simplified regional geological map of southwest Ghana, modified after Pigois *et al.* (2003). Orogenic gold deposits are principally located along the margins of the Birimian volcanic belts, while the palaeoplacer-type deposits occur within the overlying Tarkwaian.

(Milési *et al.*, 1991; Feybesse *et al.*, 2006). The disagreement was addressed by Leube *et al.* (1990) who suggested that the sedimentary and volcanic portions are in fact lateral facies equivalents that formed broadly contemporaneously. This stratigraphy is the one that is generally accepted today. The basement to the Birimian is not exposed, but the Birimian volcanics have high initial ϵ_{Nd} values of +2 to +4, suggesting their origin was during a significantly post-Archaean, juvenile mantle-derived magmatic event (Abouchami *et al.*, 1990; Boher *et al.*, 1992; Taylor *et al.*, 1992). However, Begg *et al.* (2009) have investigated lithospheric architecture through seismic tomography and interpret that much of West Africa is underlain by Archaean lithosphere.

Volcanic Belts

Six volcanic belts have been identified in Ghana. From south to north they are the Kibi-Winneba, Ashanti, Sefwi, Bui, Bole-Navrongo and Lawra belts, the first five of which are shown on figure 2.3. They are approximately evenly spaced and vary in width between 15–40 km. The average width tends to narrow northwards. Leube *et al.* (1990) interpreted this to be a function of erosion level. Detailed stratigraphy is unknown due to the deformation of the belts (see section 2.4), although total thicknesses are thought to be in the range of 4000 – 12,000 m (Sylvester & Atttoh, 1992). The ratio between pyroclastic and lava components varies between the belts, being highest in the Ashanti belt and lowest in the Sefwi belt, which could also be a function of erosion level. An additional ‘failed’ belt, the Asankangwra belt, occurs in the centre of the Kumasi basin, near the location of the Obotan gold deposit (Hirdes *et al.*, 1993). The majority of orogenic gold deposits in Ghana are hosted in the Birimian volcanics, typically along the edges of the belts (Fig. 2.3).

The belts are primarily composed of basic to intermediate volcanics with associated volcanoclastic deposits (Leube *et al.*, 1990). Lesser amounts of rhyolitic and dacitic lavas and tuffs and interflow sediments also occur throughout and there is a distinct lack of komatiites (Milési *et al.*, 1991; Eisenlohr, 1992). The volcanics show bimodal tholeiitic (dominant) and calc-alkaline (minor) affinities and pillow structures are occasionally preserved (Abouchami *et al.*, 1990; Leube *et al.*, 1990; Milési *et al.*, 1991). Geochemical characteristics of the

volcanics are intermediate between mid-ocean ridge basalt (MORB) and island arc tholeiite (IAT) giving rise to a number of different theories regarding formation. These include a rift setting (Leube *et al.*, 1990), oceanic flood basalts Abouchami *et al.* (1990), or related to a transition from a compressional to transcurrent tectonic style (Milési *et al.*, 1991). More recent detailed trace element geochemistry has highlighted two geochemical types, LREE-depleted tholeiitic basalts/basaltic andesites formed in a back-arc basin setting, and calc-alkaline andesites and basalts/basaltic/andesites that show a subduction zone signature and likely formed in an intra-oceanic arc setting (Dampare *et al.*, 2008).

Sedimentary Basins

The Birimian volcanic belts are separated by clastic sedimentary basins, named, in order from south to north, the Cape Coast, Kumasi, Sunyani and Maluwe basins (Fig. 2.3). Sediments in the basins comprise four distinct sedimentary facies that are lateral equivalents and occur as a function of distance from the volcanic belts (Leube *et al.*, 1990). Moving inwards from the basin edge towards the centre, these are a chemical sediment facies, volcanoclastic deposits, turbidite-related wackes and argillites, which are a conformable sequence with gradational transitions between them. Along the basin margins, belt volcanics and basin sediments are intercalated, interpreted as representing coeval formation of oceanic basins between a series of volcanic arcs (Leube *et al.*, 1990). The total thickness of Birimian sediments is not known, although it is thought to exceed 2000 m (Sylvester & Attah, 1992).

Rocks of the chemical facies comprise cherts, Fe-Mg carbonates, carbonaceous rocks, sulphide-rich rocks and manganiferous cherts. In places the Mn contents are high enough to have economic value, such as at the Nutse mine, a few kilometres south of Tarkwa. These sediments often play a crucial role in the formation of orogenic gold deposits in Ghana (see section 2.7.2).

Volcanoclastic sediments are the dominant component of the Birimian basins and comprise a mix of silt- to sand-sized bipyramidal quartz, idiomorphic plagioclase and chloritized glass fragments. These are interpreted as representing reworked pyroclastic and epiclastic deposits

([Leube *et al.*, 1990](#); [Hirdes *et al.*, 1993](#)). Compositionally, the volcanoclastics exhibit immobile trace element ratios indistinguishable that are from the Birimian belt volcanics, suggestive of a common source ([Leube *et al.*, 1990](#)).

The turbidite-related wackes become thicker and progressively more important with increasing distance from the volcanic belts. Major components include chert, graphitic schist and quartz with lesser basalt fragments, while there is a complete absence of granitic fragments ([Leube *et al.*, 1990](#); [Hirdes *et al.*, 1993](#)). The rocks commonly show well developed graded bedding and evidence of submarine slumping of soft sediment.

Argillites are common throughout the Birimian basins, often interbedded between coarser clastic sediments. The overall argillite/volcanoclastic ratio, however, increases with distance from the volcanic belts ([Leube *et al.*, 1990](#); [Hirdes *et al.*, 1993](#)). They typically comprise a mix of phyllites and graphitic schists, both of which have compositions broadly similar to the volcanoclastic sediments ([Leube *et al.*, 1990](#)). These have been interpreted as ‘background’ sedimentation of wind-blown ash and very-fine clastic basinal muds between the periodic influx of coarser, more volcanic dominated turbidites ([Hirdes *et al.*, 1993](#)).

2.2.2 Tarkwaian System

Rocks of the Tarkwaian System were first differentiated from the Birimian by [Whitelaw \(1929\)](#) who applied the term ‘Tarkwaian’ to them. The stratigraphy and classification was further refined by [Hirst \(1938\)](#) and [Junner *et al.* \(1942\)](#). Tarkwaian System sediments occur only in belt domains where they overlie the older Birimian volcanics. The Tarkwaian forms a major component of both the Ashanti and Bui belts and a lesser component of the Kibi-Winneba belt. In these cases, they predominantly occur towards the centre of the belts, implying an overriding, broadly synformal structure (Fig. [2.3](#)). In contrast, smaller occurrences of the Tarkwaian in the Sefwi belt occur towards the belt margins. There are also minor occurrences in the Bole-Navrongo belt. Sediments of the Tarkwaian System are the dominant hosts to gold mineralization at Damang.

The Tarkwaian System represents a broadly upwards fining sequence of clastic sediments interpreted to infill a tensional rift basin (Strogen, 1988; Ledru *et al.*, 1994). From the base, the constituent units are the Kawere Group, Banket Series, Tarkwa Phyllite and Huni Sandstone. Geomagnetic evidence suggests the presence of a second phyllite unit within the Huni Sandstone of the Tarkwa-Damang region, known locally as the Dompim Phyllite (Abosso Goldfields Ltd., *pers. comm.*).

The Kawere Group comprises coarse-grained, poorly-sorted, polymictic conglomerates, interbedded with grits, sandstones and very minor argillites (Junner *et al.*, 1942; Strogen, 1988; Brabham, 1998). Total thickness of the Kawere Group varies from 250–700 m. The group gradually fines upwards as the conglomerate component decreases. Principal components of the conglomerate units are intermediate to granitic clasts, which are likely derived from the underlying Birimian volcanics. Lenticular and channel shaped gravel beds, graded-bedding and cross-bedding, which indicate a fluvial origin, are abundant with a palaeocurrent direction from the east (Strogen, 1988).

The Banket Series is primarily a sequence of medium- to coarse-grained sandstones and quartzites with minor grits, argillites and conglomerate horizons. The Banket Series ranges between 150–600 m and generally thickens towards the southwest, again indicating a palaeocurrent direction from the east (Sestini, 1973; Strogen, 1988). Prominent cross-bedding transitions up section to planar-bedding, indicating a progressively deepening water environment. A characteristic feature of the Banket Series is the presence of heavy mineral concentrations, including haematite, magnetite, rutile, ilmenite and zircon, along bedding planes and cross-bed foresets (Hirdes & Nunoo, 1994; Brabham, 1998; Pigois, 2003). Conglomerate horizons in the Banket Series are of particular importance since these host the palaeoplacer gold deposits of the Ashanti belt. The conglomerates are mostly composed of large, well-sorted, rounded vein quartz pebbles (90%) set in a pelitic matrix and accompanied by lesser amounts of volcanic and schistose clasts, all of which are ultimately derived from the Birimian (Sestini, 1973; Brabham, 1998).

The Tarkwa Phyllite is a series of greenish, finely laminated argillites and lesser sandstones

dominated by an assemblage of chlorite, sericite, chloritoid, magnetite and tourmaline (Junner *et al.*, 1942). It has a fairly uniform thickness between 120–150 m and represents deposition in a deeper water environment (Strogen, 1988; Hirdes & Nunoo, 1994).

The Huni Sandstone is the thickest unit of the Tarkwaian System being at least 1300 m thick. It is principally a uniform, weakly-bedded to massive, fine-grained feldspathic sandstone and represents a transition to a shallower, and occasionally fluviially influenced, environment (Strogen, 1988).

2.2.3 Granitoids

Granitoid intrusions form a major component of the Birimian terrane across Ghana and the rest of West Africa (Fig. 2.3). There are four distinct series in Ghana, the dominant Belt- and Basin-type granitoids and the minor Winneba and K-rich granites. Each of these are discussed below.

Belt-type Granitoids

The belt-type granitoids, as the name suggests, intrude the Birimian volcanic belts and were formerly known as Dixcove-type granites. They range in composition from hornblende- and biotite-bearing granites to diorites, monzogranites, syenites and even tonalites and trondhjemites (Leube *et al.*, 1990; Hirdes *et al.*, 1992; Eisenlohr & Hirdes, 1992). Individual intrusions often form composites comprising several different granitoid types. They are typically unfoliated and cross-cut regional structures. The lack of fabric development does not necessarily indicate post-deformational emplacement since the constituent mineral assemblage would not readily develop a fabric (Eisenlohr & Hirdes, 1992). Megacrysts of oligoclase–andesine plagioclase are common, are typically heavily saussuritized or sericitized and often cross-cut by actinolite. This suggests that the granitoids have been subjected to regional metamorphism and/or alteration (Eisenlohr & Hirdes, 1992). No contact metamorphic aureoles are observed, suggesting that the granitoids were either emplaced into already hot country rocks, or that any aureole has been masked by subsequent metamorphism.

Xenoliths of basalt are common and in places there is a gradational transition between coarse-grained granitoid and basalt. [Eisenlohr & Hirdes \(1992\)](#) use this to suggest a close genetic relationship between the belt-type granitoids and the Birimian volcanics during a common magmatic event. Indeed, there is a substantial body evidence that mafic enclaves and dykes in granitoid plutons are genetically related to their assembly and growth through progressive recharge of mafic, mantle-derived magma (e.g. [Reid *et al.*, 1983](#); [Hill *et al.*, 1985](#); [Didier, 1987](#); [Hill, 1988](#); [Dorais *et al.*, 1990](#))

Basin-type Granitoids

The Basin-type granitoids, formerly known as Cape Coast type, intrude the Birimian sedimentary basins as large batholiths. They are predominantly two-mica granitoids with lesser biotite-only or hornblende-bearing types. They are typically granodioritic in composition but do range from granite to monzogranite or tonalite ([Leube *et al.*, 1990](#); [Eisenlohr & Hirdes, 1992](#)). The granitoids are extensively foliated with fabrics defined by the alignment of micas, and many show compositional banding between biotite-rich and biotite-poor layers ([Eisenlohr & Hirdes, 1992](#)). Grain boundaries are smooth and commonly form triple junctions, although other evidence for quartz recrystallization or undulose extinction are rare, suggesting that the fabric was developed at high temperature during emplacement synchronous with regional deformation ([Eisenlohr & Hirdes, 1992](#)). Extensive contact metamorphic aureoles form around the intrusions with the development of mineral assemblages including garnet, staurolite, kyanite and sillimanite ([Junner, 1935](#); [Woodfield, 1966](#)). A summary comparison with the belt-type granitoids is presented in table [2.1](#).

Winneba Granite

The Winneba granite has only been recorded at one occurrence near the town of Winneba in southern Ghana. It is a porphyritic microcline-biotite monzogranite, is cut by a biotite granodiorite and is commonly foliated. The most interesting feature, however, is that it produces a Sm/Nd model age of around 2.6 Ga, which serves as the only record of an

Table 2.1: Characteristics of belt- and basin-type granitoids, after [Hirdes *et al.* \(1992\)](#)

Belt-type Granitoids	Basin-type Granitoids
Small to medium-sized plutons intruding volcanic belts	Large batholiths intruding Birimian basins
Contact aureoles limited to a few tens of metres	Extensive contact-metamorphic aureoles
Rare foliation and compositional banding	Well-developed foliation and compositional banding
Hornblende as dominant ferromagnesian mineral	Biotite as dominant ferromagnesian mineral
Metaluminous	Peraluminous
Typically dioritic to granodioritic but range from granite to diorite	Typically granodioritic but range from granite to tonalite
$\text{Na}_2\text{O}, \text{CaO}_{\text{belt}} > \text{Na}_2\text{O}, \text{CaO}_{\text{basin}}$	$\text{Rb}, \text{K}_2\text{O}_{\text{basin}} > \text{Rb}, \text{K}_2\text{O}_{\text{belt}}$
Extensive retrogression	Little alteration
Some geochemical similarity to Birimian belt volcanics	No evidence for geochemical similarity to Birimian belt volcanics

Archaean crustal component in Ghana ([Taylor *et al.*, 1992](#)).

K-rich Granites

Two outcrops of alkalic K-rich granite occur in Ghana, the Bongo granite in the Bole-Navrongo belt and the Banso granite in the northern Ashanti belt. These granites were formerly known collectively as Bongo granites ([Leube *et al.*, 1990](#)). They intrude, and therefore post-date, the Tarkwaian System sediments and are characterized by phenocrysts of pink K-feldspar.

2.2.4 Post-Tarkwaian Rocks

Three sets of post-Tarkwaian rocks are present in the Palaeoproterozoic of Ghana. One, the K-rich granites, have been discussed in the preceding section. The other two are separate series of doleritic intrusions. The earliest of these intruded the Tarkwaian as sills and subordinate dykes prior to the majority of regional deformation (Junner *et al.*, 1942). These are particularly well developed in the Ashanti belt where they mainly exploit the boundary between the Tarkwa Phyllite and the Huni Sandstone. The sills are typically 10-60 m thick, although they can be thicker, and are primarily basic in composition.

The final set of post-Tarkwaian dolerites form very long, continuous, north – south trending dykes that cross-cut Birimian and Tarkwaian structures and post-date regional metamorphism. These are extremely well defined in regional aeromagnetic maps (Fig. 2.4). Woodfield (1966) described them as unmetamorphosed, coarse-grained, ophitic olivine dolerites.

2.3 Metamorphism

Regional metamorphism of the Birimian in Ghana is typically low grade. Chl $\pm$ Bt assemblages in pelites and Chl + Act + Ep + Ab assemblages in metabasites indicate prehnite-pumpellyite to broadly greenschist facies conditions, with temperatures barely exceeding 400°C. In aureoles to large granitoid plutons, Grt + Ky assemblages indicate localized amphibolite facies conditions, with temperatures up to 500°C (Eisenlohr, 1989; Leube *et al.*, 1990; Taylor *et al.*, 1992; Hirdes *et al.*, 1992; Milési *et al.*, 1992; Eisenlohr & Hirdes, 1992; Ledru *et al.*, 1994; Mumin & Fleet, 1995). There are, however, recorded occurrences of higher grade rocks, even including partial anatexis in southeast Ghana (Milési *et al.*, 1991, 1992).

More recent, dedicated metamorphic studies of Birimian rocks from around Ghana have highlighted further, more widespread occurrences of amphibolite facies rocks, with metamorphic conditions of 500–600°C and 4-6 kbar, in the Ashanti, Kibi-Winneba and Sefwi belts (John *et al.*, 1999; Klemd *et al.*, 2002; Galipp *et al.*, 2003). John *et al.* (1999) even suggested that

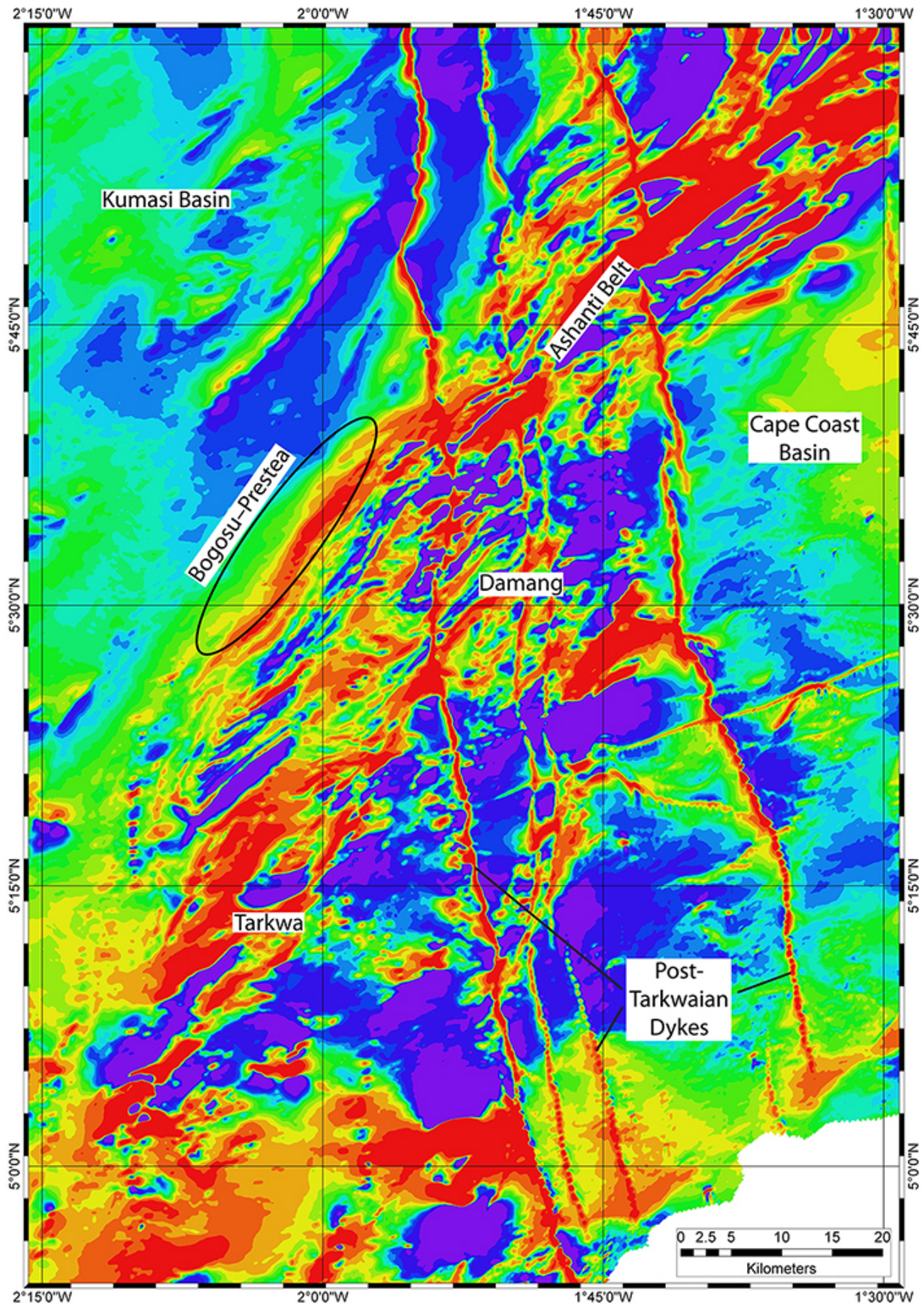


Figure 2.4: Aeromagnetic (reduced to pole) map of the southern Ashanti belt showing important lithological and structural features along with the location of key gold deposits. Data provided by Gold Fields Ltd. Colour scale is qualitative from blue (low) to red (high).

the whole of the Ashanti belt had experienced amphibolite facies metamorphism followed by a greenschist facies overprint, although this has not been verified outside of their study area in the southern most Ashanti belt.

Metamorphism of the Tarkwaian is significantly less well studied due to the lack of ferromagnesian minerals and other reliable facies indicators. The Tarkwaian is often casually described as ‘very low grade’, while petrographic studies suggest conditions up to greenschist facies (Hirdes *et al.*, 1992; Klemd *et al.*, 1993; Ledru *et al.*, 1994; Bossiere *et al.*, 1996). A more reliable indicator of metamorphic grade comes from the mafic sills and dykes that intrude the Tarkwaian of the Ashanti belt and typically comprise assemblages of Chl + Act + Ep, suggesting greenschist facies conditions similar to the Birimian (Brabham, 1998). Once again, local occurrences of amphibolite facies assemblages in aureoles surrounding igneous intrusions have been reported (Ledru *et al.*, 1994; Pigois, 2003). The only reported occurrence of amphibolite facies metamorphism of the Tarkwaian that is not directly associated with a granite aureole is at Damang (Brabham, 1998; Pigois, 2003).

2.4 Structure

Structural deformation and fabric development occurred in the Birimian and Tarkwaian rocks during the Eburnean orogeny between approximately 2130 – 1980 Ma (Abouchami *et al.*, 1990; Boher *et al.*, 1992; Eisenlohr & Hirdes, 1992; Feybesse & Milési, 1994; Alibone *et al.*, 2002; Hirdes & Davis, 2002; Feybesse *et al.*, 2006; Harcouët *et al.*, 2007). Structural features in the Birimian and Tarkwaian record either a low-strain or high-strain type (Eisenlohr & Hirdes, 1992). Low-strain deformation is dominant and manifested as a northeast-trending, sub-vertical foliation that is sub-parallel to bedding producing a sub-horizontal foliation–bedding intersection lineation. It is present in both Birimian and Tarkwaian rocks and represents a primarily coaxial pure shear deformation event. Conversely, the high strain deformation is partitioned along the margins of the volcanic belts where a northeast – southwest-trending penetrative foliation and a southwest-trending stretching lineation developed in response to a non-coaxial simple shear component (Eisenlohr & Hirdes,

1992). Further to this general scheme, individual lithological regions develop their own more detailed structural paragenesis.

Birimian Volcanic Belts

The structure of the Birimian belts is somewhat variable between the different belts. Within the Bui, Ashanti and Kibi-Winneba belts, Tarkwaian sediments occur towards the centre of the belts, suggesting a large-scale synformal structure. Whereas in the Sefwi belt, the Tarkwaian lies towards the margins, suggesting a broadly antiformal structure. Bedding is generally sub-vertical or dips steeply to the northwest. Overturned beds are common, particularly towards the northwestern margins of the belts (Eisenlohr & Hirdes, 1992). Along the northwestern margin of the Ashanti belt, Birimian volcanics are thrust over the younger Tarkwaian sediments (Junner, 1935; Leube *et al.*, 1990). Outliers of Birimian volcanics in the Kumasi basin near Obuasi, along with alternating occurrences of Tarkwaian sediments and Birimian volcanics in the southern Ashanti belt indicates that the Ashanti belt is strongly folded, with a possible southeast vergence (Leube *et al.*, 1990). A dominant foliation forms an axial planar cleavage to these folds, while a second weaker cleavage forms at an oblique angle (Leube *et al.*, 1990).

Birimian Sedimentary Basins

Birimian sediments are typically rather poorly exposed and the lack of distinctive marker beds makes determination of large-scale structure difficult (Eisenlohr & Hirdes, 1992). However, on a local scale, beds dip steeply and strike northeast – southwest. Rocks form tight to isoclinal folds with a sub-horizontal, northeast trending axes and a well-defined, sub-vertical cleavage. Systematic changes in fold vergence are rarely recorded, suggesting a lack of large-scale folding (Eisenlohr & Hirdes, 1992). Overall, deformation increases towards the margins of neighbouring volcanic belts.

Tarkwaian Sediments

The Tarkwaian sediments are generally less extensively deformed than the Birimian (Leube *et al.*, 1990) suggesting two alternatives: the Birimian was subjected to a period of deformation prior to erosion and deposition of the Tarkwaian during an earlier orogenic event or stage of orogenesis, or changing structural style as a function of rock competency, with the clastic, quartzite-rich Tarkwaian being more resistant to tight folding. Within the Bui belt, for example, the Tarkwaian sediments fill a broadly synclinal trough with relatively little internal deformation and rocks near the margins gently dipping in towards the centre (Leube *et al.*, 1990). In contrast, the Tarkwaian of the Ashanti belt forms a series of open folds, which locally intensifies towards the northwest margin of the belt.

Granitoids

As discussed above, the belt and basin-type granitoids show differing styles of structural deformation, which is primarily thought to be related to the timing of their emplacement, relative to regional deformation (Eisenlohr & Hirdes, 1992).

2.5 Geochronology

The Birimian terrane has been subject to a number of geochronological studies, particularly during the late 1980s and early 1990s. Early attempts at determining ages for the Birimian volcanics and various granitoid intrusions used whole rock Rb/Sr, Sm/Nd or Pb/Pb isochron methods (Vachette & Yace, 1969; Bassot & Caen-Vachette, 1984, Lenz in Hirdes *et al.*, 1987; Abouchami *et al.*, 1990; Taylor *et al.*, 1992) and were consequently rather poorly constrained. By 1992, more precise U/Pb dating of zircons, and similar minerals, was being employed. These (1992 and later) dates are given in table 2.2, plotted, where possible, on figure 2.5 and summarized in figure 2.6.

The first estimate of the timing of Birimian volcanism in Ghana was the Sm/Nd whole rock

Table 2.2: Summary table of geochronology (1992 and later) from across Ghana.

Age (Ma)	Method	Constrains	Reference
Birimian Volcanics			
2158 $\pm$ 5	U/Pb zircon	Formation	Feybesse <i>et al.</i>, 2006
2189 $\pm$ 1	U/Pb zircon	Formation	Hirdes & Davis, 1998
1978 $\pm$ 37	Ar/Ar hornblende	500°C cooling	Feybesse <i>et al.</i>, 2006
2006 $\pm$ 40	Ar/Ar hornblende	500°C cooling	Feybesse <i>et al.</i>, 2006
Birimian Sediments			
2135 $\pm$ 5	U/Pb detrital zircon	Sedimentation	Davis <i>et al.</i>, 1994
Belt Granitoids			
2172 $\pm$ 7	U/Pb zircon	Formation	Brownscombe, 2009
2151 $\pm$ 7	U/Pb zircon	Formation	Brownscombe, 2009
2200 $\pm$ 4	U/Pb zircon	Formation	Feybesse <i>et al.</i>, 2006
2161 $\pm$ 4	U/Pb zircon	Formation	Feybesse <i>et al.</i>, 2006
2159 $\pm$ 4	U/Pb zircon	Formation	Feybesse <i>et al.</i>, 2006
2137 $\pm$ 10	U/Pb zircon	Formation	Yao & Robb, 1999
2173.9 $\pm$ 1.5	U/Pb zircon	Formation	Oberthür <i>et al.</i>, 1998
2179 $\pm$ 2	U/Pb zircon	Formation	Hirdes <i>et al.</i>, 1992
2172 $\pm$ 2	U/Pb zircon	Formation	Hirdes <i>et al.</i>, 1992
2092 $\pm$ 4	U/Pb titanite	Metamorphism	Oberthür <i>et al.</i>, 1998
Basin Granitoids			
2113 $\pm$ 13	U/Pb zircon	Formation	Brownscombe, 2009
2107 $\pm$ 9	U/Pb zircon	Formation	Brownscombe, 2009
2105 $\pm$ 3	U/Pb zircon	Formation	Oberthür <i>et al.</i>, 1998
2106.3 $\pm$ 1.7	U/Pb zircon	Formation	Oberthür <i>et al.</i>, 1998
2090 $\pm$ 1	U/Pb zircon	Formation	Davis <i>et al.</i>, 1994
2116 $\pm$ 2	U/Pb zircon	Formation	Hirdes <i>et al.</i>, 1992
2088 $\pm$ 1	U/Pb zircon	Formation	Hirdes <i>et al.</i>, 1992
2105.4 $\pm$ 4.7	Pb/Pb titanite	Metamorphism	Oberthür <i>et al.</i>, 1998
2086 $\pm$ 4	Pb/Pb titanite	Metamorphism	Oberthür <i>et al.</i>, 1998
2098 $\pm$ 7	Pb/Pb titanite	Metamorphism	Oberthür <i>et al.</i>, 1998
2103.9 $\pm$ 1.7	U/Pb monazite	Metamorphism	Oberthür <i>et al.</i>, 1998
K-rich Granitoids			
2096.8 $\pm$ 1.9	Pb/Pb titanite	Formation	Oberthür <i>et al.</i>, 1998
Tarkwaian Sediments			
2133 $\pm$ 4	U/Pb detrital zircon	Sedimentation	Pigois, 2003
2154.5 $\pm$ 2.1	U/Pb detrital zircon	Sedimentation	Oberthür <i>et al.</i>, 1998
2132 $\pm$ 3	U/Pb detrital zircon	Sedimentation	Davis <i>et al.</i>, 1994
2029 $\pm$ 4	Ar/Ar biotite	300°C cooling	Pigois, 2003
2034 $\pm$ 4	Ar/Ar biotite	300°C cooling	Pigois, 2003
2063 $\pm$ 9	U/Pb xenotime	Mineralization	Pigois, 2003

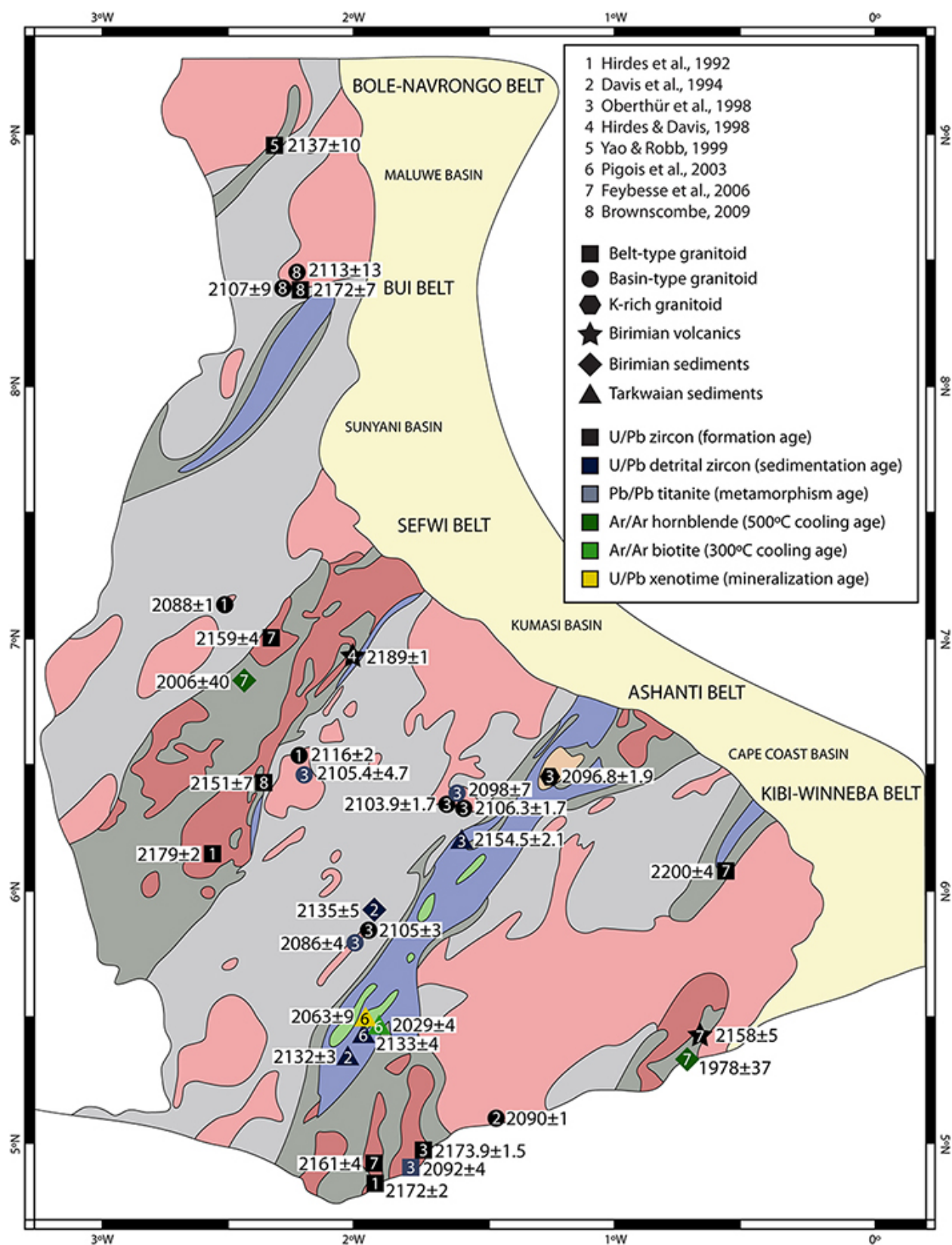


Figure 2.5: Simplified regional geological map of SW Ghana as per figure 2.3, plotted with reliable, 1992 and later, geochronological data.

isochron age of 2166 ± 66 Ma, presented by [Taylor *et al.* \(1992\)](#), which combined samples from a number of different belts. More recent U/Pb zircon ages provide more reliable estimates of 2189 ± 1 Ma in the Sefwi belt ([Hirdes & Davis, 1998](#)) and 2158 ± 5 Ma in the Kibi-Winneba belt ([Feybesse *et al.*, 2006](#)).

A best estimate for the maximum age of Birimian sedimentation in Ghana is 2135 ± 5 Ma, based on the detrital zircon study of [Davis *et al.* \(1994\)](#). In the Ivory Coast however, ages as young as 2098 ± 11 Ma have been determined ([Boher *et al.*, 1992](#)). The minimum age of sedimentation is constrained by the intruding Basin-type granitoids (see below). Within the constraints of the limited data, this age range does not overlap with that for the Birimian volcanics. However, as described in section 2.2.1, the two groups are interpreted to have been deposited broadly coevally.

Birimian volcanic belts were intruded by the Belt-type granitoids between 2137 ± 10 Ma and 2200 ± 4 Ma ([Hirdes *et al.*, 1992](#); [Oberthür *et al.*, 1998](#); [Yao & Robb, 1999](#); [Feybesse *et al.*, 2006](#); [Brownscombe, 2009](#)), with the majority falling between 2172 – 2179 Ma. This indicates overlapping formation with the Birimian mafic volcanism, possibly suggesting a comagmatic origin. Younger ages have been recorded from equivalents in the Ivory Coast (2152 ± 2 Ma and 2104 ± 2 Ma) and Senegal (2080 ± 1 Ma) ([Hirdes *et al.*, 1992](#); [Hirdes & Davis, 2002](#)). Peak metamorphism of the Belt-type granitoids in Ghana is estimated to have occurred at 2092 ± 2 Ma based on a U/Pb titanite age provided by [Oberthür *et al.* \(1998\)](#). [Yao & Robb \(1999\)](#) noted an approximate younging of ages from the southeast towards the northwest, which they interpreted as representing progressive accretion of volcanic belts.

The Basin-type granitoids are up to 90 Ma younger than the Belt-type granitoids, with intrusion ages between 2088 ± 1 and 2116 ± 2 Ma, and are likely related to crustal thickening and melting ([Hirdes *et al.*, 1992](#); [Davis *et al.*, 1994](#); [Oberthür *et al.*, 1998](#); [Brownscombe, 2009](#)). Metamorphism is likely at a similar time to the Belt-type granitoids, with a U/Pb titanite age of 2086 ± 4 Ma ([Oberthür *et al.*, 1998](#)).

Whole rock Pb/Pb and Rb/Sr isochrons constrain the age of intrusion of the Winneba granite rather poorly to 2173^{+107}_{-115} Ma and 2024 ± 159 Ma respectively ([Taylor *et al.*, 1992](#)).

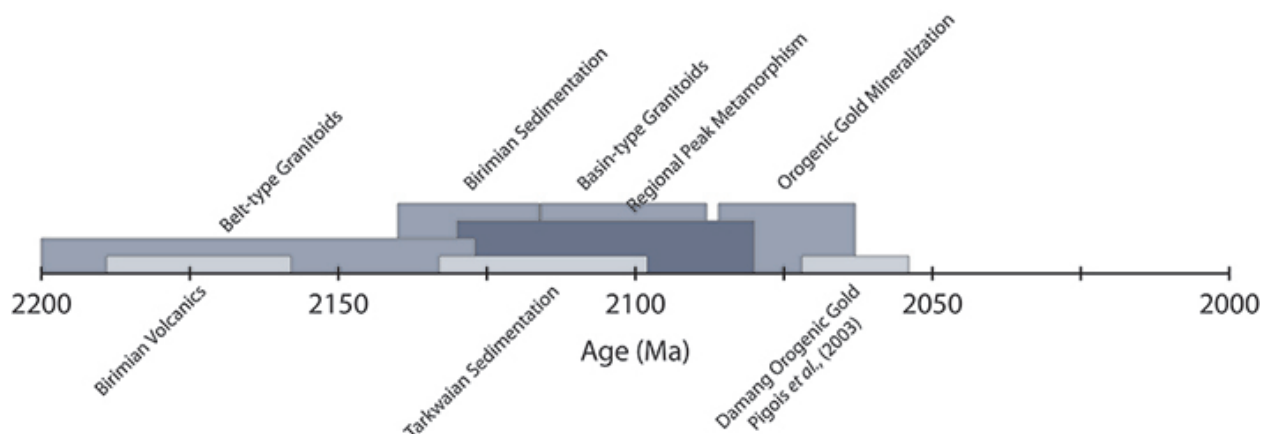


Figure 2.6: Summary of existing geochronology data in Ghana.

Intriguingly, these samples produce Sm/Nd model ages of 2.6 and 2.59 Ga, which indicates a crustal contribution at least as old as this, thereby providing the only direct evidence for an Archaean basement in Ghana. The K-rich Bansa granite in the northern Ashanti belt has a Pb/Pb titanite age of 2097 ± 2 Ma and provides the best estimate of formation age for this series of granitoids (Oberthür *et al.*, 1998).

The maximum age of Tarkwaian sedimentation has been determined through the detrital zircon studies of Davis *et al.* (1994) and Pigois (2003) in the southern Ashanti belt. Both studies are consistent, giving ages of 2132 ± 3 and 2133 ± 4 Ma respectively. A minimum age constraint on sedimentation is provided by the intrusion of the Bansa granite at 2097 ± 2 Ma. This age range is similar to that for the Birimian sediments and younger than the Birimian volcanics.

Estimates of the timing of orogenic gold mineralization are mostly determined indirectly, based on metamorphic/hydrothermal minerals, such as rutile (e.g. 2086 ± 4 Ma, Oberthür *et al.* (1998)), xenotime (2063 ± 9 Ma, Pigois *et al.* (2003)), or biotite (2029 ± 4 Ma, Pigois *et al.* (2003)). These are the youngest ages determined in the Birimian terrane of Ghana, post-dating all lithological units as well as being marginally younger than the best estimate for regional peak metamorphism at 2092 ± 2 Ma (Oberthür *et al.*, 1998).

2.6 The Eburnean Orogeny in the Context of the Birimian

The Eburnean orogeny is the period of tectonic activity that led to the metamorphism and structural deformation described above (Leube *et al.*, 1990; Eisenlohr & Hirdes, 1992; Milési *et al.*, 1992; Davis *et al.*, 1994; Feybesse & Milési, 1994; Ledru *et al.*, 1994; Allibone *et al.*, 2002; Hirdes & Davis, 2002; Feybesse *et al.*, 2006). One of the earliest models for the geodynamic setting of the Birimian terrane during the Eburnean orogeny suggested two separate phases of compressional deformation, with Tarkwaian sedimentation occurring between them (Leube *et al.*, 1990). A similar two stage evolution was described by Allibone *et al.* (2002). The earlier event in these cases may correlate with the older Tangaeen orogeny of Burkina Faso (Hein, 2010).

This model has now generally been superseded by others that invoke a single progressive deformational event during regional northwest – southeast compression. In these models, the degree of deformation increases continuously over time, thereby explaining the additional deformation experienced by the Birimian compared to the Tarkwaian (Eisenlohr & Hirdes, 1992; Feybesse & Milési, 1994). There is, however, strong geochronological evidence for diachronous deformation across the Man Shield, with deformation progressing from the southeast to the northwest (Davis *et al.*, 1994; Feybesse & Milési, 1994; Hirdes & Davis, 2002). The main compressional event was subsequently followed by a period of major transcurrent tectonism, particularly along the margins of the volcanic belts, creating the shear zones that host orogenic gold mineralization in Ghana (Milési *et al.*, 1992; Feybesse *et al.*, 2006).

Although the Tarkwaian System is typically considered to have been deposited in a rifted basin, Ledru *et al.* (1994) described this as a rifted foreland basin that was subsequently overthrust and buried as part of the same continued orogenic event, as evidenced by similar, although less intense, structural styles.

2.7 Gold Mineralization in Ghana

The Palaeoproterozoic terranes of the West African Craton have been exploited for gold and other mineral resources for hundreds of years. Today, they form one of the richest gold provinces in the world. Indeed, Ghana was previously known as the Gold Coast and, with an annual production of around 2.7 Moz (Brown *et al.*, 2009), is currently the second largest producer of gold in Africa, after South Africa, and ninth largest in the world.

Gold mineralization in southwest Ghana is typically of two contrasting styles (Leube *et al.*, 1990; Milési *et al.*, 1991; Mumin *et al.*, 1994; Oberthür *et al.*, 1997; Goldfarb *et al.*, 2001; Allibone *et al.*, 2002; Wille & Klemm, 2004) (Fig. 2.3):

1. Palaeoplacer-type sedimentary gold deposits hosted in Banket Series conglomerates of the Tarkwaian System (e.g. Tarkwa, Iduapriem, Teberebie), and
2. Orogenic (hydrothermal) gold deposits hosted in shear zones along the margins of the Birimian volcanic belts (e.g. Prestea, Bogosu, Obuasi)

These deposit styles are discussed in more detail below.

Additional minor gold mineralization styles of increasing economic importance have been documented across Ghana. Examples include the granite-associated hydrothermal deposit at Ayanfuri (Yao *et al.*, 2001; Allibone *et al.*, 2004; Brownscombe, 2009). These typically occur within the volcanic belts and are thought to represent the same regional mineralization event as the shear zone-hosted deposits.

2.7.1 Palaeoplacer Gold Deposits

Palaeoplacer-type gold deposits occur within the Tarkwaian System sediments of the southern Ashanti belt. Gold is hosted in coarse quartz-pebble conglomerates of the Banket Series. These stratiform conglomerate horizons are locally termed ‘reefs’, in concordance with terminology applied to the Witwatersrand palaeoplacer deposits in South Africa. Sedimentary

structures within the Banket Series infer deposition in a series of coalescing alluvial fans and braided streams with the conglomerates representing tectonic interruptions ([Sestini, 1973](#); [Strogen, 1988](#); [Brabham, 1998](#)).

In the conglomerates, large, rounded, well-sorted pebbles of white vein quartz predominate. In general, the horizons increase in thickness and number towards the southwest, from Damang towards Tarkwa. Heavy mineral concentrations are common along cross-bedding fore sets, comprising haematite, magnetite, rutile, ilmenite, zircon and gold. Underground mining operations, such as the now abandoned Abosso mine, commonly recovered gold grades of over 10 g/t, while more recent open pit operations can exploit lower gold grades below 1 g/t.

The distribution of gold within the conglomerates is strongly related to sedimentary structures ([Sestini, 1973](#)), although [Bray \(1928\)](#) and [Bergene \(1944\)](#) suggest considerable reworking and remobilization of gold and other hydrothermal minerals has occurred. This remobilization led some early workers to suggest a purely hydrothermal origin ([Bray, 1928](#)), although this has now been broadly discounted. Instead a description as a ‘modified palaeoplacer’ is more appropriate ([Milési *et al.*, 1991](#)). Gold occurs as free gold amongst the pebbles and is rarely contained within the quartz pebbles themselves.

The source of gold for the Tarkwaian palaeoplacers was originally thought to be the Birimian shear zone-hosted gold ([Junner *et al.*, 1942](#); [Kesse, 1984](#); [Hirdes *et al.*, 1988](#)). Later structural analysis, however, highlighted that Birimian orogenic gold deposition post-dates deformation that affected the Tarkwaian palaeoplacers. This precludes the Birimian-hosted gold being a source ([Milési *et al.*, 1991](#); [Eisenlohr, 1992](#)). Analysis of fluid inclusions in the conglomerate vein-quartz pebbles highlight the dominance of CO₂–N₂-rich inclusions. These are very similar to those involved in orogenic gold deposit formation and imply that the Tarkwaian palaeoplacers were derived from a pre-Eburnean orogenic gold province ([Klemm *et al.*, 1993](#)). Furthermore, since the source of the Tarkwaian sediments was the deforming Birimian volcanic arcs (see sections [2.2.2](#) and [2.4](#)), these must also be the ultimate source of Tarkwaian gold. However, despite early inaccuracies, no pre-Tarkwaian orogenic gold

deposits have been found in the Birimian of Ghana.

2.7.2 Orogenic Gold Deposits

Orogenic gold deposits in Ghana are generally hosted by large-scale, sub-vertical graphitic shear zones along the margins of the Birimian volcanic belts. In these shear zones, gold is contained within steeply dipping quartz veins and in massive disseminated sulphide deposits (Oberthür *et al.*, 1997; Allibone *et al.*, 2002). Despite being Palaeoproterozoic in age, these deposits share a number of characteristics with Archaean greenstone-hosted gold deposits, such as those of the Superior craton of Canada and the Yilgarn craton of Western Australia (Goldfarb *et al.*, 2001, 2005; Dubé & Gosselin, 2007). The Birimian deposits have also been categorised as turbidite-hosted deposits, due to the high proportion of volcanoclastic and sedimentary material in the host rocks, similar to the Lachlan fold belt of southeast Australia, the Otago schist belt of New Zealand and the Slave Province of Canada (Berge, 2011).

Within Ghana, orogenic gold deposits of varying sizes are known from all Birimian belts. The largest and greatest number of deposits occur along the northwestern margin of the Ashanti belt, including the Bogosu–Prestea gold field and the giant Ashanti deposit at Obuasi (containing in excess of 1700 t of Au (Allibone *et al.*, 2002)) (Fig. 2.3). Other major deposits occur in the Sefwi belt, such as the Bibiani and Yamfo deposits. Major deposits like these often correlate with structural features. For example, where the regional-scale hosting shear zones exhibit left-hand dilational jogs and/or intersect minor cross-cutting faults (Allibone *et al.*, 2002). Indeed, these minor faults commonly bound mineralized zones or focus mineralization at high-angle intersections with bedding (Allibone *et al.*, 2002). Large-scale shear zones that host gold mineralization are clearly identified from regional geophysics, which have commonly been used during early-stage exploration efforts. The Bogosu–Prestea goldfield, for example, coincides with a steep geomagnetic gradient along the edge of the Ashanti belt (Fig. 2.4).

Gold shows a bimodal distribution, occurring in two distinct mineralogical settings that show

a close spatial association, suggestive of a genetic link:

1. Quartz vein-hosted gold mineralization.

Massive or laminated quartz veins host free-milling gold. The veins are generally 0.2–5 m wide and occasionally form multiple sets intercalated with highly sulphidized wall rock ([Appiah, 1991](#); [Oberthür *et al.*, 1996, 1997](#); [Allibone *et al.*, 2002](#)). Pyrite and arsenopyrite are common, as are small amounts of Cu-, Sb-, Pb-, Zn-sulphides. Ore bodies and hosting quartz veins dip steeply and trend sub-parallel to the regional strike of the host lithologies ([Oberthür *et al.*, 1996](#)). Veining is typically concentrated in brittle fractures along controlling faults ([Allibone *et al.*, 2002](#)). Selvages of sulphidized and carbonated wall rock commonly surround quartz-veins, which frequently contain carbonate.

2. Disseminated sulphide gold mineralization.

‘Invisible’ refractory gold is associated with intensely sulphidized wall rocks containing arsenopyrite and pyrite, with lesser pyrrhotite and marcasite ([Oberthür *et al.*, 1996, 1997](#)). Gold is primarily hosted in arsenopyrite, although a pyrrhotite host is more common at Bogosu ([Oberthür *et al.*, 1997](#)). Such disseminated sulphide concentrations often show a strong spatial association with graphitic mylonitized host rocks as well as the surrounding relatively undeformed wall rocks ([Allibone *et al.*, 2002](#)).

Compositional variation in carbonates and zonation of gold within sulphide minerals indicate that both of these styles formed coevally as a result of several episodic fluid infiltration events ([Mumin & Fleet, 1995](#); [Oberthür *et al.*, 1997](#)). Hydrothermal alteration in both cases results in the development of graphite along the hosting faults, and wall rock alteration containing ankerite, sericite and the sulphides ([Appiah, 1991](#); [Mumin *et al.*, 1994](#); [Mumin & Fleet, 1995](#); [Allibone *et al.*, 2002](#)). Subsequent to initial formation, [Mumin *et al.* \(1994\)](#) describe the extensive remobilization of invisible gold out of auriferous arsenopyrite.

Stable isotope ratios are similar between quartz and carbonates from gold-bearing quartz veins and the surrounding host rocks ([Oberthür *et al.*, 1996](#)). This has been interpreted

as evidence that the mineralizing fluid interacted extensively with the Birimian sediments, particularly the reduced carbonaceous (graphitic) chemical sediment facies. These rocks may therefore have had a significant effect on ore formation, as is seen at many orogenic gold deposits globally (Leube *et al.*, 1990; Goldfarb *et al.*, 2005). Conversely, Oberthür *et al.* (1994) and Mumin *et al.* (1996) suggested that CO₂-rich fluid derived during progressive regional metamorphism rose slowly through the host rocks, allowing for extensive equilibration of isotopic values. This was then followed by episodic rapid expulsion of deep-seated fluid accompanied by CO₂–H₂O phase separation that resulted in sulphide and gold precipitation.

Fluid inclusions from gold-bearing quartz veins highlight a broad range of different fluid types responsible for mineralization. Schmidt Mumm *et al.* (1997) report the dominance of dilute (<6 eq.wt% NaCl) CO₂>>H₂O±N₂±CH₄ fluids across Ghana. Wille & Klemm (2004) describe the relative abundance of H₂O-rich fluids at the Abawso deposit, with CO₂-dominant inclusions being the result of post-entrapment modification, such as grain-boundary migration. Temperatures of mineralization determined from fluid inclusions vary between different deposits, ranging from close to 200°C at Konongo to as much as 400°C at Ashanti (Schmidt Mumm *et al.*, 1997; Yao *et al.*, 2001; Wille & Klemm, 2004). Similarly, estimated pressures vary between 0.1–5.5 kbar.

Further temperature constraints on the mineralization event have been calculated by Mumin & Fleet (1995) using carbonate thermometry. This study suggests that mineralizing fluids were in the range 140–360°C. Furthermore, compositions of carbonate in the host rocks at Bogosu–Prestea, coupled with an average geothermal gradient, suggest that mineralization occurred at pressures of around 1–2 kbar (Mumin *et al.*, 1996).

In summary, mineral equilibria and fluid inclusions suggest that the likely conditions of mineralization across Ghana were in the range 300–450°C and 2–5 kbar (Blenkinsop *et al.*, 1994; Oberthür *et al.*, 1994; Mumin & Fleet, 1995; Schmidt Mumm *et al.*, 1997).

More recent studies of fluid inclusions from elsewhere in the Man Shield, and particularly Mali, have demonstrated an even broader range of fluid types responsible for the province's gold deposits, including saline (<20 eq.wt% NaCl) and hypersaline (<40 eq.wt% NaCl)

fluids, likely representing some magmatic influence ([Hammond *et al.*, 2011](#); [Treloar *et al.*, 2011](#))

A final subcategory of orogenic gold deposit found in Ghana is the granitoid-hosted sulphide stockwork, such as that found at Ayanfuri ([Yao *et al.*, 2001](#); [Brownscombe, 2009](#)). These deposits are primarily found in the eastern Kumasi basin and comprise a brittle-fracture quartz vein stockwork surrounded by pervasive sulphide-rich hydrothermal alteration. Other mineralogical and fluid inclusion characteristics are comparable with those described above for the shear zone-hosted deposits.

2.8 Geology of the Damang Region

The Damang deposit, and its associated satellite deposits, are hosted in Tarkwaian System sediments of the southern Ashanti belt and centres on the Damang anticline (Figs. [2.7](#) and [2.8](#)). The Tarkwaian sediments unconformably overly, or are faulted against, Birimian volcanoclastics in the core of the anticline. The local stratigraphy of Damang has been described by [Junner *et al.* \(1942\)](#) and [Brabham \(1998\)](#). Detailed petrographic descriptions of the lithologies are made in chapter [3](#).

The Birimian rocks in the core of the Damang anticline are predominantly volcanoclastics with minor clastic siltstones and black shales. Geophysical evidence suggests that mafic volcanics may be present towards the south of the Damang camp. The Tarkwaian sediments at Damang are similar to those elsewhere in Ghana, as described in section [2.2.2](#). In summary, the base of the Tarkwaian Group is represented by a coarse pebble–boulder conglomerate, the Kawere Group, which fines upwards to a coarse sandstone. This is overlain by the economically important quartzitic Banket Series, a sequence of cross- to planar-bedded quartzites and subarkoses. Contained within the Banket Series are four quartz-lithic-pebble conglomerate horizons that host high grades of palaeoplacer-type gold mineralization. The Tarkwa Phyllite is a sequence of finely laminated metamorphosed mudstones and siltstones (metapelites) containing a mid – upper amphibolite facies metamorphic mineral assemblage. The upper most unit of the Tarkwaian Group exposed in the Damang region is the Huni

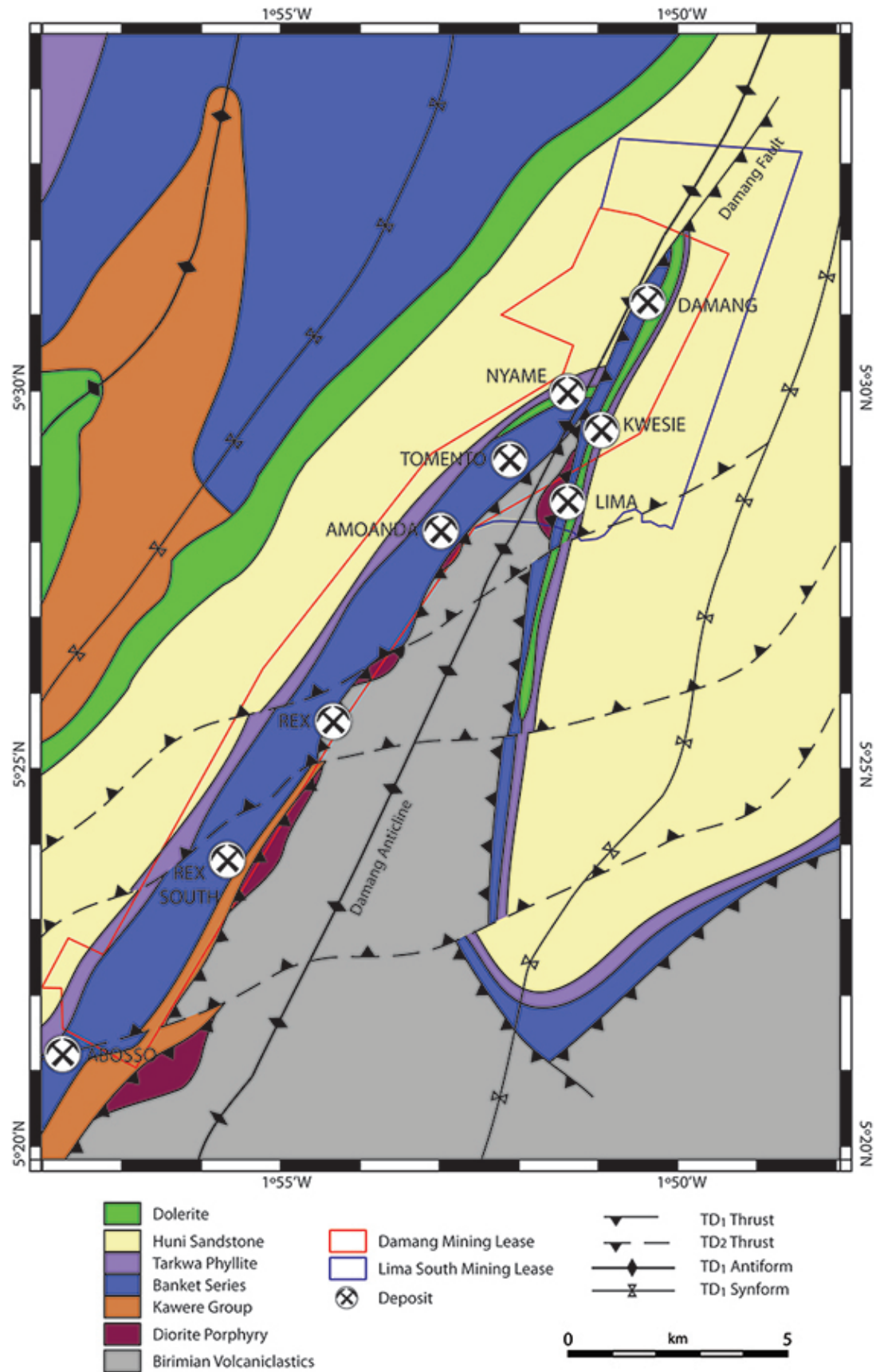


Figure 2.7: Simplified geological map of the Damang region (modified from [Tunks *et al.* \(2004\)](#) and unpublished Abosso Goldfields Ltd. data), illustrating the regional structure and location of Damang mine and its satellite deposits. Fold and thrust terminology from [Tunks *et al.* \(2004\)](#).

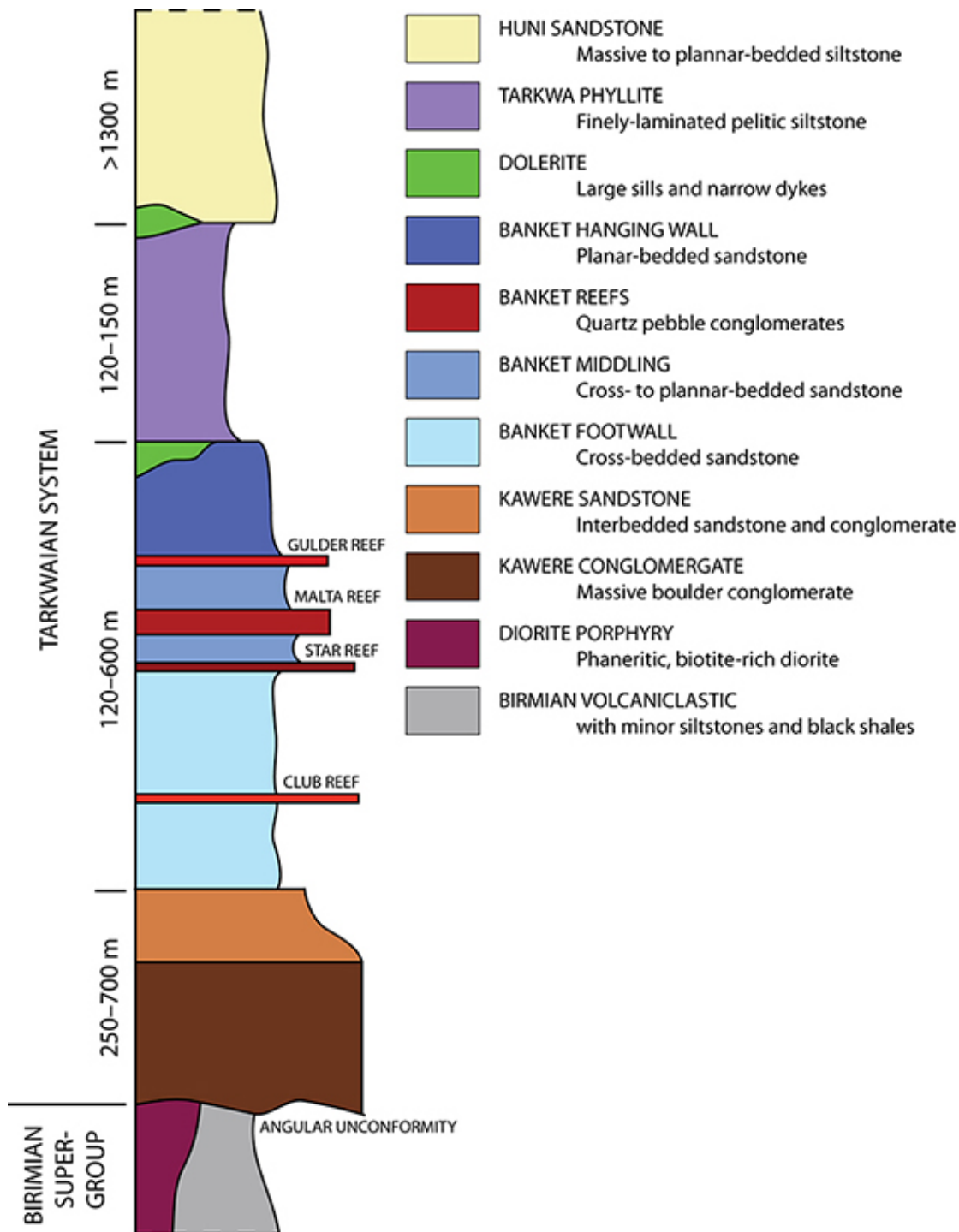


Figure 2.8: Simplified, schematic stratigraphic column of the Damang region.

Sandstone, a thick succession of massive feldspathic siltstones, sandstones and arkoses.

The Damang region is intruded by two distinct igneous bodies. The first, a series of mafic sills and dykes, intrude the upper Tarkwaian stratigraphy, commonly along boundaries of the Tarkwa Phyllite. The sills are semi-conformable with the host stratigraphy and thus their emplacement has previously been interpreted to have occurred prior to regional folding.

However, this does not agree with new observations, as will be discussed in section 3.11. These intrusions range from gabbros to diorites and are now characterized by an amphibole-dominated mineral assemblage. The second intrusion type is the so-called Diorite Porphyry, which occurs primarily near the contact between the Birimian and Tarkwaian. The exact age relationship of this unit is unknown although it is generally assumed to be Birimian. However, a post-Tarkwaian age could have significant implications for the region's gold mineralization. This will be addressed in chapter 5.

2.8.1 Structure at Damang

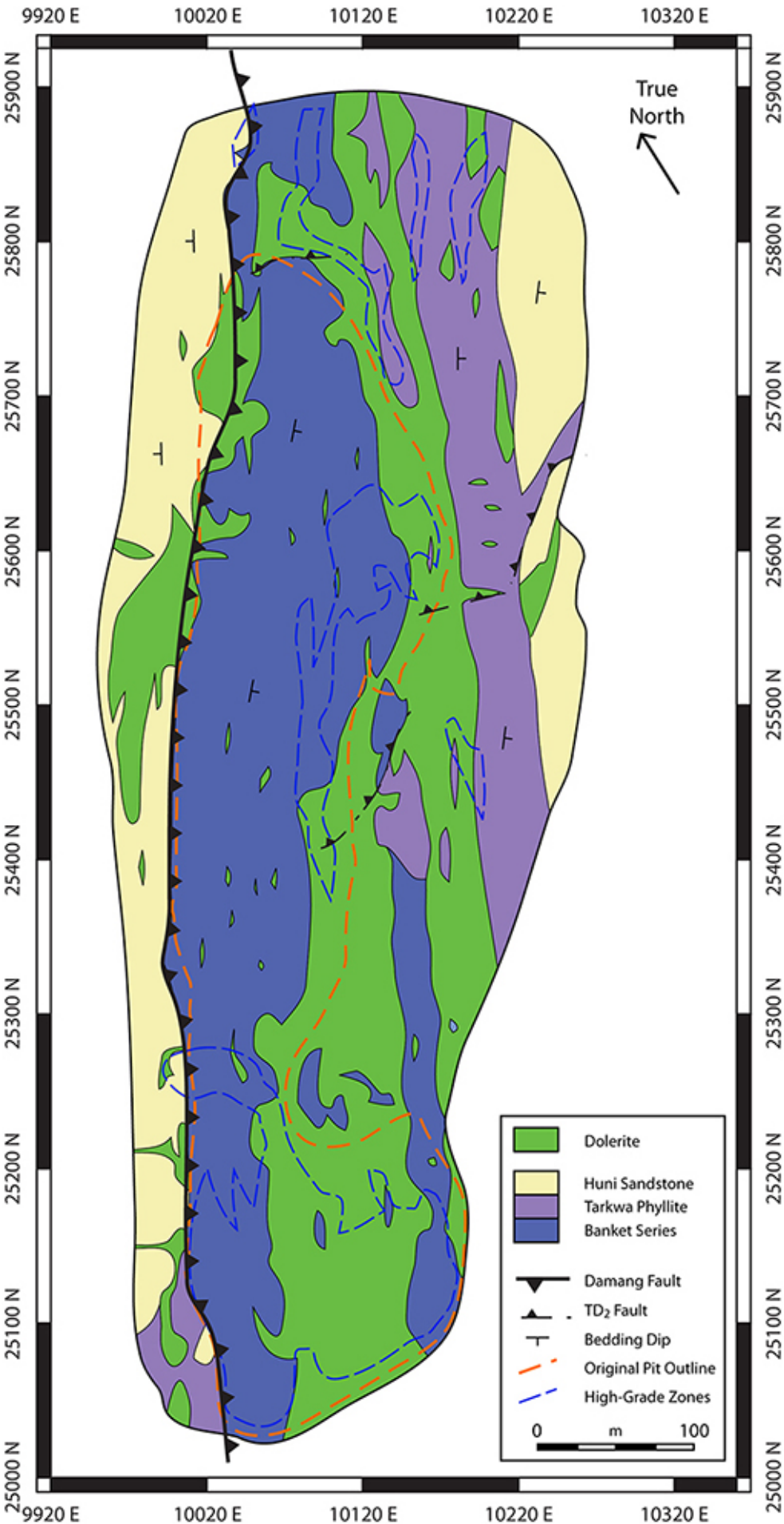
The Tarkwa–Damang region is folded into a series of northeast – southwest orientated, northeast plunging anticlines and synclines, cut by several generations of faults. Detailed structural mapping by Tunks *et al.* (2004) identified four phases of compressional deformation, termed TD₁–TD₄. This general scheme is summarized here.

TD₁ is responsible for the macroscopic, up-right, northeast – southwest trending, shallowly north-northeast – northeast plunging folds, produced during regional northwest – southeast orientated compression (Fig. 2.7). Previous measurements of fold axis orientation lie between 31/045 (Pigois, 2003) and 12/032 (Tunks *et al.*, 2004). TD₁ folding is accompanied by approximately northeast-trending, steeply-dipping faults that develop in the hinge zones of major folds. The Damang fault, which dips steeply to the east, is the most important of these in the Damang region. The relative displacement of rocks at Damang indicates the western limb of the Damang anticline has been downthrown implying a reverse sense of motion (Figs. 2.9 and 2.10). This TD₁ phase provides the framework for all later deformation in the area as well as the large-scale control on Damang's orogenic gold mineralization.

TD₂ is represented by the numerous east-northeast – west-southwest-trending thrust faults and minor west-southwest – east-northeast-trending folds, produced during a north-northwest – south-southeast orientated compressional event.

East-southeast – west-northwest directed compression during TD₃ resulted in minor low

Figure 2.9:
Simplified geological map of the main Damang pit at the 840 RL, constrained from pit mapping and grade-control drilling. The original pit outline is that that existed prior to commencement of the Damang Pit Cutback in 2005, which primarily exploited palaeoplacer style mineralization in the Banket Series conglomerates. Coordinates follow the AGL grid (see Appendix A).



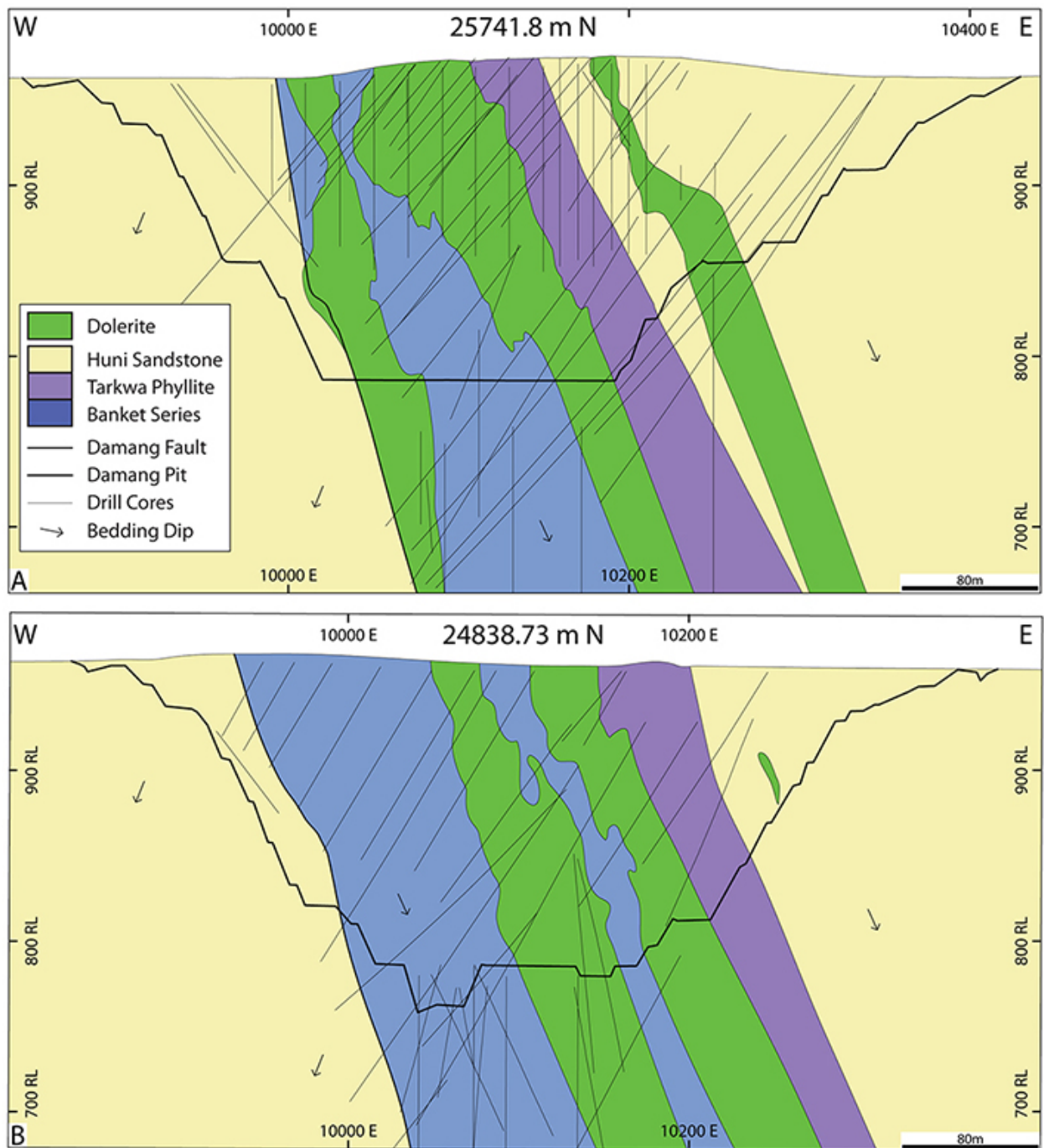


Figure 2.10: Geological cross-sections through the Damang Pit based on exploration drilling. Drill cores are projections from all drilling within 25 m of the line of section. A, 25741.8 m N; B, 24838.73 m N, according to the AGL grid (see appendix A). Section B lies in the southern portion of the Damang pit, beyond that shown in figure 2.9.

amplitude folding, small-scale brittle thrust faulting and an extensive brittle ‘fault-fracture mesh’ of sub-horizontal, extensional quartz veins. These veins host the economic hydrothermal gold at Damang. It is common for successive deformation events to reactivate earlier structures, particularly when stress directions are comparable between events. Such reactivation of cross-cutting fault sets results in compressional and dilational zones that further

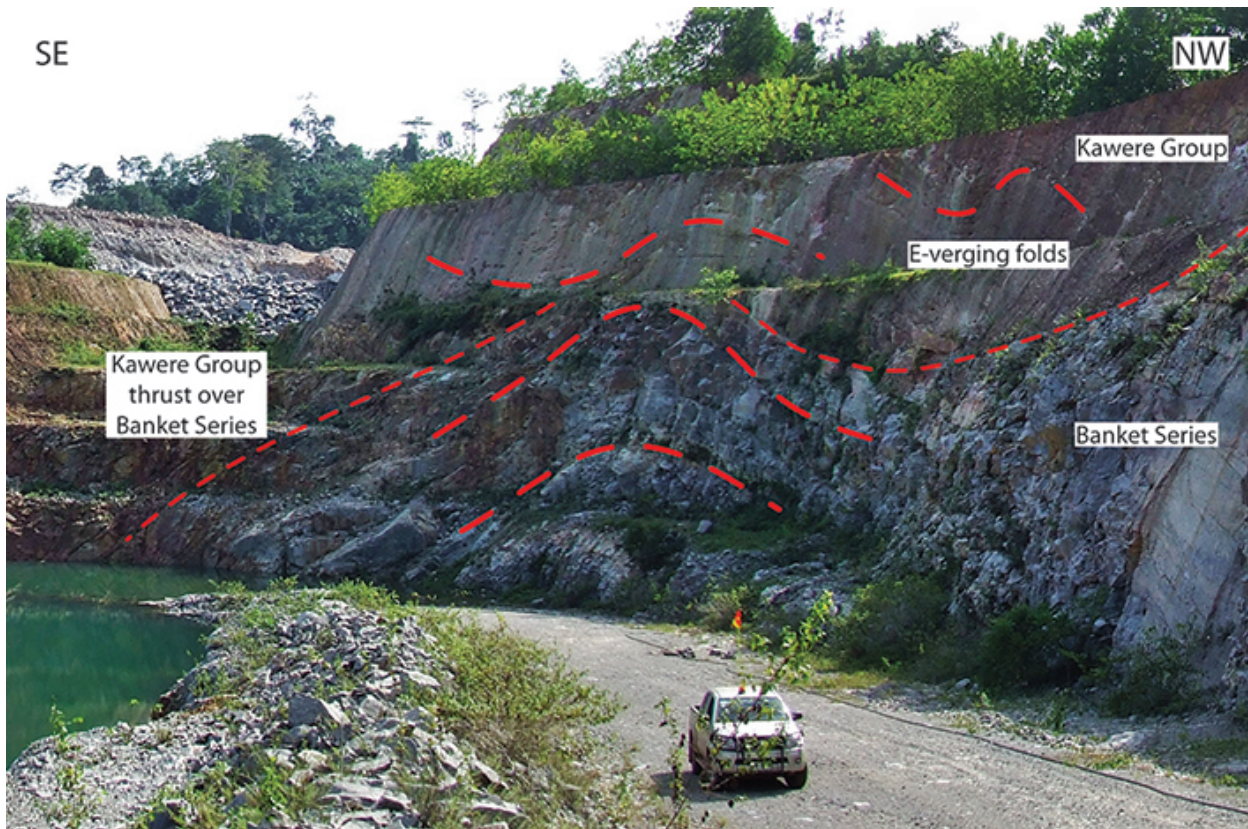


Figure 2.11: Photograph of small-scale folding at Amoanda pit. Multiple folds (large dash) show a slight eastwards vergence, suggesting a genetic association with regional folding. Here, the Kawere Group can be seen thrust (small dash) over the Banket Series prior to folding. Note vehicle for scale.

enhances the development of the extensional vein array during TD₃.

TD₄ is represented by minor brittle strike-slip dominated faulting along pre-existing faults that offsets mineralized zones and disrupts vein networks.

New Observations

The overall structural scheme described above is supported by observations made as part of this study, with the addition of an early pre-TD₁ event. These observations will now be described here with a final structural history summarized in table 2.3.

The Banket Series and Kawere Group rocks in the Amoanda pit contain a number of eastward-verging, pit-scale folds (Fig. 2.11). Measurements of bedding orientation around these folds, combined with measurements made in the Damang pit, are shown in figure 2.12.

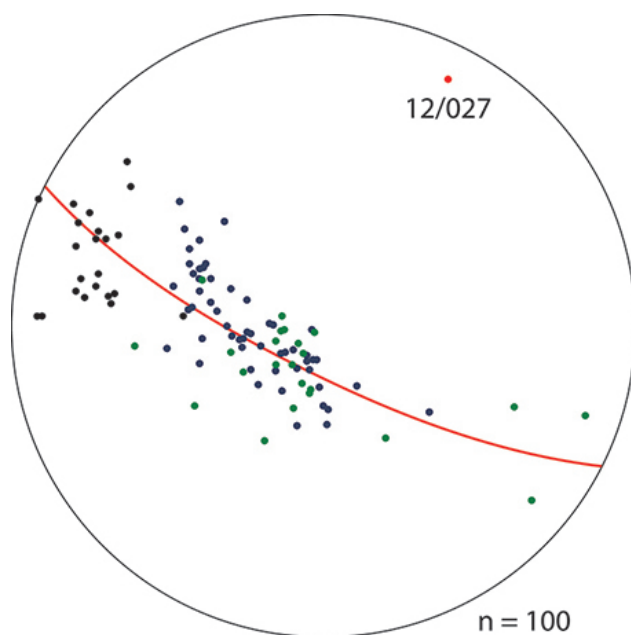


Figure 2.12: Lower-hemisphere equal-area stereonet showing poles to bedding measured at Damang and Amoanda. Black points are measurements made in the Damang pit, blue points are measurements of the Banket Series at Amoanda and the green points are those made on the Kawere Group at Amoanda. The red line represents the best fit cylindrical π -girdle, while the red point is the pole to this great circle, giving a regional fold axis trend of 12/027. The implication is that the small-scale folds observed at Amoanda are parasitic to the regional folding and formed coevally.

Since all bedding measurements fall along a common great circle, it can be inferred that these small-scale folds are parasitic to the regional-scale TD_1 fold and formed coevally. The orientation of fold axes resulting from this folding event is 12/027, consistent with previous estimates.

Figure 2.11, however, also shows that at Amoanda, the Kawere Group conglomerates overlie the Banket Series quartzites. Cross-bedding in the Banket Series serves as a way up criterion and show that the beds are not overturned. This implies that the Kawere Group has been thrust over the Banket Series. The thrust contact can be traced throughout the Amoanda pit and since it is now folded, as in figure 2.11, must have formed prior to TD_1 folding. This earlier, pre- TD_1 , phase of deformation has not previously been recognized at Damang and is the only adjustment made to the existing structural history of [Tunks *et al.* \(2004\)](#).

The TD_1 Damang fault is currently difficult to identify in exposures in the main Damang pit itself. It has, however, been intersected in a number of recent (mid-2011) deep drill holes beneath the Damang pit. These have shown that the Damang fault exists as a discrete brittle fracture surrounded by a less than 1 m wide zone of plastically deformed and foliated rocks that grade outwards into undeformed rock (Fig. 2.13). In the Damang pit area the Damang fault generally appears in the lower sections the Huni Sandstone, although the majority of ductile strain appears to be concentrated within a narrow zone of Tarkwa Phyllite, which

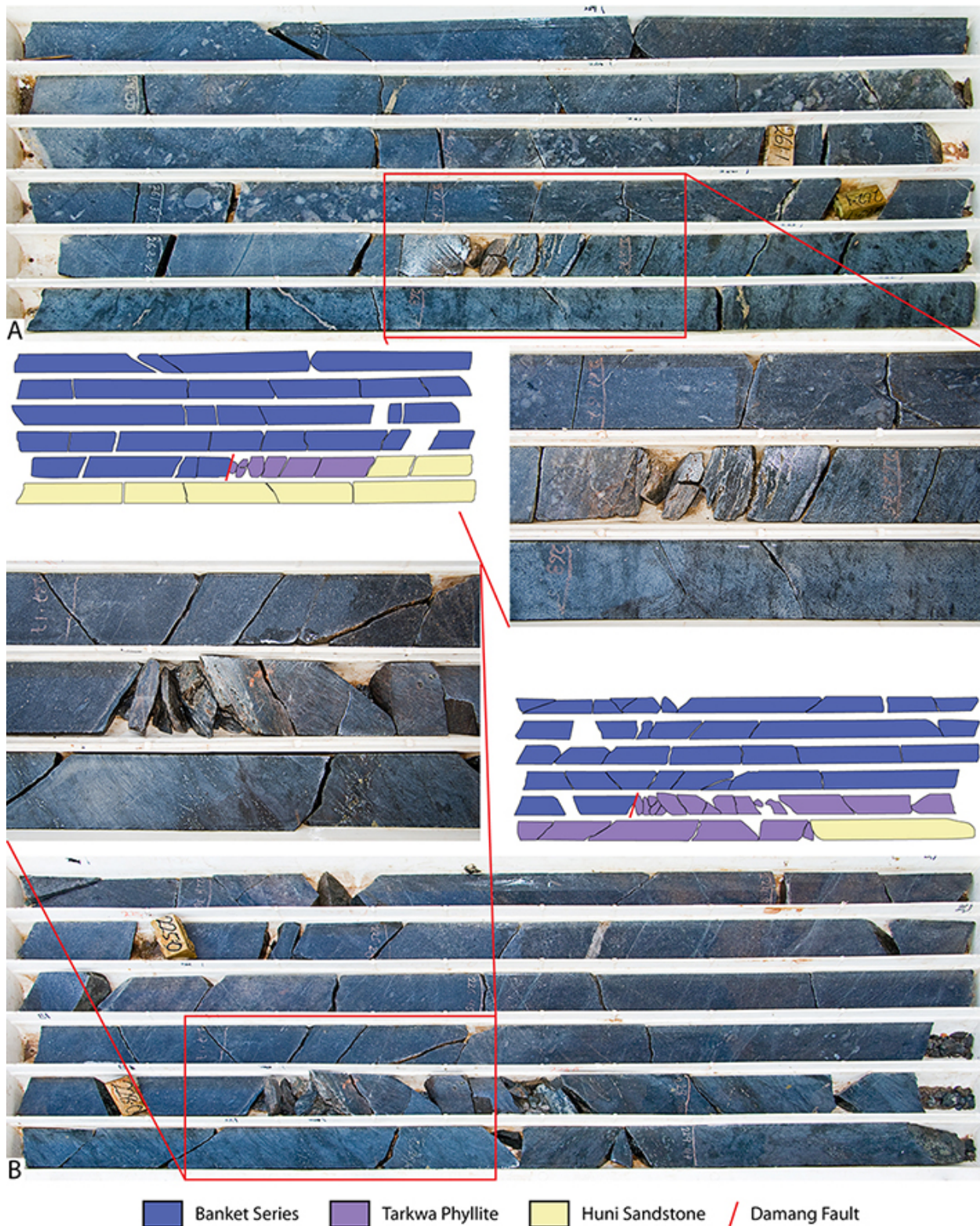


Figure 2.13: Photographs of intersections of the Damang Fault in drill core from beneath the Damang pit. (A) core DRC1758D, (B) core DRC1777D. The Damang Fault can be identified as a discrete fracture surrounded by a 1 m wide zone of ductile deformation. Core trays are 1 m long.

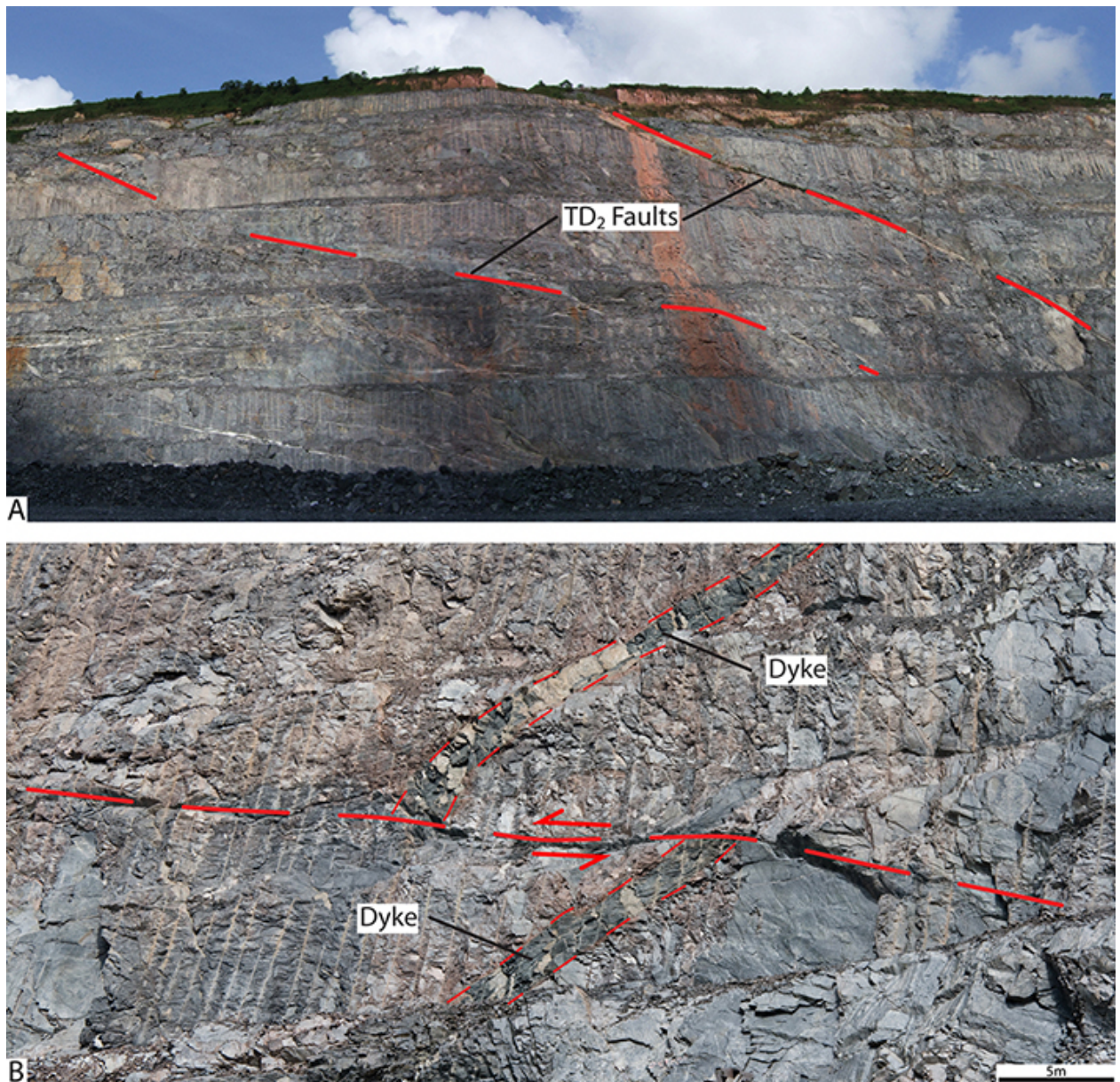


Figure 2.14: Photographs of northwards dipping TD₂ faults in the western wall of the Damang pit (A). Cross-cutting relationships with dolerite dykes indicate a reverse sense of motion (B).

grades upwards into the Huni Sandstone. Narrow, deformed carbonate veinlets are common within 10 cm of the fault. These features suggest that the Damang fault must have formed close to the brittle–ductile transition in order to generate such a well-defined fault surface accompanied by a narrow zone of ductile deformation in the surrounding rocks.

TD₂ faults are best observed in the Damang pit where they can be seen in the east and west walls of the pit (Fig. 2.14A). Due to the steeply dipping monotonous Huni Sandstone that forms the walls to the Damang pit, there are no distinctive marker horizons and determining sense of motion is difficult. However, where the faults cross-cut dolerite dykes, the relative

Figure 2.15: Photograph of a TD₂ fault in the eastern wall of the Damang pit. As with the Damang fault, the fault is a primarily brittle structure with a narrow, 10 cm, zone of ductile deformation with numerous carbonate veinlets.



offset indicates a reverse sense of motion on these faults (Fig. 2.14B). The nature of the TD₂ faults is rather similar to that for the Damang fault described above, suggesting they formed under similar pressure-temperature conditions (Fig. 2.15).

The most prominent manifestation of TD₃ deformation is the sub-horizontal quartz vein array. The constituent veins are present in a range of morphologies. They typically occur as narrow, high aspect ratio veins cutting across all earlier structures (Fig. 2.16A, B) but also occur locally as en échelon arrays (Fig. 2.16C) interpreted as tension gashes associated with small thrust faults. TD₃ faults themselves are also occasionally veined and/or develop splaying veins from them (Fig. 2.16D, E). The presence of such extensional quartz veins implies that fluid pressures must have at least locally exceeded lithostatic pressures, something which may have been enhanced by the strongly laminated Tarkwa Phyllite acting as a relatively impermeable cap.

TD₄ is relatively minor compared to the earlier events. It is typically represented by small-scale, brittle disruptions to TD₃ veins (Fig. 2.17).

Figure 2.16: (Following page) Photographs of the range of morphologies of TD₃ gold-bearing quartz veins at Damang. Veins typically form sub-horizontal, mutually cross-cutting sets, as at Tomento (A) and in the eastern wall of the Damang pit (B). Vein sets also develop in en échelon arrays bounded by TD₃ (and earlier) faults (C). TD₃ faults themselves may become veined (D), or develop splaying veins from them (E).



Figure 2.17: Photograph of brittle TD₄ disruption to earlier TD₃ extensional quartz veins in the western wall of the Damang pit.



Overall, in comparison with other Ghanaian orogenic gold deposits, no large-scale regional shear zone hosting structure has been recognized. Although the Damang Fault does offset stratigraphy by up to 3 km, the Damang deposit is not clearly identified as a likely target for mineralization based on regional geophysics (c.f. figure 2.4).

2.8.2 Gold Mineralization at Damang

Gold mineralization at Damang is unique amongst known deposits in Ghana with gold hosted in two distinct settings (Brabham, 1998; Pigois, 2003; Tunks *et al.*, 2004):

Table 2.3: Summary of structural episodes at Damang. Phases TD₁ – TD₄ were first described by Tunks *et al.* (2004) and are supported by this study.

Phase	Description	Evidence
pre-TD ₁	N-directed thrust faulting	Fig. 2.11
TD ₁	NE–SW trending, NE-plunging folds (Damang Anticline) and associated NE–SW trending thrust faulting (Damang Fault)	Figs. 2.7, 2.11, 2.12, 2.13
TD ₂	ENE–WSW trending thrust faults	Fig. 2.14
TD ₃	Sub-horizontal fault-fracture mesh	Fig. 2.16
TD ₄	Minor strike-slip reactivation of earlier faults	Fig. 2.17

1. Palaeoplacer style mineralization hosted in Banket Series conglomerates, as described elsewhere in Ghana (see section 2.7.1).
2. Hydrothermal orogenic gold hosted in TD₃ extensional quartz vein arrays in the Tarkwaian System sediments. This mineralization style is economically more important and overprints the earlier palaeoplacer style mineralization.

The main Damang pit itself and associated prospects occur along the east and west limbs of the Damang Anticline (Fig. 2.7), although economically exploited hydrothermal mineralization is currently restricted to the western limb. The Banket Series quartzites are the major hosts to hydrothermal gold mineralization, although high grade is also carried by the Tarkwa Phyllite and dolerite intrusions. The Huni Sandstone is relatively poorly mineralized while the Kawere Group is virtually barren. Similarly, Birimian volcanoclastics in the core of the anticline are poorly exposed and currently not known to host significant gold mineralization.

Free-milling gold occurs in the Banket Series conglomerates, within TD₃ quartz veins, and in close association with sulphides in the extensive hydrothermal alteration selvages around the veins (Fig. 2.18). Visible gold is regularly found within the quartz veins and is not restricted by host lithology or vein morphology or mineralogy.

As described above, the development of quartz vein arrays may be enhanced around intersections of different generations of regional faults (Tunks *et al.*, 2004). These intersections also likely served to promote fluid flow. Aeromagnetic data highlights the presence of broadly east – west trending structures (likely TD₂ faults) that intersect the Damang stratigraphy and Damang fault (TD₁ fault), particularly around Tomento, Amoanda and Rex (Fig. 2.19).

At the commencement of this thesis in late 2008, the known distribution of auxiliary deposits at Damang was similar to that shown in figure 2.7. However, at present (late 2011), significantly more mineralization has been discovered such that a large proportion of the western limb of the Damang anticline is now known to host economic concentrations of hydrothermal gold. In particular, the Greater Damang project has essentially doubled the strike length of known mineralization from north of the main Damang pit southwards towards Nyame,

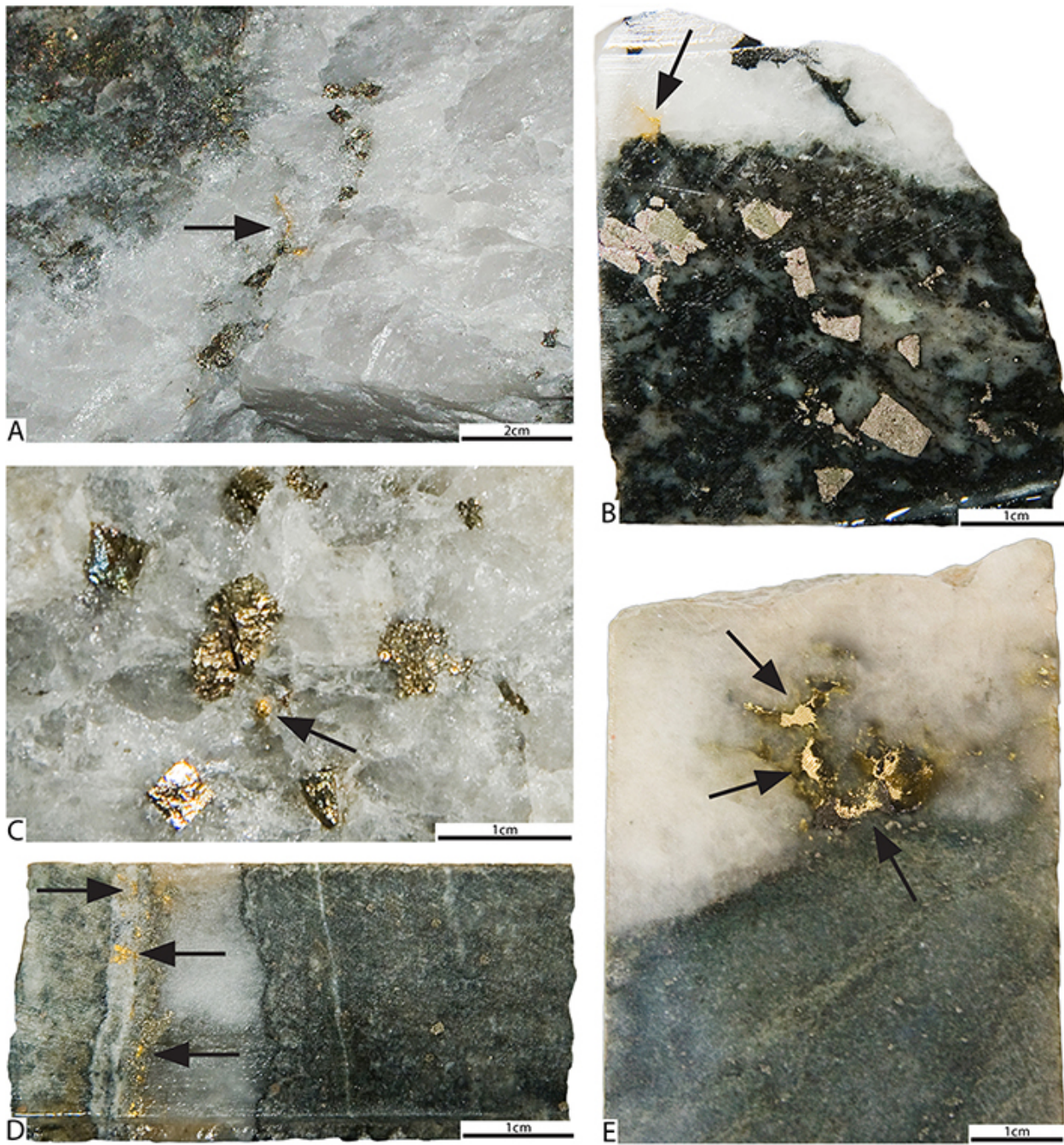


Figure 2.18: Photographs of visible gold occurring in quartz veins at Damang. Gold is not restricted by lithology or vein style: (A) Massive quartz vein in a Banket Series conglomerate. (B) Tourmaline-bearing quartz vein in a dolerite. (C) Sulphide-rich vein in a dolerite. (D) Laminated quartz vein in a dolerite. (E) Extremely rich occurrence in a massive vein in a Banket Series quartzite.

through Juno and Tamang (not shown on figure 2.7). Similarly, hydrothermal mineralization is known to extend almost continuously between Amoanda and Tomento East and from Rex South to Rex along the Damang Fault corridor and its southern extension. Ultimately, however, this suggests that intersecting fault generations are not essential to the presence of gold mineralization at Damang.

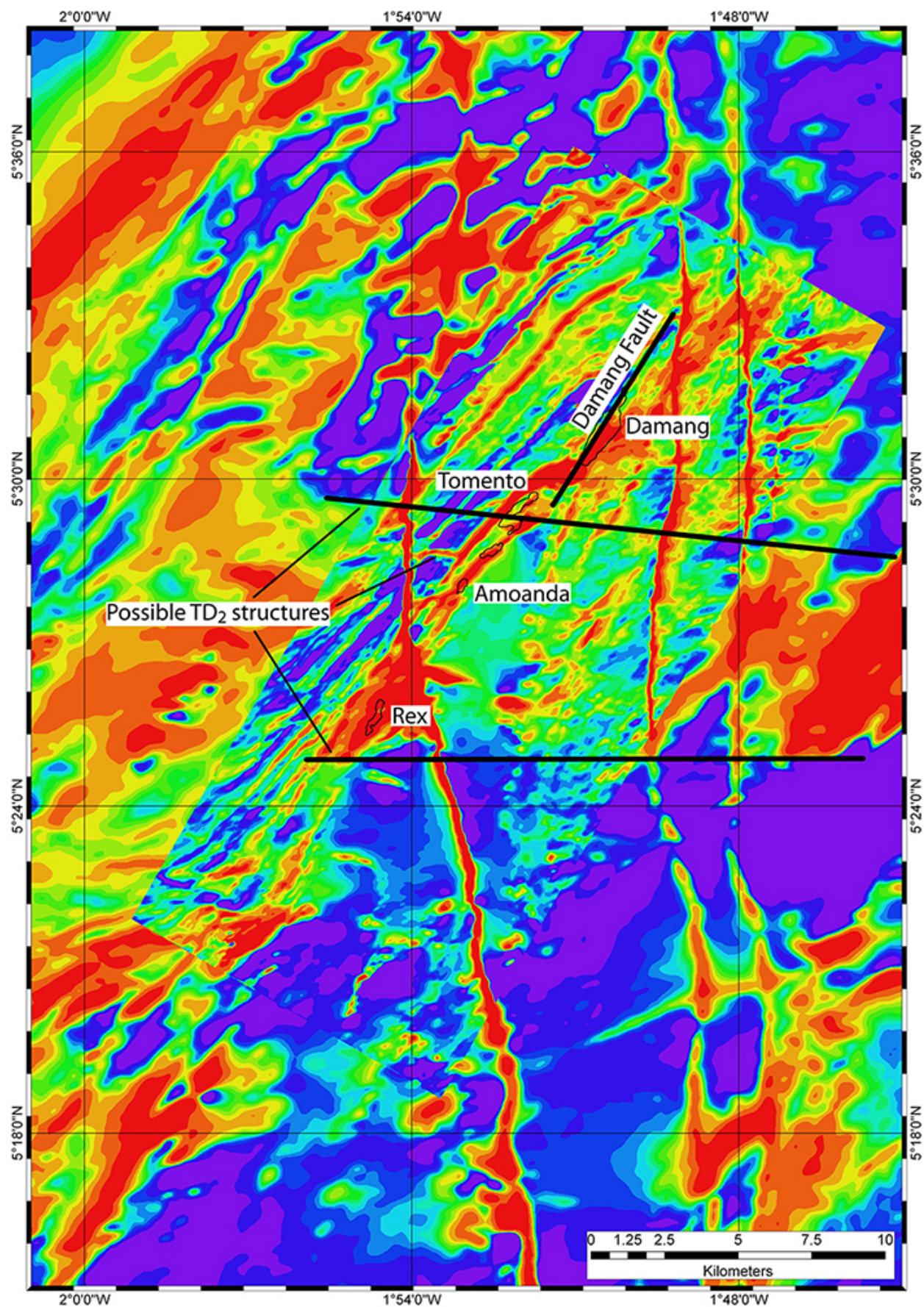


Figure 2.19: Aeromagnetic (reduced to pole) map of the Damang region showing important lithological and structural features, including the intersection of east–west trending structures with the Damang anticline near Tomento, Amoanda and Rex. Current pit outlines are shown. Data provided by Gold Fields Ltd. Colour scale is qualitative from blue (low) to red (high).

2.9 Geology of the Tarkwa Region

Throughout this thesis occasional reference or comparisons will be made to the Tarkwa mine, which lies some 30 km southwest of Damang. Tarkwa mine, also owned by Abosso Goldfields Ltd., exploits palaeoplacer style mineralization from Banket Series conglomerates and is the largest such mine in West Africa. Quartz veining and associated hydrothermal gold mineralization is not present at Tarkwa, and as such it provides a worthwhile comparison to Damang.

The Tarkwaian stratigraphy and overall TD₁ structure is similar to that seen at Damang (Fig. 2.20). The Banket Series conglomerates, however, are significantly more developed at Tarkwa, comprising as many as seven distinct horizons, and the beds generally dip and plunge at shallower angles. This has allowed Tarkwa to be mined through a series of large, yet relatively shallow open pits. This shallow dip, however, results in virtually no drilling intersections of the upper and lower Tarkwaian stratigraphy and as such, discussion is limited to the Banket Series, which is also the primary host of gold mineralization at Damang.

2.9.1 Quartz Veining at Tarkwa

Although Tarkwa does not host an extensively developed quartz vein network like that at Damang, one localized occurrence of quartz-sulphide-tourmaline veining does occur in the Pepe pit, in the northeastern part of the Tarkwa mine complex. Despite some mineralogical similarities with veining at Damang (see chapter 3) the structural setting is rather different, with the Tarkwa veining related to flexural slip between a conglomerate unit and a highly competent Banket Series quartzite package above. It does, however, still provide an interesting point of comparison, that will be addressed later.

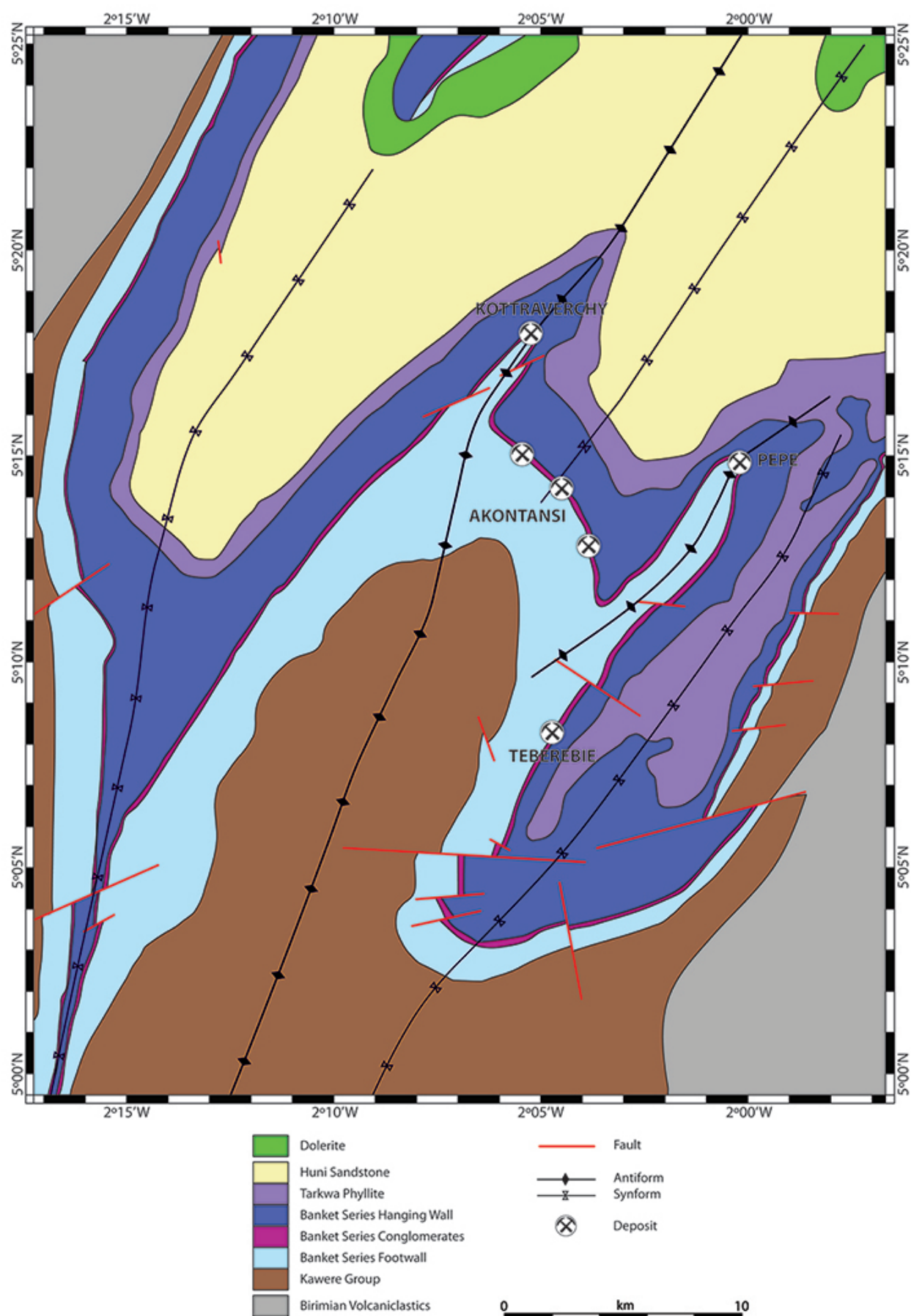


Figure 2.20: Simplified geological map of the Tarkwa region (modified from unpublished Abosso Goldfields Ltd. data), illustrating the regional structure and location of the constituent deposits at Tarkwa mine.

Chapter 3

Petrography, Petrology and Geochemistry



Exploration 24/7.

One of the primary aims of this thesis is to investigate the petrology and geochemistry of the Damang system, particularly with regards to the hydrothermal alteration events and the lithological and geochemical controls on gold mineralization. Furthermore, these fundamental petrographic and geochemical observations place absolute constraints on any and all subsequent analysis and it is thus vital to characterise them accurately. The initial sections of this chapter deal with the basic petrography of the range of lithologies at Damang, covering both unmineralized varieties and those altered during major hydrothermal events. Two such events have been identified previously; the so-called ‘Pink Haematite’ alteration and gold-associated alteration ([Brabham, 1998](#); [Pigois, 2003](#)). This is followed by summary discussions of the individual hydrothermal events, using mineral chemical and whole-rock geochemical data to characterise them and constrain likely alteration reactions/mechanisms. Finally, a full structural and hydrothermal paragenesis for Damang is provided that draws on data discussed in both this chapter and section 2.8 of chapter 2. The locations and petrographic details of analysed samples are given in appendix A.

3.1 The Birimian Supergroup

The Birimian Supergroup in the core of the Damang anticline is poorly exposed and not drilled extensively. The best exposure and most numerous drill intersections occur at Amoanda (Fig. 2.7), where the Birimian is faulted against the mineralized Banket Series quartzites. In the Damang region, the Birimian predominantly comprises volcanoclastic sediments with interbedded siltstones and black shales. A true extrusive volcanic component is rarely, if ever, encountered, although regional geophysical data suggests that mafic volcanics may be present towards the southern end of the Damang camp (Abosso Goldfields Ltd. unpublished data).

3.1.1 Birimian Volcaniclastics

The Birimian volcaniclastics make up the majority of the Birimian Supergroup at Damang and essentially comprise a pale- to medium-grey, fine- to medium-grained, massive quartz-

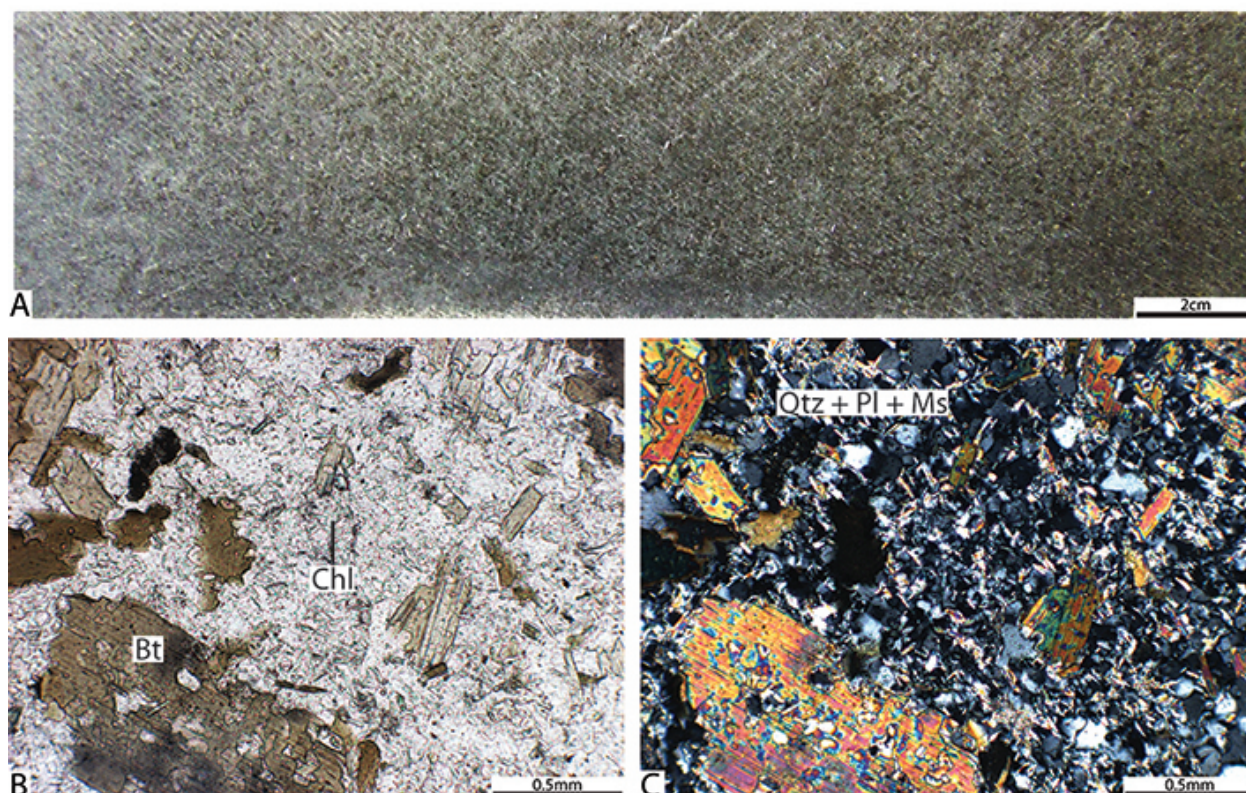


Figure 3.1: Photograph and photomicrographs of the Birimian Volcaniclastic from Amoanda. A massive, medium-grey quartz-rich rock overprinted by coarse biotite flakes. Sample AWABv1 (A) in plane-polarized light (B) and between crossed polars (C).

rich rock (Fig. 3.1A). In thin section, a matrix of quartz, with lesser feldspar, muscovite, very fine, randomly orientated chlorite flakes and trace ilmenite is overprinted by coarse biotite flakes (Fig. 3.1B, C).

3.1.2 Birimian Sediments

The Birimian at Damang contains a number of quartz-feldspar siltstones and fine-grained carbonaceous black shales interbedded with the volcaniclastics. Of these, the black shales are of particular interest due in part to their potential as a fluid and gold source for the mineralization event (Large *et al.*, 2011), but also for the microstructures exhibited.

The Birimian black shales are dark-grey to black, fine-grained quartz-rich sediments displaying irregular alternations between more and less carbonaceous material, interpreted as bedding (Fig. 3.2A). This fabric is overprinted by randomly orientated needles of white arsenopyrite and coarse biotite flakes (Fig. 3.2A, B). The rock is commonly crenulated, which

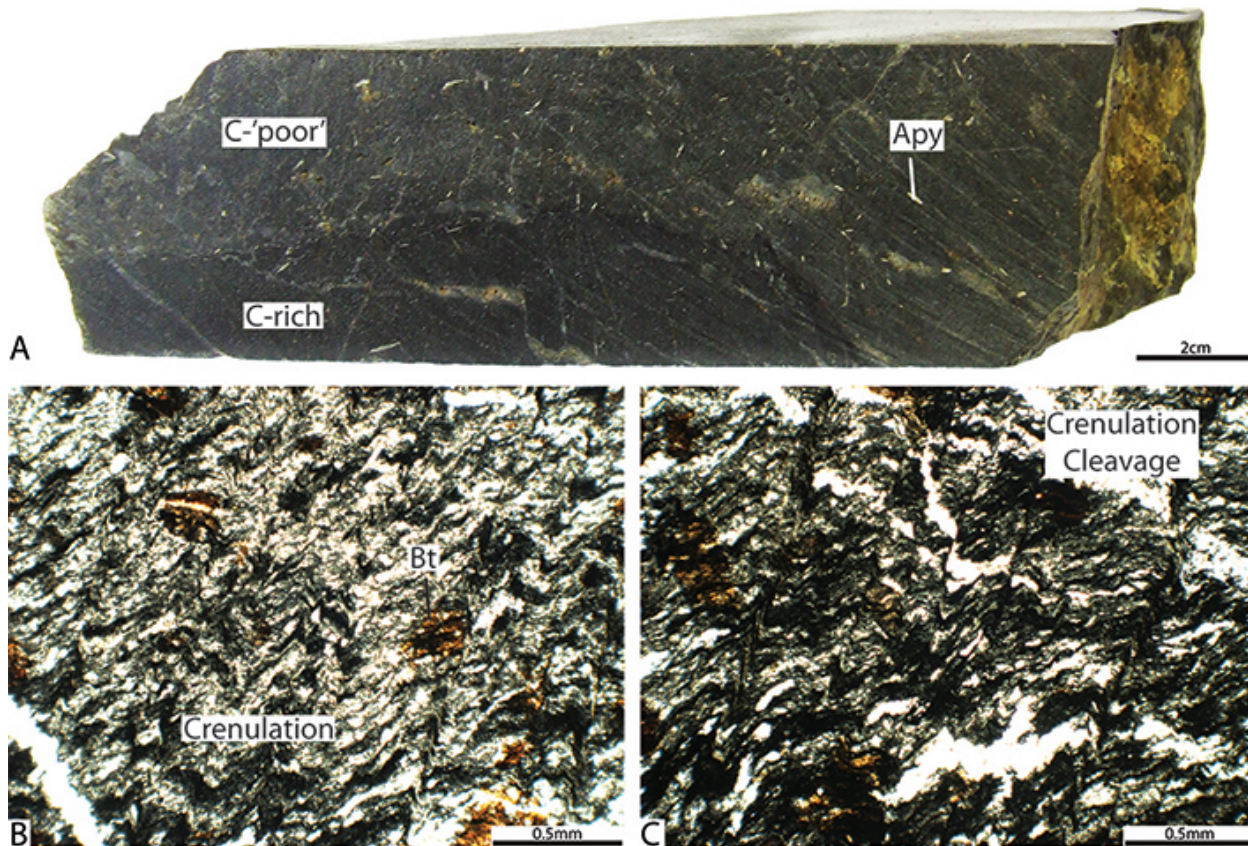


Figure 3.2: Photograph and photomicrographs of the Birimian Black Shale from Amoanda. A dark-grey to black, fine-grained sediment with randomly orientated needles of arsenopyrite (A). Is commonly crenulated (B), which occasionally develops into a penetrative crenulation cleavage (C). Sample AWABbs1 (A) in plane-polarized (B, C) light.

regularly develops into a penetrative crenulation cleavage (Fig. 3.2C). These crenulations are asymmetrical, implying either a simple shear component to the deformation, or a vergence related to larger-scale folding. Such microstructural deformation is not seen in the Tarkwaian rocks at Damang (see below), even in peltitic rocks of a similar rheology, and implies a period of Birimian deformation prior to Tarkwaian sedimentation (c.f. section 2.4). The randomly orientated arsenopyrite post-dates the crenulation deformation. Arsenopyrite is commonly associated with orogenic gold mineralization at other Ghanaian deposits and may imply a similar link at Damang.

3.1.3 Diorite Porphyry

The so-called Diorite Porphyry is a mottled medium-grey, coarse-grained, diorite/quartz-diorite comprising dominant biotite and feldspar with lesser quartz (Fig. 3.3A). The term

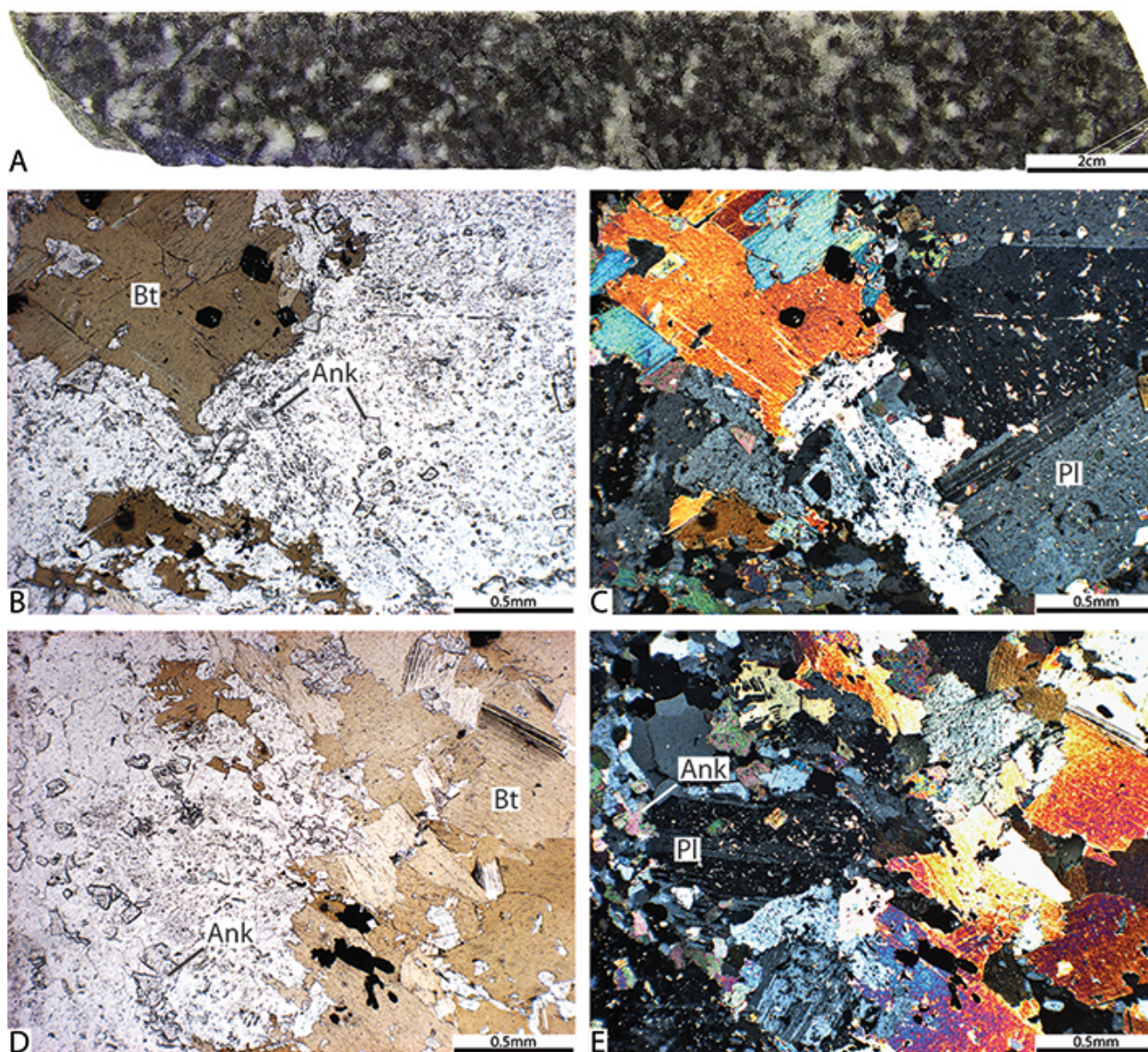


Figure 3.3: Photograph and photomicrographs of the Diorite Porphyry from Amoanda. A massive, phaneritic diorite/quartz-diorite (A) primarily comprising coarse biotite flakes and randomly orientated, interlocking plagioclase feldspar laths. Abundant ankerite rhombs occur throughout and suggest extensive, pervasive alteration. Sample AWADi2 (A) in plane-polarized light (B) and between crossed polars (C), sample AWADi1 in plane-polarized light (D) and between crossed polars (E).

‘porphyry’ is inaccurate as the rock possesses a massive phaneritic texture, not a porphyritic one. However, in order to maintain consistency with previous studies and mine terminology at Damang, the name Diorite Porphyry will still be applied to this unit.

The Diorite Porphyry displays an igneous texture of coarse biotite flakes amongst randomly orientated, interlocking plagioclase feldspar crystals with lesser quartz and additional minor chlorite and ilmenite (Fig. 3.3B–E). Numerous small, euhedral carbonate (ankerite) rhombs are dispersed throughout the rock. These are ubiquitous in all examples of the Diorite

Porphyry, suggesting they could be an original feature. Or, more likely, they may represent a pervasive hydrothermal alteration. As will be discussed for other lithologies below, both Pink Haematite and gold-associated alteration result in the development of extensive ankerite. Although the Diorite Porphyry does not appear to undergo Pink Haematite alteration and is rarely, if ever, mineralized, the observed carbonate could be the manifestation of one or both of these hydrothermal events. Indeed, the euhedral and well-annealed habit of the ankerite rhombs points towards a pre-peak-metamorphic origin, consistent with the Pink Haematite alteration event (see section 3.11).

3.2 The Kawere Group

The Kawere Group is a series of pebble–boulder conglomerates with intercalated sandstones that become dominant up section (Fig. 3.4). The conglomerates contain moderately sorted, sub-angular clasts of fine-grained siliceous lithic fragments and lesser vein quartz set in a fine-grained matrix. It is easily weathered and fresh examples are rarely seen. The Kawere Group does not contain many quartz veins and is not known to host gold mineralization at Damang. Indeed, in certain locations, such as at Amoanda, the Kawere Group acts as a barrier to mineralization in surrounding rocks. This is likely due to the rheologically weak nature of the unit, caused by the abundant matrix, which restricts brittle fracturing and veining. There is no recognisable high-temperature hydrothermal alteration in this unit.

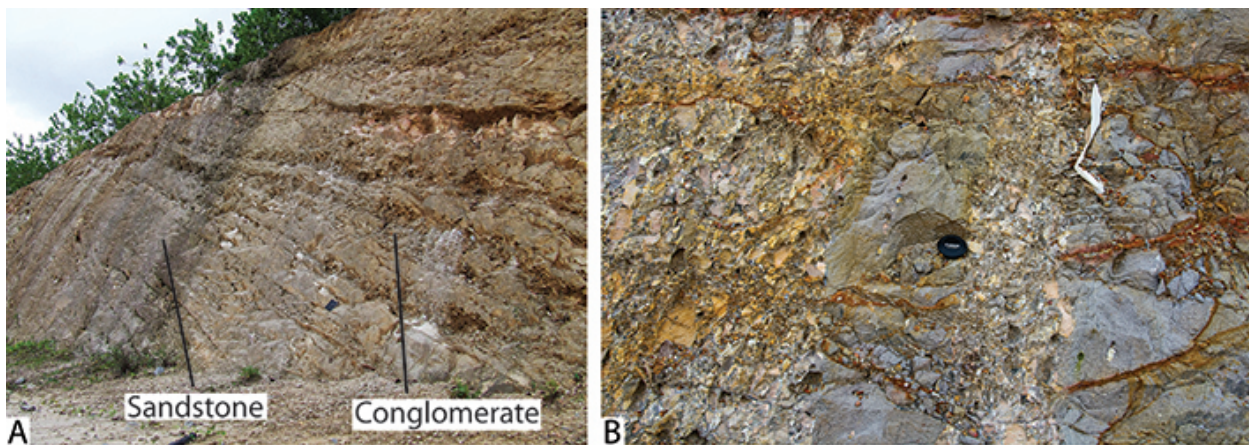


Figure 3.4: Outcrop photographs of Kawere Group sandstones and conglomerates from Amoanda (A) and Rex (B). Black notebook in A is 22 cm in height.

3.3 The Banket Series

The Banket Series comprises a sequence of pale- to medium-grey, medium- to coarse-grained sandstones and quartzites characterized by heavy mineral concentrations along bedding planes and cross-bed foresets (Fig. 3.5). These heavy minerals are dominated by magnetite, with additional ilmenite, rutile, haematite and zircon. A Banket 'Footwall' is defined as lying beneath a package of conglomerate horizons, with a Banket 'Hanging Wall' lying above. The Footwall uniformly shows fluvial, trough cross-bedding, while the Hanging Wall typically exhibits a planar bedding morphology.

However, magnetite and other Fe-oxides in the Banket Series quartzites commonly show a complex alteration/remobilization pattern. This is particularly well displayed in highly weathered quartzite, such as at Tomento, where there exists a helpful colour contrast between the oxides and the host rock. Here, the intense weathering of the feldspar component in the Banket Footwall has destroyed much of the textural evidence for original cross bedding although its presence is still locally defined by magnetite (now likely haematite). In addition to the cross bedding, Fe-oxides occur in at least three other very different styles:

1. Highly irregular seams occurring throughout the rock (Fig. 3.6A, D, E).
2. Straight, cross-cutting seams that are moderately less resistant to weathering and form prominent ridges. These occur on a range of scales from several centimetres to a few metres (Fig. 3.6B).
3. Rare cross-cutting stylolitic horizons (Fig. 3.6C).

Such a variety of morphologies suggests an extensive remobilization history. These remobilized forms are difficult to identify in fresh Banket Series quartzite as the darker colour of the fresh rock masks the Fe-oxides. Detailed examination of fresh drill core, however, can result in such varieties occasionally being identified (Fig. 3.6D), particularly where Pink Haematite alteration provides a colour contrast (Fig. 3.6E). The presence of such features in fresh rock implies that this is not a recent weathering feature.

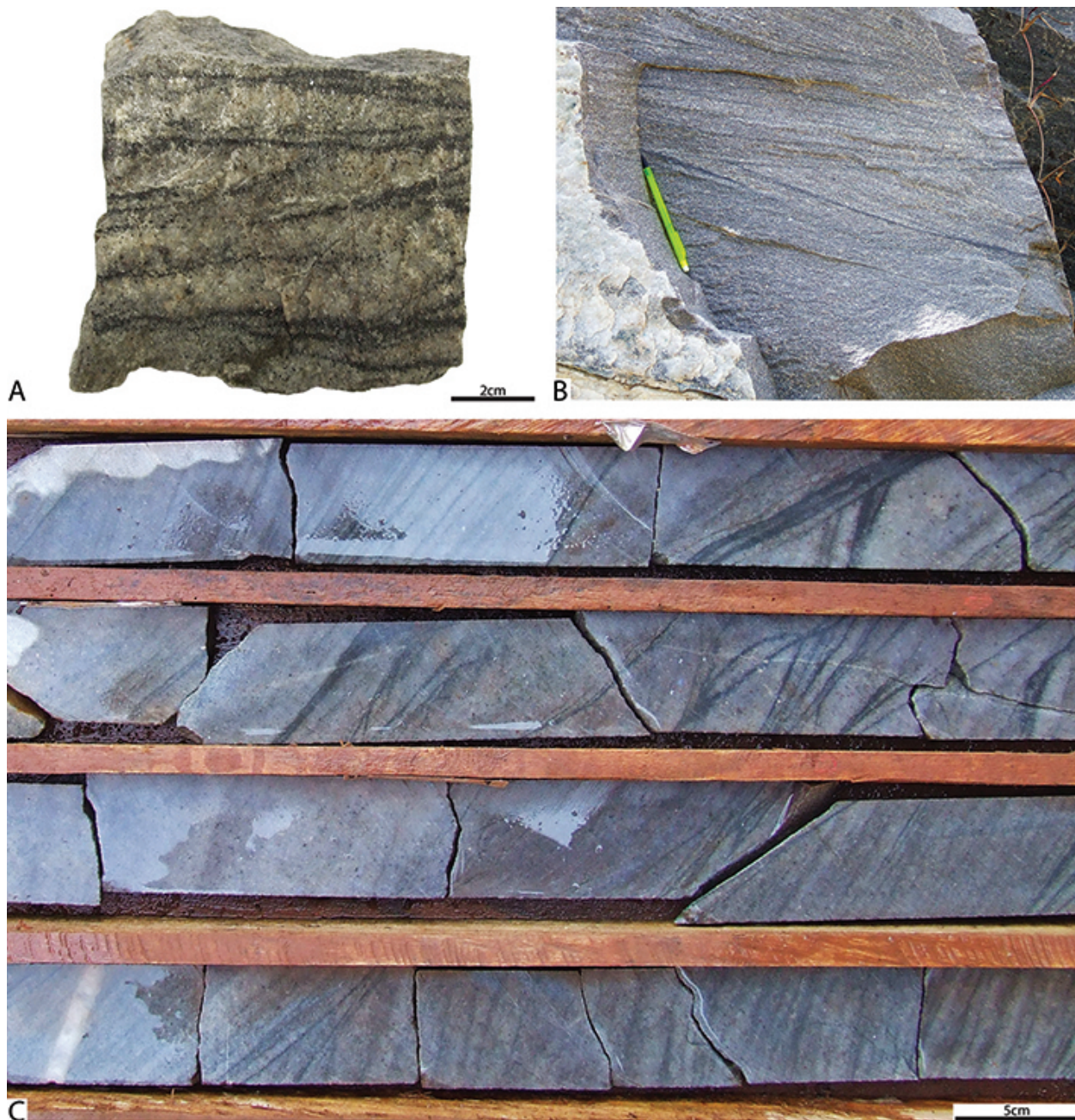


Figure 3.5: Photographs of unaltered Banket Series quartzites from Damang and Amoanda. Pale-to medium-grey quartzites with heavy mineral concentrations along bedding planes and cross-bed fore sets. Sample AWDBf1 (A), outcrop in Amonada pit (B), core AMD113 (C).

Banket Series quartzites typically comprise a matrix of fine muscovite, quartz and feldspar with minor metamorphic biotite, chlorite and trace garnet surrounding coarse, rounded, moderately–well sorted clasts of quartz with lesser feldspar grains and quartz-lithic fragments (Fig. 3.7A-D).

The Banket Series at Damang contains four distinct conglomerate horizons (see section 2.8). Three of these form a package that can be correlated between the east and west limbs of the

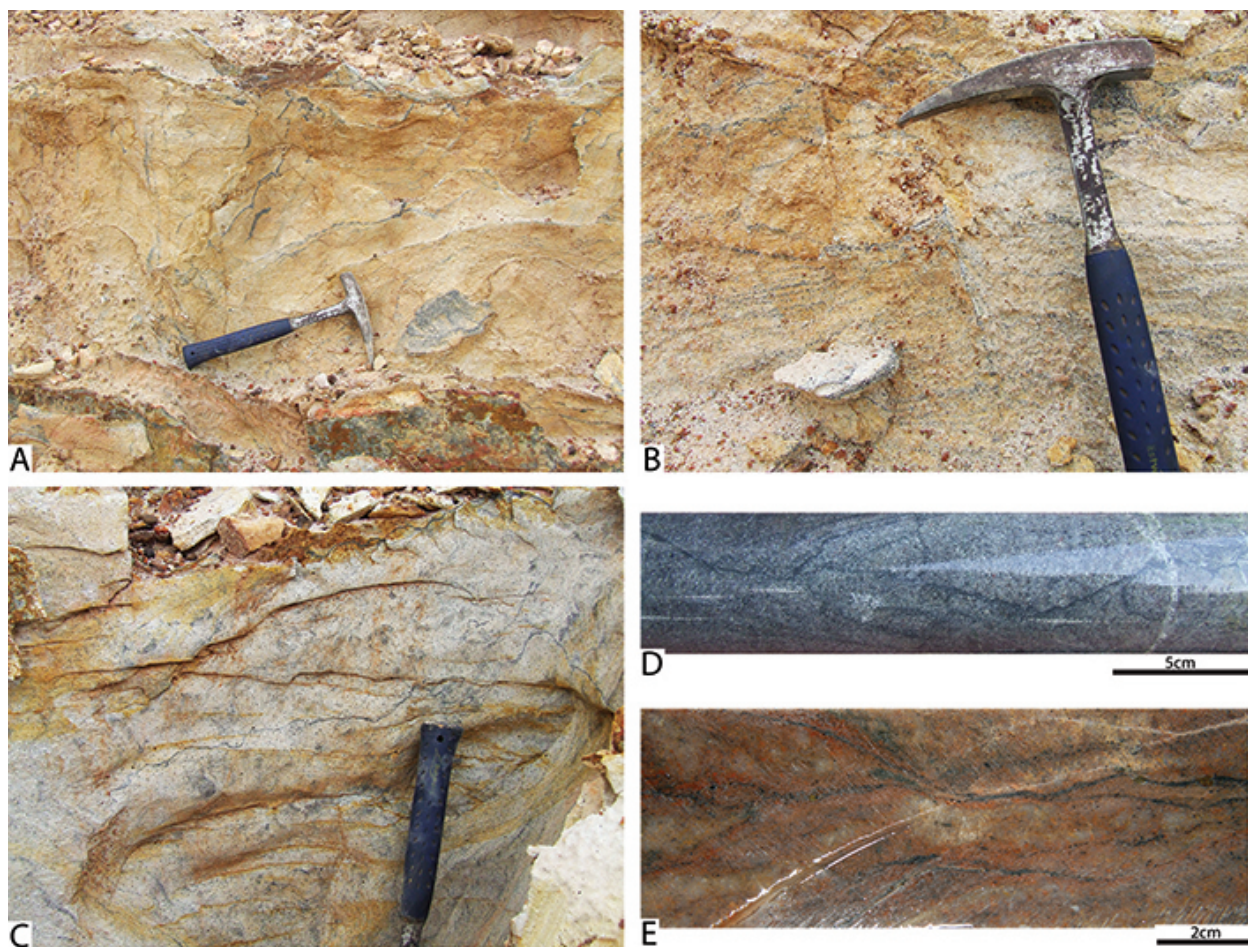


Figure 3.6: Photographs of Fe-oxide remobilization in Banket Series quartzites. Occurrences as irregular seams (A, D, E), straight, cross-cutting seams (B) and rare cross-cutting stylolitic horizons (C). Weathered outcrop at Tomento (A-C), fresh core WLRC221D (D), Pink Haematite altered core DRC423D (E)

the Damang anticline and along strike to Tarkwa mine (Brabham, 1998). These horizons, or ‘reefs’, are individually named although these differ between the east and west limbs of the Damang anticline. For the purposes of clarity, west limb terminology will be used throughout this thesis, with east limb terminology given in parentheses in this paragraph only. In stratigraphic sequence, these three reefs are the lower Star (Lima) Reef, the middle Malta (Kwesie 1) Reef, which often develops as multiple narrow horizons with intercalated quartzite lenses, and the upper Gulder (Kwesie 2) Reef. These reefs become less mature up-sequence, with the Star Reef almost completely comprising vein quartz clasts, while the Gulder Reef contains a large proportion of lithic clasts (Fig. 3.8). Gold grade correlates with maturity and is highest in the Star Reef. The fourth, so-called Club Reef, occurs sporadically within the Banket Footwall, particularly around Rex, and is petrographically similar to the Star Reef. The multiple reefs at Tarkwa mine are petrographically similar to

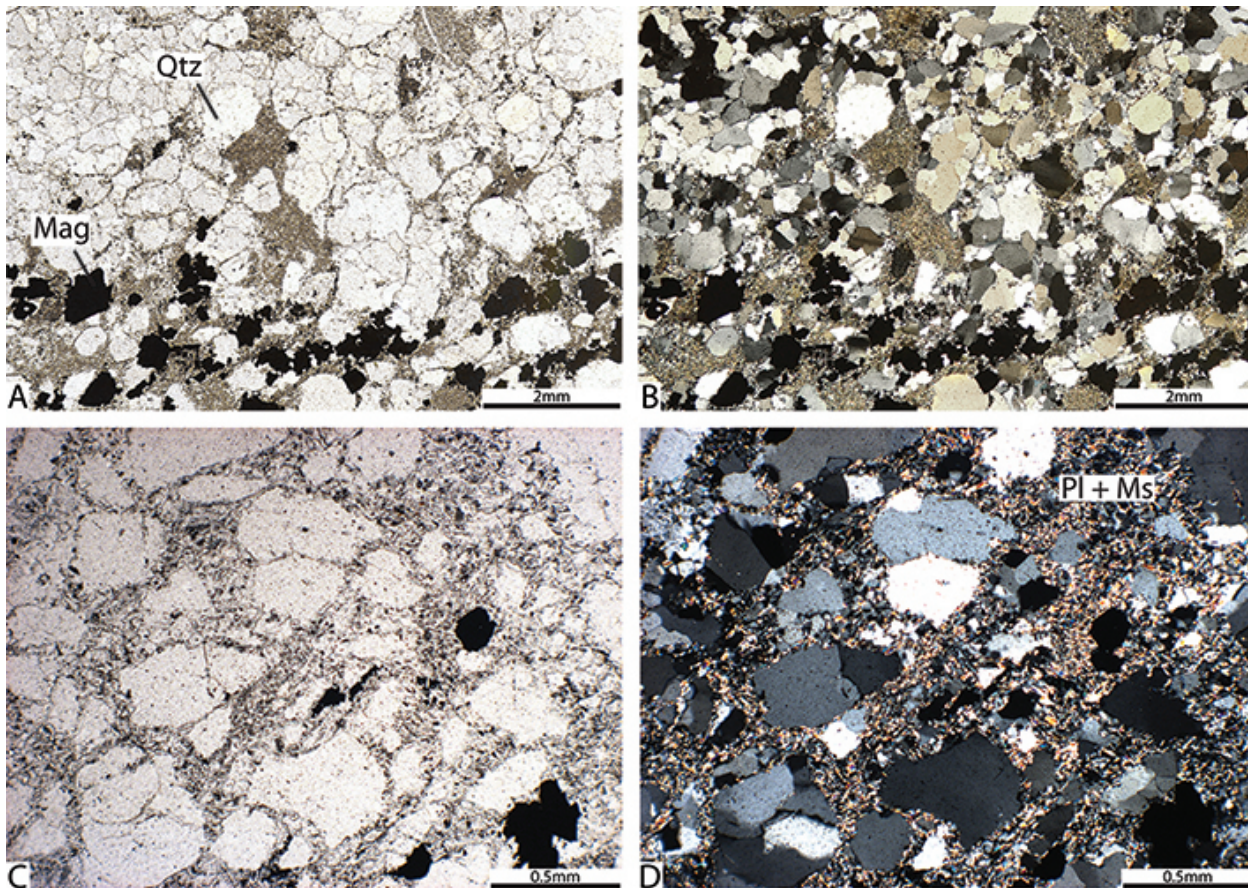


Figure 3.7: Photomicrographs of unaltered Banket Series quartzites. Predominant quartz clasts sit amongst a matrix of fine muscovite, quartz and feldspar. C and D show an unusually high proportion of matrix and may in fact include small amounts of hydrothermal muscovite (c.f. Fig. 3.11). Sample AWABf1 in plane-polarized light (A) and between crossed polars (B), sample AWDBf1 in plane-polarized light (C) and between crossed polars (D).

those at Damang, although the predominant Fe-oxide phase (in the Banket Series quartzites also) is haematite rather than magnetite (Pigois, 2003).

Clasts in the Banket Conglomerates are predominantly quartz or quartz-lithic, with lesser amounts of lithic clasts comprising finer quartz, feldspar and mica (Fig. 3.9A, B). Occasional clasts of micaceous schist are present that show a similar crenulation – crenulation-cleavage to that developed in the Birimian sediments at Damang (Fig. 3.9C, D). Metamorphic garnet and staurolite develop extensively within lithic clasts and pelitic matrix domains (Fig. 3.9E-G). The garnet, in particular, commonly grows to fill spaces between quartz clasts. These phases are therefore interpreted to post-date Tarkwaian sedimentation and are not derived from earlier metamorphism of the Birimian. In contrast, Banket Series conglomerates at Tarkwa do not contain any amphibolite facies metamorphic minerals, either as

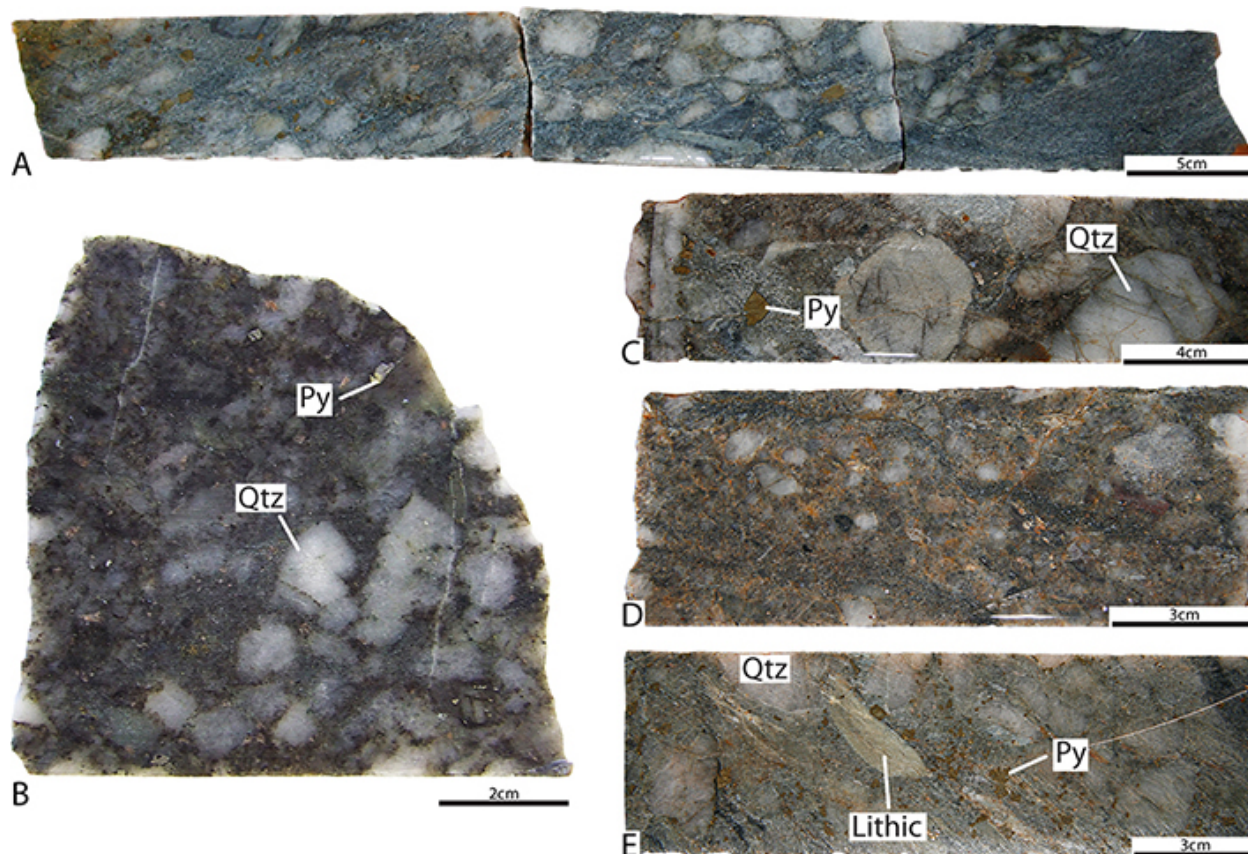
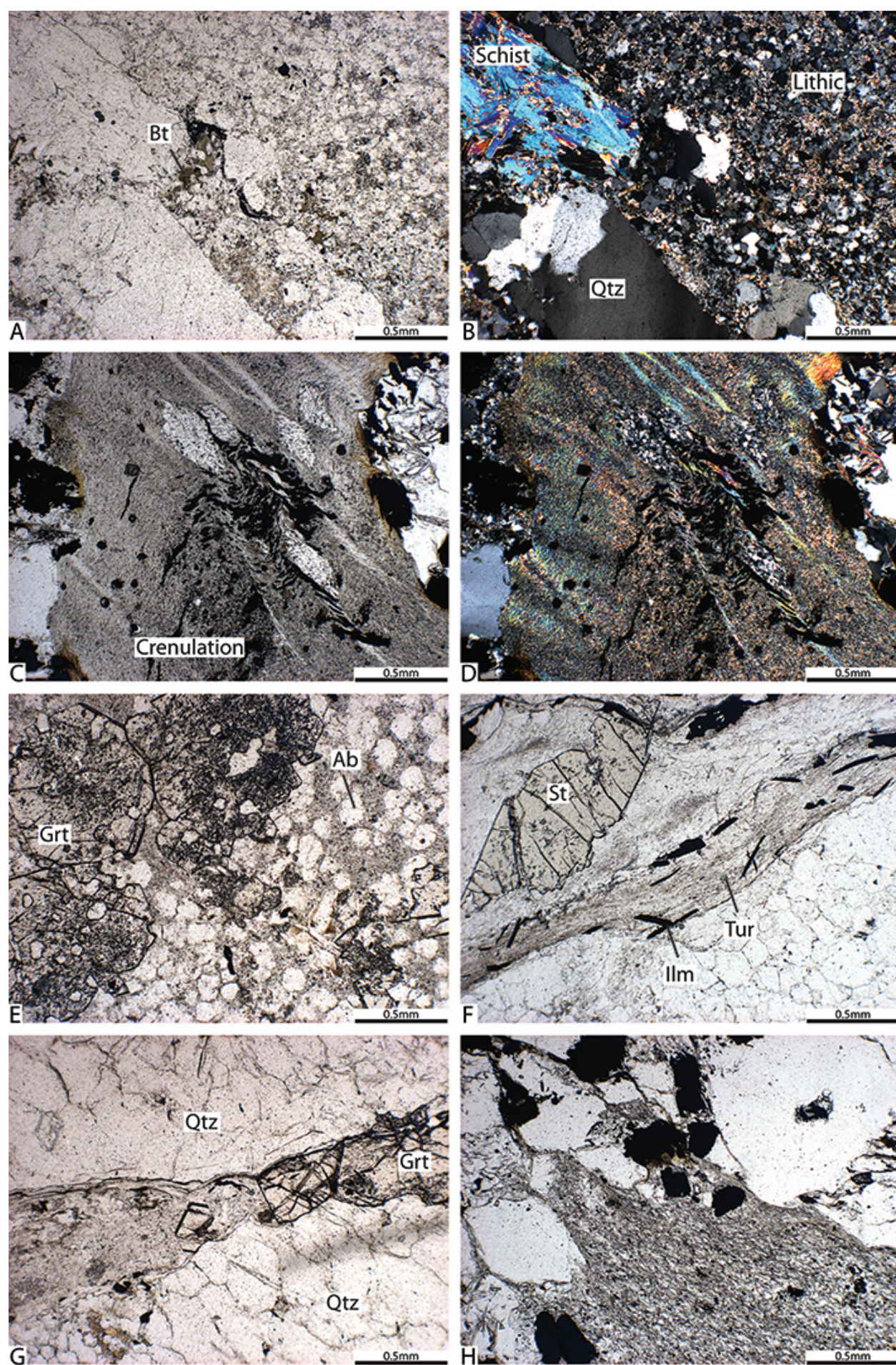


Figure 3.8: Photographs of Banket Series conglomerates from Damang. The lower Star Reef is dominated by coarse, rounded vein-quartz clasts (A-C), while the Malta Reef contains a greater proportion of lithic clasts (D-E).

porphyroblasts or matrix phases, suggesting that the Tarkwa region experienced lower grade peak metamorphism (Fig. 3.9H).

Pink Haematite altered Banket Series quartzites are pinkish-grey in colour but retain the characteristic heavy mineral concentrations along sedimentary bedding planes (Fig. 3.10A-C). A fine haematite dusting throughout the minor feldspar component produces the pink colour. This is, however, difficult to see, even using a scanning electron microscope, where occasional ‘Fe-spots’ are visible that are too fine to analyse precisely (Fig. 3.10G). This

Figure 3.9: (Following page) Photomicrographs of unaltered Banket Series conglomerates from Damang and Tarkwa. Predominant quartz, along with lithic and occasional schistose clasts set in a pelitic matrix (A, B). Certain schistose clasts exhibit a crenulation deformation, similar to that seen in the Birimian sediments at Damang (C, D). Metamorphic garnet and staurolite are common in pelitic matrix zones and lithic clasts (E-G), although these are absent from samples from Tarkwa (H). Sample AWDCm1 in plane-polarized light (A) and between crossed polars (B), sample AWTC1 in plane-polarized light (C) and between crossed polars (D), samples AWDCs1 (E, G), AWDCs3 (F) and AWTC2 (H), all in plane-polarized light.



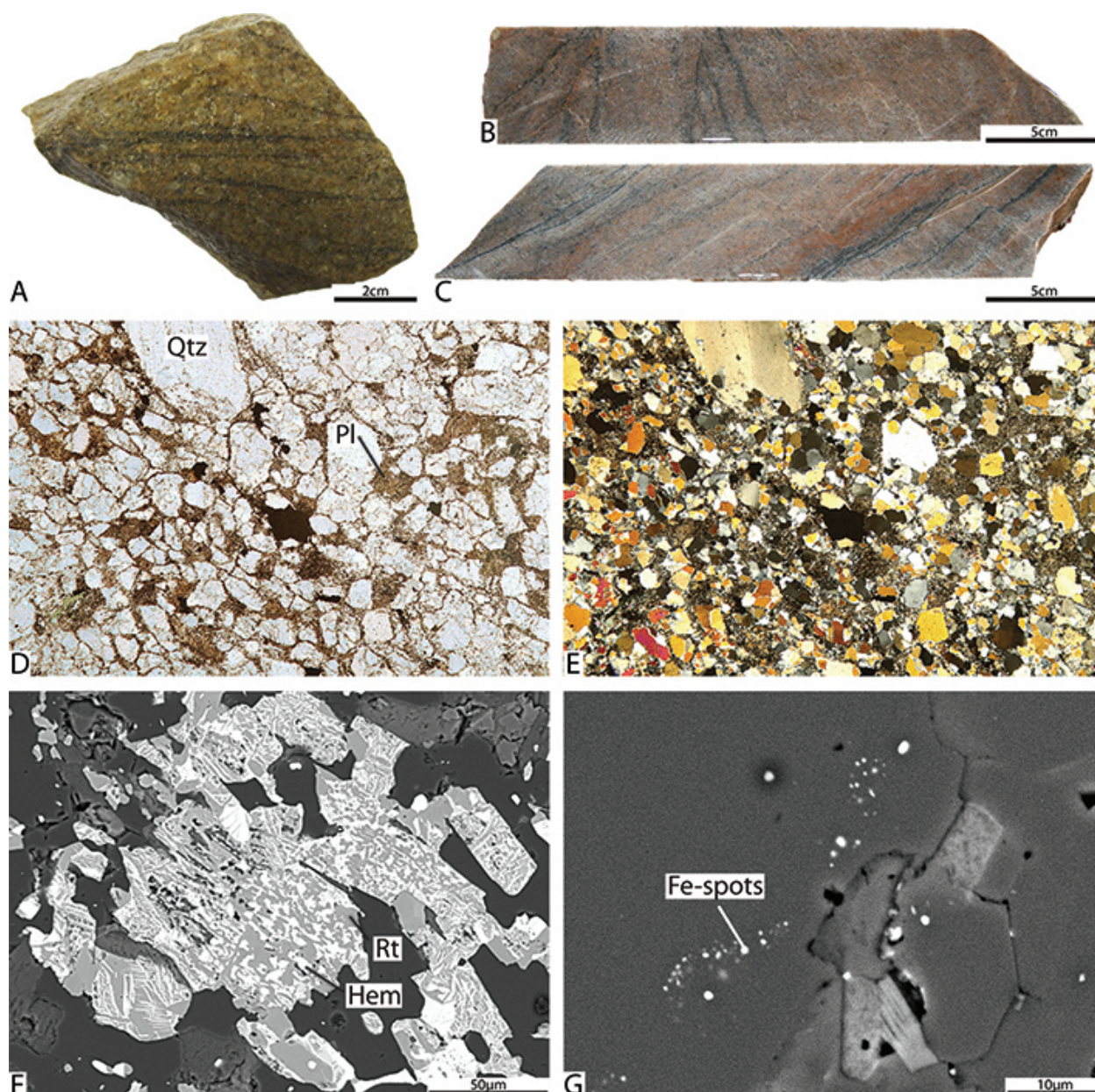


Figure 3.10: Photographs and photomicrographs of Pink Haematite altered Banket Series quartzites. Hand sample and core photographs (A-C) showing typical pink coloration and retention of sedimentary features. The pink colour results from a fine haematite dusting to feldspar grains and is accompanied by the development of chlorite, calcite, ankerite and muscovite (D, E). While magnetite becomes completely haematitized, ilmenite commonly exsolves to an intergrowth of rutile and haematite, as seen in backscattered electron images (F). The fine haematite dusting in feldspar is difficult to make out, even in backscattered electron images, showing only as occasional ‘Fe-spots’ (G). Sample AWABf2 in plane-polarized light (D) and between crossed polars (E), sample AWDBf4 (F, G). The yellow colour of quartz in E is due to the thin sections being too thick.

is accompanied by the development of chlorite, calcite and lesser ankerite, with additional muscovite (Fig. 3.10D, E). All magnetite is completely haematitized and ilmenite commonly exsolves to an intergrowth of rutile and haematite (Fig. 3.10F).

Mineralized Banket Series quartzites develop a paler colour overall along with extensive

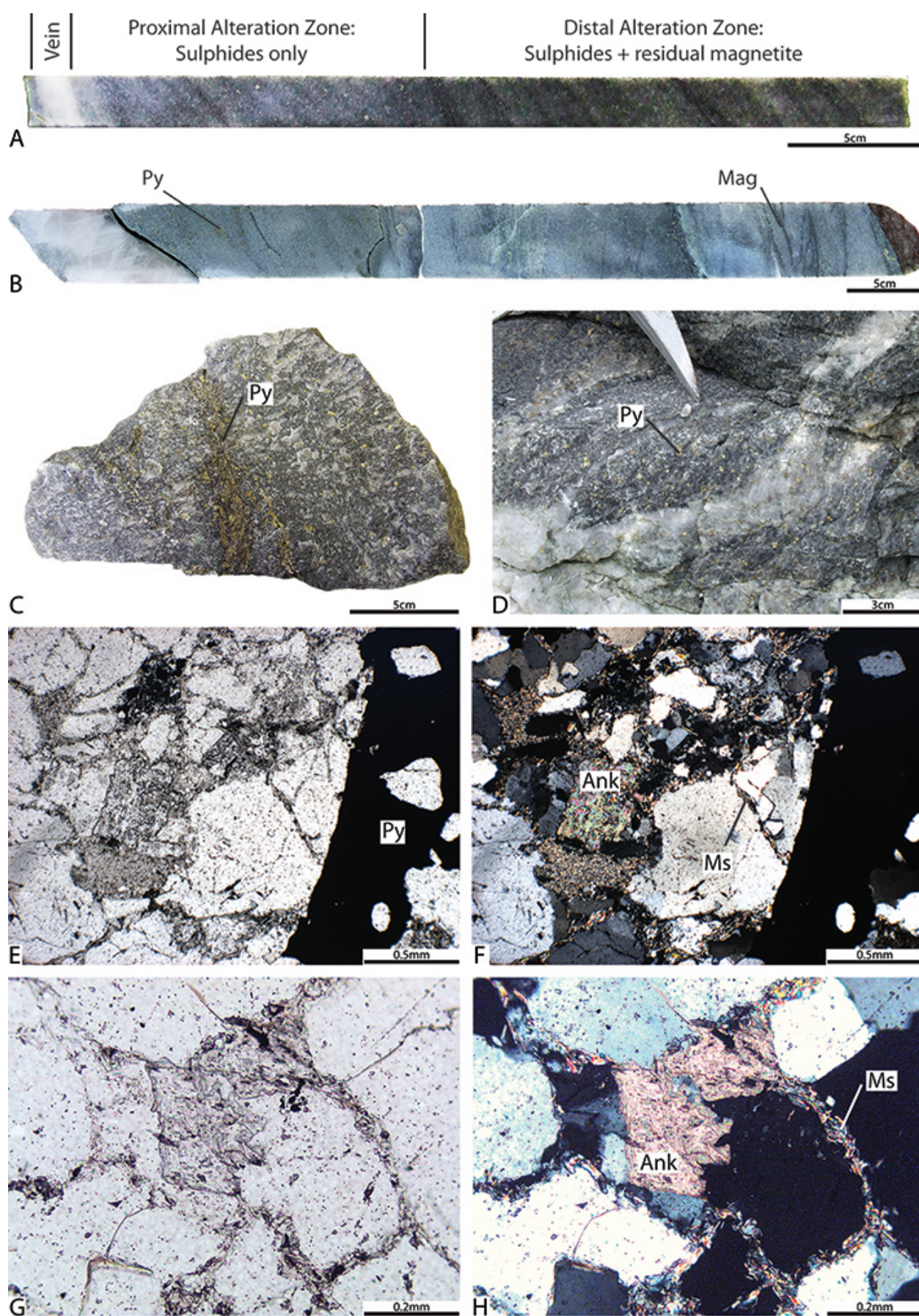


Figure 3.11: (Previous page) Photographs and photomicrographs of mineralized Banket Series quartzites. Quartzites develop extensive pyrite and a paler colour overall. A distal alteration zone still contains remnant magnetite, while a zone most proximal to the quartz vein contains no magnetite (A, B). Pyrite occurs either by selectively replacing magnetite along bedding planes (C) or as disseminations (D). This is accompanied by extensive ankerite and muscovite along grain boundaries (E-H). Core AMD113 (A), core ANRC017D (B), grab sample from the Damang pit (C) outcrop in the Damang pit (D), sample AWABf3 in plane-polarized light (E, G) and between crossed polars (F, H).

coarse pyrite, which occurs both as disseminations and as preferential replacement of magnetite along bedding planes (Fig. 3.11A-D). As such, the mineralized quartzites retain their sedimentary structures. This pyritization is progressive and a function of distance from the quartz vein, such that proximal and distal zones can be defined. The proximal zone comprises a complete replacement of magnetite, while within the distal zone, remnant magnetite is still present amongst the pyrite (Fig. 3.11A, B). The quartzites develop a matrix assemblage containing widely distributed ankerite in addition to a significantly increased modal proportion of fine muscovite, which preferentially grows along quartz and feldspar grain boundaries and produces the paler colour seen in hand sample (Fig. 3.11E-H).

3.4 Tarkwa Phyllite

The Tarkwa Phyllite is a finely laminated, pale- to medium-grey metapelite (Fig. 3.12). Typical Tarkwa Phyllite also has a speckled appearance with clearly identifiable, sub-millimetre-scale garnet and staurolite porphyroblasts along with black oxide phases. The laminations are somewhat irregular with pinch and swell features, indicative of an original sedimentary origin.

In thin section, a typical Tarkwa Phyllite is seen to comprise a porphyroblastic amphibolite facies mineral assemblage of Chl-Bt-Grt-St-Mag $\pm$ Ky $\pm$ Cld amongst a fine Qtz-Ms-Ab-Chl-Tur-Ilm matrix (Figs. 3.13 and 3.14). Porphyroblasts are randomly orientated, cross-cutting sedimentary laminations. They are generally euhedral with staurolite showing excellent cruciform twinning and garnets an equidimensional hexagonal shape (Fig. 3.14A-D). Biotite flakes are more irregularly shaped. Chlorite shows two distinct morphologies;

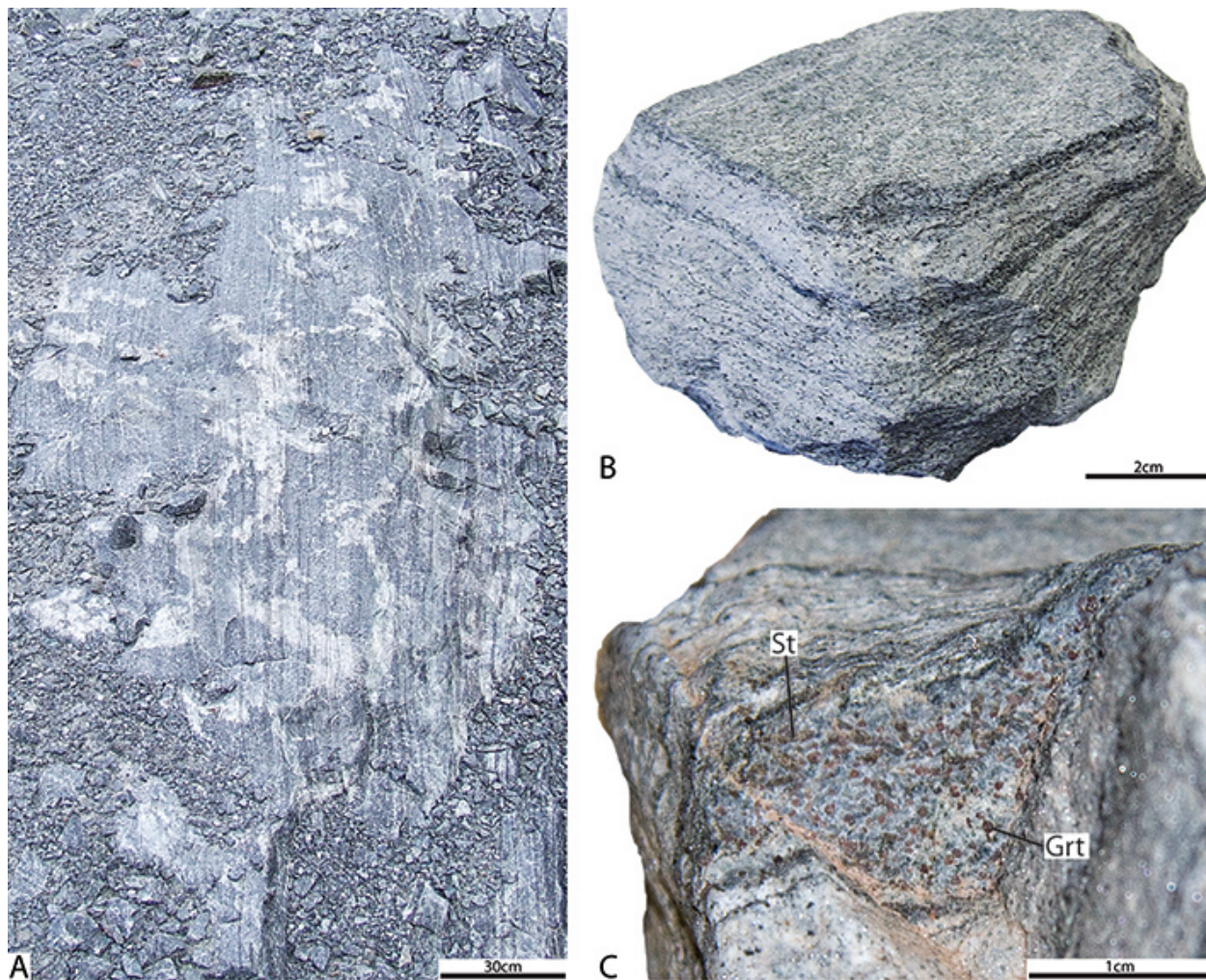


Figure 3.12: Pit outcrop (A) and hand specimen (B, C) photographs of unaltered, ‘typical’ Tarkwa Phyllite from Damang. Laminations are present on a mm- to cm-scale (A) and sub-mm-scale porphyroblast phases give a sparsely speckled appearance (B). Identifiable porphyroblast phases include garnet and staurolite (C).

as small, narrow flakes amongst the matrix and as coarse subhedral porphyroblasts. There is commonly a close spatial association between chlorite and biotite porphyroblasts, and in many cases individual biotite porphyroblasts are partially chloritized (Fig. 3.14A, B). Magnetite occurs as coarse-grained, sub-spherical compound grains dispersed throughout the matrix.

The matrix comprises small equigranular quartz and albite overprinted by fine decussate muscovite flakes and ilmenite grains (Fig. 3.14A, B). Occasionally, matrix feldspar forms together in coarse, elliptical, sub-mm zones, that are likely pseudomorphs after albite porphyroblasts (Fig. 3.13). The matrix also contains abundant dark blue-green tourmaline needles, making up as much as 20% of the matrix in certain samples. These are aligned in

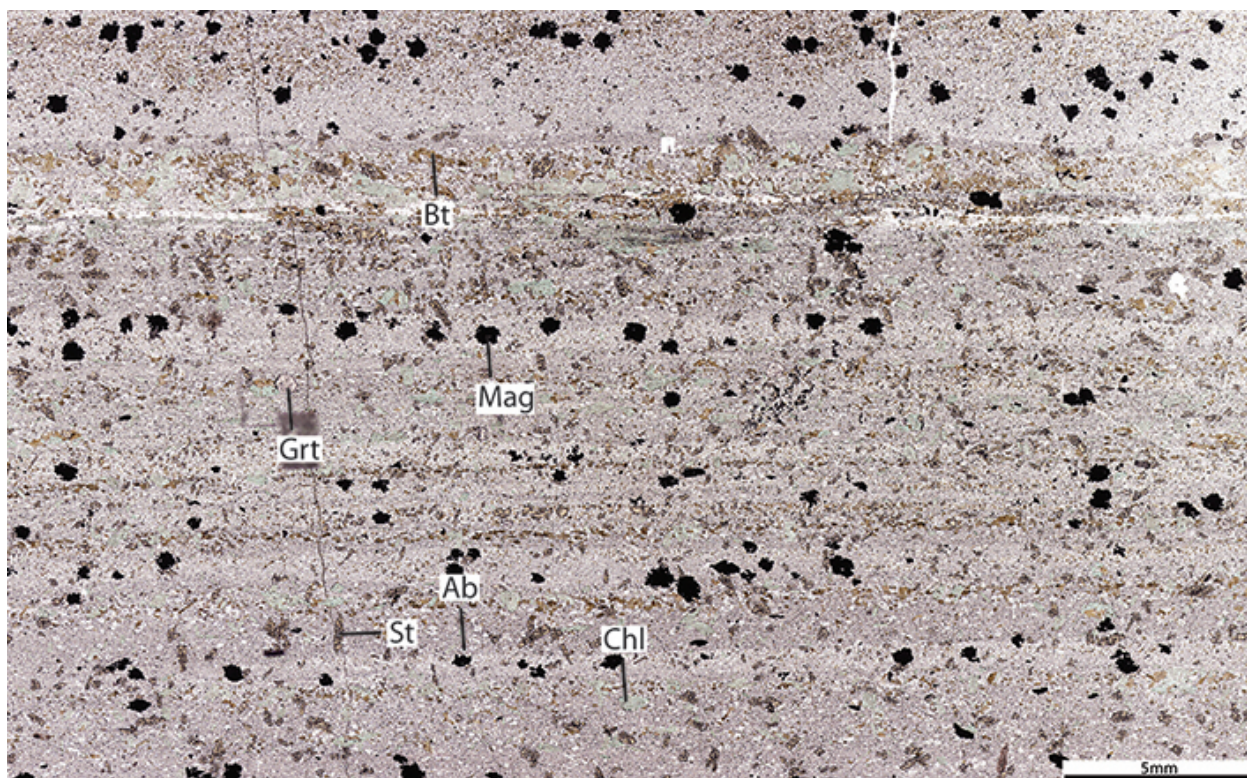


Figure 3.13: Whole slide scan of Tarkwa Phyllite sample AWDP1, illustrating the laminated fabric that is distinguished by varying sizes and proportions of metamorphic phases.

the plane of the laminations, suggesting that the boron content may be pre-metamorphic in origin (Fig. 3.15A, B). The matrix also contains abundant sub-10 μm monazite grains, which will be described and discussed further in chapter 5.

Other phases that are occasionally present include kyanite, such as in sample AWDP4 (Fig. 3.14D). This is interpreted to be a function of bulk composition since that sample is highly aluminous with little Fe, Mg and Ca, which restricts the growth of biotite and garnet, favouring staurolite and kyanite. Sample AWDP4 is mineralized although the presence of kyanite is not thought to be related to the mineralization process. The effect of bulk composition variation is important on a lamination-by-lamination scale, as the laminations are defined by variations in proportions of metamorphic minerals, resulting in the colour variations seen in hand specimen. Similarly, sample AWDBm1 contains a ‘seam’ of coarse garnets that appear to have nucleated in a thin layer of abundant very fine-grained apatite (Fig. 3.14E). These garnets contain inclusions of chloritoid, which has not been identified in other samples. Additionally, they partially envelope staurolite at their margins, indicating growth of garnet at least contemporaneous with staurolite.

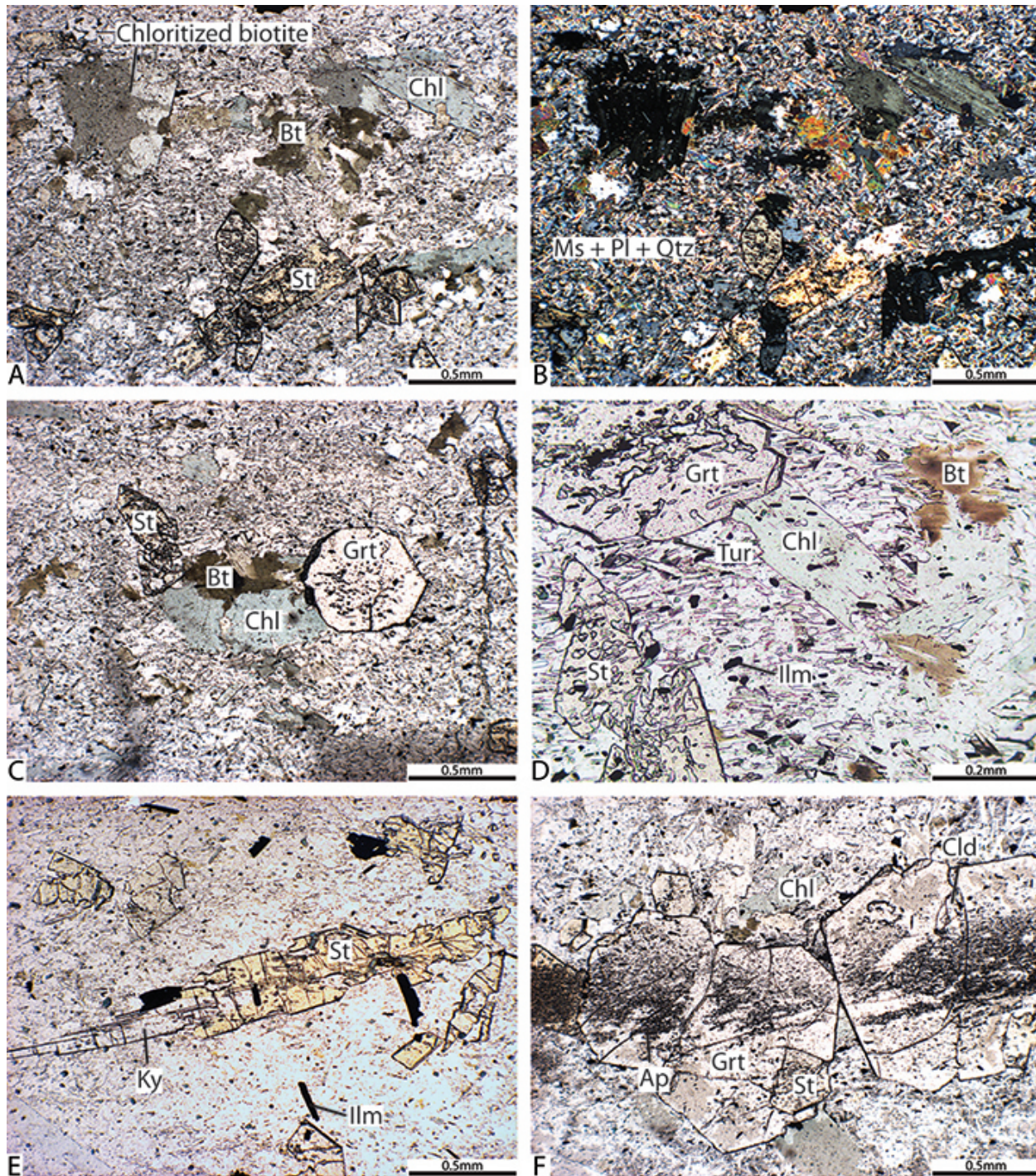


Figure 3.14: Photomicrographs of unaltered Tarkwa Phyllite from Damang. Abundant staurolite is euhedral and often twinned (A–D). Minor garnet is similarly euhedral (C, D) while biotite flakes are commonly chloritized (A, B). Samples occasionally contain kyanite (E) and chloritoid (F) depending upon local bulk composition variations. Sample AWDP1 in plane-polarized light (A, C) and between crossed polars (B), sample AWDBm1 in plane-polarized light (D, F). sample AWDP4 in plane-polarized light (E).

The Tarkwa Phyllite contains rare tourmalinite horizons that comprise between 30–80% tourmaline, along with quartz, muscovite and minor garnet (Fig. 3.15C, D). These tourmalinites are dark grey in colour and the majority contain gold mineralization. Tourmaline needles

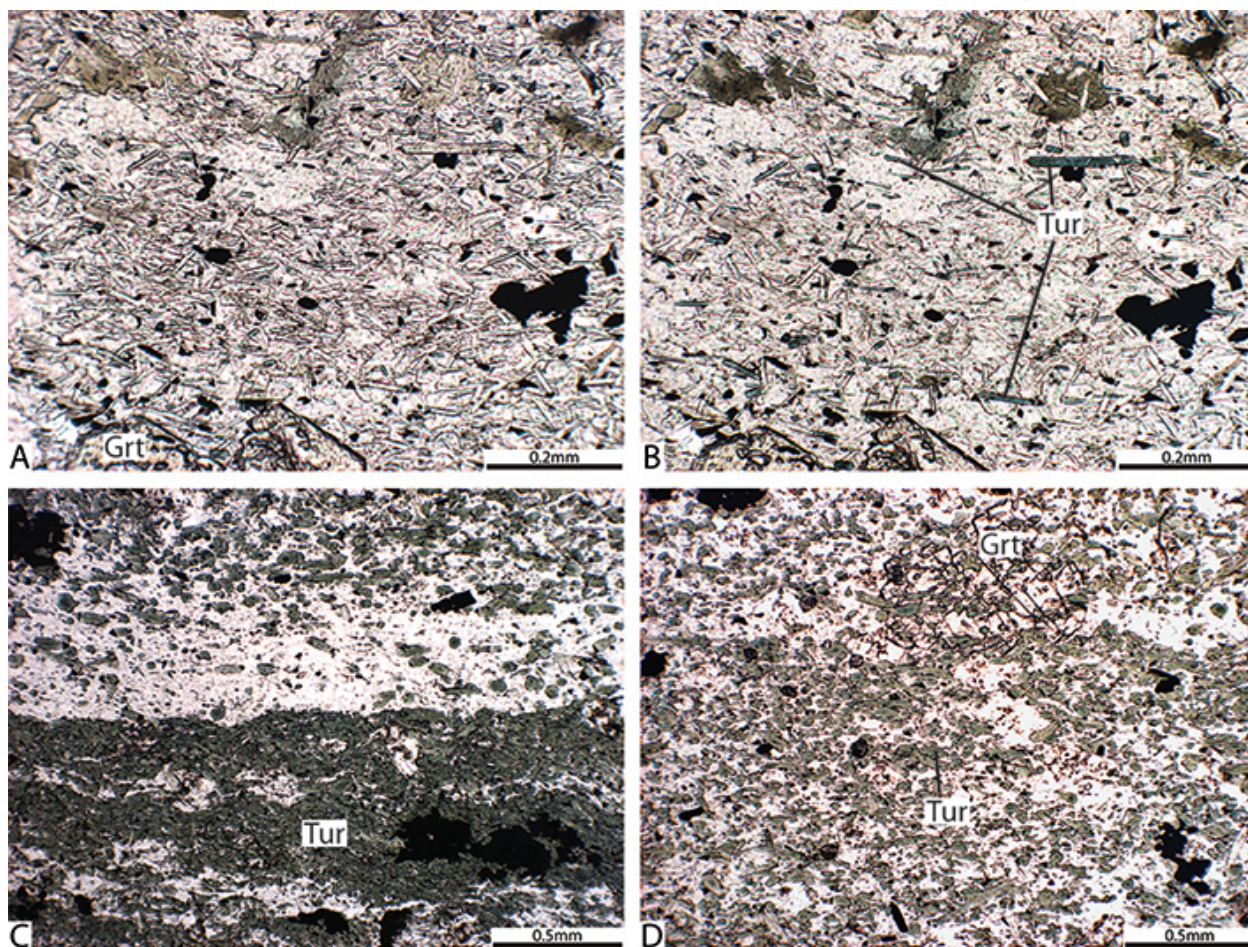


Figure 3.15: Photomicrographs of tourmaline in unaltered Tarkwa Phyllite from Damang. It occurs as a matrix phase in ‘typical’ Tarkwa Phyllite (A, B) and in mineralized tourmalinites (C, D). In both cases tourmaline needles are aligned in the plane of the laminations (needles seen side-on, orientated E–W, or as end-on cross-sections) and are included within metamorphic minerals, indicating a pre-metamorphic origin. Sample AWDP1 (A, B) shown in 90° orientations to enhance tourmaline visibility when the ordinary ray is aligned with the polarizer (B). Sample AWDP6 (C, D) showing almost 100% tourmaline layers (C) and tourmaline as inclusions in garnet (D).

are again aligned within the plane of the laminations, suggesting a pre-metamorphic and pre-mineralization origin. This is further confirmed by the presence of tourmaline inclusions within garnet, which are continuous with the matrix fabric (Fig. 3.15D). Discrete tourmalinite horizons such as these have been recognized in metamorphic terranes across the world (Slack *et al.*, 1984). The tourmalinites at Damang are particularly comparable to those documented in the Namaqualand Metamorphic Complex, Bushmanland, South Africa. Here, they display an identical mode of occurrence and compositional range (see section 3.7) and are hosted in a similar volcano-sedimentary succession, deposited in a rifted setting (Willner *et al.*, 1990; Willner, 1992). The Namaqualand tourmalinites have been interpreted to have formed during submarine sediment-hosted exhalative processes (Willner, 1992), which also

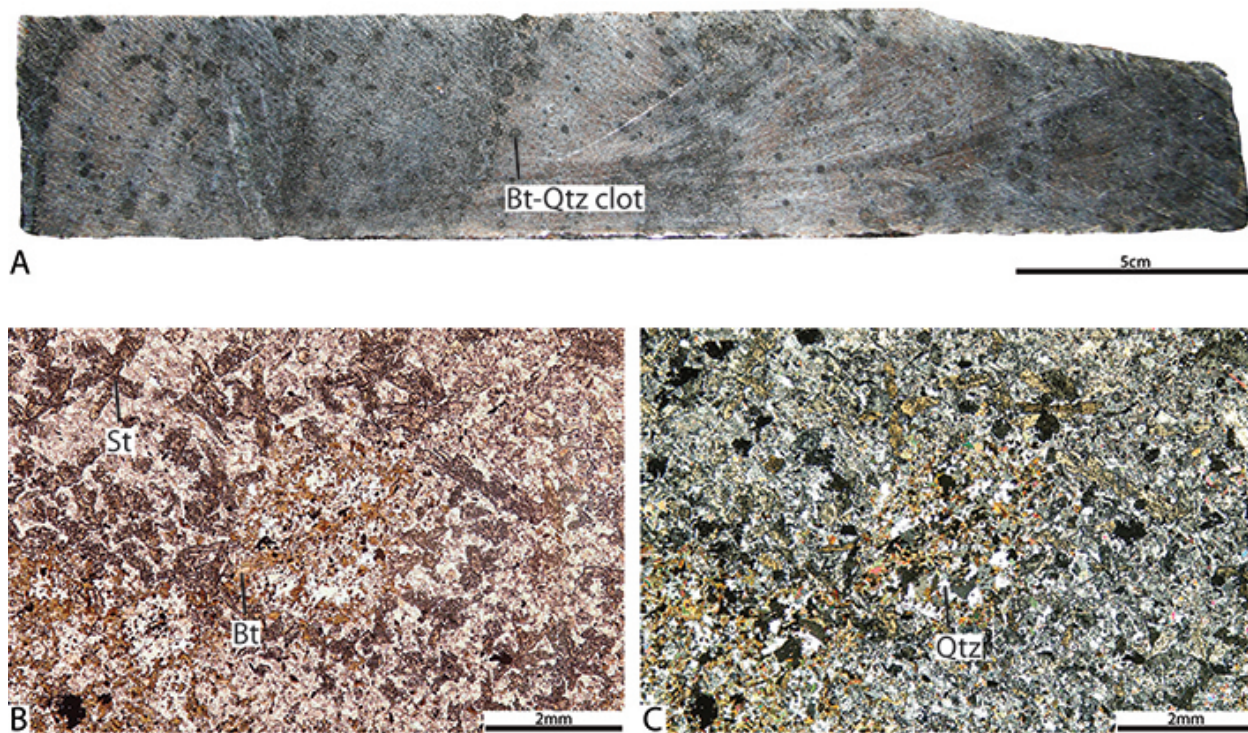


Figure 3.16: Core photograph and photomicrographs of hornfelsed Tarkwa Phyllite from Damang. Overall characteristics are similar to unaltered Tarkwa Phyllite, but include the presence of 2–5 mm biotite-rimmed clots of quartz that are distributed throughout. DRC220D (A), sample AWDP5 in plane-polarized light (B) and between crossed polars (C).

explains the common association between tourmalinites and sediment-hosted exhalative base metal and tungsten deposits such as Broken Hill, Australia (Slack *et al.*, 1984; Plimer, 1984, 1987). While no such deposits have been discovered in Ghana, this link is still intriguing.

Another important variety of Tarkwa Phyllite is locally termed ‘hornfelsed’ and occurs in association with some, although not all, of the dolerite intrusives. Hornfelsed Tarkwa Phyllite is similar to that described above but contains additional biotite, with lesser chlorite. The additional biotite commonly occurs as a rim surrounding spherical clots of quartz (Fig. 3.16). The precise origin of these features is unclear. Superficially, they resemble cordierite spots and could indeed represent pseudomorphs after some contact metamorphic mineral, although the pure biotite–quartz products make this difficult to determine.

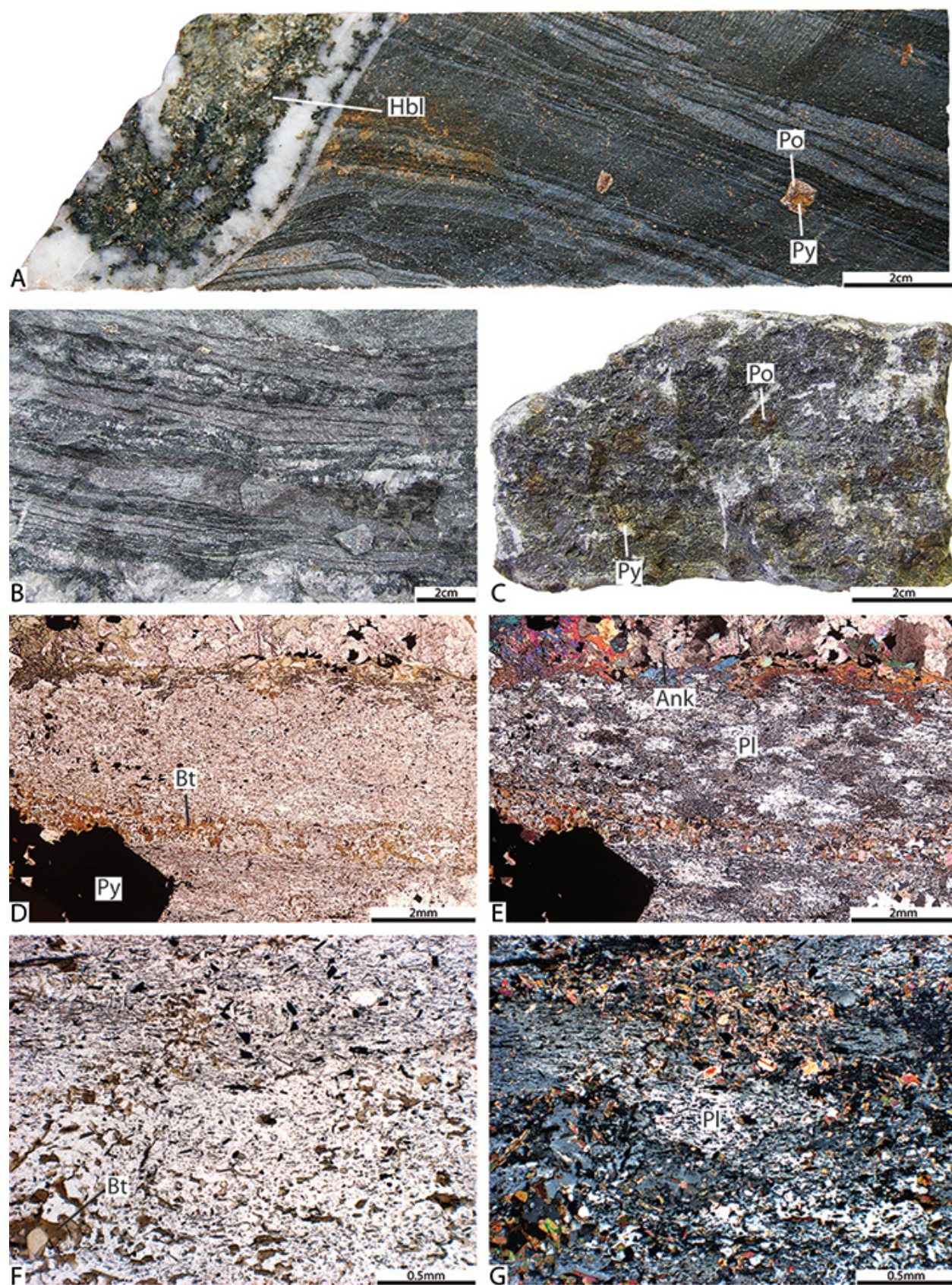
The Tarkwa Phyllite does not readily undergo Pink Haematite alteration, unlike the more quartzo-feldspathic Banket Series and Huni Sandstone.

Mineralized Tarkwa Phyllite takes on a darker grey colour but retains its distinctive lam-

inated fabric (Fig. 3.17A-C). Coarse pyrite and pyrrhotite are extensively developed and overprint the laminations. Mineralized samples are often cut by numerous narrow, discordant or bedding-parallel veins with a range of mineralogies (see section 3.9.1). Progressive destruction of metamorphic minerals during alteration produces an assemblage of coarsely recrystallized, interlocking feldspar poikiloblasts with inclusions of muscovite and abundant ankerite along with tourmaline, ilmenite and accessory remnant staurolite, biotite and other metamorphic minerals (Fig. 3.17D-G). The coarse sulphides clearly cross-cut the fabric but generally exhibit rather irregular crystal edges that are rimmed by fine, irregular biotite.

Recent drill intersections of the Damang Fault (section 2.8.1) suggest that it commonly develops within the Tarkwa Phyllite, immediately below the base of the Huni Sandstone. The rocks in these zones contain abundant ankerite and recrystallized feldspar (Fig. 3.18A-E), typical of Tarkwa Phyllite that has undergone gold-associated alteration. There is, however, a notable lack of sulphide minerals. Sedimentary laminations are modified by tectonic deformation to produce microfolds (Fig. 3.18G) along with microstructural shear bands, biotite-fish and sigma-type biotite porphyroblasts (Fig. 3.18F) indicating a simple-shear component, consistent with fault movement. There is also a brittle component to this otherwise plastic deformation, with diffuse ‘fractures’ forming in hinge zones between more competent fold limbs (Fig. 3.18H). These structural features are overprinted by metamorphic garnets (Fig. 3.18G). These intersections indicate that deformation associated with movement on the Damang Fault occurred during prograde metamorphism, between the growth of biotite and garnet. Extensive carbonate alteration suggests that the Damang Fault acted as a fluid pathway during the mineralization event. However, the lack of sulphidation resulted in no economic gold mineralization within the fault zones examined.

Figure 3.17: (Following page) Photographs and photomicrographs of mineralized Tarkwa Phyllite from Damang, which is notably darker than the unmineralized variety (A-C). While still laminated, the dark colour can sometimes obscure the laminations somewhat (C). Coarse, cross-cutting sulphides are common, as are discordant or bedding-parallel veins (A-C). The matrix is now dominated by coarsely recrystallized poikiloblastic feldspar containing inclusions of muscovite, biotite, ankerite and relic metamorphic minerals (D-G). Sulphide phases are coarse-grained and overprint the fabric (D, E). Biotite-quartz and actinolite-ankerite veins are common (D, G). DRC1394D (A), outcrop in Damang pit (B) and sample AWDP4 (C), sample AWDP2 in plane-polarized light (D, F) and between crossed polars (E, G).



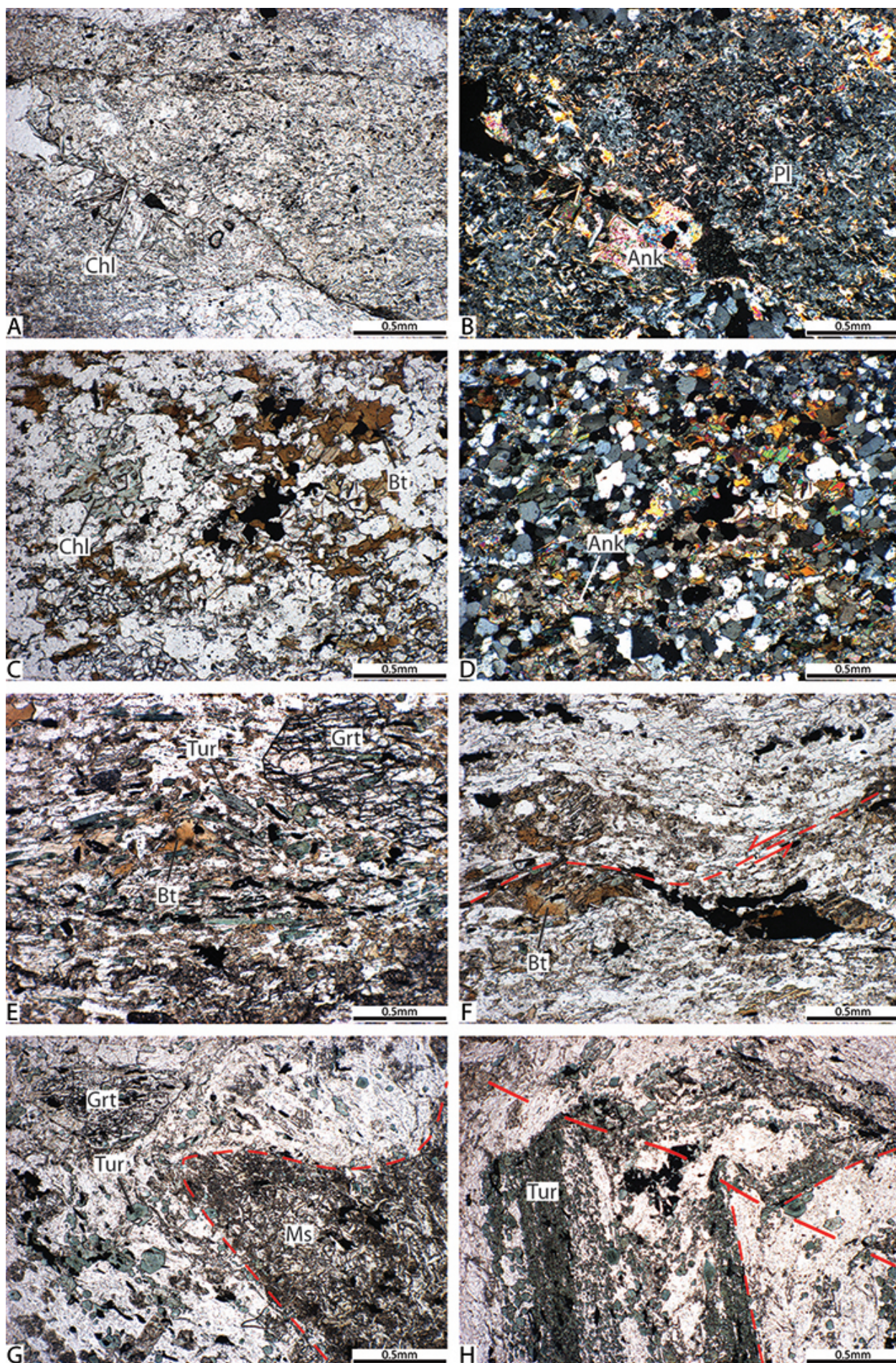


Figure 3.18: (Previous page) Photomicrographs of the Damang Fault zone, hosted in the Tarkwa Phyllite. Fault zone rocks are similar to typical Tarkwa Phyllite and exhibit gold-associated alteration with the development of ankerite and extensive recrystallized feldspar (A-E). Sedimentary laminations are modified by tectonic deformation producing shear bands and biotite-fish (F) and microfolding (G, H). These features are overprinted by later metamorphic garnet growth (G). Deformation retains a brittle component with fractures developing in hinge zones of microfolds (H). Sample AWDF3 in plane-polarized light (A) and between crossed polars (B), sample AWDF1 in plane-polarized light (C, E, F) and between crossed polars (D) light, sample AWDF2 in plane-polarized light (G, H).

3.5 Huni Sandstone

The Huni Sandstone is a fine- to medium-grained, dark grey, arenaceous metasediment (Fig. 3.19A). Although generally massive and unstructured, weak laminations are occasionally visible (Fig. 3.19B). It is arkosic, comprising an interlocking mass of quartz and feldspar, the latter making up about 35% of the rock on average (Fig. 3.19C, D). Disseminated biotite, chlorite and carbonate are common along with lesser amounts of very fine-grained muscovite in interstitial clots or along grain boundaries (Fig. 3.19E, F). Small amounts of poorly-formed garnet along with ilmenite and magnetite are also typically present (Fig. 3.19E, F).

Pink Haematite altered Huni Sandstone is typically pinkish-grey in colour and retains the texture of the unaltered variety. In certain locations however, the alteration can become extremely intense with the rock developing a distinctive, bright pink-orange colour and taking a sugary, crystalline texture (Fig. 3.20A, B). A very fine haematite dust throughout the feldspar crystals is responsible for the pink coloration, although the feldspars also appear to be partially muscovitized (Fig. 3.20C-F). Other alteration phases include an increase in the proportion of carbonate (calcite with lesser ankerite), chlorite and chloritized biotite. Any magnetite in the rock becomes partially or completely haematitized (Fig. 3.20G, H).

The Huni Sandstone is typically rather poorly mineralized compared to the Banket Series and Tarkwa Phyllite, although it is heavily veined and does occasionally develop disseminated pyrite around quartz veins (Fig. 3.21). There are no other distinctive alteration features related to gold mineralization. On a large scale, however, there appears to be a greater degree of quartz veining in broadly Pink Haematite altered Huni Sandstone, as can be seen in the western wall of the main Damang pit. (Fig. 3.22).

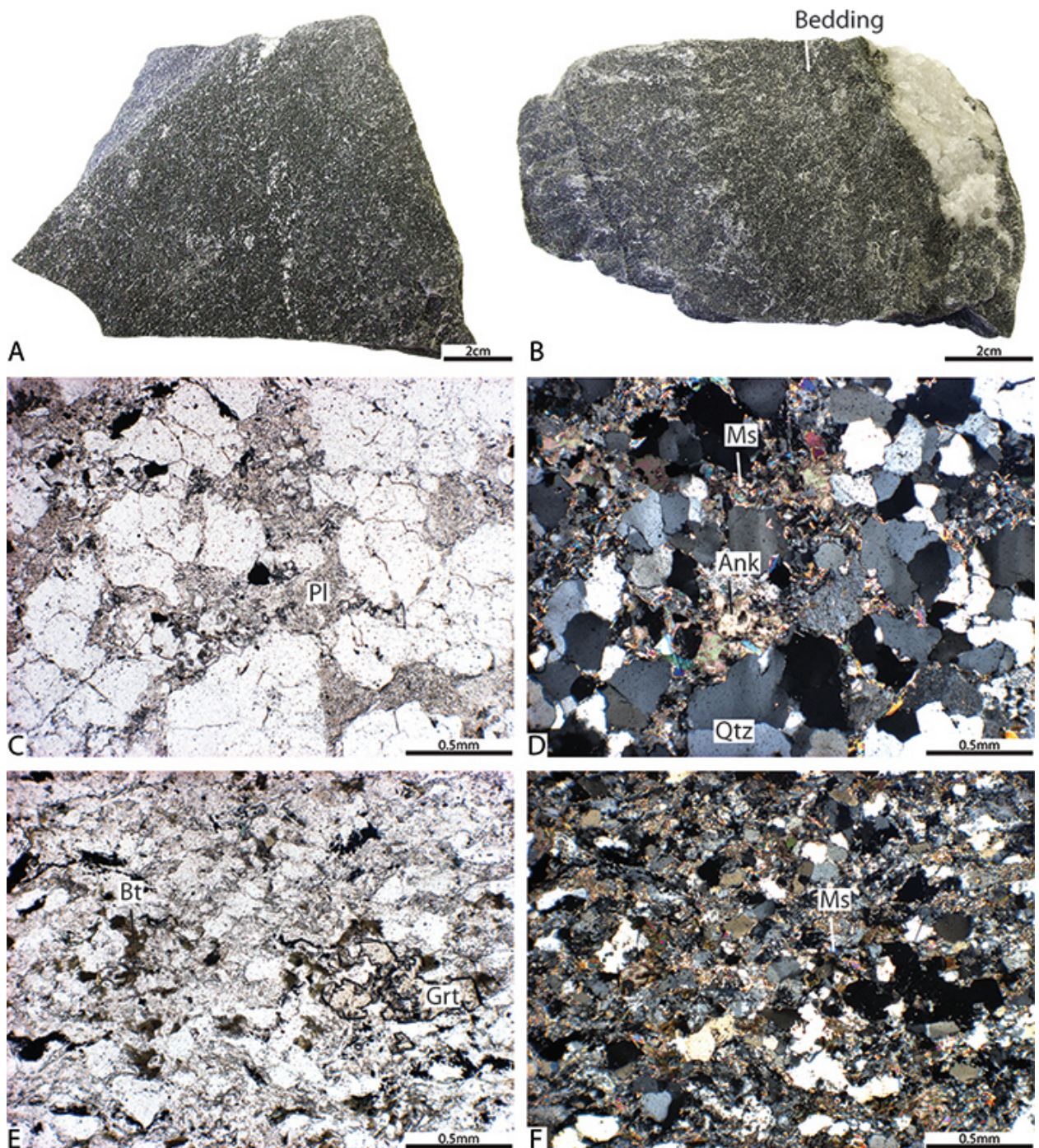


Figure 3.19: Photographs and photomicrographs of unaltered Huni Sandstone from Damang. It is typically massive and dark grey in colour (A), although occasionally displays faint laminations (B). Quartz and feldspar grains are set in a carbonate-muscovite matrix (C, D) along with small amounts of garnet, biotite and chlorite. Sample AWDH1 (A) in plane-polarized light (C) and between crossed polars (D), sample AWDH4 in plane-polarized light (E) and between crossed polars (F).

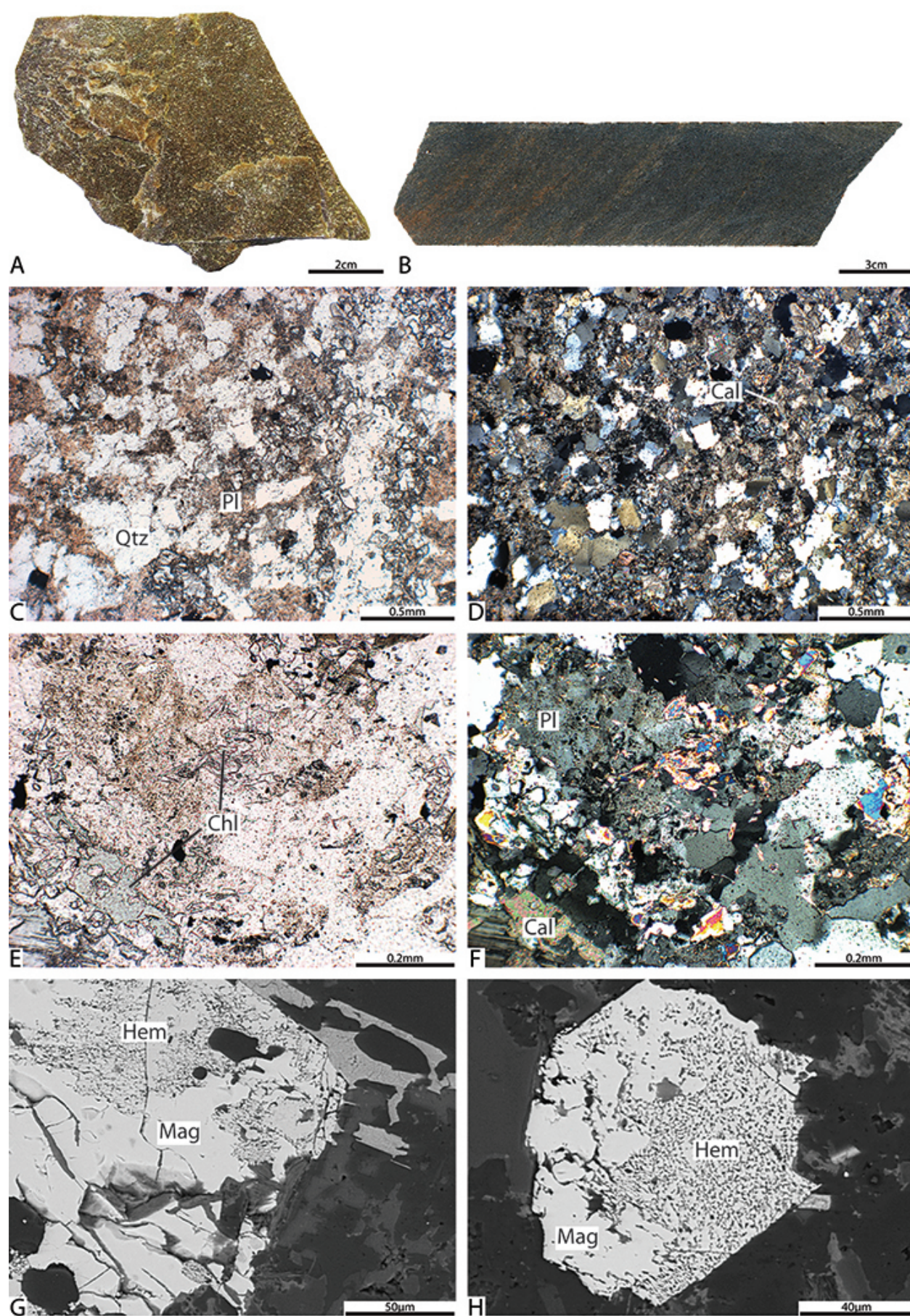


Figure 3.20: (Previous page) Photographs and photomicrographs of Pink Haematite altered Huni Sandstone. Altered rocks take on a characteristic orange-pink colour and develop a silicified texture (A). Alteration occurs favourably in coarse grained horizons (B). The pink colour is due to heavily haematite dusted feldspar and occur with abundant carbonate and rare quartz-carbonate veinlets (C-F). Magnetite becomes haematitized (G, H). Sample AWDH3 (A) in plane-polarized light (C) and between crossed polars (D) and backscattered electron images (G, H), sample AWDH2 in plane-polarized light (E) and between crossed polars (F).



Figure 3.21: Hand specimen photograph of sulphide-bearing, barren Huni Sandstone from Damang.



Figure 3.22: Dense mineralized quartz veining in the western wall of the Damang pit correlates with a broad zone of Pink Haematite altered Huni Sandstone.

3.6 Dolerite Intrusions

The 2–20 m thick doleritic intrusions are dark green in colour, due to significant proportions of amphibole, along with varying amounts of feldspar. The proportion of feldspar is used as a discriminator between mafic ‘dolerites’ and ‘diorites’, the latter of which have a distinct mottled appearance and contain as much as 50–60% feldspar plus quartz. This distinction has recently been implemented by exploration geologists at Damang, although, there exists a complete observable range between dolerites and diorites (Fig. 3.23), so they will be collectively termed ‘dolerites’ throughout this thesis. Observed contacts between dolerite and diorite are either knife-sharp boundaries (Fig. 3.23E) or intermingled ‘chilled’ zones with a notably finer grain size (Fig. 3.23F). These are only ‘apparent’ contacts, however, as will be discussed below. Texturally they are all massive, irrespective of their mineralogy, intrusion size or structural location. Intrusions are medium- to coarse-grained although many of the narrower dykes (such as those visible in the west wall of the Damang pit) are much finer grained.

Unaltered intrusions are dominated by blue-green hornblende and plagioclase feldspar with lesser epidote, chlorite, ilmenite $\pm$ magnetite $\pm$ rutile. Original igneous textures are commonly present with euhedral hornblende and large, twinned plagioclase laths (Fig. 3.24A-D), although this is affected by a metamorphic overprint (Fig. 3.24E-H). Epidote most commonly occurs as tiny inclusions within hornblende while the chlorite is likely a retrograde alteration product.

In addition to a greater proportion of feldspar, dioritic intrusions are distinguished by the presence of quartz (Fig. 3.24C-D) and commonly, intergrowths of quartz and feldspar (Fig. 3.24G-H). Oxide phases (Mag + Ilm $\pm$ Rt) are notably more common in the dioritic end-members.

Pink Haematite altered intrusives develop a distinctive pink colour with poorly-discernible mineralogy in hand specimen (Fig. 3.25A). Within some particularly large Pink Haematite zones, the alteration can transition into and out of a green colour that is interpreted to

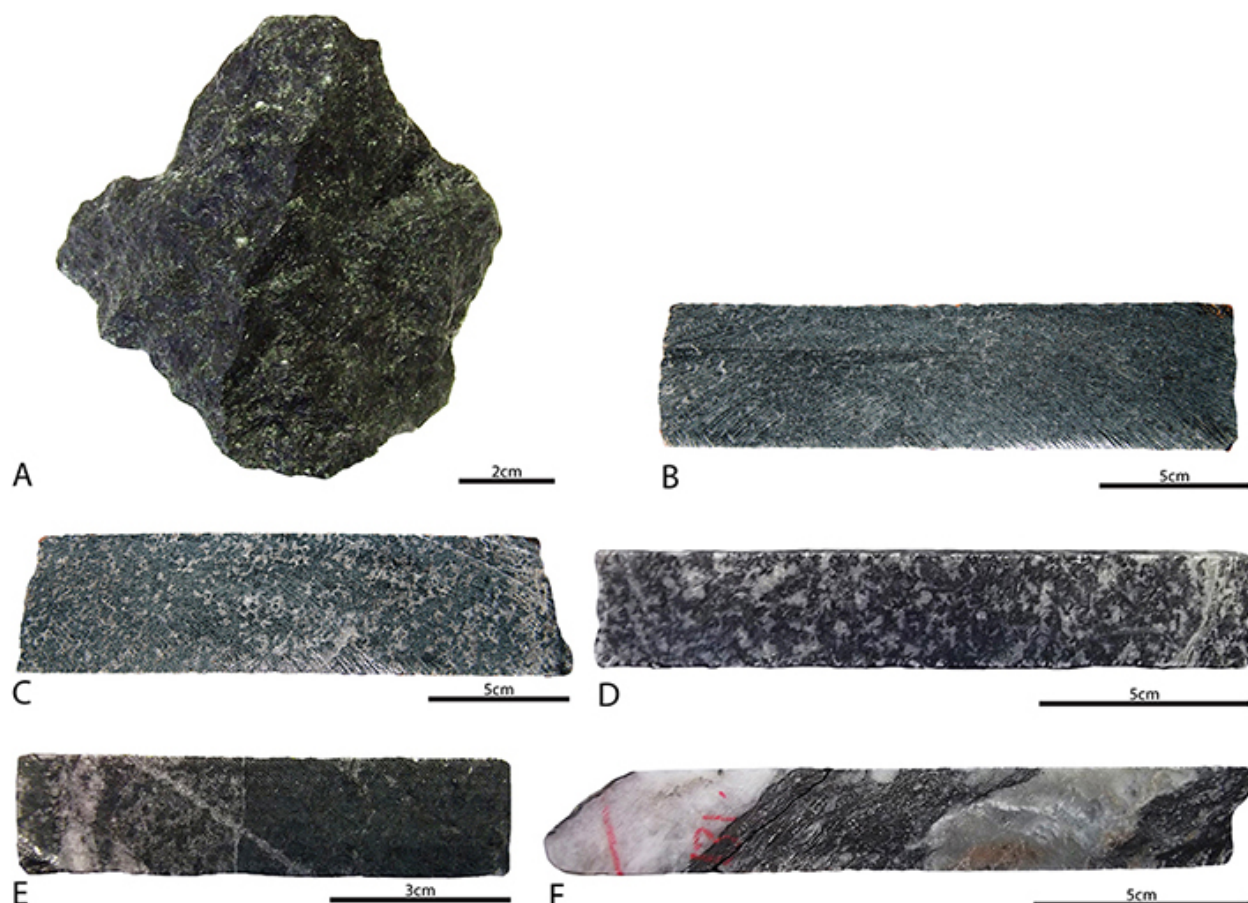
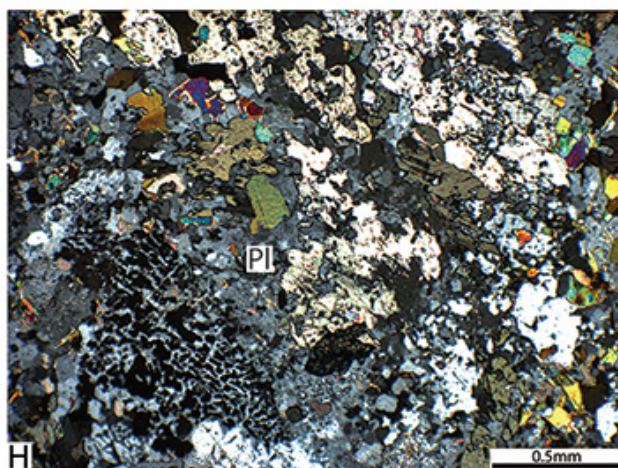
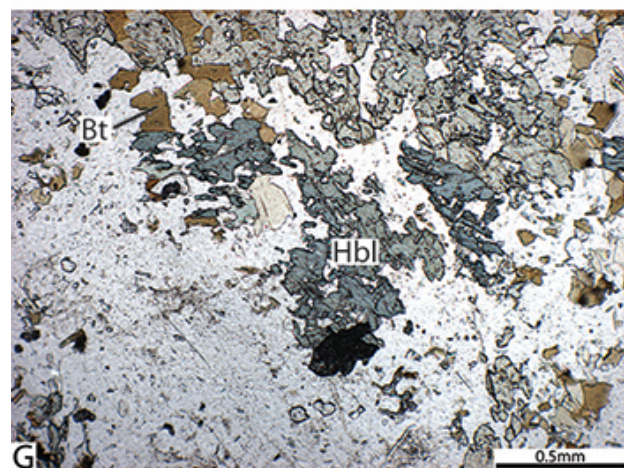
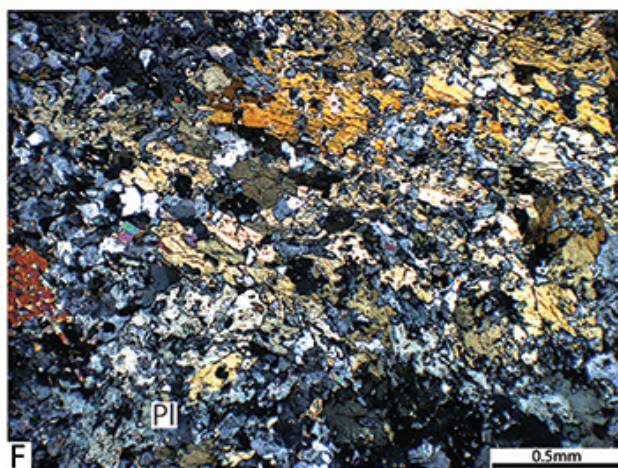
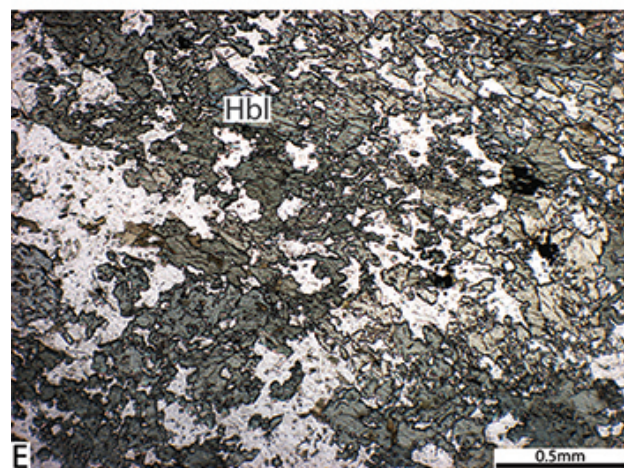
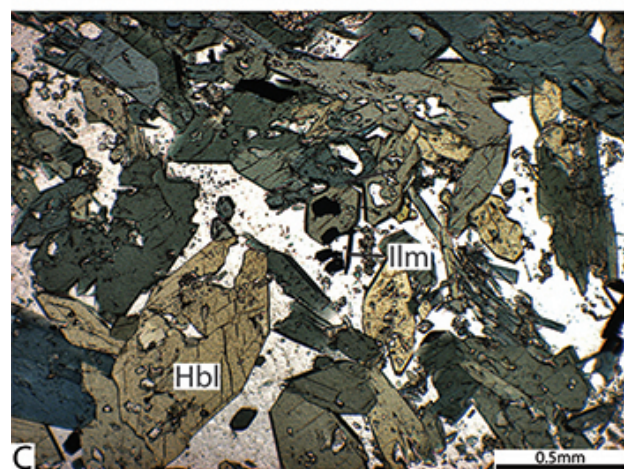
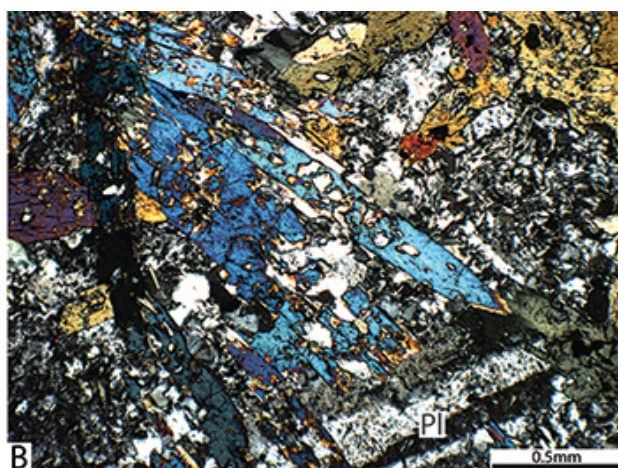
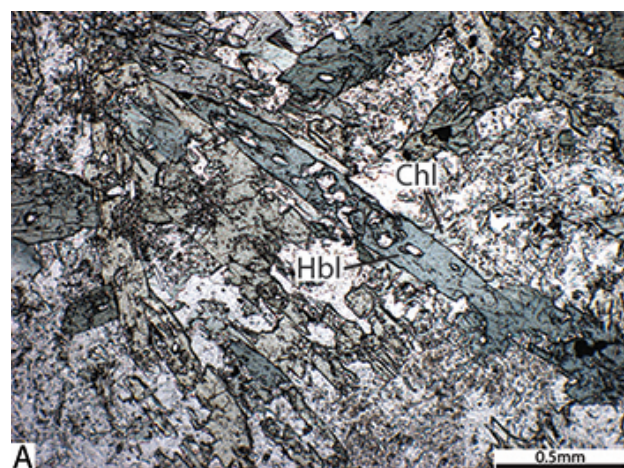


Figure 3.23: Hand specimen photographs of unaltered mafic intrusives from Damang. (A) Coarse-grained mafic dolerite. Sample AWDDO3. (B) A dolerite with visible grey feldspar. Sample AWDDo15. (C) An intermediate dolerite showing a mottled appearance. Sample AWDDo16. (D) A 'diorite'. Sample PS02. (E) Apparent contact between 'dolerite' and 'diorite'. Sample PS09. (F) Apparent 'chilled' contact between dark mafic dolerite and paler mottled diorite. Sample PS04.

be a function of oxidation state (Fig. 3.25B). The pink colour is the result of both a fine haematite dusting to feldspar crystals and the growth of extensive pink carbonate (predominantly calcite with additional ankerite) (Fig. 3.25C, D). Additional minor alteration phases include patchily developed chlorite and occasional muscovite. This alteration accompanies a complete destruction of amphibole. The rock does, however, retain aspects of the original igneous texture with visible feldspar laths.

Figure 3.24: (Following page) Photomicrographs of unaltered mafic intrusives from Damang. Sample AWDDo3, a coarse-grained mafic dolerite in plane-polarized light (A) and between crossed polars (B). Sample PS02, a coarse-grained diorite in plane-polarized light (C) and between crossed polars (D). Sample PS09, a medium-grained mafic dolerite in plane-polarized light (E) and between crossed polars (F). Sample AWDDo1, a partially altered diorite showing intergrowths of quartz and feldspar in plane-polarized light (G) and between crossed polars (H).



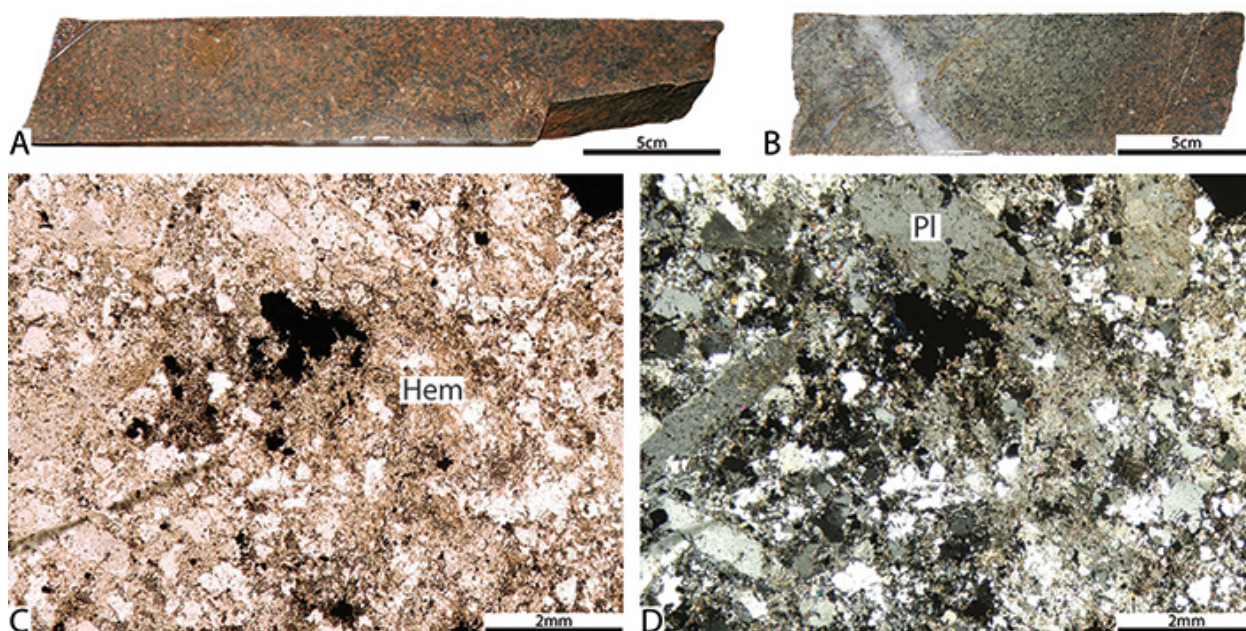


Figure 3.25: Photographs and photomicrographs of Pink Haematite altered dolerite intrusives showing the distinctive pink colour (A) and a transition into a green alteration zone (B). The pink colour is produced by a haematite dusting of feldspars and extensive development of pink carbonate with additional minor chlorite. Sample AWDDo12 (A) in plane-polarized light (C) and between crossed polars (D), sample AWDDo11 (B).

Mineralized intrusives become variably ‘bleached’ through alteration, resulting in a paler colour overall. Alteration assemblages comprise variable amounts of $Bt + Ank + Qtz + Po \pm Chl \pm Py \pm Tur$, coupled with a complete destruction of hornblende and most oxide phases (Figs. 3.26, 3.27 and 3.28). The proportions of each alteration phase are strongly controlled by original mineralogy and results in a range of observed colours from grey (predominant) through brown to green (Fig. 3.26A–F). For example, dolerites, with a large proportion of amphibole, typically alter to a very biotite-rich assemblage, while diorites tend to become more dominated by carbonate. Similarly, tourmaline only occurs as a major alteration phase in cases where it also occurs within the enclosed vein (Fig. 3.27B). Sulphides are significantly more abundant in the diorites compared to the dolerites. The predominant sulphide phase is pyrrhotite although lesser pyrite also occurs, often rimmed by pyrrhotite. In addition, narrow bands of carbonate-rich alteration commonly occur within larger intrusive bodies (Fig. 3.26G–H). A gradual transition exists between the metamorphic and hydrothermal assemblages (Figs. 3.27 and 3.28G–H). Alteration phases are patchily developed and do not preserve the texture of the original rock.

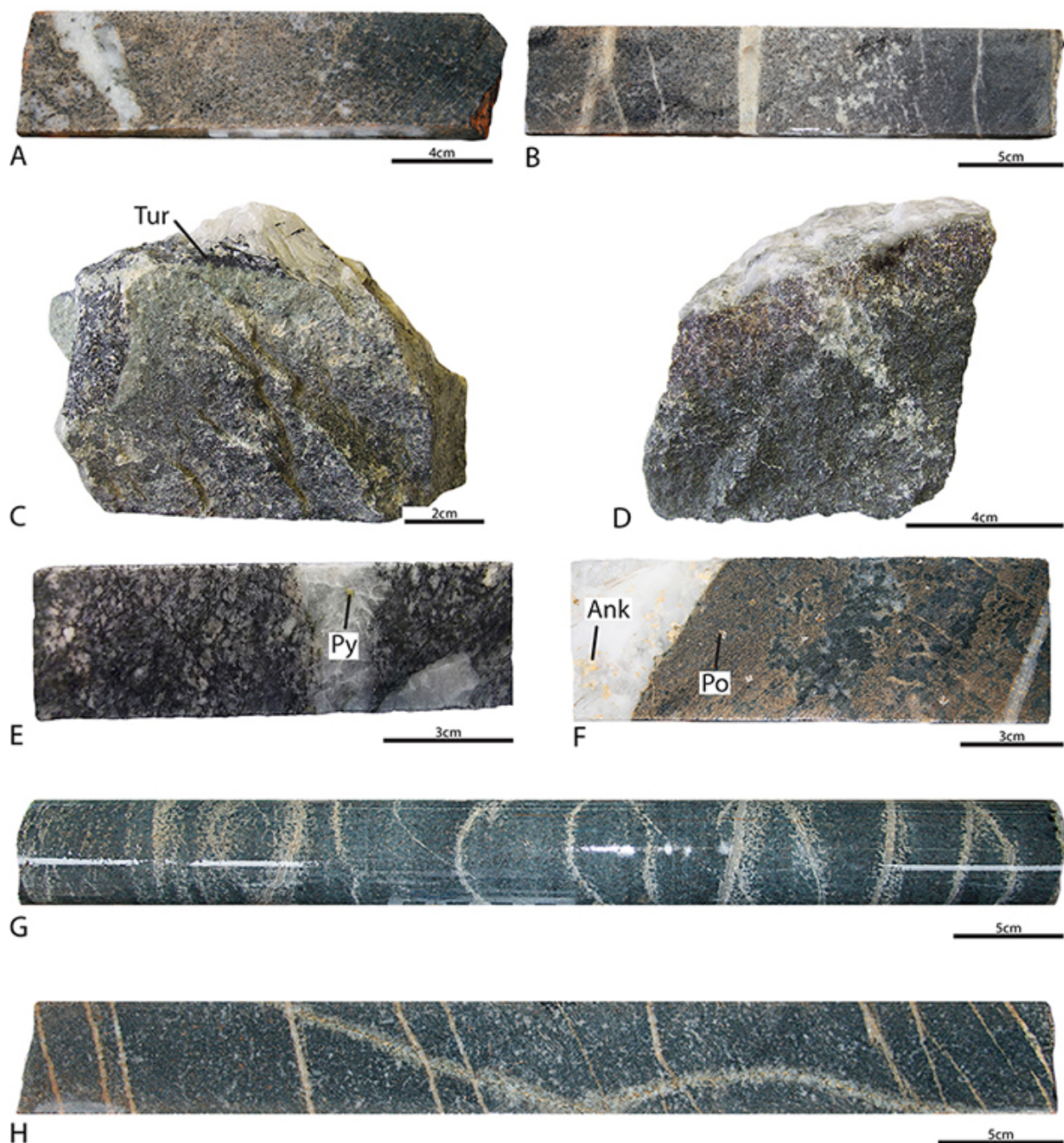


Figure 3.26: Hand specimen photographs of mineralized mafic intrusives from Damang. (A) Carbonate dominated 'brown' alteration of a dolerite. (B) Quartz-feldspar dominated 'grey' alteration of a diorite. (C) Green carbonate-chlorite-tourmaline alteration of an intermediate dolerite. Sample AWDDo4. (D) Biotite dominated alteration of a dolerite. Sample AWDDo6. (E) Advanced 'grey' alteration with sparse pyrite close to a quartz vein. Sample PS08. (F) Patchy carbonate-biotite alteration with dispersed pyrrhotite and pyrite. Sample AWDDo13. (G-H) Multi-generation, narrow carbonate alteration veinlets.

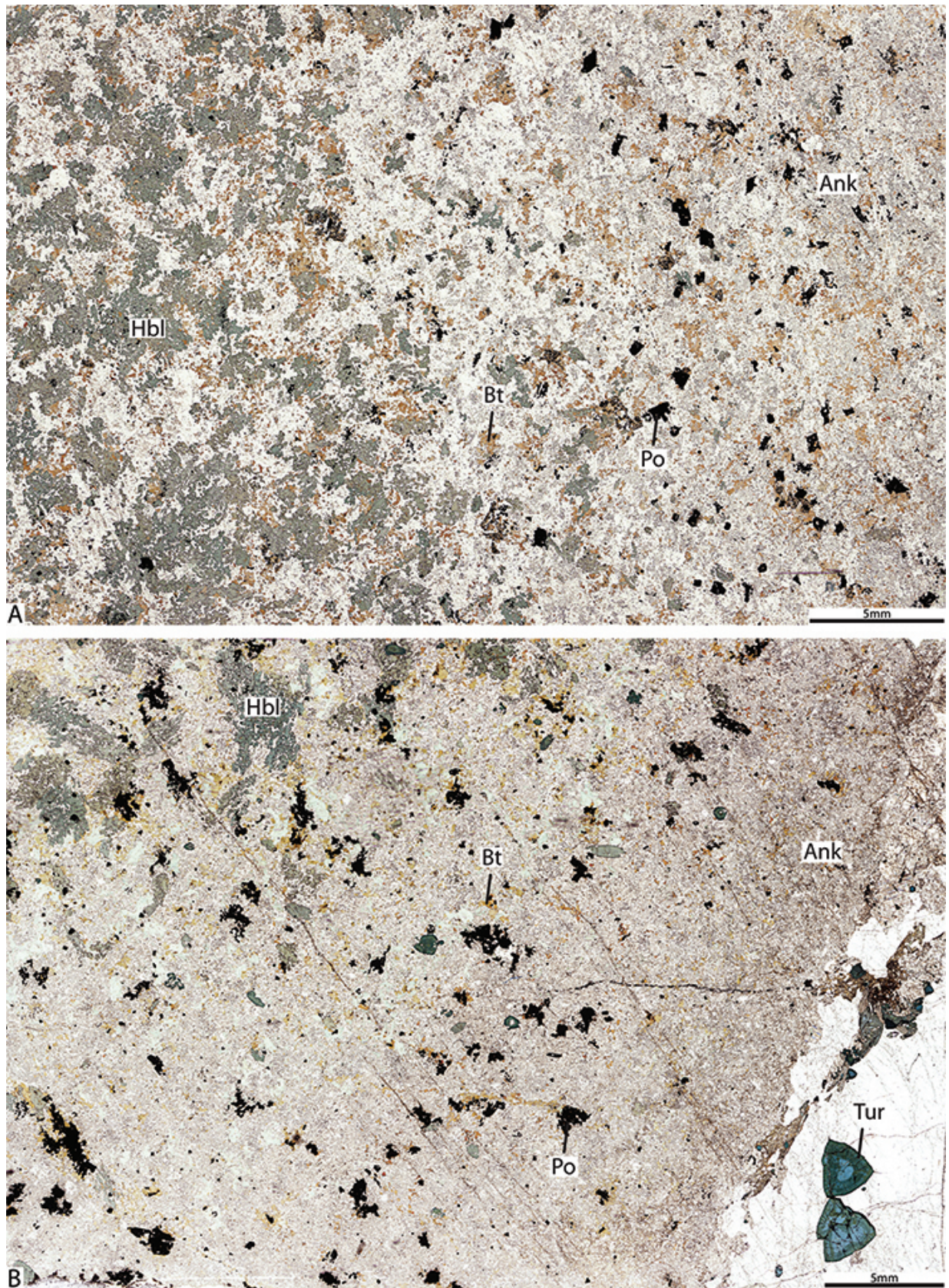


Figure 3.27: Whole slide scans of partially mineralized mafic intrusives from Damang, showing the destruction of amphibole and the formation of biotite, carbonate and sulphides. Alteration progresses from left to right with a quartz-tourmaline vein visible in (b). Sample AWDDo1 (A) and AWDDo4 (B).

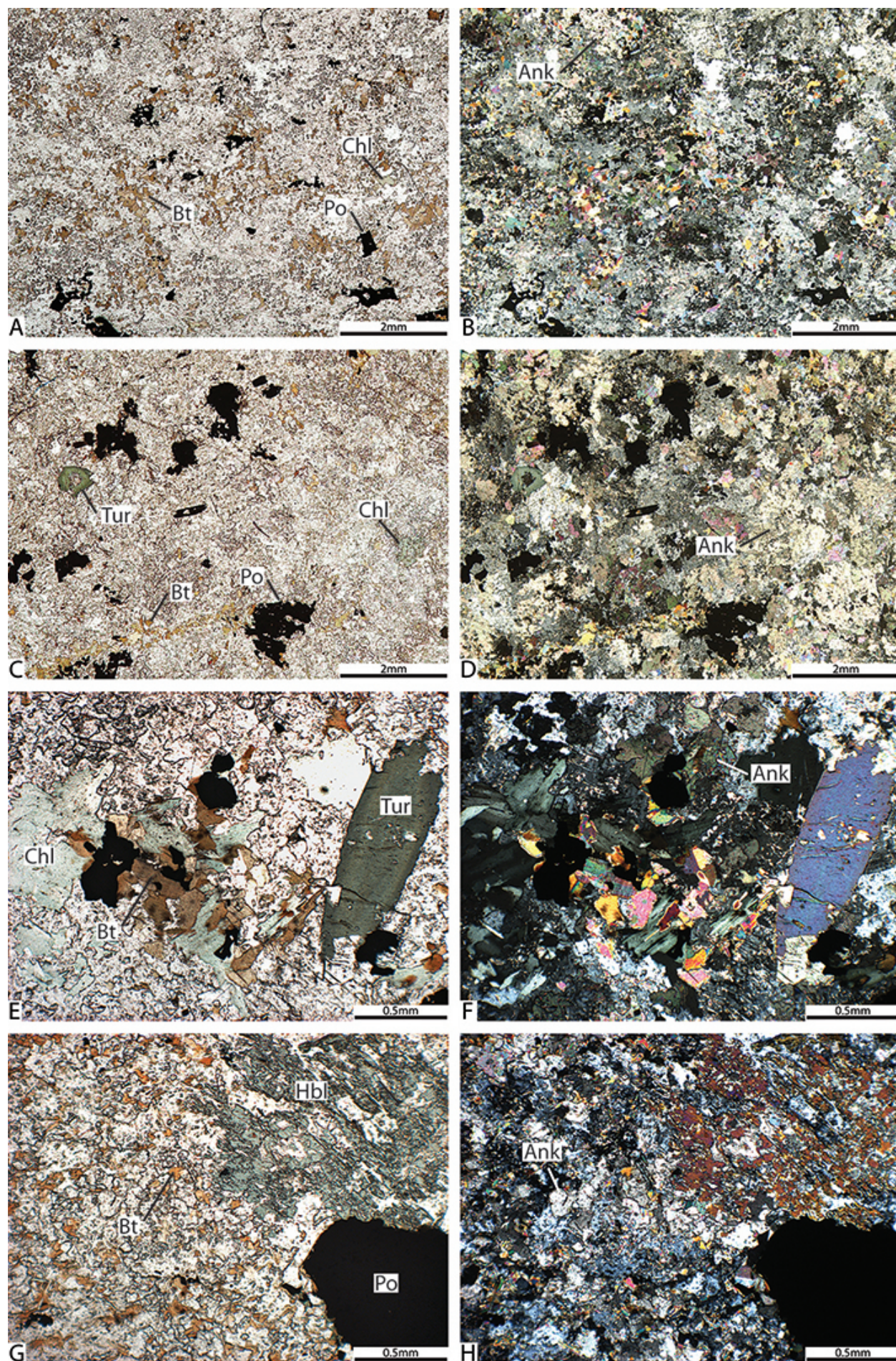


Figure 3.28: (Previous page) Photomicrographs of mineralized mafic intrusives from Damang. Sample AWDDo1, a wholly altered mafic dolerite in plane-polarized light (A) and between crossed polars (B). Sample AWDDo4, a wholly altered mafic dolerite in plane-polarized light (C and E) and between crossed polars (D and F), showing patchily developed chlorite and tourmaline (E and F). Sample AWDDo13, a partially altered dolerite with relic hornblende in plane-polarized light (G) and between crossed polars (H).

Mineralogical and compositional variations between intrusions occasionally result in the development of other alteration minerals. For example, in sample AWDDo2 alteration does not destroy amphibole but instead converts hornblende into a calcium-free amphibole (cummingtonite), along with the usual carbonate and biotite (Fig. 3.29).

In thin section, more information is revealed about the apparent contacts between dolerites and diorites (Fig. 3.23E, F). Photomicrographs of these contact zones are shown in figure 3.30 and reveal that these zones are intensely altered. Specifically, the ‘contact’ seen in sample PS09 (Fig. 3.23E) is an artefact of the alteration, existing between unaltered and highly altered dolerite (Fig. 3.30A, B). Similarly, sample PS04 (Fig. 3.23F) is strongly altered with extensive ankerite (Fig. 3.30C, D). This is intermingled with quartz-rich zones, which could represent parts of a dioritic lithology or alteration associated quartz veining. In any case, the carbonate alteration has destroyed the original texture such that it is now very difficult to interpret the true relationships.

3.6.1 Dolerite Geochemistry

Whole-rock, major and trace element geochemical data were obtained for the dolerite intrusions to classify them and ascribe them to a tectonic setting. Thirty eight samples from this study, Brabham (1998) and Pigois (2003), collected from open pit and drill core, were used. Samples include both unaltered and altered lithologies, although all vein material was removed prior to analysis. Data from this study were obtained using X-ray fluorescence (XRF) techniques at the Open University and University of Manchester (App. B).

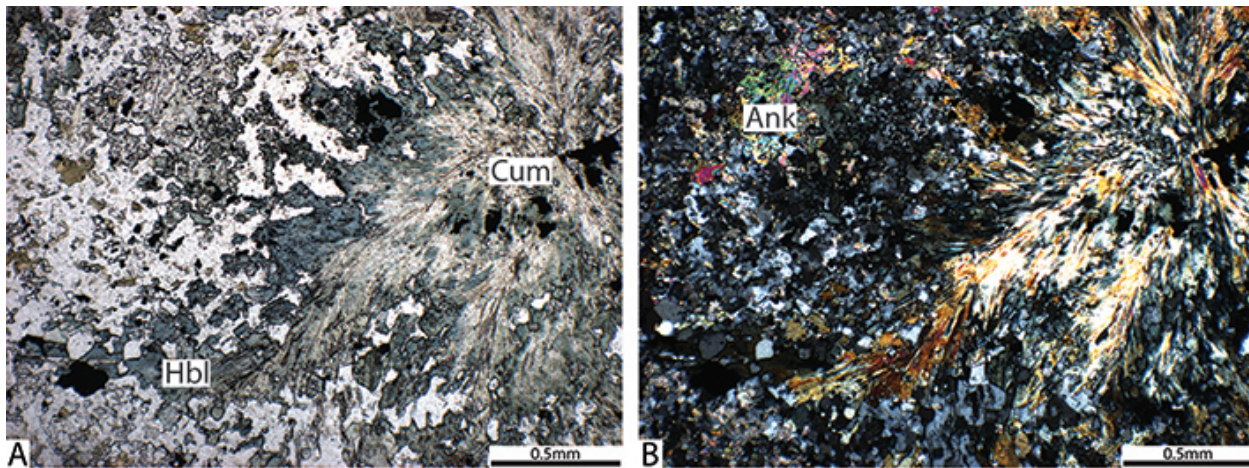


Figure 3.29: Photomicrographs of an altered dolerite dyke showing the conversion of blue-green hornblende to feathery, pale-brown cummingtonite + carbonate + biotite. Sample AWDDo2, in plane-polarized light (A) and between crossed polars (B).

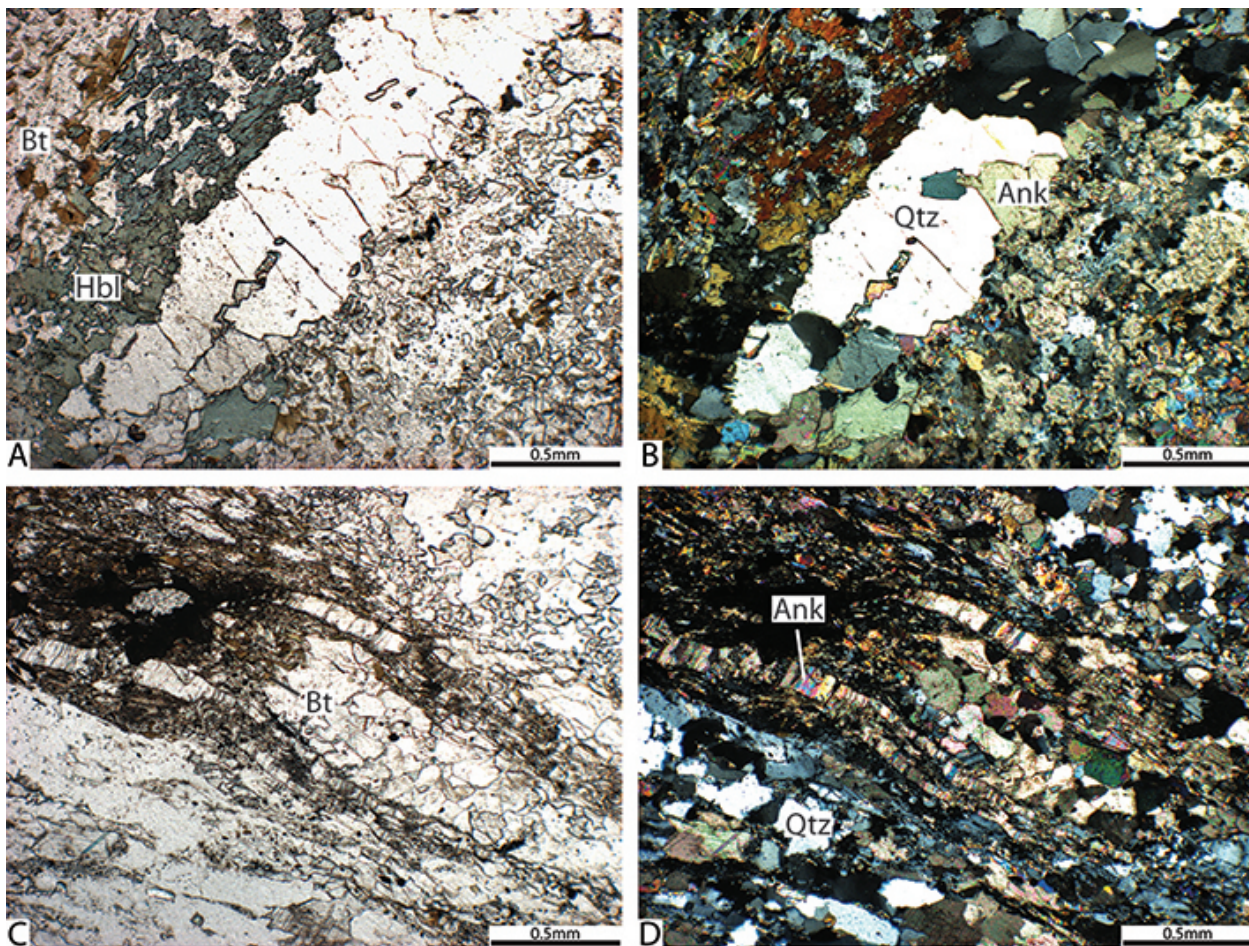


Figure 3.30: Photomicrographs of apparent contact zones between dolerite and diorite. Sample PS09, unaltered dolerite on the left with intense alteration on the right giving the appearance of a diorite. In plane-polarized light (A) and between crossed polars (B). Sample PS04, intermingled 'chilled' contact zone masked by extensive carbonate formation. In plane-polarized light (C) and between crossed polars (D).

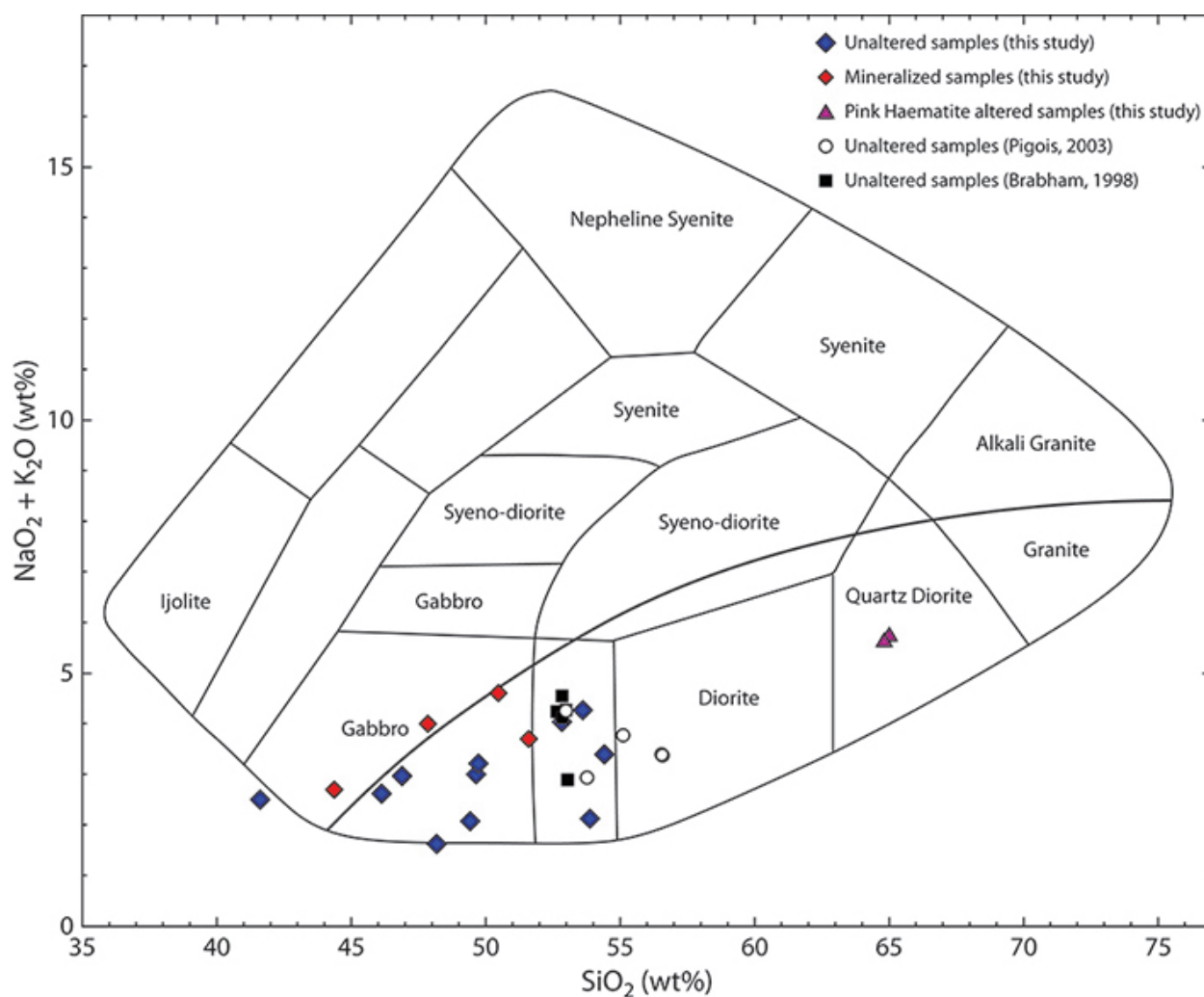


Figure 3.31: The total alkali vs. silica igneous rock classification diagram of [Cox *et al.* \(1979\)](#), adapted by [Wilson \(1989\)](#) for plutonic rocks. The heavier curved line distinguishes between the alkalic (above) and sub-alkalic (below) series, after [Miyashiro \(1978\)](#). Altered samples from this study are shown for reference only.

Major element classification

A total alkali ($\text{Na}_2\text{O} + \text{K}_2\text{O}$) vs. silica (SiO_2) (TAS) diagram is commonly used in classifying igneous rocks as it can display a wide range of common terrestrial compositions and can distinguish between the two major magma series, alkalic and sub-alkalic ([Cox *et al.*, 1979](#)). Figure 3.31 shows that the Damang dolerites are sub-alkalic and typically have compositions between between gabbro and diorite. An alternative major element classification diagram is that of [De La Roche *et al.* \(1980\)](#), which utilizes a wider range of elements than the TAS plot and where, again, the majority of Damang dolerites are classified as gabbro–diorite (Fig. 3.32).

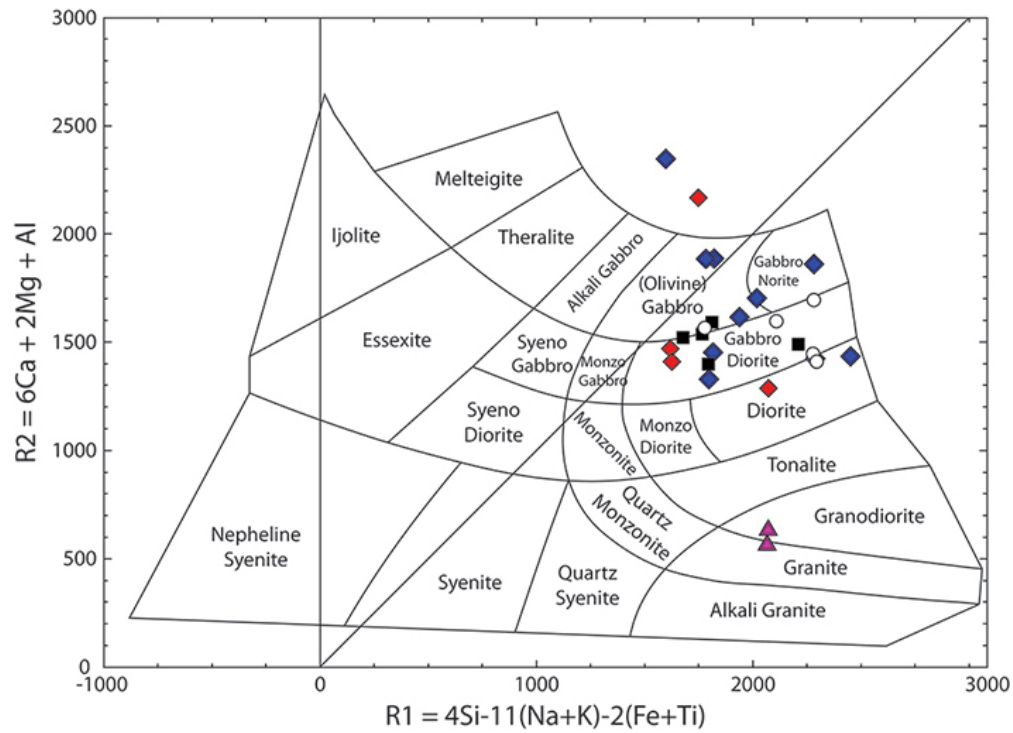


Figure 3.32: The R_1R_2 major element igneous rock classification diagram of De La Roche *et al.* (1980). Altered samples from this study are shown for reference only. Symbols as per figure 3.31.

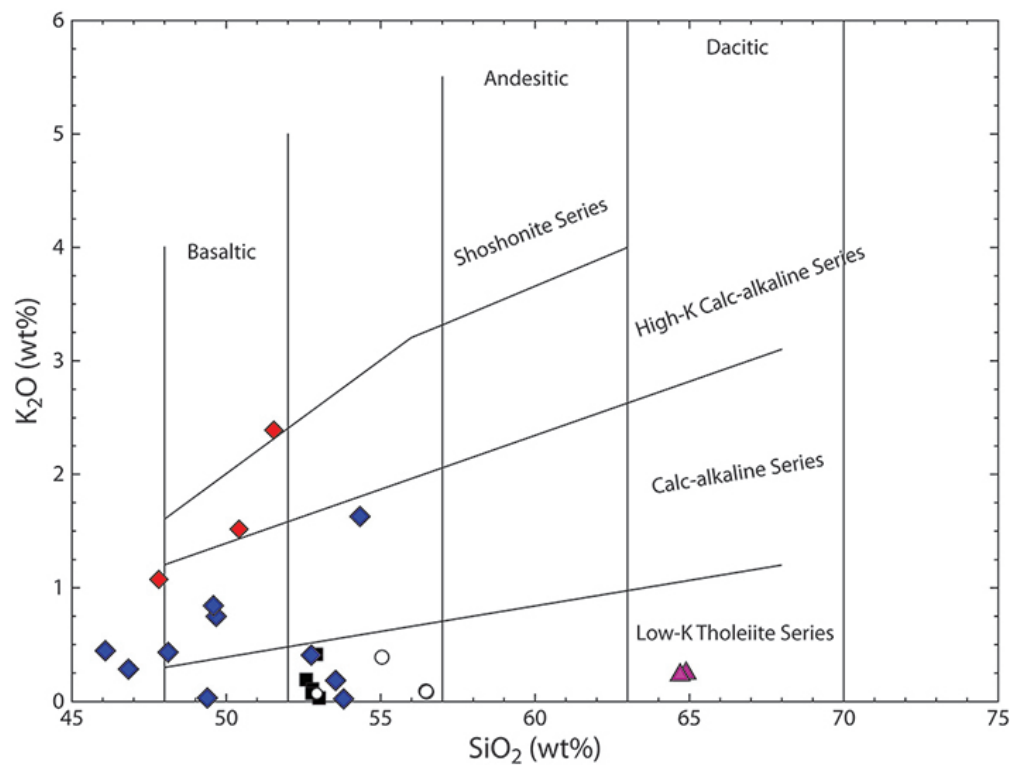


Figure 3.33: A K_2O vs. SiO_2 diagram, used to discriminate between calc-alkaline and tholeiitic magma series, after Peccerillo & Taylor (1976). Altered samples from this study are shown for reference only. Symbols as per figure 3.31.

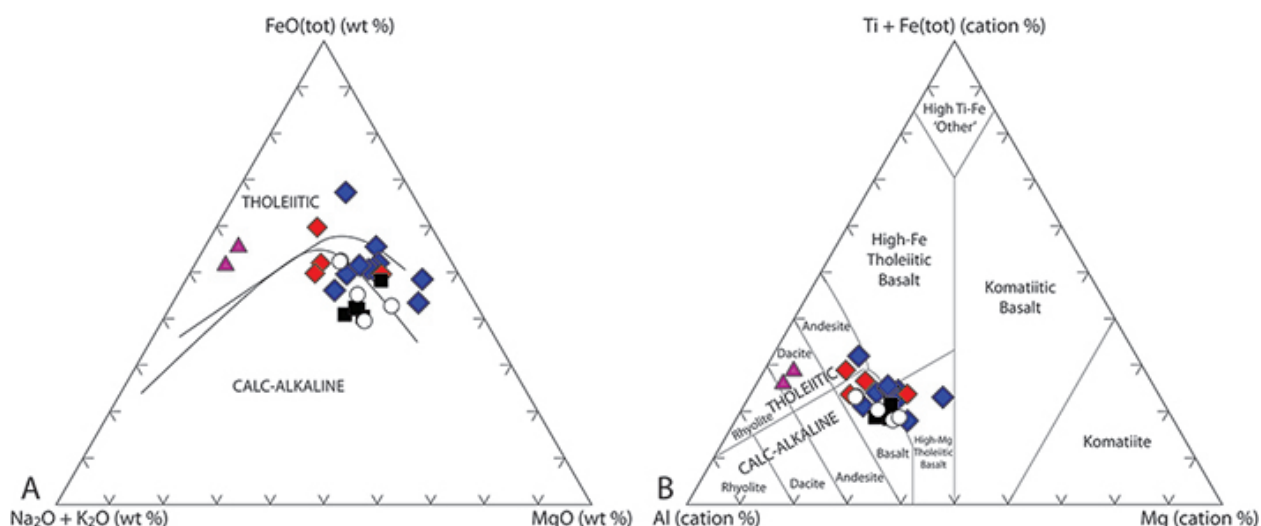


Figure 3.34: Alkali-FeO-MgO plot with boundaries between tholeiite and calc-alkaline series after Kuno (1968) and Irvine & Baragar (1971) (A). Jensen cation plot of Jensen (1976) (B). Altered samples from this study are shown for reference only. Symbols as per figure 3.31.

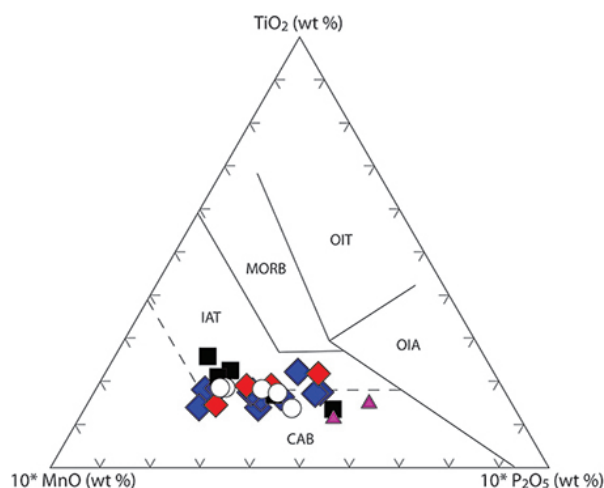


Figure 3.35: The $\text{TiO}_2\text{-MnO-P}_2\text{O}_5$ tectonic discrimination diagram of Mullen (1983). Altered samples are shown for reference only. Symbols as per figure 3.31.

The sub-alkalic magma series can be further subdivided into the calc-alkaline and tholeiite series. Tholeiites are typically low in K and therefore the two series can be discriminated using a K_2O vs. SiO_2 plot (Peccerillo & Taylor, 1976). Figure 3.33 indicates that the Damang dolerites are generally consistent with the tholeiitic series, with some overlap into the calc-alkaline series.

A more distinctive characteristic of the tholeiitic magma series however, is that it shows significant Fe enrichment during the early stages of fractionation (Arculus, 2003). This can be represented on a number of diagrams including alkali-FeO-MgO and (Fe+Ti)-Al-Mg ternary plots (Kuno, 1968; Irvine & Baragar, 1971; Jensen, 1976) (Fig. 3.34). The Damang dolerites, however, fall on the boundary between the two series.

Minor element classification diagrams are also useful for petrotectonic discrimination. In particular, the TiO_2 - MnO - P_2O_5 ternary diagram of [Mullen \(1983\)](#) can distinguish between five distinct tectonic environments for basaltic rocks. Figure 3.35 shows that the Damang dolerites straddle the gradational boundary between island arc tholeiites (IAT) and calc-alkaline basalts (CAB).

Trace element classification

Care must be taken, however, as many major and minor elements are mobile during metamorphic and hydrothermal events. As a result, discrimination based on immobile trace elements, particularly Y and Zr, as well as Ti, are commonly used, allowing for the interpretation of altered samples ([Pearce & Cann, 1973](#)). There are however, certain conditions under which even these elements may become mobile. The mobility of trace elements can be tested using the method of [Cann \(1970\)](#), whereby two elements are plotted against each other such that moderately–highly incompatible and immobile elements from a cogenetic suite of samples should form an array with a log-slope close to unity ([Hastie *et al.*, 2007](#)). Such test diagrams are presented in figure 3.36. Nb is included in this figure as it is another typically immobile trace element that can be used for testing. The data of [Brabham \(1998\)](#) exhibit significant scatter and do not form an array with a slope close to unity. In addition, Ti in Pink Haematite altered samples from this study appear to show a certain amount of mobility and fall off the array of other data. As such, the implications of these data points in subsequent diagrams should be used with caution.

The Ti-Zr-Y ternary plot of [Pearce & Cann \(1973\)](#) is used to discriminate between oceanic and within plate basalts (WPB) and figure 3.37 illustrates that the Damang dolerites are of oceanic affinity, although there is significant overlap between CAB, mid-ocean ridge basalts (MORB) and IAT fields. This can be addressed with the Ti-Zr bivariate plot, however, as seen in figure 3.38, the Damang dolerites, while predominantly within the CAB field, do overlap into the IAT field. Finally, a Ti-Zr-Sr ternary diagram can also be used to distinguish between oceanic basalts (Figure 3.39). Sr is mobile in hydrothermal fluids so only unaltered samples are applicable, but the Damang dolerites plot predominantly within the CAB field.

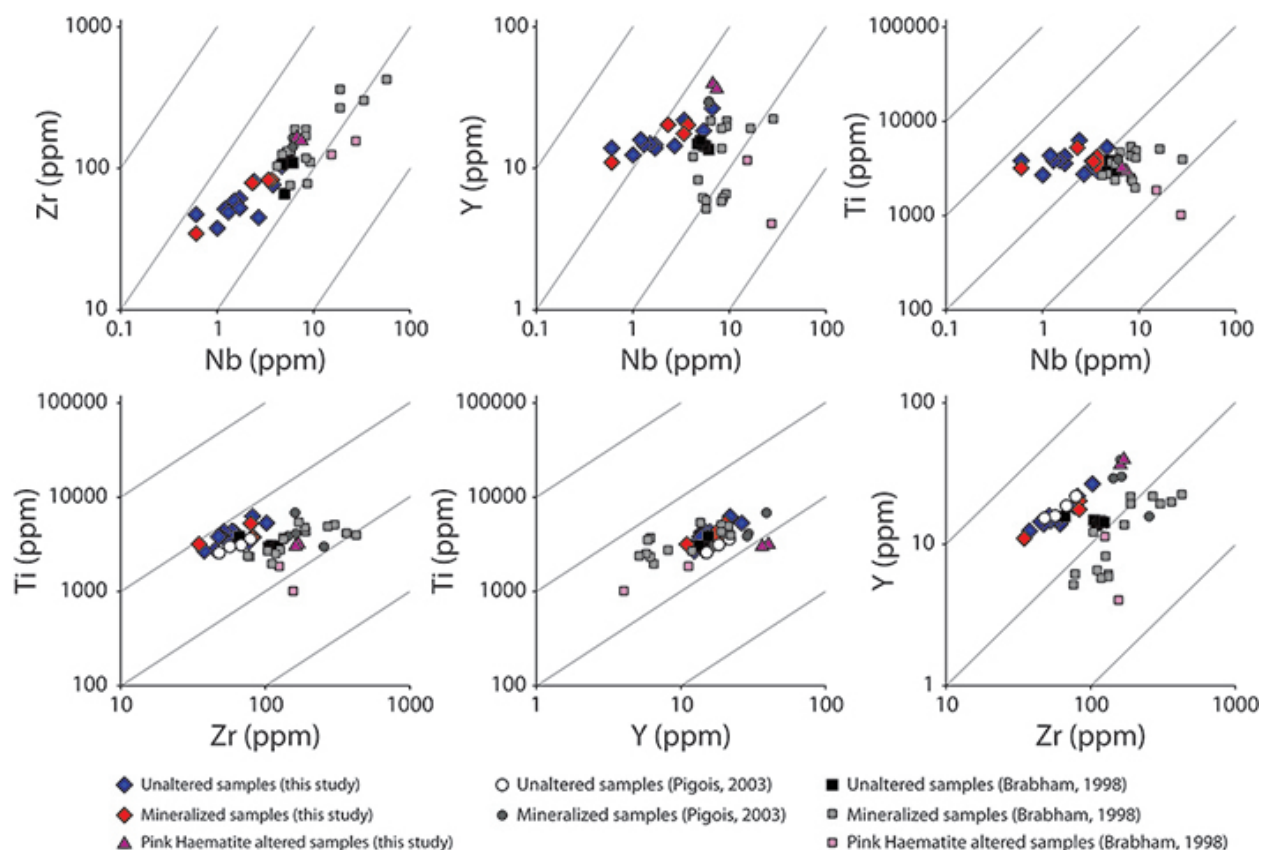


Figure 3.36: Testing mobility of supposedly 'immobile' trace elements in the dolerite intrusions after [Cann \(1970\)](#). Perfectly immobile elements with variation due solely to fractionation in a cogenetic suite of rocks should plot along an array with a slope close to unity. It can be seen that the trace element data of [Brabham \(1998\)](#) do not plot along such a slope and their implications in subsequent diagrams should be used with caution.

3.6.2 Petrogenesis of the Damang Dolerites

Whole rock geochemical analysis of the Damang dolerite intrusions indicates that they are sub-alkaline, low-K, gabbro-diorites with a transitional tholeiite – calc-alkaline affinity. While tholeiitic magmas form in a wide range of tectonic settings, the calc-alkaline series occurs exclusively in association with convergent plate margins. Such a setting is therefore invoked for the Damang dolerites. This suggests a link with the underlying Birimian tholeiitic – calc-alkaline volcanic arc rocks.

Figure 3.37: The Ti-Zr-Y tectonic discrimination diagram of [Pearce & Cann \(1973\)](#) used for discriminating between oceanic and within plate basalts. Symbols as per figure 3.36.

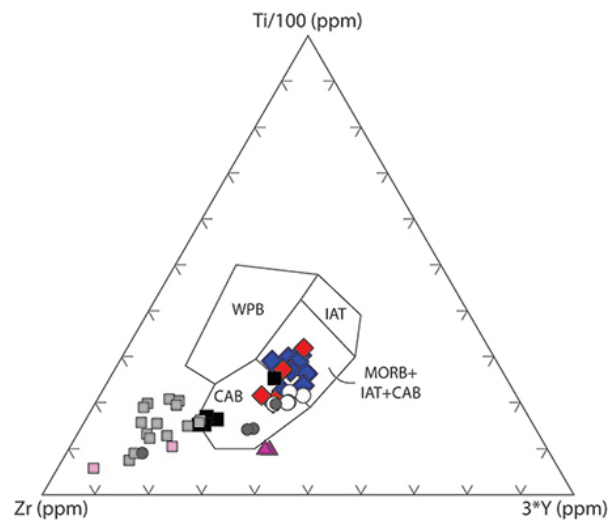


Figure 3.38: The Ti-Zr tectonic discrimination diagram of [Pearce & Cann \(1973\)](#) used for discriminating between oceanic basalts. Symbols as per figure 3.36.

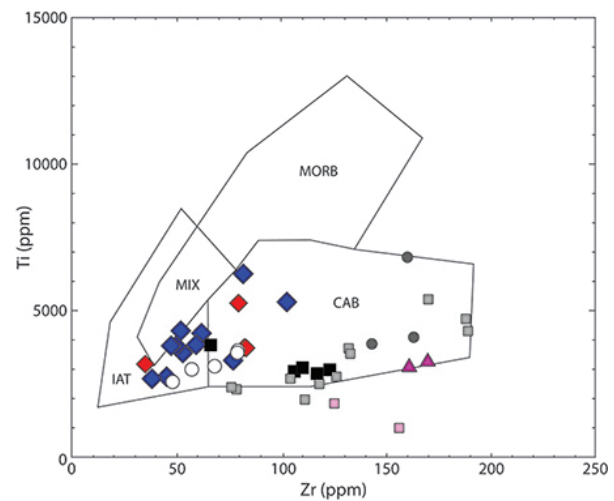
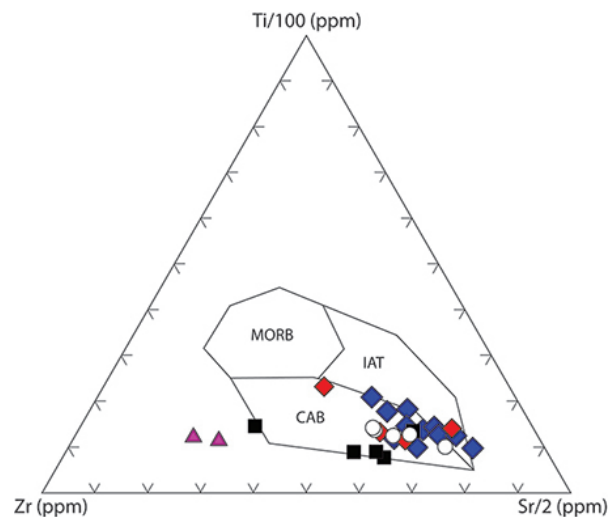


Figure 3.39: The Ti-Zr-Sr tectonic discrimination diagram of [Pearce & Cann \(1973\)](#) used for discriminating between oceanic basalts. Altered samples from this study are shown for reference only. Symbols as per figure 3.31.



3.7 Mineral Chemistry

Mineral compositions were analysed using a JEOL JSM-840A scanning electron microscope (SEM), fitted with an Oxford Instruments Isis 300 energy-dispersive analytical system (EDS), located in the Department of Earth Sciences, University of Oxford. Standard analytical conditions comprised a 20 kV accelerating voltage, 5 nA beam current that was monitored regularly, and a live counting time of 100 seconds. Elemental calibrations were made against a range of natural and synthetic standards, a ZAF correction procedure was used, and the count rate was calibrated approximately every 1–2 hours using a pure cobalt metal standard.

Mineral chemical data was collected for the majority of phases in each analysed sample, although discussion here will be limited to those that show significant and/or important variations or trends as a function of hydrothermal alteration. The chemistry of other phases that do not show compositional trends associated with hydrothermal alteration will be discussed in later sections where it is relevant to a specific application. For example, the composition of biotite will be discussed in chapter 4 with regards to geothermometry. Additional comments for other phases are given in appendix C.

Plagioclase Feldspars

Banket Series quartzites show a broad range of compositions from pure albite to An_{30} in both unmineralized and Pink Haematite altered samples (Fig. 3.40A). This range may reflect an original detrital spread, possibly modified during regional metamorphism. By contrast, mineralized samples show two or more distinct compositional clusters in the ranges An_{0-4} and An_{25-30} , or An_{25-30} and An_{40-50} , indicating two coexisting, compositionally different plagioclases and the presence of a miscibility gap (specifically termed the peristerite gap, when present between albite and oligoclase (Crawford, 1966)). The width of any miscibility gap is a function of temperature and thus here, can provide a useful constraint on the temperature of alteration. This will be addressed further in chapter 4.

Plagioclase analyses from samples of the Tarkwa Phyllite show a number of distinct clus-

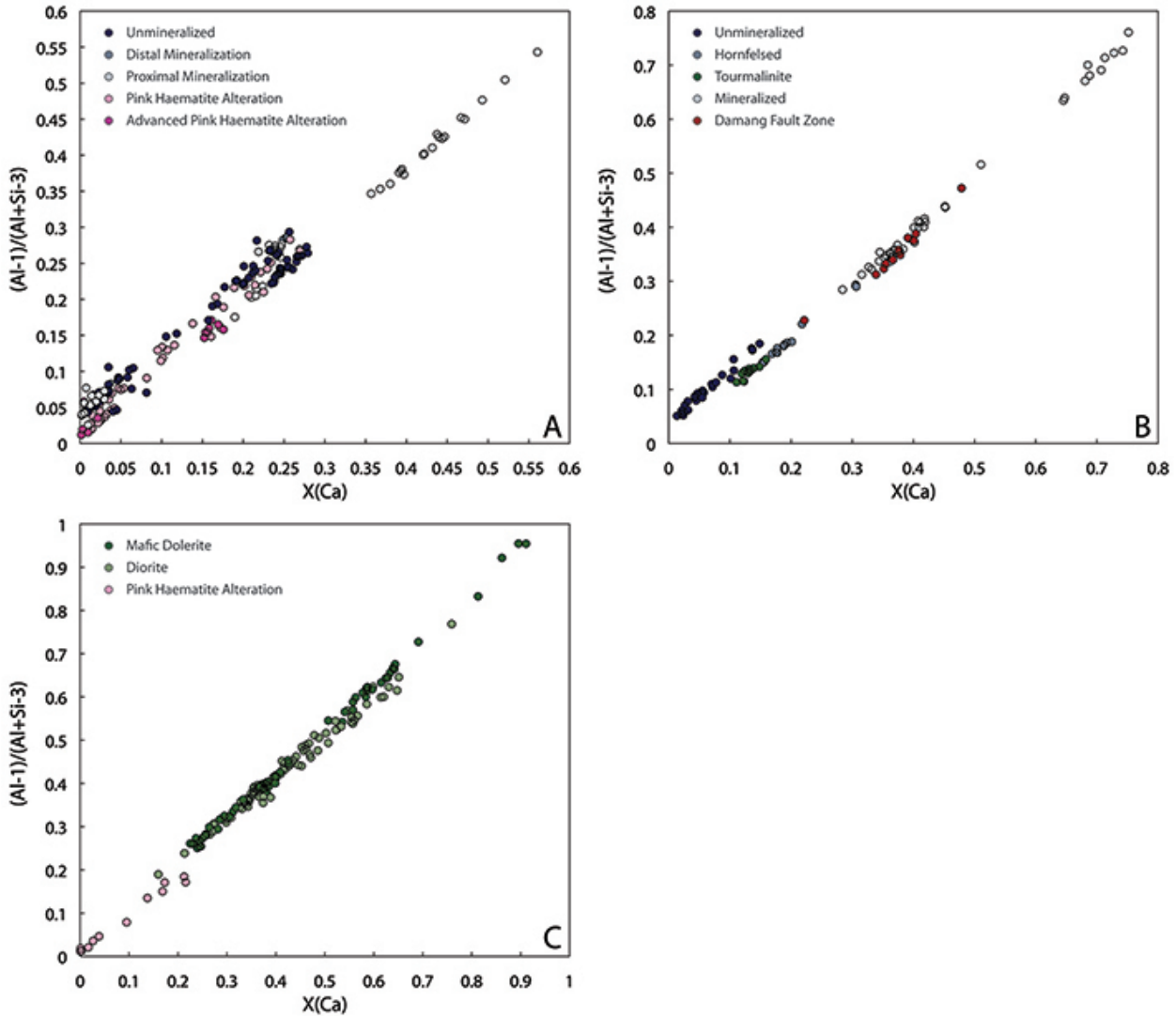


Figure 3.40: Compositional data for plagioclase feldspars from the Banket Quartzites and Huni Sandstone (A), Tarkwa Phyllite (B) and Dolerite Intrusives (C). The Banket Quartzites show a range of compositions, which develop into one or more distinct clusters as a result of mineralization or advanced Pink Haematite alteration. Mineralized Tarkwa Phyllite samples show a trend towards more calcic compositions. Dolerites cover a broad range of compositions, becoming more sodic as a result of Pink Haematite alteration. The apparent parallel trends of data points in A and B are due to a recalibration of the instrument during the course of this thesis. This has resulted in a slight lowering of the measured Al/Si ratio while leaving the $X(Ca)$ unaffected. Although error bars are not shown for the sake of clarity, these two calibrations are within error of each other.

ters (Fig. 3.40B). ‘Typical’ Tarkwa Phyllite predominantly contains albite, with a compositional range of An_{2-15} . Hornfelsed samples and tourmalinites contain oligoclase in the range An_{10-22} , with the tourmalinite samples falling at the upper end of this range. By contrast, mineralized samples primarily fall in the compositional range An_{30-50} , with one particular sample (AWDP2) containing a further cluster around An_{70} , indicating the presence of a ‘labradorite’ miscibility gap. The albite-rich nature of the unaltered samples is unexpected given the mid–upper amphibolite facies assemblage, but as will be shown in chapter 4, this

is related to the retrograde chloritization described in section 3.4. Samples from within the Damang Fault zone overlap with mineralized samples, consistent with the fault zone having been affected by the same alteration event.

Dolerite intrusives show a very broad range of compositions, primarily ranging between An_{20-70} with a dominant cluster at An_{30-40} (Fig. 3.40C). While not well-defined, there is a weak trend towards more sodic compositions in dioritic intrusives and more calcic compositions in the more mafic dolerites, as would generally be expected. There is no clear trend between mineralized and unmineralized samples, while Pink Haematite altered samples trend towards more sodic compositions, with a dominant cluster close to pure albite. The total compositional range of Pink Haematite altered intrusives is comparable to that seen in the Banket Quartzites.

Chlorites

Chlorites are common in all lithologies and alteration styles at Damang, and due to their relatively complex chemistry can provide information on a range of different geologic events. Common ‘tri-octahedral’ chlorites comprise a structure of alternating tetrahedral-octahedral-tetrahedral (talc-like) and brucite-like layers, giving a net composition for the Mg-rich end-member (clinochlore) of $Mg_3(Mg_2Al)[Si_3Al]O_{10}(OH)_8$ (Deer *et al.*, 1992). Other chlorite compositions can be considered through cation exchange mechanisms acting on this clinochlore composition. Other tri-octahedral end-members are the Fe-bearing chamosite ($Fe_3(Fe_2Al)[Si_3Al]O_{10}(OH)_8$) and Mn-bearing pennantite ($Mn_3(Mn_2Al)[Si_3Al]O_{10}(OH)_8$), which involve simple Fe-Mg-Mn exchange. A tschermak substitution ($Al_2Si_{-1}Mg_{-1}$) gives the composition of a notional ‘amesite’ ($(Mg_2Al)(Mg_2Al)[Si_2Al_2]O_{10}(OH)_8$) end-member. Amesite *sensu stricto* is in fact a serpentine mineral. Another group of chlorites, the di-octahedral chlorites, contain a pyrophyllite-type sheet in place of the talc-like sheet as a result of the exchange vector $Al_2\Box_1Mg_{-3}$, producing sudoite ($\Box Al_2(Mg_2Al)[Si_3Al]O_{10}(OH)_8$), where $\Box$ represents a site vacancy.

Chlorite compositional data from Damang are presented in figure 3.41. In all unmineral-

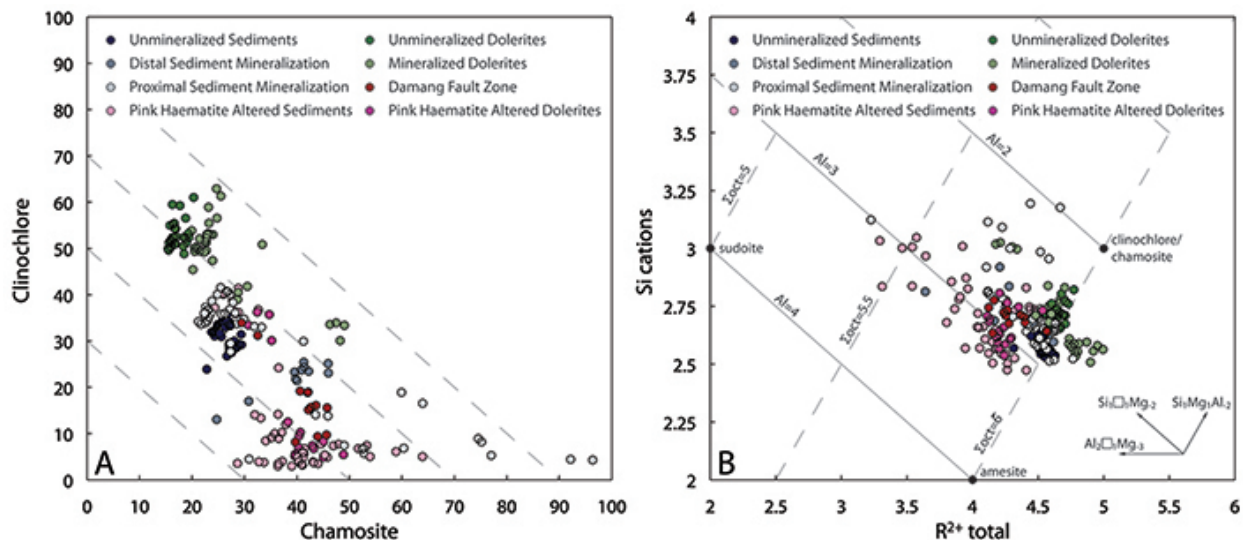


Figure 3.41: Compositional data for chlorites from sedimentary rocks (Banket Quartzites, Tarkwa Phyllite, Huni Sandstone) and dolerite intrusives. A shows the proportion of clinochlore and chamosite end-members, with both mineralized and Pink Haematite altered samples for both sediments and dolerites showing a trend towards more Fe-rich (chamosite) compositions, with Pink Haematite altered samples also showing a trend towards increased Al content. Contours represent approximately constant Al content. B shows a plot of Si cations versus total divalent cation content after Wiewora & Weiss (1990). Solid and dashed contours represent constant Al content and constant octahedral cation content respectively. Chlorites typically cluster well along the clinocllore/chamosite – amesite join, while haematite altered samples fall to the left, indicating a sudoite exchange vector.

ized/unaltered samples, compositions cluster tightly with a composition falling along the clinocllore/chamosite – amesite vector (Fig. 3.41B). Sedimentary rocks (including the Banket Quartzite, Tarkwa Phyllite and Huni Sandstone) and the dolerite intrusives are distinguishable only by the differing Fe/Mg ratios, with the dolerites being richer in Mg (Fig. 3.41A).

Both gold-associated and Pink Haematite alteration drive chlorites towards more Fe-rich compositions, although this is significantly more extensive for Pink Haematite alteration (Fig. 3.41A). For mineralized samples this trend occurs with negligible change in total Al content, whereas the Pink Haematite alteration trend falls well-below the appropriate constant-Al contour, suggesting an increase in total Al content. This is supported by figure 3.41B, which shows that Pink Haematite altered samples have indeed undergone a component of sudoite exchange. Figure 3.41B also shows that for both alteration styles, scatter is introduced towards increased Si content, suggesting small components of both sudoite and inverse-amesite exchange. Chlorites from the Tarkwa Phyllite in the Damang Fault zone

are intermediate between the two alteration styles and may represent contributions from both events since the Damang Fault would have been a major fluid pathway during both hydrothermal episodes.

Amphiboles

Amphiboles are even more compositionally complex than the chlorites and again, it is simplest to consider the range of possible compositions as a function of a number of different cation exchange mechanisms (Deer *et al.*, 1992). Common end-members in the monoclinic, non-calcic and calcic amphiboles, along with exchange mechanisms relative to tremolite, are given in table 3.1.

Amphiboles are a major constituent of the dolerite intrusives at Damang and in the majority of cases show a good clustering close to the composition of common hornblende (Fig. 3.42). As described in section 3.6, sample AWDDo2 contains hornblende coexisting with cummingtonite. As shown in figure 3.42, an apparent compositional trend exists between the typical hornblende and pure cummingtonite. This trend is likely related to lattice-scale interlayering of the two end-members. Amphiboles also occasionally occur within mineralized veins in the Tarkwa Phyllite. Here they also show a compositional range varying between hornblende and an actinolitic (tremolite–ferro-actinolite) composition, again interpreted to relate to interlayering. Both of the compositional trends observed in samples AWDDo2 and AWDP2

Table 3.1: Common amphibole cation exchange mechanisms, relative to tremolite. The edenite exchange vector can also include K in place of Na.

Mineral	Composition	Exchange Mechanism
Tremolite	$\text{Ca}_2\text{Mg}_5\text{Si}_8\text{O}_{22}(\text{OH})_2$	
Ferroactinolite	$\text{Ca}_2\text{Fe}_5\text{Si}_8\text{O}_{22}(\text{OH})_2$	$\text{Fe}_1\text{Mg}_{-1}$
Cummingtonite	$\text{Mg}_2\text{Mg}_5\text{Si}_8\text{O}_{22}(\text{OH})_2$	$\text{Mg}_2\text{Ca}_{-2}$
‘Hornblende’	$\text{Ca}_2(\text{Mg}_4\text{Al})(\text{Si}_7\text{Al})\text{O}_{22}(\text{OH})_2$	$\text{Al}_2\text{Si}_{-1}\text{Mg}_{-1}$
Tschermakite	$\text{Ca}_2(\text{Mg}_3\text{Al}_2)(\text{Si}_6\text{Al}_2)\text{O}_{22}(\text{OH})_2$	$\text{Al}_4\text{Si}_{-2}\text{Mg}_{-2}$
Edenite	$\text{NaCa}_2\text{Mg}_5(\text{Si}_7\text{Al})\text{O}_{22}(\text{OH})_2$	$\text{Na}_1\text{Al}_1\text{Si}_{-1}$
Pargasite	$\text{NaCa}_2(\text{Mg}_4\text{Al})(\text{Si}_6\text{Al}_2)\text{O}_{22}(\text{OH})_2$	$\text{Na}_1\text{Al}_3\text{Mg}_{-1}\text{Si}_{-2}$

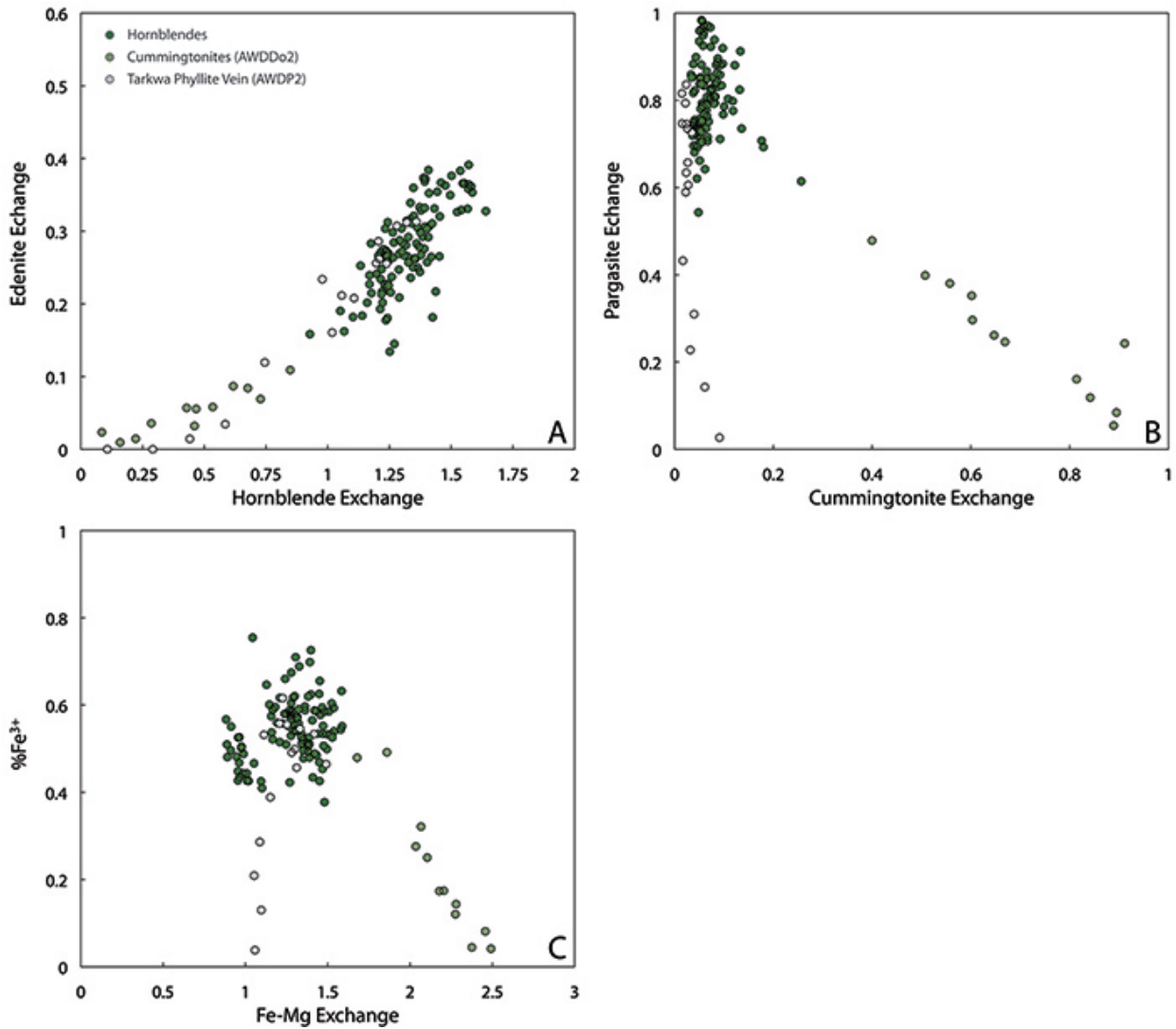


Figure 3.42: Compositional data for amphiboles from dolerite intrusives and mineralized veins in the Tarkwa Phyllite. The origin of each plot represents pure tremolite. The majority of amphiboles are common hornblendes that cluster tightly. Samples AWDDo2 and AWDP2 contain other amphiboles trending towards pure cummingtonite and actinolite respectively. Exchange mechanisms are those described in table 3.1. Cation recalculation and Fe^{3+} determination after (Holland & Blundy, 1994).

are towards effectively pure end-members, with all alkali elements, Fe^{3+} -content and in the case of AWDDo2, Ca, being completely removed. The AWDDo2 cummingtonites also show a trend towards a higher Fe/Mg ratio.

Tourmalines

Tourmaline has a general formula of $\text{XY}_3\text{Z}_6\text{B}_3\text{Si}_6\text{O}_{30}(\text{OH})$ where X may be Na, Ca, K or $\square$, Y is Al, Li, Fe, Mg, Mn or Fe^{3+} and Z is Al, Mg, Mn, Fe^{3+} or Cr (Deer *et al.*, 1992). As such they can show a great deal of compositional variation that is too complex to discuss in detail

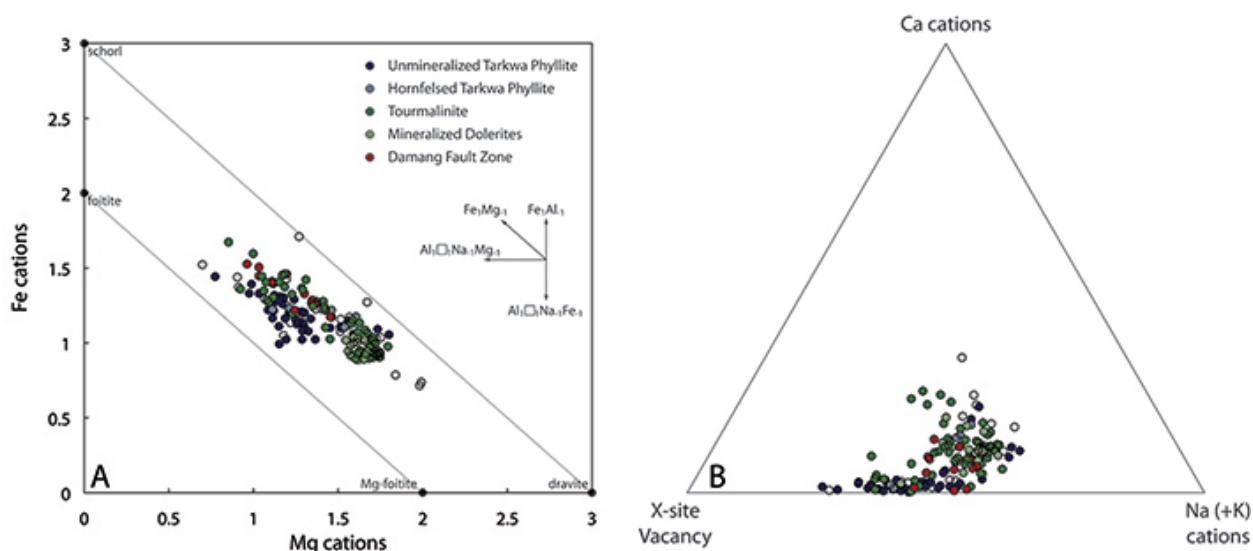


Figure 3.43: Compositional data for tourmalines from the Tarkwa Phyllite and mineralized dolerites. A primarily investigates Fe-Mg exchange, with the directions of other exchange vectors also shown. B shows X-site occupancy. A clear distinction exists between detrital/metamorphic tourmalines in the Tarkwa Phyllite and hydrothermal tourmaline in the dolerites, the latter of which are much more tightly clustered.

here. A simple series of common tourmalines is the schorl(Fe)–dravite(Mg) series, given by the composition $\text{Na}(\text{Mg},\text{Fe})_3\text{Al}_6(\text{BO}_3)_3[\text{Si}_6\text{O}_{18}]\text{O}_3(\text{OH})$, which serves as a base composition to consider variation from.

Tourmalines are common components of both the Tarkwa Phyllite and mineralized dolerites (see section 3.4 and 3.6). Pre-metamorphic tourmalines in the Tarkwa Phyllite, including tourmalinite horizons and samples from the Damang Fault zone, show a broad range of Fe/Mg ratios with an Al content slightly higher than for schorl–dravite, which is balanced by X-site vacancies (Fig. 3.43). These compositions are similar, although slightly less aluminous, than those that occur in Namaqualand, South Africa and in association with sedimentary exhalative deposits such as Broken Hill, Australia (Plimer, 1984; Willner, 1992), which supports a similar origin for the tourmaline at Damang. By contrast, mineralized dolerites and certain analyses from the Tarkwa Phyllite tourmalinites show a much more tightly constrained Fe/Mg ratio towards the Mg-rich end of the total observed range. X-site occupancy also changes in these samples to include a greater amount of Ca, which due to charge balance restrictions, suggests that these tourmalines may also contain a small amount of monovalent cations such as Li, lesser amounts of trivalent cations such as Fe^{3+} or a substitution of Al for Si. This tight compositional clustering accords with a hydrothermal origin,

possibly related to B-remobilization from the pre-metamorphic tourmalines, as is seen in the Namaqualand tourmalinites ([Willner, 1992](#)).

3.8 Summary Discussion of Pink Haematite Alteration

Hydrothermal alteration assemblages associated with the Pink Haematite alteration are broadly consistent between lithologies, comprising the formation of extensive calcite, with lesser ankerite, along with Fe-rich chlorite, albite, quartz and a haematitization of magnetite. This is coupled, in the case of the dolerite intrusives, with a complete destruction of hornblende. This broad similarity and apparent lack of host-rock control suggests a moderately high fluid/rock ratio. There is no veining associated with this alteration event, which suggests fluid movement was through the grain boundary network, which would thereby increase the effective fluid/rock ratio, compared to fluid moving primarily along fracture networks.

Isocon diagrams are a simple way of illustrating the mass balance equation of [Gresens \(1967\)](#) between an altered rock and the original. They are therefore useful for assessing the geochemical signature (element enrichments and depletions) of hydrothermal alteration events ([Grant, 1986](#)). Whole-rock major element data used to construct the isocon diagrams in this study are determined from X-ray fluorescence analysis (App. B). In order for isocon diagrams to be meaningful, however, there needs to be a high degree of certainty that the altered rock had an original bulk composition that is comparable to the unaltered rock. For this reason, the dolerite intrusions are the most appropriate for this purpose as unaltered and altered samples can be taken from a single intrusion (sill or dyke), which likely had a broadly homogeneous bulk composition, particularly when compared to the primary compositional variation present in sedimentary rocks.

The isocon diagram constructed for Pink Haematite alteration is shown in figure [3.44](#). Al_2O_3 was chosen to define the isocon as it is likely the least mobile of the major elements. Figure [3.44](#) shows that Pink Haematite alteration results in a significant increase in Na and Si, with decreases in Mg and Ca. Interestingly, Fe shows no variation, therefore Fe must be recycled between various phases in the rock. Volatiles (LOI; loss on ignition) display an

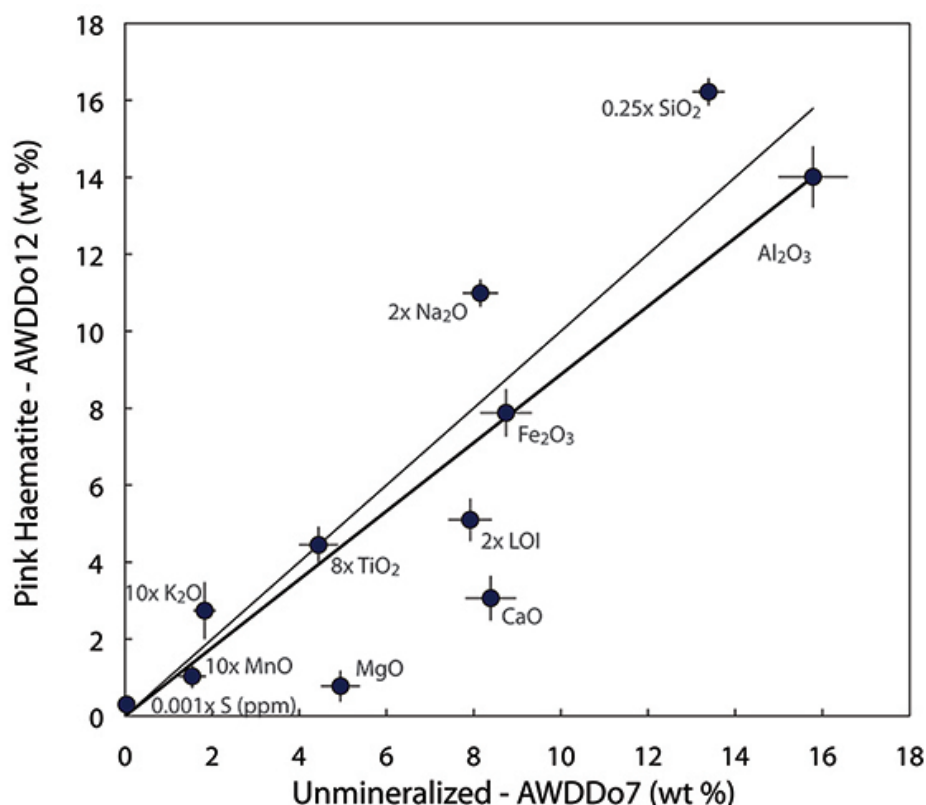
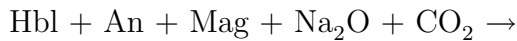


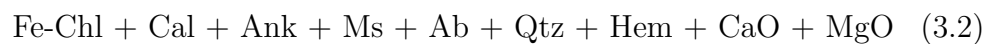
Figure 3.44: Major element isocon plot for Pink Haematite altered dolerite intrusives, showing enrichment in Na and Si with depletions in Mg, Ca and volatiles (LOI). The heavy line represents the line of constant concentration, the isocon. The lighter line represents the line of constant volume, which shows that this alteration results in an overall volume decrease. Representative error bars given at 1%.

overall decrease, which is somewhat unexpected given the petrographic observations. The clear addition of CO_2 must therefore be balanced by a loss of H_2O . The magnitude of this decrease varies between samples however, implying that volatile content is the most variable compositional parameter. Interestingly, there is a significant overall volume decrease as a result of this alteration. There is, however, no veining or obvious fracturing associated with it. It is inferred therefore that the volume decrease is accommodated by a micro-porosity developed in feldspars, which is intimately linked with the haematite dusting.

Comparing this data along with petrographic observations and mineral chemical data allows for the determination of generalized, descriptive alteration reactions. Considering the dolerite intrusives, a reaction of the form:



can be constructed. Whereas the Banket Series quartzites and Huni Sandstones undergo a reaction more akin to:



In all cases, ferromagnesian minerals are generally more Fe-rich in altered rocks, due to the loss of Mg. The addition of Si and the production of quartz during alteration explains the apparent silicification exhibited by Pink Haematite altered samples. Furthermore, this silicification would serve to change the rheology of the rocks by making them more brittle and therefore susceptible to fracturing. This is inferred to explain the large-scale relationship between Pink Haematite alteration and mineralized veining, as described in section 3.5 and shown in figure 3.22.

The alteration style overall can be described as a quartz-carbonate-chlorite alteration and from the haematitization of magnetite, is facilitated by an oxidising fluid. There is a close temporal and spatial relationship between the Pink Haematite alteration and the dolerite intrusions (see section 3.11) at Damang. It is therefore interpreted that this fluid is magmatic in origin and is genetically related to this magmatic event.

3.9 Summary Discussion of Gold-Associated Quartz Veining and Hydrothermal Alteration

3.9.1 Mineralized Quartz Vein Mineralogy and Textures

The structural setting and physical morphology of gold-bearing quartz veins has been discussed in section 2.8.1. Mineralogically they are primarily quartz-sulphide veins, although other phases such as feldspar, tourmaline and particularly carbonate may be present. The presence of veining and the type and nature of vein mineralogy is strongly controlled by host lithology. While Banket Series quartzites and the Huni Sandstone are dominated by massive quartz-pyrite veins, the dolerite intrusives and Tarkwa Phyllite commonly contain one or more additional phase and are occasionally laminated (Fig. 3.45). In particular, veins cutting the Tarkwa Phyllite often contain significant ankerite, biotite and/or actinolitic amphibole (Fig. 3.17A, B), while those hosted in the dolerite intrusives typically contain significant tourmaline, ankerite and/or biotite (Fig. 3.26C, F). Sulphide minerals are regularly accompanied by trace amounts of other phases including chalcopyrite and various tellurides, particularly Bi-bearing tellurides, and scheelite. This is true for wall rock alteration as well as vein mineralogy.

3.9.2 Gold-Associated Hydrothermal Alteration

Hydrothermal alteration occurs in selvages around quartz veins, although the width of the alteration zone is not proportional to the width of the vein. Alteration assemblages associated with gold mineralization are also strongly controlled by host rock lithology as described above and summarized in table 3.2 and figure 3.46. However, general trends are clearly recognisable, such as the development of sulphide minerals and ankeritic carbonate along with either muscovite or biotite (in more Fe-rich rocks), coupled with the complete destruction of magnetite. These occur in varying proportions and may be associated with lesser amounts of chlorite and/or tourmaline. This relationship between alteration assemblage and

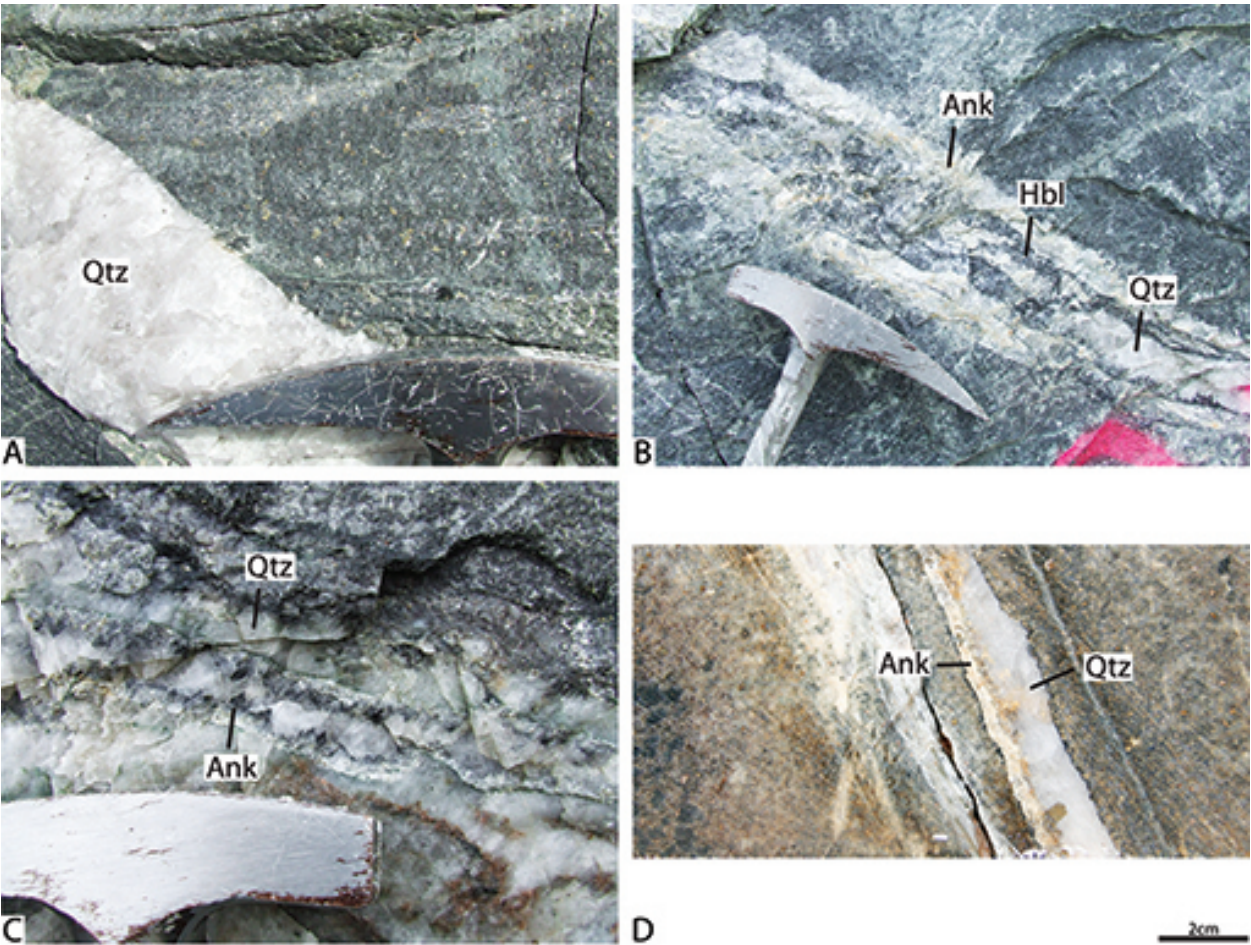


Figure 3.45: Photographs of mineralized veins illustrating the range of mineralogies and textures. Massive quartz-pyrite veins (A) dominate and are the only type to occur within the Banket Series and Huni Sandstone. The Tarkwa Phyllite and Dolerite Intrusives also contain a variety of laminated veins (B-D), indicating multiple progressive crack-and-seal events, with phases of quartz-rich and carbonate-rich precipitation. Ferromagnesian minerals including tourmaline, chlorite, biotite and hornblende also occur (B, see also figure 3.17)

host rock composition implies an internally buffered, low fluid/rock ratio system.

An isocon diagram for gold-associated alteration of a dolerite intrusive is shown in figure 3.47. Al_2O_3 was again chosen to define the isocon as the most immobile major element. This plot shows that during alteration the rock undergoes significant enrichment in K and S with a lesser increase in volatiles (both H_2O and CO_2), which agrees directly with petrographic observations. Other changes include a slight increase in Fe, with marginal decreases in Na,

Table 3.2: Summary of gold-associated alteration assemblages in sediments and dolerites.

Lithology	Hydrothermal Minerals	Consumed Minerals
Sediments	$\text{Ms} + \text{Ank} + \text{Py} + \text{Chl} \pm \text{Bt} \pm \text{Po}$	Mag
Dolerite Intrusives	$\text{Bt} + \text{Ank} + \text{Po} \pm \text{Chl} \pm \text{Py}$	Hbl + Mag

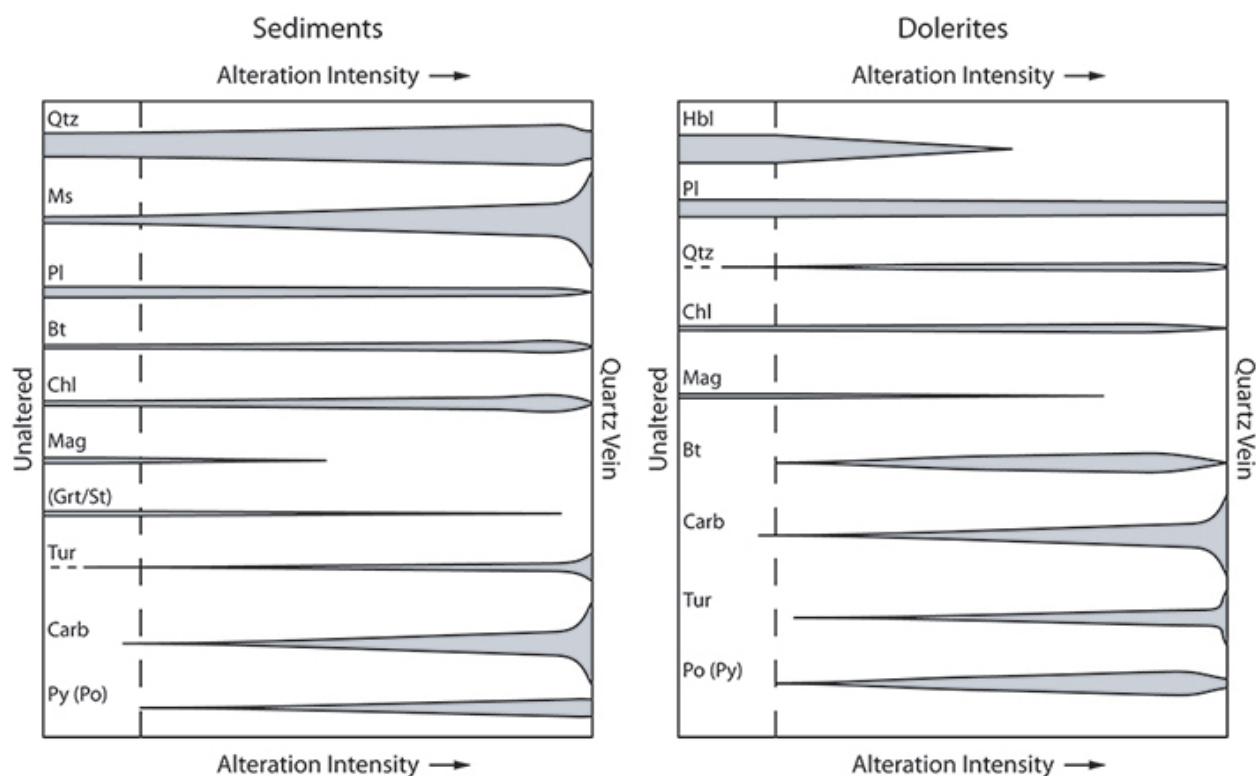
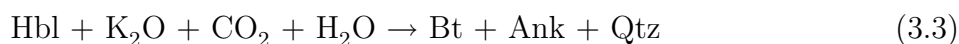


Figure 3.46: Schematic summary of gold-associated alteration assemblages in sediments and dolerites. The dashed vertical line marks the first appearance of sulphide phases and associated gold mineralization. Dashed horizontal lines represent phases that may or may not be present in the unaltered lithology. Phase names in parentheses indicate phases that may or may not be present at all.

Mg and Si. Overall however, there is little or no change in the other major elements, implying that these are conserved. This is particularly noticeable when compared with figure 3.44. It is also noteworthy that this alteration occurs at constant volume, again in contrast to the Pink Haematite alteration.

Comparing this data with petrographic observations and mineral chemical data allows for the determination of generalized, descriptive alteration reactions, depending upon lithology. With regards to the dolerite intrusives, a general reaction exists:



Ca content and Fe/Mg ratios are broadly conserved between phases across this reaction and imbalances in Al can be accounted for by the inclusion of minor phases such as chlorite and/or tourmaline.

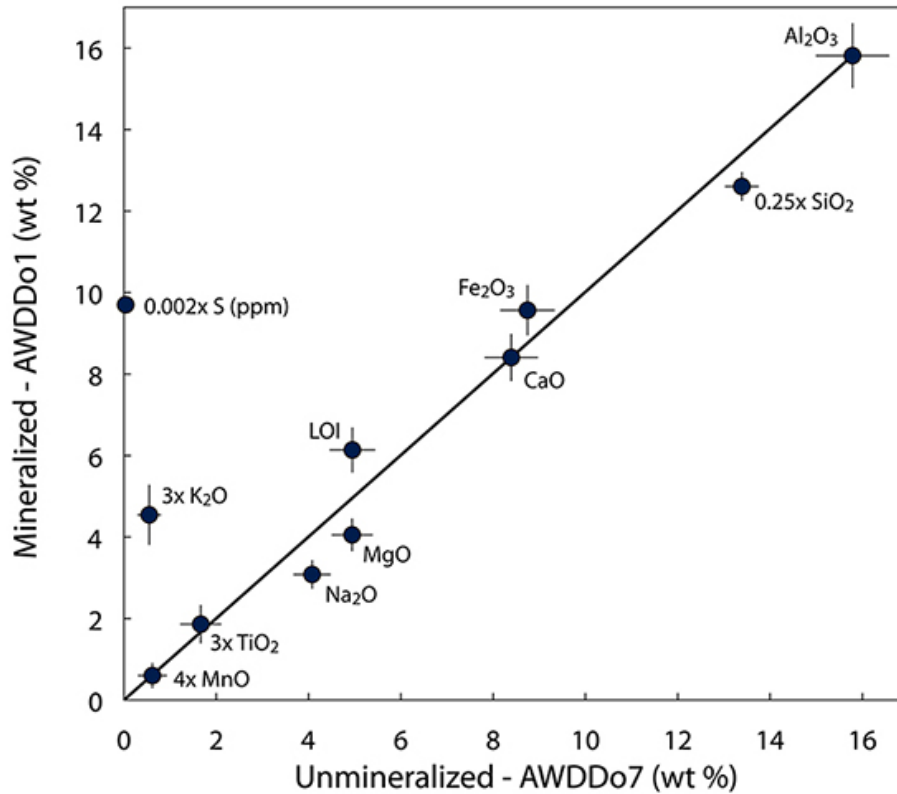


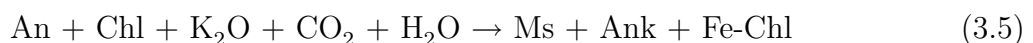
Figure 3.47: Major element isocon plot for mineralized dolerite intrusives, showing significant enrichment in K, S and volatiles (LOI). The heavy line represents the line of constant concentration, the isocon. The line of constant volume in this case is parallel to the isocon, indicating no volume change during alteration. Representative error bars given at 1%.

Variations in intrusive bulk composition, however, can lead to variations of this reaction. For example, considering sample AWDDo2 (see section 3.6), which includes cummingtonite as an alteration product suggests a reaction of the form:



In this case, the produced cummingtonite is more Fe-rich than the original hornblende, as well as being devoid of Al, which is necessarily balanced by a magnesian chlorite. The amount of biotite produced is significantly less than that seen for other mineralized dolerites and serves to highlight the effect of host rock composition on the alteration assemblage.

A generalized reaction for the sedimentary rocks can be given as:



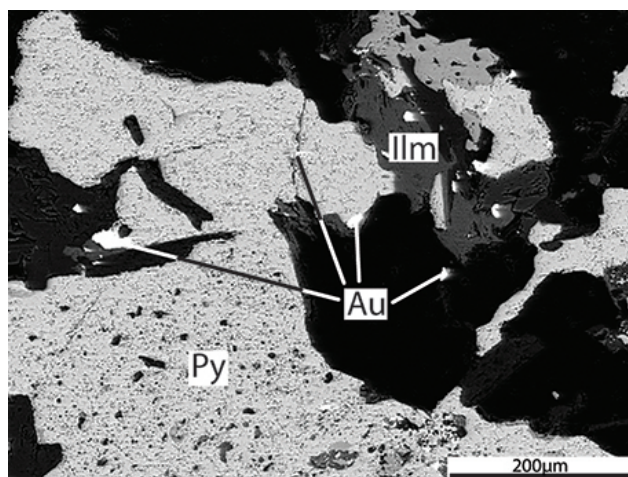
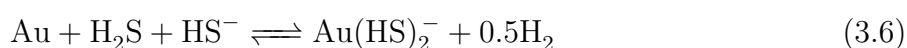


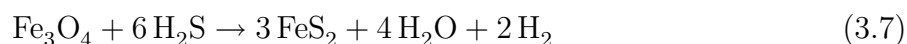
Figure 3.48: Back scattered electron image of gold associated with sulphides in a mineralized Banket Series quartzite, sample AWABf3.

The Fe/Mg ratio of chlorite in unmineralized rocks is slightly lower on average than that of the produced ankerite, suggesting that the majority of Mg is sourced from chlorite (and possibly biotite), with the excess Fe therefore driving the composition of chlorite in mineralized rocks towards chamosite. The consumption of anorthite (in plagioclase) in this reaction must not be on a scale that has a major effect on average plagioclase composition within the rock. Extra Ca may be supplied, particularly in the Tarkwa Phyllite, from the consumption of garnet.

In the types of fluid under consideration here, gold is most commonly transported as a gold-bisulphide complex, according to the following equilibrium (Phillips & Powell, 2010):



Wall rock sulphidation is a reducing process and proceeds through the following, magnetite consuming, reaction:



The sulphide product here is pyrite, although the same principle applies to the formation of pyrrhotite in the more Fe-rich rocks at Damang. The consumption of H_2S in reaction 3.7 serves to drive reaction 3.6 to the left, thereby precipitating Au. This genetic relationship between gold and sulphide phases is evidenced by their intimate spatial association in mineralized rocks (Fig. 3.48).

The abundance of magnetite in the Banket Series, Tarkwa Phyllite and dolerite intrusives therefore, is inferred to explain why these lithologies are so well mineralized in contrast to the Huni Sandstone, which typically contains significantly less magnetite. This is supported by the interaction between gold-associated alteration and the Pink Haematite alteration. Mineralized veins commonly cut through zones of Pink Haematite alteration but only occasionally result in the formation of minor pyrite (and gold), which is interpreted to be a function of the amount of magnetite (and possibly coarse-grained reactive haematite) remaining in the rock.

Overall, the gold-associated alteration can reasonably be described as a sulphidation, carbonation, potassic alteration event by a reducing, CO₂-bearing aqueous fluid. This is the same as that seen at many other orogenic gold deposits, not only in Ghana but globally (see section 1.2; Mikucki & Ridley, 1993; Ridley & Diamond, 2000; Goldfarb *et al.*, 2001, 2005; Dubé & Gosselin, 2007). Similarly the ubiquity of carbonate minerals in all alteration assemblages is indicative of a CO₂-rich mineralizing fluid, which accords with the features of many Birimian-hosted hydrothermal gold deposits across Ghana (Schmidt Mumm *et al.*, 1997; Yao *et al.*, 2001; Wille & Klemm, 2004; Feybesse *et al.*, 2006). This compositional similarity with fluids responsible for other orogenic gold deposits, particularly in Ghana, has significant implications for the Damang deposit. Specifically, the mineralization event at Damang is interpreted to be the same one that produced the Birimian-hosted deposits.

3.10 Summary of Other Hydrothermal Activity

In addition to the two major hydrothermal alteration events discussed above, Damang has experienced other hydrothermal activity. Remobilization of Fe-oxides in the Banket Series has already been discussed in section 3.3, and undoubtedly required the action of some (likely diagenetic) fluid.

At extensively weathered locations, such as Tomento and Rex, the Banket Series contains a high concentration of both bedding parallel and cross-cutting vein-like structures. The ‘veins’ are 1-2cm in thickness and are now composed entirely of a white clay; presumed to be

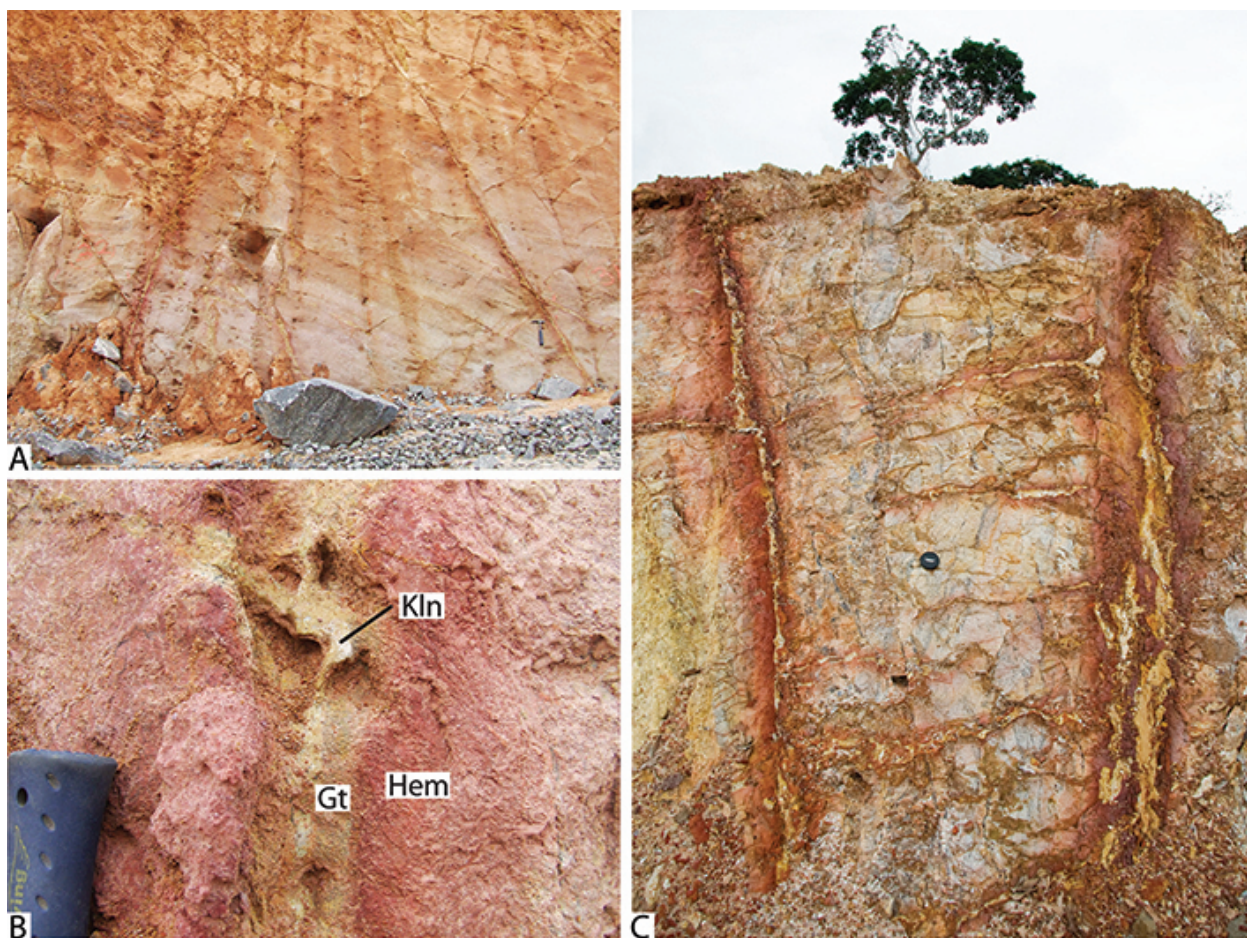


Figure 3.49: Photographs of clay ‘veining’ at Tomento (A, B) and Rex (C). Subsets occur both parallel and discordant to bedding (A). Narrow veins contain white kaolinite, surrounded by an alteration selvage of inner goethite and outer haematite (B).

kaolin. These likely formed through advanced argillic alteration associated with very high fluid/rock ratios. A broad selvage up to 10 cm wide is equally divided between an inner yellow, goethitic alteration zone and an outer red haematite alteration zone that fades with increasing distance border the veins (Fig. 3.49). There are two sub-generations of these clay veins, with the first set occurring parallel to bedding within the Banket Series quartzites. These are cut by a second strongly discordant set that have a more intense alteration halo. Both sets cut across sedimentary structures within the quartzites. No apparent precursor or ‘fresh’ version of these veins has been identified in fresh rock, suggesting that they represent recent weathering during which meteoric or other waters have infiltrated along fracture sets and bedding planes.

A third identifiable hydrothermal event is represented by a set of narrow (up to 10 cm), sub-vertical, broadly east–west orientated quartz veins. This set is common across the

entire Damang camp and are completely barren. These will be discussed further in terms of a paragenetic sequence in section 3.11.

3.11 Structural and Hydrothermal Paragenesis at Damang

The various igneous, metamorphic and hydrothermal events described above, coupled with the structural history outlined in section 2.4, must now be placed into a coherent paragenetic sequence, providing a relative chronology that will be constrained with absolute timings in chapter 5.

The earliest recognisable event to affect the rocks at Damang is the remobilization of Fe-oxides in the Banket Series quartzites (Fig. 3.6, section 3.3). This has been described above and likely relates to diagenesis of the sediments early in the region's history. Although it is best seen in weathered rock, the presence of such features in fresh rock indicates that it is not a recent weathering feature.

The first of the structural events, involved north-directed, sub-bedding parallel, thrust faulting, as seen at Amoanda (see section 2.8.1 and Fig. 2.11). This event has only been recorded at one location and so its regional extent cannot be ascertained. However, it does have an important consequence for later mineralization. At Amoanda the Kawere Group has been thrust above the Banket Series and subsequently acted as a barrier to gold mineralization, restricting mineralization to the underlying Banket Series.

The first demonstrably regional structural event, TD₁, occurred prior to any igneous, hydrothermal or peak-metamorphic events. This was followed by a major igneous-hydrothermal-structural event with intrusion of the dolerite dykes and sills and Pink Haematite alteration broadly coeval with TD₂ deformation. The dykes observed in the main Damang pit clearly cross-cut the regional fold trend (Fig. 3.50A). Although the sills are sub-parallel to the present-day bedding orientation, they are interpreted to have intruded sub-vertically by exploiting the steep dip of the bedding at Damang following TD₁ folding.

There is a strong relationship between the dolerite intrusions and the Pink Haematite al-

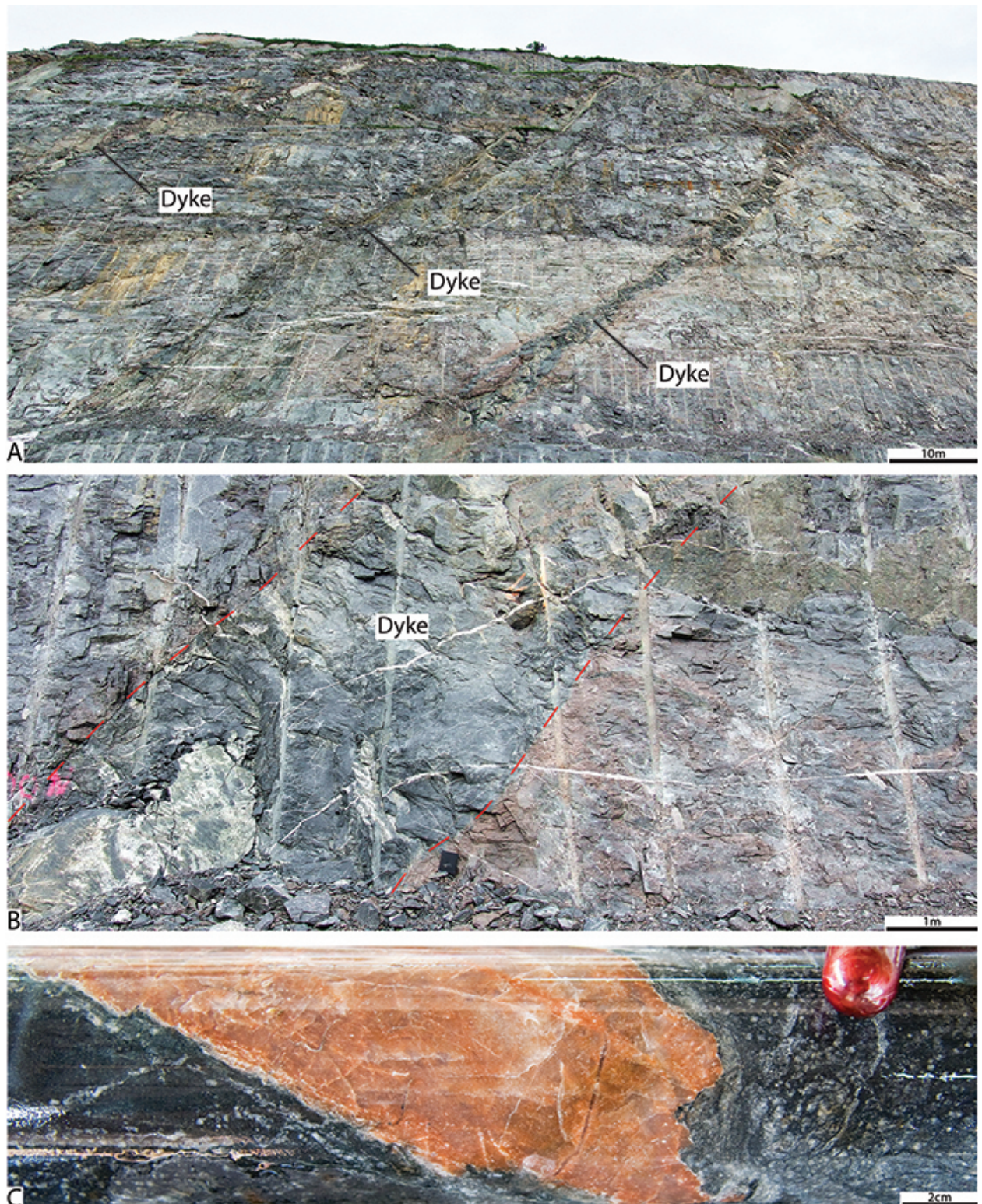


Figure 3.50: Photographs of Pink Haematite alteration associated with dolerite intrusives at Damang. The western wall of the Damang pit is cut by numerous narrow dolerite dykes (A). These typically display a Pink Haematite alteration aureole along their margins (B). Certain intrusives also contain xenoliths of Pink Haematite altered sediments (C).

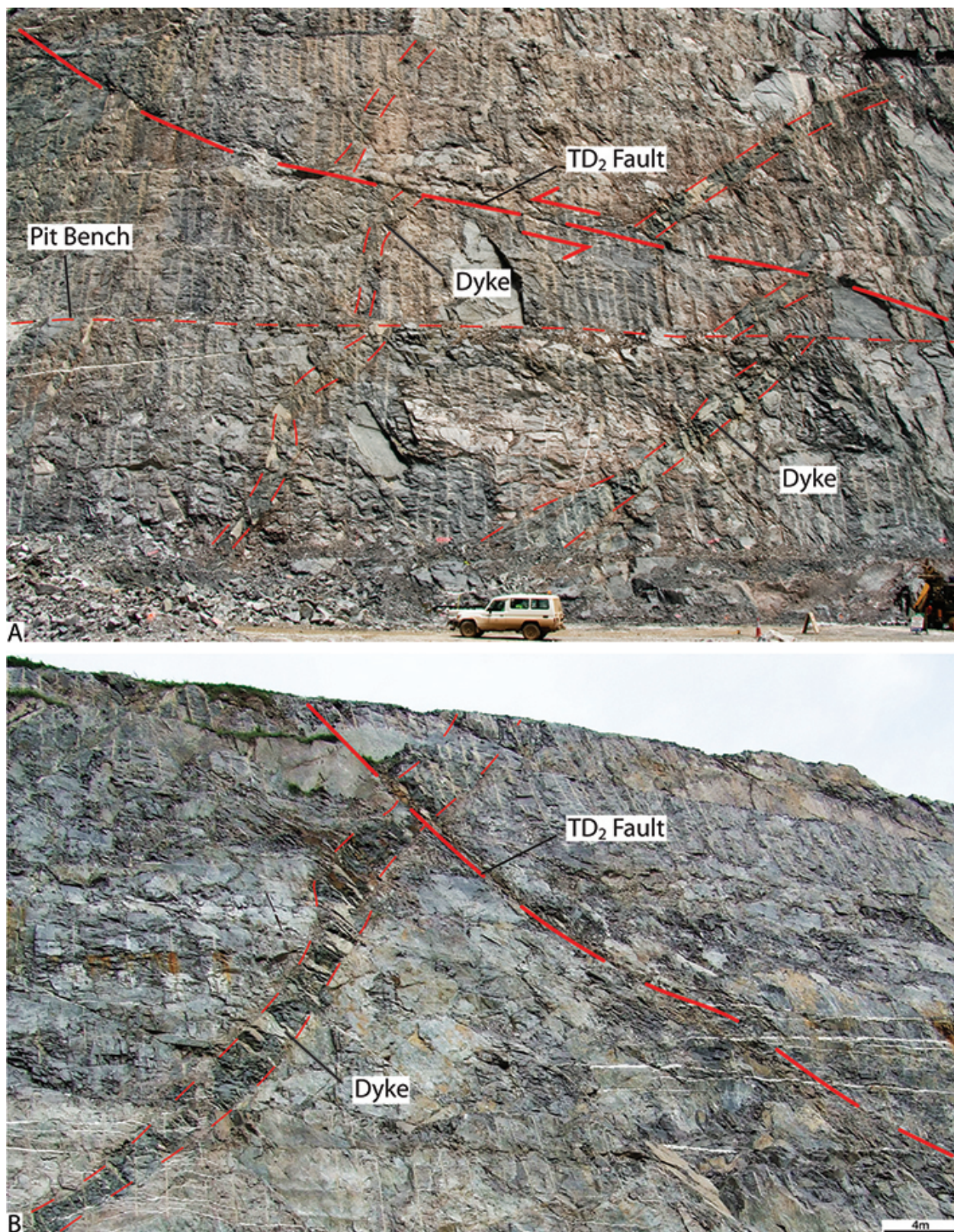


Figure 3.51: Photographs of TD₂ faults cross-cutting dolerite intrusives in the western wall of the Damang pit. There is a progressive decrease in offset moving southwards (left) up structurally upwards (A), implying coeval faulting and intrusion. Where the fault reaches the top of the pit, the offset is essentially zero (B).



Figure 3.52: Photograph of unmineralized quartz veins cross-cutting Pink Haematite alteration at Amoanda.

teration that commonly occurs in an aureole bordering the dykes and decreases in intensity with distance from the dykes (Fig. 3.50B). This would suggest a genetic relationship between the fluid and the intrusives. However, the dolerite intrusives are themselves regularly haematite altered, suggesting that at least in certain cases, the hydrothermal fluid outlasted dyke intrusion. Furthermore, there are examples of intensely Pink Haematite altered sediments as xenoliths within dolerite intrusives, implying intrusion post-dating alteration (Fig. 3.50C). Ultimately, these observations indicate that Pink Haematite alteration and dolerite intrusion was broadly contemporaneous. It is therefore likely that the Pink Haematite fluid is magmatic in origin and related to this regional igneous event.

Examination of cross-cutting relationships between the dolerite dykes and TD₂ faults in the western wall of the Damang pit shows that while the faults cross-cut all dykes, the amount of offset exhibited by each dyke differs systematically, decreasing structurally upwards and southwards (Fig. 3.51). This suggests progressive intrusion of this particular dyke set while TD₂ faults were active.

This is followed by formation of the first generation of quartz veins. As described above, these

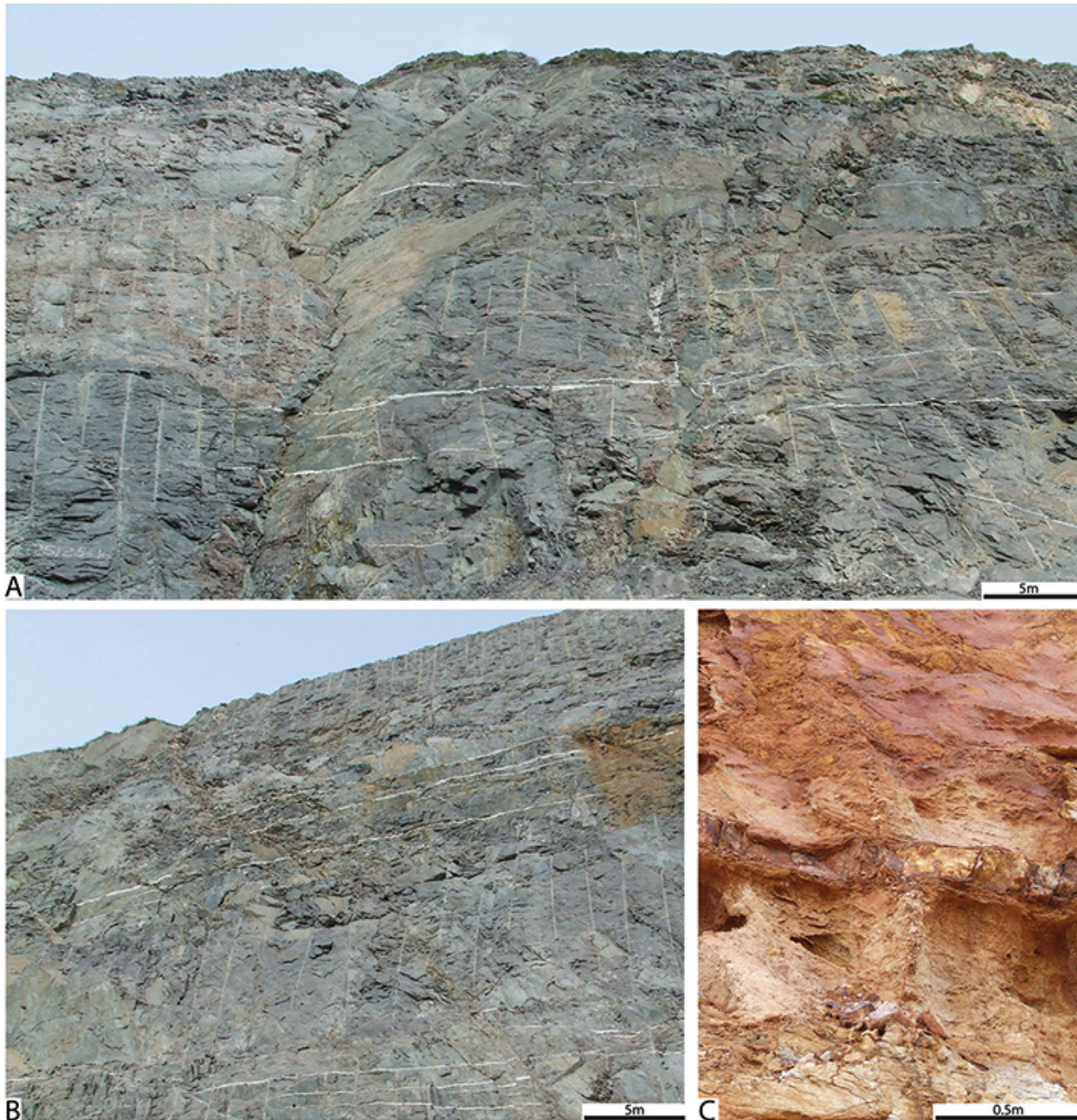


Figure 3.53: Photographs of mineralized quartz veins cross-cutting earlier structures at Damang (A, B) and Tomento (C). Veins clearly cross-cut dolerite dykes (A), TD₂ faults (A, B) and unmineralized quartz veins (C). See also figures 3.50B and 2.16B.

are narrow, sub-vertical, orientated approximately east–west and are completely unmineralized. In outcrop these veins can clearly be seen to cross-cut Pink Haematite alteration (Fig. 3.52). The specific origin of these veins is difficult to ascertain since there is no associated hydrothermal alteration or other igneous or structural event occurring simultaneously.

Peak regional metamorphism must also post-date dyke intrusion since they are now overprinted with an amphibolite facies mineral assemblage. The relative sequence of metamor-



Figure 3.54: Photograph of unmineralized quartz veins cross-cutting clay veining at Tomento. Although the argillic alteration is interpreted to be recent, it appears to exploit an early fracture set.

phism and the unmineralized quartz veins however, is unknown. It is certainly possible that these veins are related to fluid generation during metamorphism, although this cannot be verified.

The last major event to affect the Damang region is TD₃ deformation and formation of mineralized, flat-lying quartz veins plus minor faulting. These clearly cross-cut all earlier structures, and vein and alteration episodes (Fig. 3.53). Similarly, the hydrothermal alteration associated with these veins overprints metamorphic assemblages in all affected lithologies, supporting a post-peak metamorphic origin for the gold mineralization at Damang.

Finally, the clay ‘veining’ observed in weathered Banket Series quartzites at Tomento and Rex formed during recent weathering. Although the alteration itself is interpreted to be recent, the veins are commonly cross-cut by the earlier unmineralized quartz veins, implying the argillic alteration is exploiting fracture sets that pre-date the formation of the unmineralized veins (Fig. 3.54).

A schematic depiction of the paragenesis discussed in this section is presented in figure 3.55.

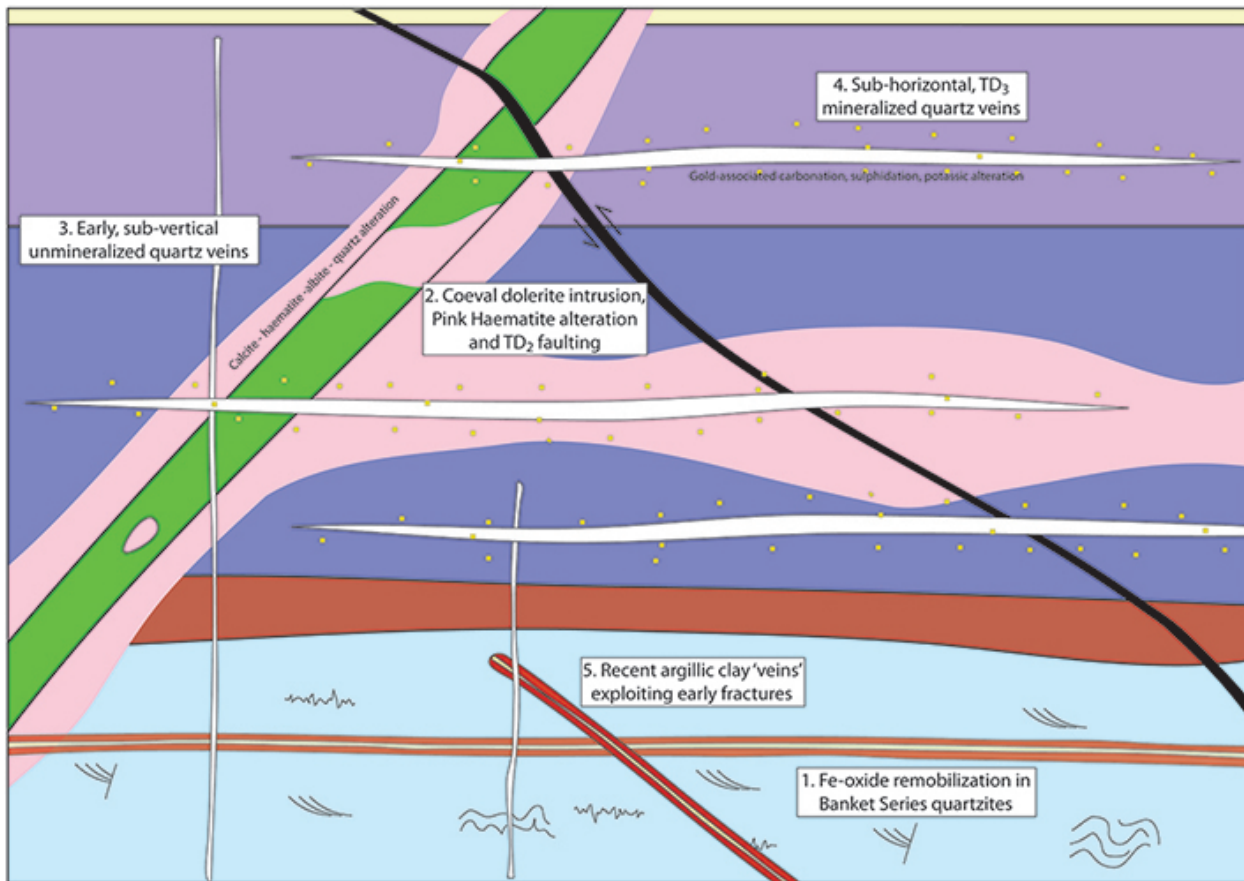


Figure 3.55: A simplified, schematic illustration of the structural and hydrothermal paragenesis at Damang. Pre-TD₁ and TD₁ structural episodes are not shown. Time approximately increases vertically upwards. Stratigraphy colours as per figure 2.8.

Chapter 4

Metamorphism and Metasomatism



Coreyard companions, the Agama lizards.

Modern petrological studies, as with all aspects of the earth sciences, are interested in not only qualitatively understanding observations but in quantifying them. This requires a fundamental appreciation for thermodynamics and the role that it plays in controlling the stability of minerals, mineral assemblages and metamorphic reactions. The study of thermodynamics as a science in its own right began in the mid to late 17th century with the desire to develop and then improve the efficiency of heat engines. The term ‘thermodynamics’ however, was not coined until 1849, by Lord Kelvin.

Thermodynamic principles were first applied to geological problems in the early parts of the 20th century (see introduction chapter of [Kern & Weisbrod \(1967\)](#)). However, it wasn’t until the 1950s that work focussing primarily on the thermodynamics of rocks and minerals were published ([Ramberg, 1952](#); [Thompson, 1955](#); [Laffitte, 1957](#); [Korzhinskii, 1959](#)). Throughout the 1960s, the application of thermodynamics was expanded and became a common part of igneous and metamorphic petrology text books ([Turner & Verhoogen, 1960](#); [Turner, 1968](#)), with the first dedicated ‘thermodynamics for geologists’ appearing shortly after ([Kern & Weisbrod, 1967](#); [Wood & Fraser, 1977](#); [Powell, 1978](#)). The biggest limitation to these early applications was the paucity of data, which existed only for standard-state, pure end-members of some of the most common mineral phases ([Turner, 1968](#)). The first attempts to compile useful thermodynamic data sets for petrological applications utilized primarily calorimetric data ([Robie & Waldbaum, 1968](#); [Robie *et al.*, 1979](#)) while more recent data sets (and those typically used today) employ phase equilibrium data ([Helgeson *et al.*, 1978](#); [Holland & Powell, 1985](#); [Berman, 1988](#)). These self-consistent data sets allow for the calculation of phase relationships to extract meaningful temperatures and pressures from observed assemblages. This is the study of thermobarometry.

Thermobarometry is of huge importance in understanding and quantifying large-scale, regional tectonic processes. Much of the Ashanti Belt, and indeed the other Birimian belts of Ghana, are typically considered to be of broadly greenschist facies metamorphism (see section [2.3](#)). However, simple petrography of rocks from Damang, particularly the Tarkwa Phyllite, indicate significantly higher grade, amphibolite facies metamorphism. It is important to accurately constrain the conditions of peak metamorphism in order to assess any effect that

elevated pressure and/or temperature may have had on subsequent gold mineralization at Damang.

In order to fully integrate the Damang deposit into the regional metallogenic and tectonic framework and compare it to other orogenic gold deposits in Ghana, pressure, temperature and fluid composition conditions of hydrothermal alteration associated with the gold mineralization must also be calculated. The same thermobarometric principles apply in this case, with slight adaptations for dealing with open, fluid-rich systems.

4.1 Principles of Thermobarometry

Thermobarometry is the study of pressure and temperature (P-T) conditions experienced by rocks and their mineral assemblages. Conventional techniques typically use single equilibria, calibrated with experimental or empirical data. More recently, computer programs, such as THERMOCALC (Powell & Holland, 1988), have provided significant improvements to this by combining multiple reactions with large thermodynamic data sets. These internally consistent data sets provide a single, optimized set of thermodynamic properties for each phase and are able to reproduce a large body of experimental observations. Continued advances in THERMOCALC and similar programs have continued to improve the ability to apply such data sets for practical purposes.

4.1.1 Conventional Thermobarometry

Stable mineral equilibria are fundamentally controlled by the thermodynamics of the system, in which temperature and pressure are the most important intensive variables. Compositional data of the mineral phases involved in any particular reaction can be used to define the position of equilibrium. At a specific pressure (P), temperature (T) and mineral composition (X) the Gibbs free energy of a reaction (ΔG_R) at equilibrium is zero:

$$\Delta G_R(P, T, X) = 0 \quad (4.1)$$

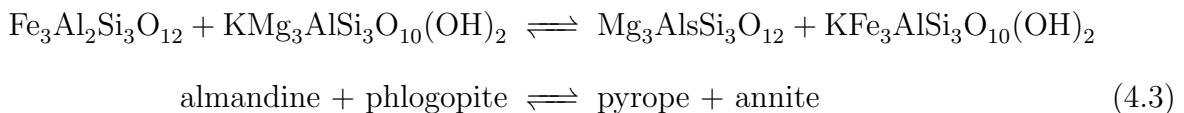
The full thermodynamic relationship between pressure, temperature and mineral composition can be expressed as follows:

$$\Delta G_R(P, T, X) = 0 =$$

$$\Delta H^0 + \int_{298}^T \Delta C_p^0 dT + \int_1^P \Delta V^0 dP - T \left(\Delta S^0 + \int_{298}^T \frac{\Delta C_p^0}{T} dT \right) + RT \ln K_{eq} \quad (4.2)$$

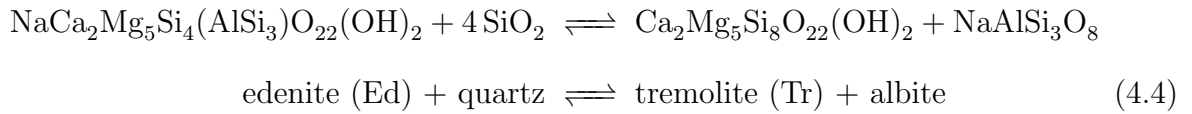
where ΔH^0 , ΔS^0 , ΔV^0 and ΔC_p^0 are the standard state (298 K, 1 bar) enthalpy, entropy, volume and heat capacity of the reaction respectively and the integral limits indicate adjustment of the data from this standard state (Spear, 1995). The equilibrium constant (K_{eq}) is the activity product of the reaction products divided by the activity product of the reaction reactants and can be calculated from mineral compositional data. Values of ΔH , ΔS , ΔV and ΔC_p can be experimentally determined. Ultimately this allows for the calculation of the equilibrium temperature at a selected pressure, or vice versa. The use of two or more independent equilibria allows for two reactions of the form of equation 4.2 to be solved simultaneously.

The suitability of any particular reaction for use as a geothermometer or geobarometer is dependent upon the values of ΔH , ΔS and ΔV for that reaction. Those with a relatively large ΔH and ΔS and small ΔV are relatively temperature sensitive and pressure insensitive and are therefore effective geothermometers. Such reactions commonly involve cation exchange between two mineral phases, for example the exchange of divalent Fe and Mg between biotite and garnet, which is commonly used for amphibolite facies metapelites:

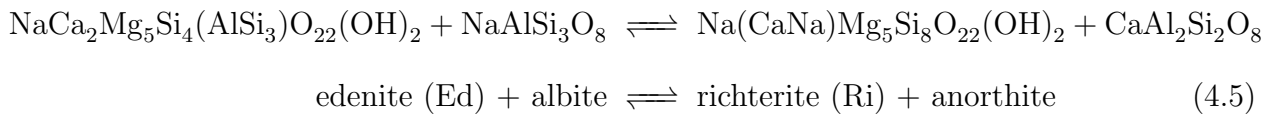


Many cation exchange equilibria in compositionally and crystallographically complex mineral reactions involve coupled substitutions of ion pairs. Examples of such equilibria include the hornblende-plagioclase thermometers of Holland & Blundy (1994). The first involves the $(\text{NaAl})_1(\square\text{Si})_{-1}$ exchange vector, where $\square$ represents a lattice vacancy, and can only be

applied to silica-saturated samples:



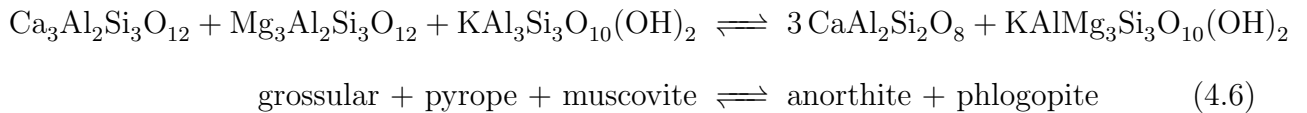
The second involves the $(\text{NaSi})_1(\text{CaAl})_{-1}$ exchange vector and can be applied to both silica-saturated and silica-undersaturated samples:



Both of these can be applied to the Damang dolerite intrusions as many of the samples do contain free quartz and are therefore silica-saturated.

An alternative thermometry method is that of [Henry *et al.* \(2005\)](#), which determines equilibrium temperatures from the Ti content of biotite in graphitic, peraluminous metapelites that contain ilmenite or rutile.

By contrast, reactions with small ΔH and ΔS and a large ΔV are relatively sensitive to pressure but insensitive to temperature and are effective geobarometers. Such reactions commonly involve net transfer with the production and consumption of mineral species, for example the garnet-plagioclase-muscovite-biotite (GPMB) barometer:



The above equation can also be written for the Fe-bearing end-members of garnet (almandine) and biotite (annite).

4.1.2 THERMOCALC Average P-T

A single P-T estimate for a rock can be calculated using the computer program THERMOCALC (Powell & Holland, 1988) in average P-T mode, which employs as many equilibria as are possible based on an input of measured phase compositions. This set of independent reactions is combined with an internally consistent dataset of fundamental thermodynamic properties for phases of petrological interest (Holland & Powell, 1998) to solve for P-T conditions. Care must be taken, however, as is the case with conventional thermobarometry, to only input those phases and phase compositions that represent equilibrium peak conditions. The inclusion of retrogression products or pre-peak relics affects the set of defined reactions and serves to give inaccurate and imprecise P-T estimates.

4.1.3 THERMOCALC Pseudosections

A pseudosection is a phase diagram showing the stability fields of mineral assemblages that are at thermodynamic equilibrium, i.e. they satisfy equation 4.2, as a function of two independent variables. Pseudosections are calculated for a single, specified bulk rock composition. A new diagram must therefore be calculated for each sample being investigated. Pseudosections can be constructed using THERMOCALC in phase diagram mode in combination with an internally consistent thermodynamic data set (Powell *et al.*, 1998).

For a system of C components containing P phases, where for each phase, k , exists e_k end-members, and therefore $e_k - 1$ compositional variables, at equilibrium there exists a total number of end-members, S , given by:

$$\sum_{k=1}^P e_k = S \quad (4.7)$$

and the total number of compositional variables given by:

$$\sum_{k=1}^P (e_k - 1) = S - P \quad (4.8)$$

There exists a total of ${}^SC_{(C+1)}$ reactions between the S end-members, of which $S - C$ are independent, i.e. they are not linear combinations of other reactions (Powell & Holland, 1988). There are therefore $S - C$ variables that can be solved for. At equilibrium, each of these reactions satisfies equation 4.2, and the total number of variables is $S - P + 2$, i.e. the number of compositional variables plus pressure plus temperature. The degrees of freedom in the system is then given by the number of variables ($S - P + 2$) minus the number of constraints ($S - C$), which equates to:

$$P + F = C + 2 \quad (4.9)$$

where F is the variance of the system. This is the Gibbs phase rule. All phase diagrams are constructed by considering the variance of the mineral assemblage under inspection, and the Gibbs phase rule shows that, for a given system with a certain number of components, an assemblage with a higher number of phases has a lower variance. This usually allows it to be constrained more accurately and reliably.

In addition, for a closed system with a known bulk composition, there exists a mass balance equation for each system component, C , of the form:

$$M(C_i) = \sum_{k=1}^P X(P_k) \cdot M(C_i^k) \quad (4.10)$$

where $M(C_i)$ is the total mole proportion of component i , $X(P_k)$ is the modal proportion of phase k and $M(C_i^k)$ is the mole proportion of component i in phase k . This provides an additional C constraints, while the unknown modal proportions of the P phases are additional variables.

The total number of constraints is therefore $S - C + C$, and the total number of variables is $S - P + 2 + P$, making the total variance (variables - constraints) 2. This implies that for the system described here, with a specified bulk composition, the equilibrium state is completely defined by two independent variables (P , T , $X(P_k)$ or $M(C_i^k)$). This is Duhem's Theorem (Prigogine & Defay, 1954)

However, T-X(CO₂) pseudosections involve a mixed-component fluid phase (binary H₂O–CO₂), the true composition of which is unknown. Consequently, the rock does not have a fully specified bulk composition. For an open system such as this, Korzhinskii’s version of the phase rule is more appropriate (Korzhinskii, 1959, cited in Yardley (1989), p. 124):

$$P_s + F = C_i + 2 \quad (4.11)$$

where P_s = number of solid phases, C_i = number of ‘inert’, or in this case, non-volatile components. T-X(CO₂) pseudosections therefore show the equilibrium solid mineral assemblage that is stable with varying compositions of fluid.

The difference in variance (variables - constraints) between closed (Gibbs) and open (Korzhinskii) systems is outlined in Table 4.1. In this case, the total variance of 3 under Korzhinskii’s phase rule implies that three variables must be specified in order to determine the equilibrium state of the system with a partially specified bulk composition. As T-X(CO₂) diagrams are isobaric, then two of the remaining remaining variables (T, $X(P_k)$ or $M(C_i^k)$) must be set. Furthermore, variance under Korzhinskii’s phase rule increases by 1 for each additional perfectly mobile (volatile) component. For example, a ternary component fluid would result in $C - 3$ mass balance constraints, making the total variance 4. Therefore, Korzhinskii’s phase rule can be applied to any open system.

The simplest application of a pseudosection is to identify the variable space in which a given mineral assemblage occurs. For example, the pressure-temperature range of an observed assemblage of interest. Additionally, such assemblage fields can be contoured for mineral composition or mode. The intersection of these contours, or isopleths, thereby constrains the specific location within the field. Furthermore, the calculated sequence of assemblage and mineral composition changes along a P-T or T-X path can be directly compared to observed reaction microstructures and zoned refractory minerals. This allows for details of the evolution of a rock to be determined; something that is not possible with conventional thermobarometry. Finally, deviation of an observed mineral assemblage, mineral compositions and mineral modes from the calculated pseudosection may serve to highlight where

Table 4.1: Comparison of closed and open system behaviour. A closed system is one in which the bulk composition is fully specified and for which the Gibbs phase rules applies. An open system is one that is open to a binary (in this case) fluid and for which Korzhinskii's variation of the phase rule is more appropriate.

Closed system		Open system	
Variables			
2	P and T	2	P and T
$S - P$	Phase composition variables	$S - P$	Phase composition variables
P	Modal proportions of phases	$P - 1$	Modal proportions of solid phases (P_s) - modal amount of fluid cannot be determined
Constraints			
$S - C$	Independent reactions	$S - C$	Independent reactions
C	Mass balance per system component	$C - 2$	Mass balance per inert system component (C_i) - masses of volatile components (H_2O and CO_2) are unknown and cannot be balanced
Variance = Variables - Constraints			
2		3	

disequilibrium processes have occurred.

4.1.4 Uncertainties in Thermobarometric Calculations

Uncertainties on thermobarometric calculations are introduced at virtually all stages of the calculation process. These uncertainties may be random or systematic, quantifiable or qualitative. Four of the most common and important sources of uncertainty include (Powell, 1985; Hodges & McKenna, 1987; Waters, 2004; Powell & Holland, 2008; Whitney, 2011):

1. 'Geological' error. These include a range of systematic, unquantifiable errors associated with the identification of phases and/or phase compositions that are potentially incorrectly interpreted to represent metamorphic equilibrium. This is particularly relevant for zoned minerals or where multiple generations of a mineral exist in a rock. Careful petrographic observations may be able to satisfactorily discount certain retrogressive phases, while rim compositions of zoned minerals can be selected, which partly reduces this uncertainty. How-

ever, even this cannot account for partial re-equilibration of a given assemblage during slow uplift and cooling, resulting in an underestimation of ‘peak’ conditions ([Hodges & McKenna, 1987](#)).

2. Analytical uncertainties. Uncertainty produced through microprobe analyses of mineral compositions are derived from counting statistics, ZAF correction procedures and standardization. These are readily quantified and propagated through thermobarometric calculations ([Hodges & McKenna, 1987](#)). Such errors can be reduced by averaging mineral analyses for a statistically homogeneous population. Additional systematic errors may be introduced through compositional variation of a phase within a sample, which are also more difficult to quantify. A further important source of uncertainty arises from formula recalculation. In particular, the determination of ferric iron content in minerals, which cannot be measured directly ([Powell, 1985](#); [Powell & Holland, 2008](#)). In certain minerals, such as garnet, stoichiometry and charge balance can be used to provide a reasonable estimate, while for more complex minerals, and particularly hydrous phases, this is not possible. This uncertainty is rather difficult to quantify.

3. Calibration uncertainties. Calibrations of end-member equilibria are made by best-fitting a regression line between bracketed phase equilibrium experiments. This regression has associated, quantifiable uncertainties that can be systematically propagated through thermobarometric calculations ([Powell, 1985](#); [Hodges & McKenna, 1987](#)), especially if the calibrations for multiple equilibria are derived from an internally consistent data set.

4. Activity-composition model uncertainties. These uncertainties relate to the choice of formulation for a phase solution model ([Waters, 2004](#)). Many minerals exhibit non-ideal behaviour for which major corrections are required to achieve sensible activities from measured compositions. The nature and magnitude of these corrections may not be known and/or may not extrapolate to the temperatures and pressures of interest. These uncertainties are very difficult to quantify and will apply systematically to all equilibria involving any given phase.

Since many of these uncertainties are partially or totally unquantifiable, any quoted error on a pressure-temperature estimate should be treated as a minimum ([Powell & Holland,](#)

2008). Overall, many commonly used conventional geothermometers and geobarometers have an associated quantifiable uncertainty of around ± 50 °C and ± 1 kbar (2σ) (Powell & Holland, 2008; Whitney, 2011). In contrast, thermobarometric estimates determined from pseudosections are generally better constrained with quoted uncertainties on the order of ± 20 °C and ± 0.5 kbar (2σ) (Powell & Holland, 2008), although this primarily relates to the calibration uncertainty. Uncertainties determined from pseudosection calculations are typically smaller than for conventional techniques since the size of an assemblage field, defined by the presence or absence of certain phases, places additional constraints and can often be smaller than the uncertainty on a conventional thermobarometer (Powell & Holland, 2008). This is despite the uncertainty on the field boundaries themselves.

Finally it is important to consider absolute versus relative uncertainty when interpreting the results from a number of samples. All systematic errors, particularly points 3 and 4 above, and those in point 2 not related to microprobe procedures, will apply similarly to all samples. These, therefore, do not affect the ability to resolve two results obtained with the same calibration. As such, the relative uncertainty between multiple results may be significantly less than the absolute errors on each result (Worley & Powell, 2000; Waters, 2004).

4.2 Pseudosection Calculations

Pseudosections were calculated using THERMOCALC 3.33 (Powell & Holland, 1988, updated October 2009) with the updated thermodynamic data set 5.5 (Holland & Powell, 1998, updated November 2003) in subsets of the model system MnO-Na₂O-CaO-K₂O-FeO-MgO-Al₂O₃-SiO₂-CO₂-H₂O-TiO₂-Fe₂O₃ (MnNCaKFMASCHTO).

4.2.1 Activity-Composition Relationships

Activity-composition (a - x) relationships for solid solution minerals are from Holland & Powell (1998) (staurolite, cordierite, epidote), Holland *et al.* (1998) (chlorite), White *et al.* (2000)

(magnetite), [White *et al.* \(2002\)](#) (spinel), [Holland & Powell \(2003\)](#) (feldspars), [White *et al.* \(2003\)](#) (carbonates), [Diener *et al.* \(2007\)](#) (amphiboles, modified by [Diener & Powell \(2011\)](#)), [Green *et al.* \(2007\)](#) (diopside, modified by [Diener & Powell \(2011\)](#)), and [White *et al.* \(2007\)](#) (melt). The ilmenite models of [White *et al.* \(2005\)](#) and [White *et al.* \(2000\)](#) were used in Mn-bearing and Mn-free systems respectively. The biotite models of [White *et al.* \(2005\)](#) and [White *et al.* \(2007\)](#) were used in Mn-bearing and Mn-free systems respectively. The garnet model is an asymmetric formulation of D.J. Waters (*pers. comm.*) with ferric iron and dependent almandine. Albite was included as a pure end member so as to allow for the calculation of the peristerite gap. An ideal binary H_2O – CO_2 fluid was used in CO_2 -bearing systems. In all other cases a H_2O activity of 1 was assumed. The solution models listed here are given in appendix F. Additional pure phases included calcite, andalusite, kyanite, sillimanite, sphene, rutile and quartz.

White Mica a-x Relationships

The most up-to-date *a-x* model for white mica (muscovite and paragonite) is that of [Coggon & Holland \(2002\)](#), which was used for the majority of pseudosection calculations involving muscovite. However, when used in calculations of samples of the Tarkwa Phyllite, this model predicts muscovite compositions that contain unrealistically high proportions of Na (often around 50%) and fall well inside the expected miscibility gap between muscovite and paragonite. This is partly a function of the high $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratio of these rocks, but is also related to the design of the *a-x* model itself. This model was optimized for high pressure white micas by introducing a pressure dependency to account for the low Na content of phengitic muscovites. While it is still applicable to most rocks across a range of pressure conditions, it is clearly not suited to highly sodic micaceous rocks at low–intermediate pressures. Therefore, for calculations of the Tarkwa Phyllite an older muscovite model was used, which is derived from [Holland & Powell \(1990\)](#), with the addition of Na and Fe, and can be obtained from [Holland \(2009\)](#). Paragonite was modelled using a simple, ideal binary (Na-K) solution combined with Darken’s Quadratic Formalism (DQF, [Powell \(1987\)](#)).

4.2.2 Choice of Model System

The choice of model system used for pseudosection calculations has significant implications for the results. The inclusion, or not, of certain chemical components can have a dramatic effect on the stability ranges of many minerals and all calculations should therefore be carried in as large a system as possible (White *et al.*, 2007; Powell & Holland, 2008; Pattison & Tinkham, 2009). In this study, metapelites were modelled in the system MnO-Na₂O-CaO-K₂O-FeO-MgO-Al₂O₃-SiO₂-H₂O-TiO₂-Fe₂O₃ (MnNCaKFMASHTO), unaltered metadolerites in Na₂O-CaO-FeO-MgO-Al₂O₃-SiO₂-H₂O-TiO₂-Fe₂O₃ (NCaFMASHTO) and mineralized lithologies in Na₂O-CaO-K₂O-FeO-MgO-Al₂O₃-SiO₂-CO₂-H₂O-TiO₂-Fe₂O₃ (NCaKFMASCHTO).

4.2.3 Bulk Composition Determination

Sample bulk compositions used for calculating pseudosections have been derived by combining average measured mineral compositions with modal proportions of phases. Modal proportions were initially estimated using an optical microscope, before pixel-counting greyscale backscattered electron images to determine accurate proportions. This approach allows for the best estimate of the local equilibrating model system that can then be compared directly to observed phase modes and compositions. Other methods of bulk composition determination, such as X-ray fluorescence (XRF) analysis, result in the loss of textural information and make an assumption that the whole of the XRF sample represents the same equilibrium as can be observed in a single thin section. XRF analysis introduces further issues such as the contribution of elements from phases, part of whose compositions lie outside of the model system. The bulk composition would therefore need to be projected from these phases into the modelled system. This, however, assumes that the same phases are responsible for holding the minor elements not in the model system, and thereby sequestering the relevant major elements, throughout the P-T-X history of interest, which is not always certain.

Further to this, each average mineral composition has been converted to an ‘ideal analysis’

appropriate for the THERMOCALC mineral a - x model being used (Powell & Holland, 2008, Appendix D). This involves a least squares difference minimisation between the measured phase composition and the idealised phase composition. This is necessary as the mineral a - x models used by THERMOCALC have defined stoichiometry and sets of substitution relationships that closely approximate the real minerals. To illustrate this, consider staurolite, whose a - x relationship uses the formula $(\text{Fe,Mg,Mn})_4\text{Al}_{18}\text{Si}_{7.5}\text{O}_{44}(\text{OH})_4$. It is possible, and indeed likely, that the true composition will contain minor, yet measurable, amounts of other elements such as Ti. It is not sensible to include this Ti in the average mineral composition used to calculate a rock's bulk composition since THERMOCALC would not be able to accommodate it in the modelled staurolite and would instead place it another phase. This could have a knock-on effect in terms of modelled phase compositions and modes. Bulk compositions used for pseudosection calculation are given in Table 4.2 with further details provided in appendix D..

4.3 Regional Metamorphism P-T Results

4.3.1 Conventional Thermobarometry and THERMOCALC Average P-T Results

Results of THERMOCALC average P-T calculations and conventional thermobarometry are presented in tables 4.3 and 4.4 respectively. Further details of THERMOCALC average P-T results, including end-member activities, are presented in appendix E. Ti-in-Bt results are

Table 4.2: Bulk compositions used in THERMOCALC pseudosection calculations.

Sample	Lithology	(Mol. %)									
		SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O	TiO ₂	MnO	Fe ₂ O ₃
AWDP1	Phyllite	62.39	16.00	0.15	3.76	10.04	2.99	2.15	0.48	0.20	1.85
AWDBm1	Phyllite	68.55	15.59	0.41	2.88	5.86	2.95	2.69	0.44	0.09	0.54
AWDDo3	Dolerite	52.47	13.32	13.46	11.08	6.11		2.94	0.58		0.03
AWDDo1	Dolerite	58.62	9.66	10.63	8.56	6.84	1.44	3.10	1.15		0.0033
AWABf3	Quartzite	83.94	7.16	1.87	1.67	1.82	1.84	1.30	0.39		0.00056
AWDDo2	Dolerite	50.20	10.84	10.58	13.66	10.56	0.14	3.28	0.37		0.36
AWDP4	Phyllite	61.44	22.78	3.08	1.79	2.10	4.95	3.86			

presented in figure 4.1. Tarkwa phyllite samples show consistently good agreement in both temperature (557–602°C) and pressure (6.7–7.2 kbar) across the different calibrations and calculation methods. Sample AWDF1 is a Tarkwa Phyllite from within the Damang Fault zone and due to extensive alteration, the calculated pressure and temperature conditions are unlikely to represent regional metamorphism. Sample AWDCs1, a Banket series conglomerate, exhibits reasonably good correlation in temperature and pressure, although the Grt-Bt thermometer gives much higher temperatures and the calculated THERMOCALC average pressure is very low. These discrepancies can be ascribed to partial small-scale disequilibrium between different matrix domains in the sample. Mineralized sedimentary lithologies show an increase in Ti content, resulting in higher calculated temperatures, although these are not interpreted to be real temperatures. Ti-in-Bt temperatures are also given for the mineralized dolerite intrusives and sample AWDP2, although these are outside of the lithological range of the calibration and are shown for reference only. Hbl-Pl thermometer temperatures for the dolerite intrusives are consistent amongst themselves (643–683°C), but significantly higher than those for the sedimentary lithologies. In contrast, THERMOCALC average P-T conditions for sample AWDDo3 are more similar to the sedimentary lithology estimates. Overall, conventional thermobarometry and THERMOCALC average P-T methods suggest peak metamorphic conditions of 550–590°C and 6.5–7 kbar. These results are significantly higher grade than the established interpretation of regional greenschist facies metamorphism.

Table 4.3: THERMOCALC average P-T results. Temperatures in °C, pressures in kbar. σ represents $\pm 1\sigma$.

Sample	Lithology	P	σ	T	σ	correl. ¹	fit ²	N _R ³
AWDP1	Tarkwa Phyllite	6.9	1.5	584	18	0.383	1.36	9
AWDBm1	Tarkwa Phyllite	6.7	1.3	578	17	0.376	1.31	9
AWDF1	Tarkwa Phyllite	6.0	1.6	615	129	0.866	0.49	4
AWDCs1	Banket Conglomerate	3.6	3.4	550	37	0.878	1.74	7
AWDBf6	Banket Quartzite	6.8	1.2	580	20	0.359	1.64	12
AWABv1	Birimian Volcaniclastic	-	-	501	20	-	1.41	4
AWDDo3	Dolerite Intrusive	7.5	2.2	594	64	0.963	1.31	4

¹correlation between uncertainties on P and T

²fit = $\sqrt{\text{MSWD}}$

³N_R = Number of independent reactions

Table 4.4: Conventional thermobarometry results. Temperatures in °C, pressures in kbar.

Sample No.	Lithology	Ti in Bt ^a	Thermometry				Hbl-Pl ^c			Barometry GPMB ^d		
			9% Fe ³⁺	Grt-Bt ^b	15% Fe ³⁺	Ave	Ed-Tr	Ed-Ri	Ave	Mg	Fe	Ave
AWDP1	Tarkwa Phyllite	580	582	570	576		-	-	-	7.7	6.6	7.2
AWDP2	Tarkwa Phyllite	632 ¹	-	-	-		-	-	-	-	-	-
AWDP4	Tarkwa Phyllite	602	-	-	-		-	-	-	-	-	-
AWDP5	Tarkwa Phyllite	584	-	-	-		-	-	-	-	-	-
AWDBm1	Tarkwa Phyllite	572	563	551	557		-	-	-	7.4	6.2	6.8
AWDF1	Tarkwa Phyllite	631	578	563	571		-	-	-	5.6	4.3	5.0
AWDCs1	Banket Conglomerate	593	706	691	699		-	-	-	7.3	5.7	6.5
AWDBf6	Banket Quartzite	646	617	603	610		-	-	-	5.6	4.6	5.1
AWABv1	Birimian Volcaniclastic	583	-	-	-		-	-	-	-	-	-
AWDDo1	Dolerite Intrusive	633 ¹	-	-	-		672	629	651	-	-	-
AWDDo2	Dolerite Intrusive	624 ¹	-	-	-		661	624	643	-	-	-
AWDDo3	Dolerite Intrusive	-	-	-	-		699	660	680	-	-	-
AWDDo4	Dolerite Intrusive	562 ¹	-	-	-		728	638	683	-	-	-

Calibrations after [Henry *et al.* \(2005\)](#) (a), [Holdaway \(2000\)](#) (b), [Holland & Blundy \(1994\)](#) (c), [Holland & Powell \(1985\)](#) (d).

¹Outside lithological range of the calibration.

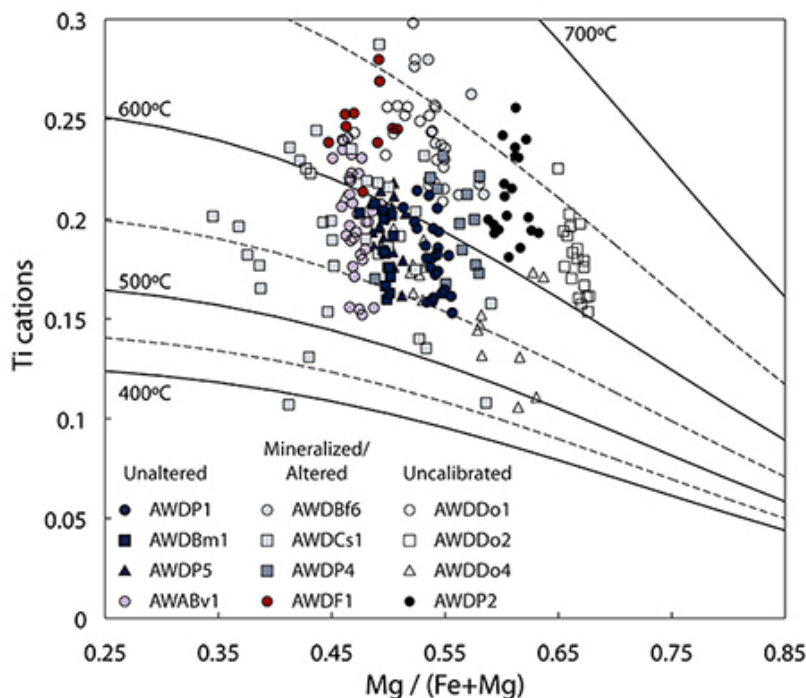


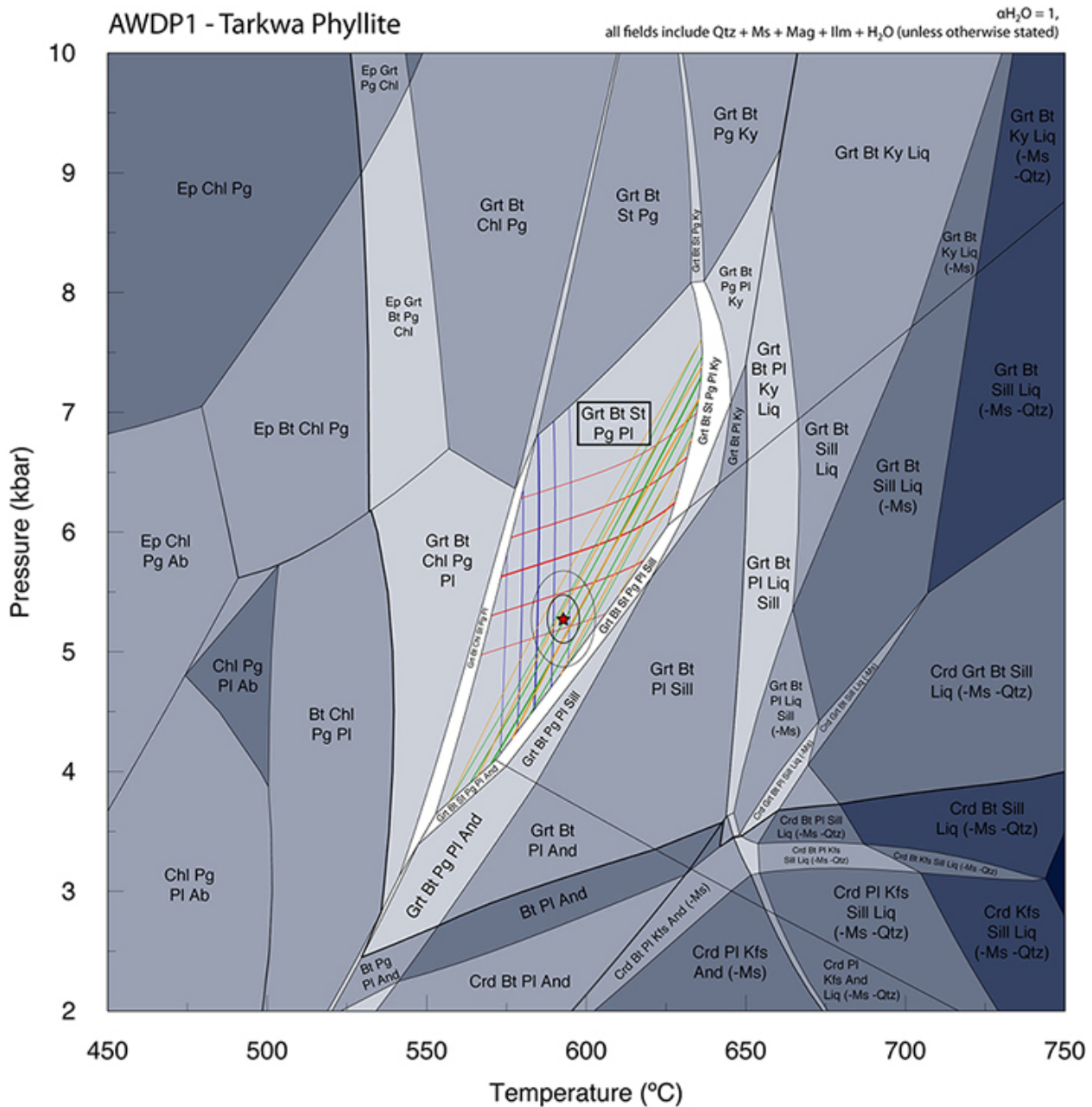
Figure 4.1: Compositional data for biotites from a range of lithologies at Damang plotted with isotherms based on the calibration of [Henry *et al.* \(2005\)](#). Unaltered sedimentary lithologies show consistent temperatures between 550–600°C. Mineralized samples generally display higher Ti contents, although the resulting higher temperatures are not considered real. Dolerite intrusives and sample AWDP2 are outside of the lithological range of the calibration and are shown for reference only.

4.3.2 Pseudosection Results

Tarkwa Phyllite Results

The Tarkwa Phyllite provides the best opportunity for constraining pressure-temperature conditions of regional peak metamorphism. The well developed amphibolite facies assemblage (Qtz-Pl-Ms-Pg-Chl-Bt-Grt-St-Ilm-Mag) comprises a large number of phases that provides a similarly large number of independent variables, which in turn can be used to reliably constrain P-T conditions. Full P-T pseudosections have been calculated for two samples of typical, unmineralized Tarkwa Phyllite, AWDP1 and AWDBm1 (figures 4.2 and 4.3 respectively). Although AWDP1 and AWDBm1 are the same lithology and contain the same overall mineral assemblage, individual mineral modes and compositions are sufficiently different that they still provide good independent tests for calculating P-T conditions.

Understandably, the calculated pseudosections have a number of features in common. Both are built around a ‘backbone’ of a typical prograde Barrovian sequence of minerals along a moderate P-T gradient. The Tarkwa Phyllite is particularly Na-rich compared to the average metapelite of [Shaw \(1956\)](#) and paragonite occurs over large proportions of the diagrams before disappearing above around 600°C (at 5 kbar). Simple reactions that eliminate



Bulk Composition (mol %):

SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O	TiO ₂	MnO	O
62.39	16.00	0.15	3.76	10.04	2.99	2.15	0.48	0.20	1.85

Figure 4.2: A P-T pseudosection for sample AWDP1. The estimated peak assemblage field (boxed) is Grt-Bt-St-Ms-Pg-Pl-Ilm-Mag-Qtz. Observed compositional and modal isopleths (blue, Grt_{Prp} = 0.116; red, Grt_{Sps} = 0.152; green, Grt_{Grs} = 0.025; yellow, St mode = 5.5%) are shown with 1 and 2σ uncertainty limits. The red star represents likely peak conditions, based on the intersection of isopleths, at 590°C and 5.25 kbar, shown with 1 and 2σ error ellipses.

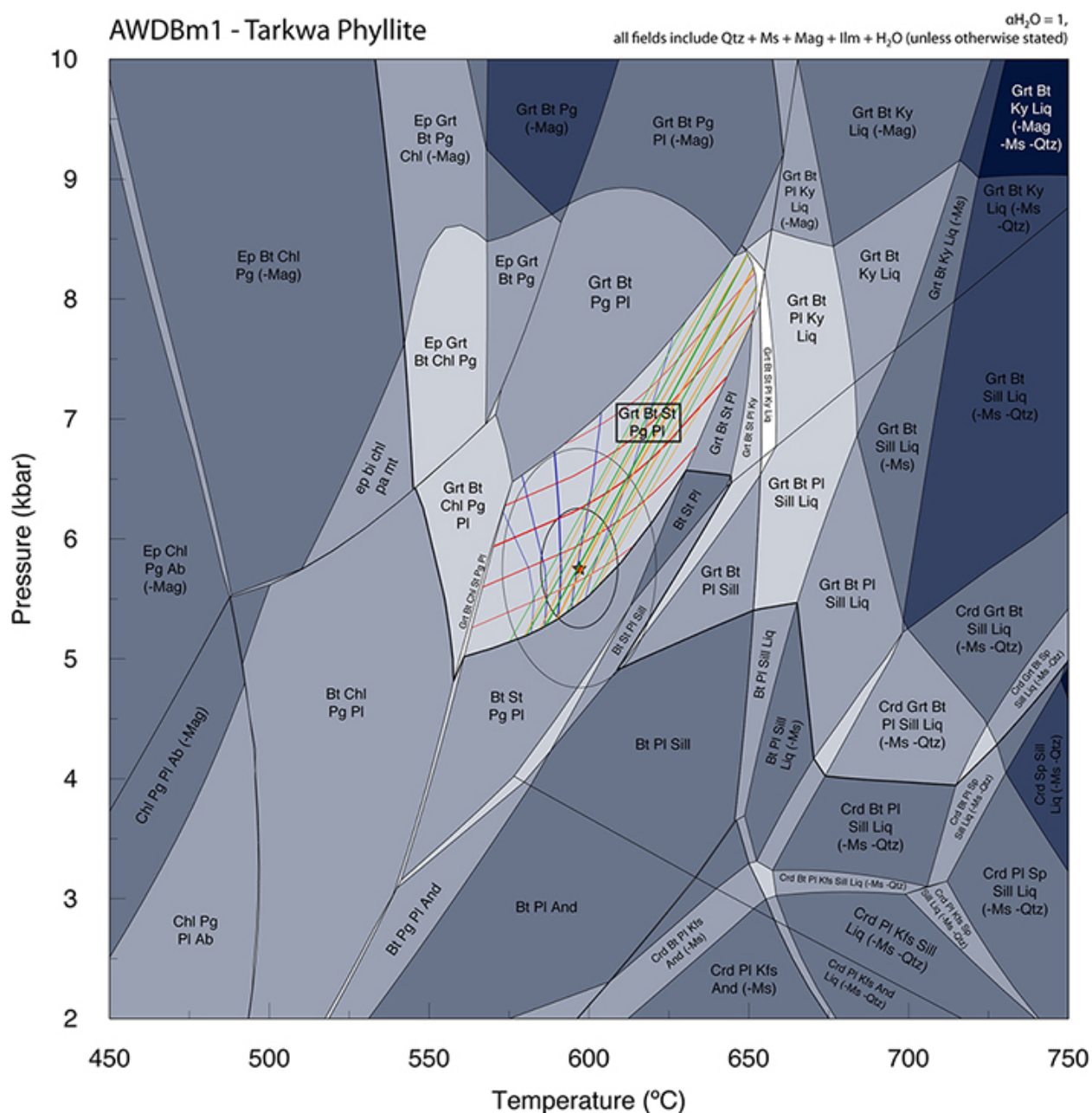


Figure 4.3: A P-T pseudosection for sample AWDBm1. The estimated peak assemblage field (boxed) is Grt-Bt-St-Ms-Pg-Pl-Ilm-Mag-Qtz. Observed compositional and modal isopleths (blue, $Grt_{Prp} = 0.0997$; red, Grt mode = 1%; green, $Grt_{Grs} = 0.0328$; yellow, St mode = 3%) are shown with 1 and 2 σ uncertainty limits. The red star represents likely peak conditions, based on the intersection of isopleths, at 597°C and 5.75 kbar, shown with 1 and 2 σ error ellipses.

paragonite include $\text{Pg} + \text{Qtz} \rightarrow \text{Ab (in Pl)} + \text{Als} + \text{H}_2\text{O}$ in AWDP1 (at low to moderate pressures), and $\text{Pg} + \text{Grt} + \text{Qtz} \rightarrow \text{Ab (in Pl)} + \text{St} + \text{H}_2\text{O}$ in AWDBm1. This latter reaction provides the upper temperature limit of the likely peak assemblage field in sample AWDBm1. Additionally, at low temperatures albite is present, coexisting with plagioclase, before disappearing at around 500°C with the reaction $\text{Ab} + \text{Chl} + \text{Ms} \rightarrow \text{Bt} + \text{Pg} + \text{Qtz} + \text{H}_2\text{O}$. Indeed, paragonite-albite pseudomorphs after albite porphyroblasts are still visible (see section 3.4).

Differences in bulk composition between AWDP1 and AWDBm1 result in subtle variations between the two pseudosections. In particular, the lower Fe^{3+} (O) content of AWDBm1 results in magnetite reacting away at high pressures as the ferric iron is accommodated by biotite (at high temperatures) and epidote (at low temperatures). Similarly, the lower MnO content of AWDBm1 results in garnet appearing later (at higher pressures) than for AWDP1, such that it provides the low pressure limit on the likely peak assemblage field for AWDBm1.

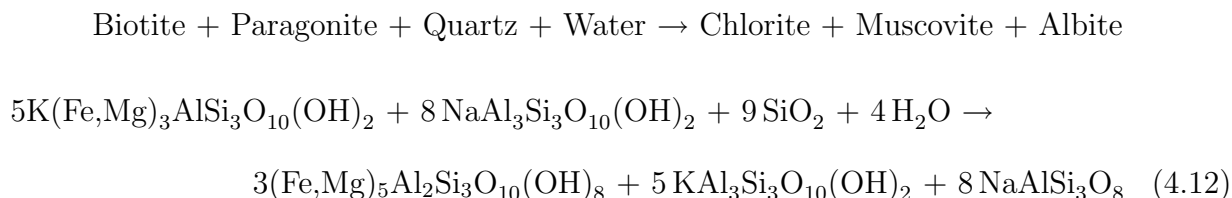
All assemblages on both pseudosections contain H_2O in excess (i.e. at least enough to produce the required hydrous phases) and as such the suprasolidus equilibria are only representative. This is because in reality, an aqueous fluid would be incorporated into any melt present. This is not a concern however, since melt equilibria are not relevant to the rocks at Damang.

Overall, the likely peak assemblage field for both AWDP1 and AWDBm1 is Qtz-Pl-Ms-Pg-Bt-Grt-St-Ilm-Mag. Calculated peak conditions for AWDP1 are 590°C and 5.25 kbar, and 597°C and 5.75 kbar for AWDBm1, which correlate well and show significant statistical overlap.

Retrograde Alteration

It can be seen from figures 4.2 and 4.3 that peak conditions occur in a field that does not include chlorite, even though it is present in the rock. This is because the chlorite appears to be a product of the retrograde alteration of biotite. These minerals have therefore not been used to define peak conditions, which is instead defined by the intersection of composition and mode isopleths of garnet and staurolite. These are refractory phases that are not involved

in the retrograde reaction. Comparisons of all calculated and observed phase compositions and modal proportions are shown in figure 4.4. Detailed inspection of these comparisons allows for a generalized retrograde reaction to be determined:



Or descriptively, as the biotite is chloritized, K is taken up by white mica, converting paragonite to muscovite, which in turn releases Na that produces albite. This reaction is illustrated through a $(\text{Al}_2\text{O}_3\text{-}3\text{K}_2\text{O-Na}_2\text{O}) - 2\text{Na}_2\text{O} - (\text{Fe,Mg})\text{O}$ projection in figure 4.5. This reaction is the reverse of the biotite-producing and albite-consuming reaction described above. The difference with this retrograde reaction however, is that albite is not produced as a separate phase but as an albitisation of plagioclase. Importantly, it is possible to account for this retrograde reaction in a system that is open only to H_2O and does not necessarily require the metasomatic exchange of other components.

Dolerite Results

In addition to the Tarkwa Phyllite samples, an unaltered dolerite (sample AWDDo3) has been modelled. Although the observed mineral assemblage ($\text{Hbl}+\text{Pl}+\text{Ep}+\text{Chl}+\text{Rt}+\text{Qtz}$) contains relatively few phases and is consequently unlikely to constrain conditions as precisely as the Tarkwa Phyllite, it is nonetheless useful to investigate a different lithology in order to confirm the results. The calculated pseudosection is shown in figure 4.6.

As is typical for amphibolite facies metabasites, assemblages above around 500°C are dominated by $\text{Hbl} + \text{Pl}$ with various minor and accessory phases, while at lower temperatures, Act , Chl and Ep are predominant. It is the distribution of calculated accessory phases that aids with constraining likely metamorphic conditions. In particular, Ti -phases show a progression at high temperatures from Ilm at low pressure to Rt at high pressure, with Sph also occurring intermittently towards higher pressures. Quartz occurs as a trace phase at

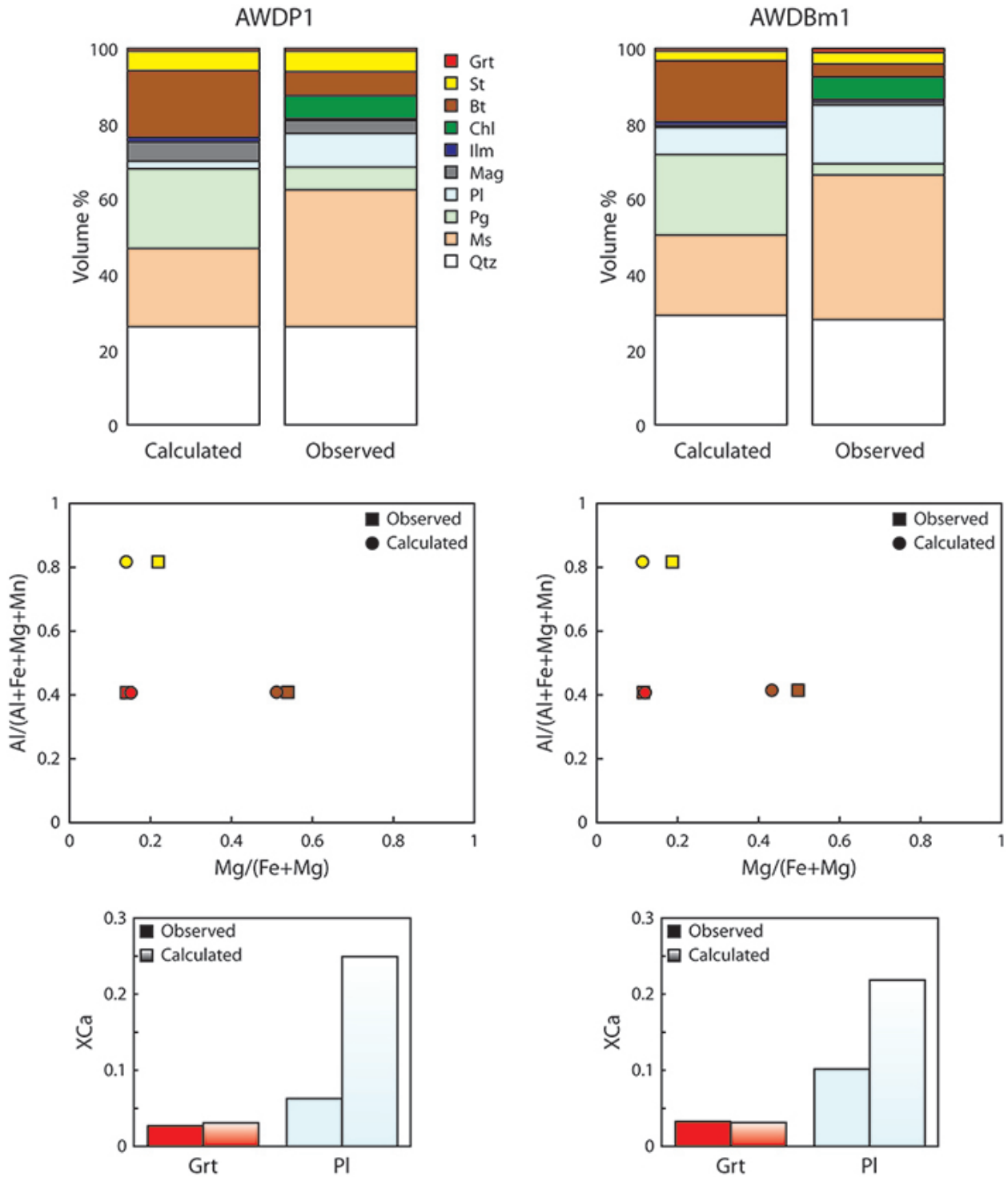


Figure 4.4: Comparisons of observed and calculated modes and phase compositions for AWDP1 (left) and AWDBm1 (right). Discrepancies between observed and modelled phase proportions are attributable to retrograde alteration. In the lower panels, XCa for garnet = $\text{Ca}/(\text{Ca}+\text{Fe}+\text{Mg}+\text{Mn})$, XCa for plagioclase = $\text{Ca}/(\text{Ca}+\text{Na}+\text{K})$.

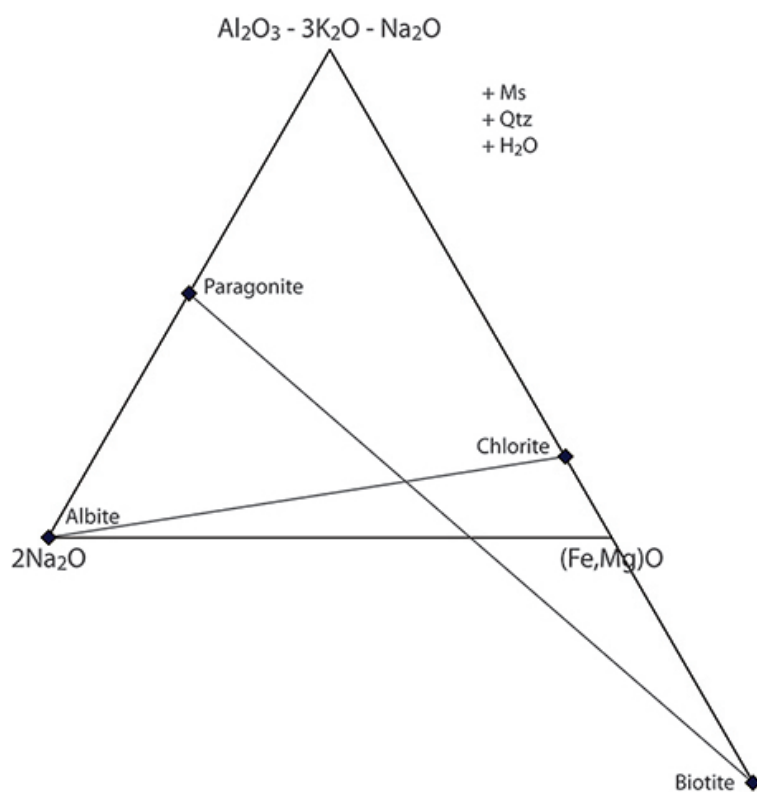
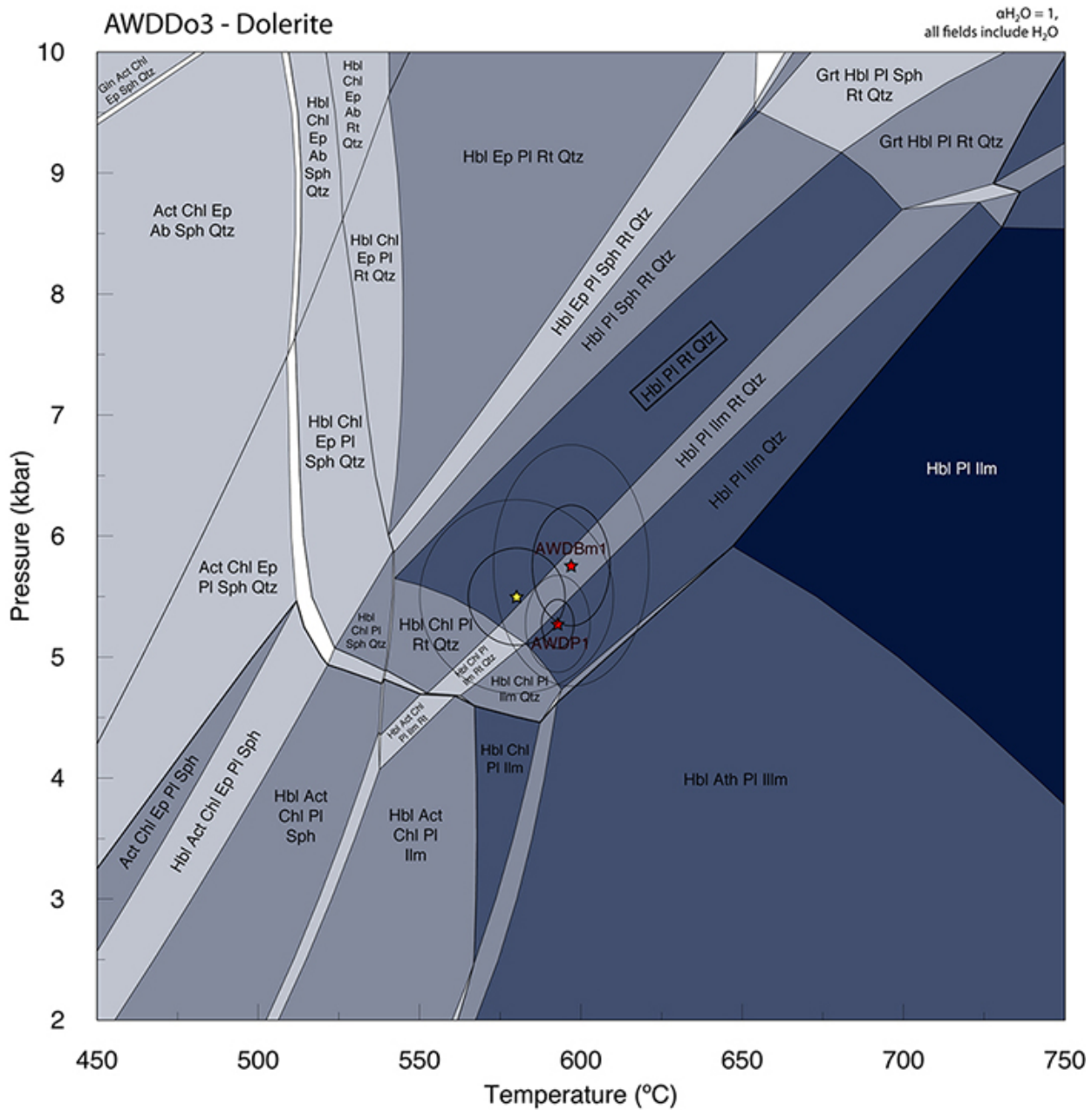


Figure 4.5: $(\text{Al}_2\text{O}_3 - 3\text{K}_2\text{O} - \text{Na}_2\text{O}) - 2\text{Na}_2\text{O} - (\text{Fe,Mg})\text{O}$ projection from muscovite (+ quartz + water) illustrating the Tarkwa Phyllite retrograde alteration reaction $\text{Bt} + \text{Pg} + \text{Qtz} + \text{H}_2\text{O} \rightarrow \text{Chl} + \text{Ms} + \text{Ab}$.

high pressures but is absent at lower pressures as hornblende becomes more tschermakitic through the cation exchange vector $\text{Mg}_1\text{Si}_1\text{Al}_{-2}$.

The calculated assemblage field is $\text{Hbl} + \text{Pl} + \text{Rt} + \text{Qtz}$ and the observed epidote is interpreted to be a relic from an earlier part of the prograde history since it only ever occurs as tiny inclusions within hornblende. Similarly, the observed chlorite is likely a retrograde alteration product. As expected, phase compositions and modal proportions do not change significantly across the field and place only weak constraints on peak conditions. However, typical calculated uncertainties, which are shown on figure 4.6, would allow for peak conditions to lie within a large portion of the field while maintaining a statistically significant correlation (i.e. overlapping 2σ error ellipses) with the better constrained Tarkwa Phyllite samples.

Comparisons between observed and calculated phase compositions and modes are shown in figure 4.7. Calculated hornblende compositions are significantly more Al-rich than observed compositions. This primarily relates to a larger calculated pargasite component ($\text{Na}_1\text{Al}_3\text{Mg}_{-1}\text{Si}_{-2}$, see section 3.7 and table 3.1) and is balanced in the model by a more calcic plagioclase feldspar. This Al discrepancy may be compounded by the presence of the



Bulk Composition (mol %):
 SiO₂ Al₂O₃ CaO MgO FeO Na₂O TiO₂ O
 52.47 13.32 13.46 11.08 6.11 2.94 0.58 0.03

Figure 4.6: A P-T pseudosection for sample AWDDo3. The estimated peak assemblage field (boxed) is Hbl-Pi-Rt-Qtz. Likely peak conditions for samples AWDP1 and AWDBm1 are shown by the red stars, as per figures 4.2 and 4.3. The yellow star and associated 1 and 2 σ error ellipses represent typical uncertainty levels on P-T determination within the peak assemblage field. Peak conditions could lie anywhere within the lower parts of the field and still show significant statistical similarity to samples AWDP1 and AWDBm1.

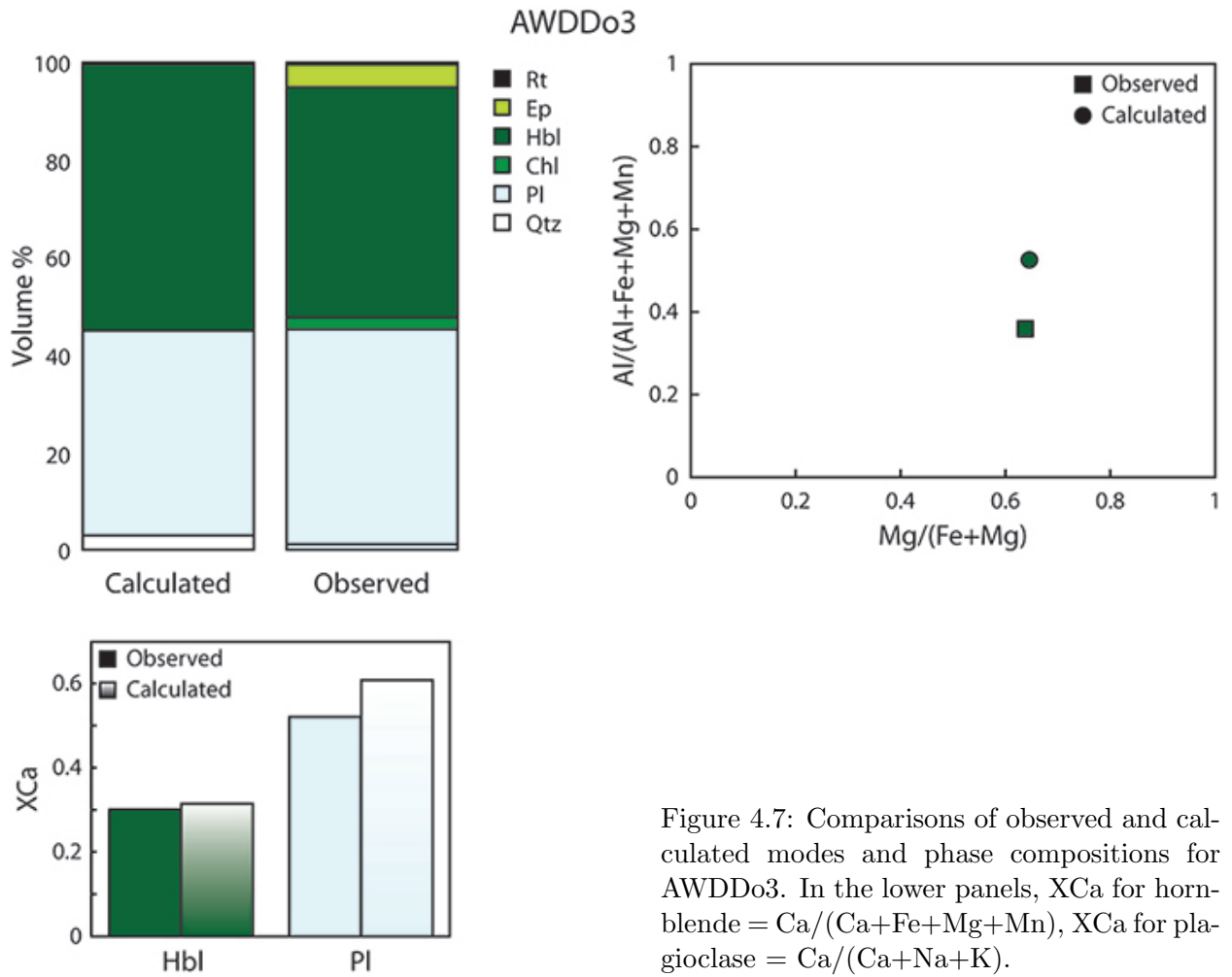


Figure 4.7: Comparisons of observed and calculated modes and phase compositions for AWDDo3. In the lower panels, XCa for hornblende = $\text{Ca}/(\text{Ca}+\text{Fe}+\text{Mg}+\text{Mn})$, XCa for plagioclase = $\text{Ca}/(\text{Ca}+\text{Na}+\text{K})$.

relic epidote and retrogressive chlorite mentioned above.

Summary of Pseudosection Results

Overall, these models are all suggesting that regional metamorphism peaked at around 595°C and 5.5 kbar, which is indeed significantly higher than the commonly described greenschist facies metamorphism of most of the Ashanti belt (Leube *et al.*, 1990; Taylor *et al.*, 1992; Hirdes *et al.*, 1992; Milési *et al.*, 1992; Eisenlohr & Hirdes, 1992; Ledru *et al.*, 1994; Mumin & Fleet, 1995). Other localised occurrences of amphibolite facies metamorphism have been described (John *et al.*, 1999; Galipp *et al.*, 2003) but these are often associated with the aureoles of large granitoids (Taylor *et al.*, 1992; Boher *et al.*, 1992), which are not evident at Damang.

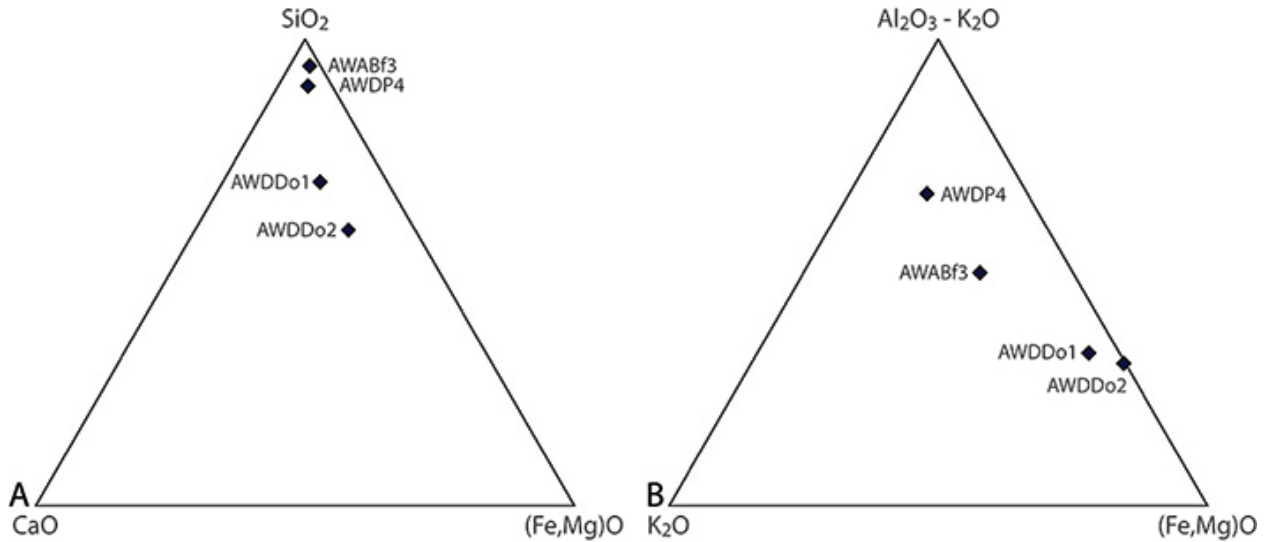


Figure 4.8: Bulk compositions of hydrothermally altered rocks modelled with THERMOCALC. Data as per table 4.2. Ternary SiO_2 – CaO – $(\text{Fe,Mg})\text{O}$ diagram (A) and ternary AKF (Al_2O_3 – K_2O)– K_2O – $(\text{Fe,Mg})\text{O}$ diagram (B) showing the distinction between silica- and mica (K, Al)-rich sediments and ferromagnesian dolerites.

4.4 Metasomatism Associated with Gold Mineralization

To investigate the temperature, pressure and fluid composition conditions that accompanied gold mineralization, four hydrothermally altered samples (AWDDo1, AWABf3, AWDDo2, AWDP4) have been extensively modelled. For each sample, two pseudosections are presented; a T- $X(\text{CO}_2)$ diagram calculated at a specific pressure and a P-T pseudosection drawn for a specific fluid $X(\text{CO}_2)$. This allows for the full range of stability fields to be assessed and, subsequently, the stability ranges of the observed mineralization-associated mineral assemblages to be explored and compared in relation to all three variables. The bulk compositions of the four modelled samples are plotted in figure 4.8. On the SiO_2 – CaO – $(\text{Fe,Mg})\text{O}$ diagram (Fig. 4.8A) the samples are clearly differentiated between the SiO_2 -rich sediments and the more ferromagnesian, mafic dolerite intrusions. Whereas on the AKF diagram (Fig. 4.8B) the samples define an array between a ferromagnesian-rich dolerite (sample AWDDo2) and a mica-dominated (K-Al-rich) sediment (sample AWDP4). It is assumed that the mineralizing H_2O – CO_2 fluid would exert an external control on the stability ranges of equilibrium mineral assemblages such that the observed assemblages in a range of compositionally different rocks should intersect in P-T- $X(\text{CO}_2)$ space.

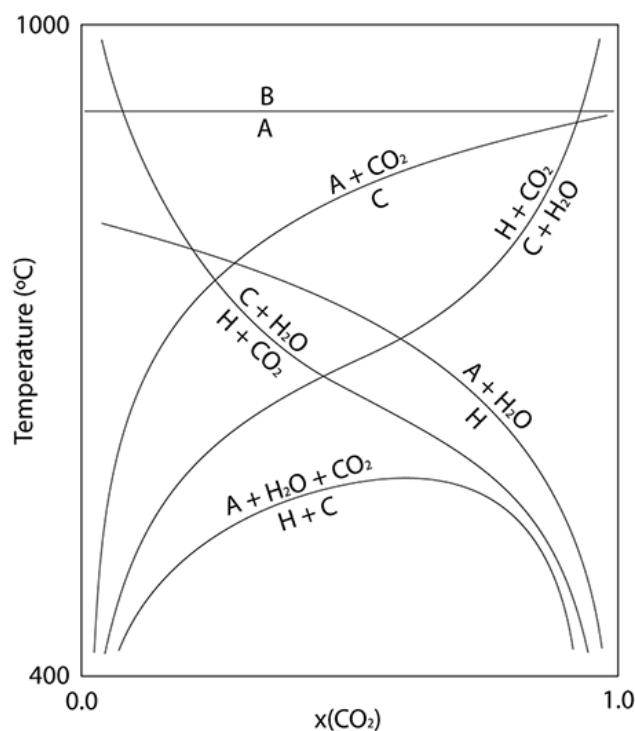


Figure 4.9: A schematic depiction of the geometries of different reaction types in T-X(CO₂) pseudosections. A and B are volatile-free phases, C is a carbonated phase and H is a hydrous phase. After Spear (1995).

The calculated pseudosections are presented in figures 4.10 and 4.11 (AWDDo1), 4.14 and 4.15 (AWABf3), 4.19 and 4.20 (AWDDo2), and 4.23 and 4.24 (AWDP4). Despite the wide range of lithologies, these pseudosections share a number of key characteristics. Overall similarities in the shape and geometries of these pseudosections are to be expected since decarbonation and/or dehydration reactions have specific geometries in T-X(CO₂) space (Fig. 4.9; Spear, 1995). Beyond this however, there are mineralogical similarities and/or systematic variations. These general features will be discussed first before the specific features of individual diagrams are considered.

General Features of Calculated Equilibria

With increasing X(CO₂) and/or increasing pressure and/or decreasing temperature, carbonate minerals become more dominant, with Fe and Mg transitioning from silicates (biotite, chlorite, etc.), coexisting with calcite, into dolomite (ankerite) and eventually siderite. The disappearance of siderite up-temperature typically occurs around 470°C, at 4 kbar and X(CO₂) around 0.6-0.7, which varies little as a function of bulk composition, and corresponds to the simplified reaction $\text{Sid} + \text{Qtz} + \text{H}_2\text{O} \rightarrow \text{Fe,Mg-silicate (Bt, Chl, etc.)} + \text{CO}_2$. By contrast, the breakdown of dolomite is much more sensitive to bulk Fe,Mg-content. Sam-

ples AWABf3 and AWDP4 both lose dolomite at a broadly similar temperature to siderite (480°C, at 4 kbar and $X(\text{CO}_2)$ around 0.6-0.7, for sample AWABf3), while the altered dolerites (AWDDo1 and AWDDo2) lose dolomite at around 525°C. This dolomite-out reaction is either accompanied by the formation of calcite (at low $X(\text{CO}_2)$ in samples AWDDo1, AWABf3 and AWDDo2) according to the simplified reaction $\text{Dol} + \text{Qtz} + \text{H}_2\text{O} \rightarrow \text{Cal} + \text{Fe,Mg-silicate (Bt, Chl, etc.)} + \text{CO}_2$, or represents total decarbonation of the rock (at higher $X(\text{CO}_2)$), corresponding to the reaction $\text{Dol} + \text{Qtz} + \text{H}_2\text{O} \rightarrow \text{Ca,Fe,Mg-silicate (Hbl, Di, etc.)} + \text{CO}_2$. Decarbonation of calcite towards low $X(\text{CO}_2)$ follows a similar reaction, $\text{Cal} + \text{Qtz} + \text{H}_2\text{O} \rightarrow \text{Ca-silicate (Act, Hbl, Pl, etc.)} + \text{CO}_2$. The exact temperature of this reaction appears to depend upon which Ca-silicate is produced, as can be seen by comparing figures 4.10, 4.14 and 4.19. Sample AWDP4 does not form calcite under any P-T- $X(\text{CO}_2)$ conditions, which is interpreted to be due to the high Al-content favouring plagioclase stability.

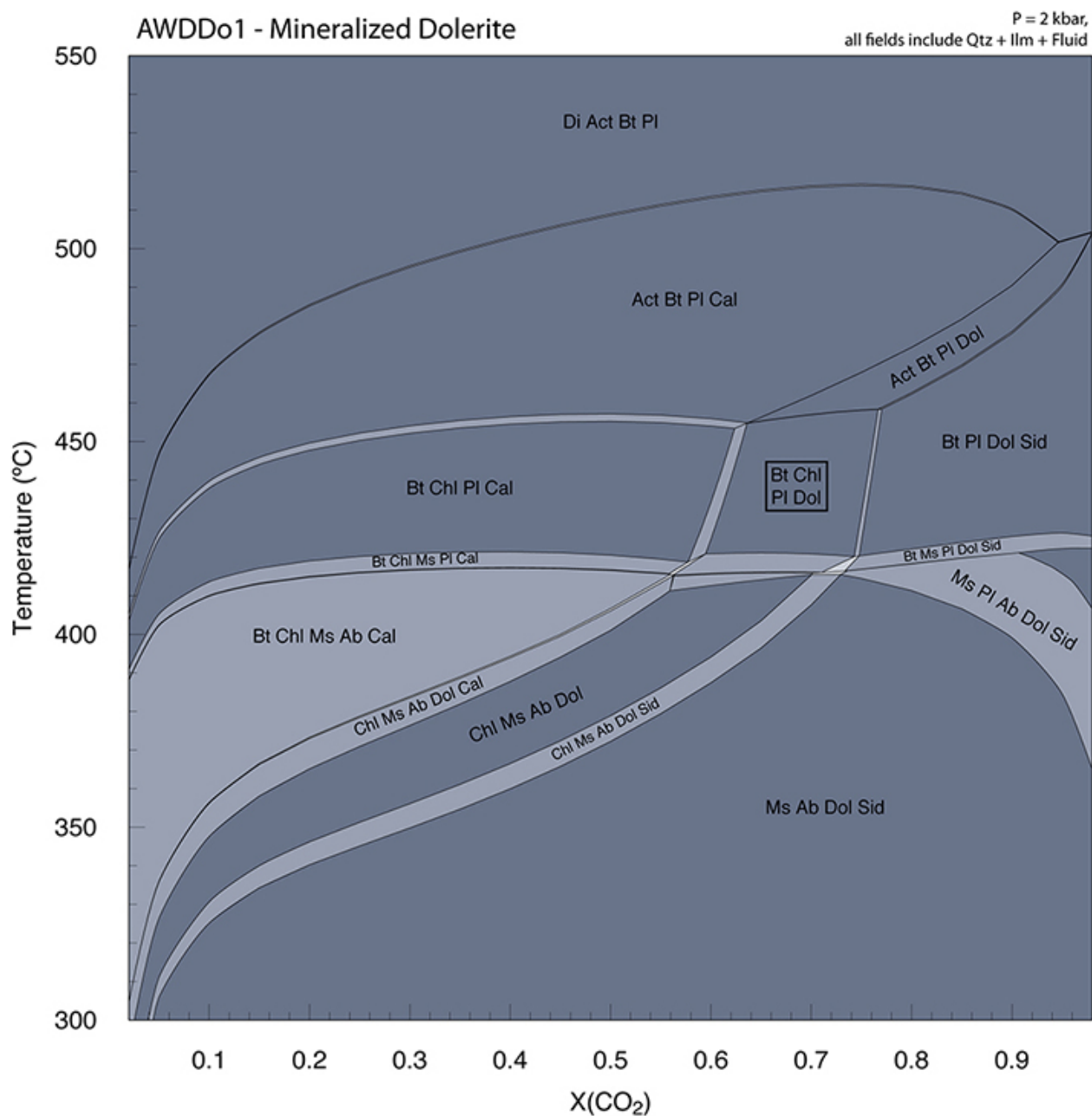
For all samples, the main K-bearing phase towards lower temperature is white mica (muscovite or paragonite), with biotite present at higher temperatures. Sample AWABf3 does, however, contain muscovite up to high temperatures, coexisting with biotite, until it reacts to andalusite. With the exception of sample AWABf3, the upper temperature limit to white mica occurs at a uniform temperature (around 440°C at 3 kbar) across a wide range of $X(\text{CO}_2)$ values, and is approximated by the reaction $\text{Ms} + \text{Chl} + \text{Cal} + \text{Qtz} \rightarrow \text{Bt} + \text{Pl} + \text{H}_2\text{O} + \text{CO}_2$.

At low temperatures, across a wide range of pressures and $X(\text{CO}_2)$, all lithologies produce assemblages dominated by dolomite, siderite, white mica and often albite. In contrast, high temperature assemblages are more sensitive to bulk composition. Sedimentary lithologies (AWABf3 and AWDP4) primarily contain biotite and plagioclase, while the dolerites (AWDDo1 and AWDDo2) contain abundant amphibole, and in the case of AWDDo1, diopside.

AWDDo1 - Mineralized Dolerite

Sample AWDDo1 is an altered and mineralized dolerite, characterised by the mineral assemblage Bt-Chl-Dol-Pl-Ilm-Qtz(-Po). Pseudosection results are shown in figures 4.10 and 4.11. The primary K-bearing phase in the observed assemblage is biotite, coexisting with plagioclase feldspar. This provides a minimum temperature constraint ($\sim 420^{\circ}\text{C}$ from figure 4.10) as at lower temperatures the assemblage Ms + Ab is stable. Higher temperature assemblages contain actinolite formed through the consumption of chlorite, which in turn provides a maximum temperature constraint ($\sim 450^{\circ}\text{C}$ from figure 4.10). The stability range of chlorite narrows with increasing $X(\text{CO}_2)$ and pressure as Fe and Mg are partitioned in siderite. The presence of dolomite (ankerite) and quartz is characteristic of greenschist facies assemblages at moderately CO_2 -rich fluid compositions. This association occurs over a relatively narrow band with positive T- $X(\text{CO}_2)$ and P-T slopes, bounded to lower and higher values of both $X(\text{CO}_2)$ and pressure by the stability ranges of calcite and siderite respectively.

The observed assemblage field is therefore reasonably well constrained in both T- $X(\text{CO}_2)$ (Fig. 4.10) and P-T (Fig. 4.11) space. Ultimately, this assemblage field defines a continuous array that trends towards higher T and P with decreasing $X(\text{CO}_2)$, as shown in figures 4.12 and 4.13. However, the high variance (low number of phases, c.f. equation 4.11) of this assemblage results in little change in composition or modal proportion of phases across this array. Precisely constraining conditions with this sample alone is therefore difficult.



Bulk Composition (mol %):

SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O	TiO ₂	O
58.62	9.66	10.63	8.56	6.84	1.44	3.10	1.15	3.3x10 ⁻²

Figure 4.10: A T-X(CO₂) pseudosection for sample AWDDo1, calculated at 2 kbar. The alteration assemblage field (boxed) is Dol-Bt-Chl-Pl-Ilm-Qtz, which is stable at conditions of 415–450°C and X(CO₂) of 0.6–0.75.

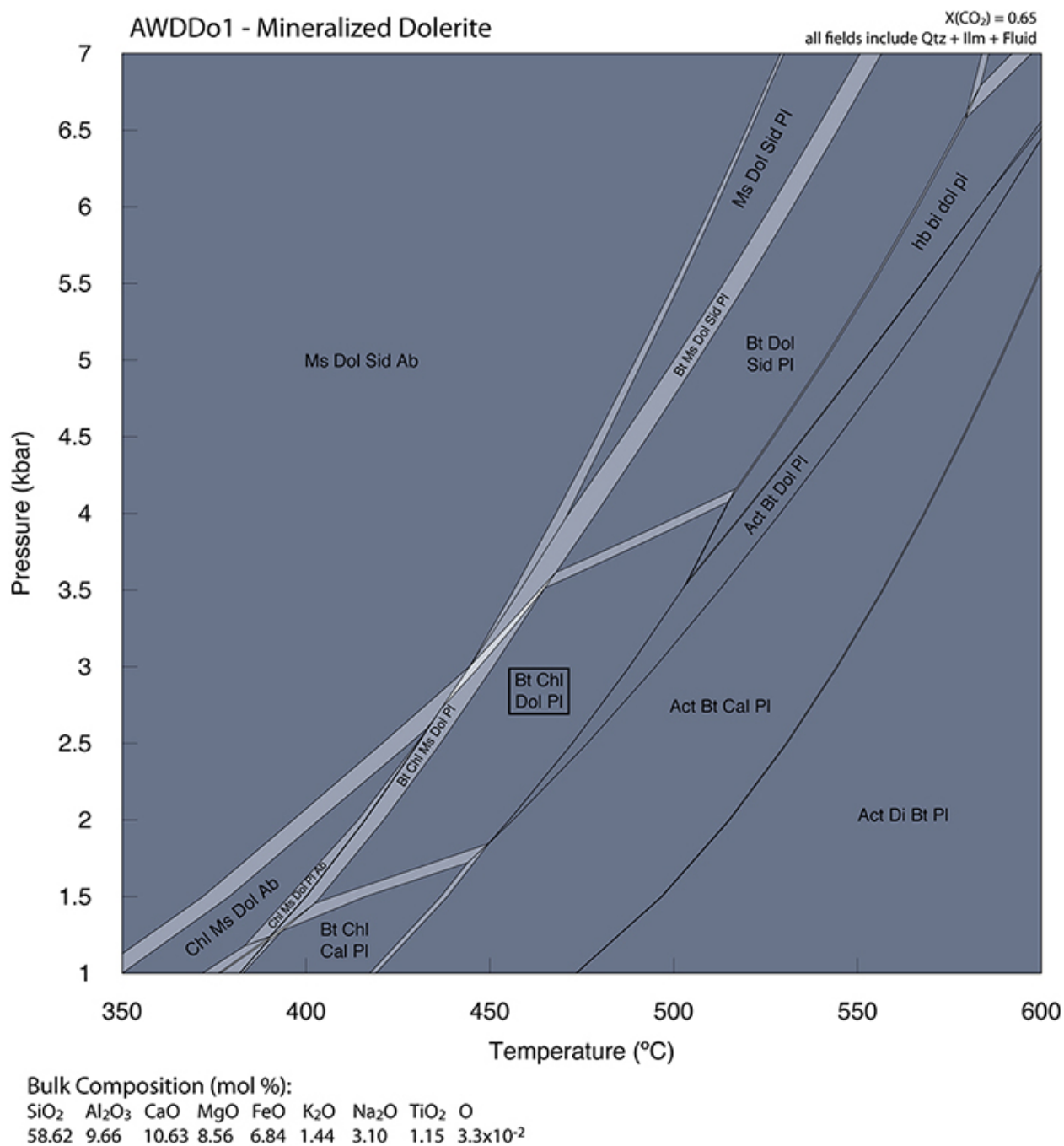


Figure 4.11: A pseudosection for sample AWDDo1, calculated at an $X(\text{CO}_2)$ value of 0.65. The alteration assemblage field (boxed) is Dol-Bt-Chl-Pl-Ilm-Qtz, which is stable at conditions of 400–500°C and 1.5–4 kbar.

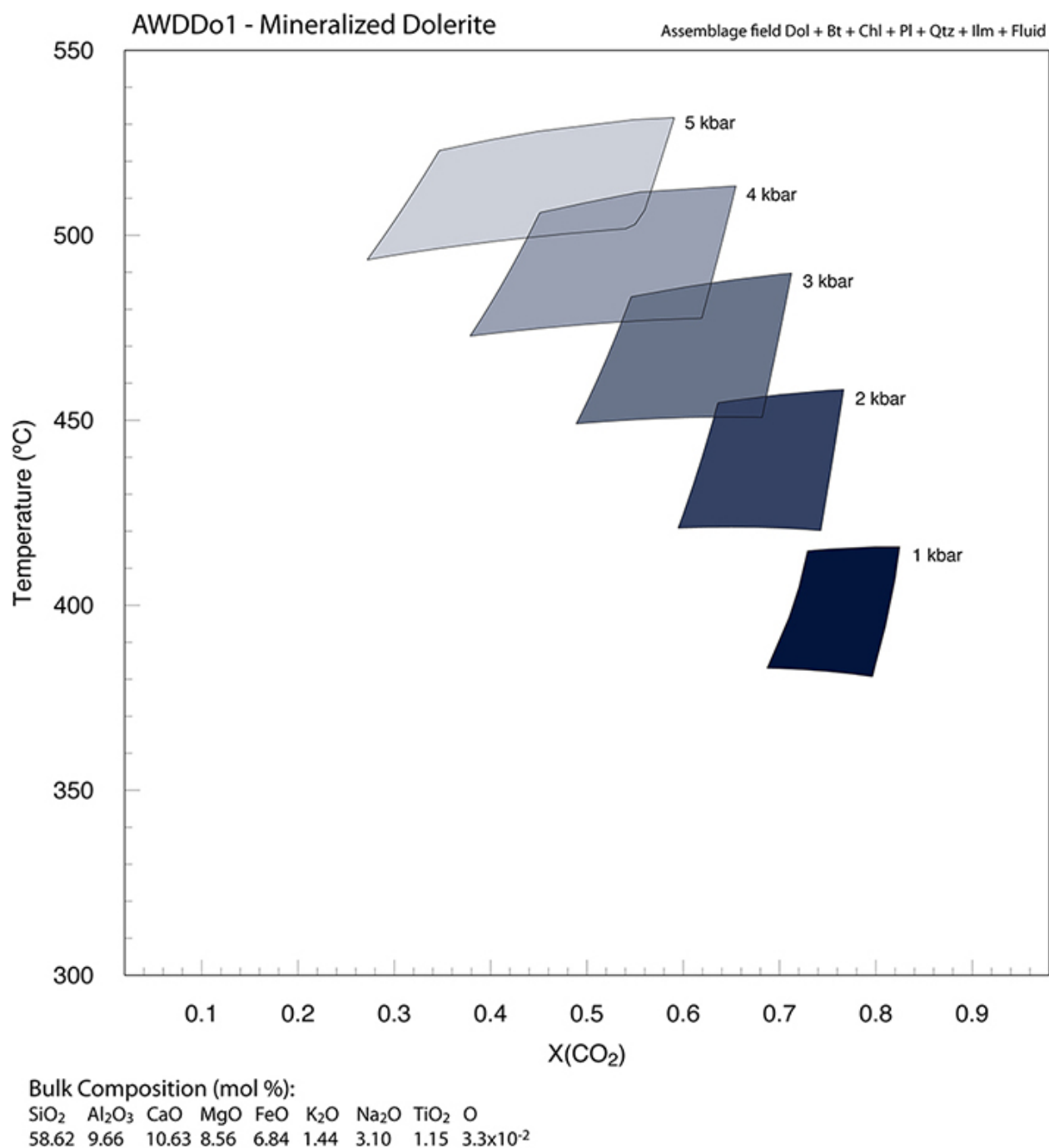


Figure 4.12: A diagram illustrating the change in position of assemblage field Dol-Bt-Chl-Pl-Ilm-Qtz as a function of pressure, in T-X(CO₂) space, for sample AWDDo1. Increasing pressure shifts the calculated field towards higher temperatures and lower X(CO₂) values.

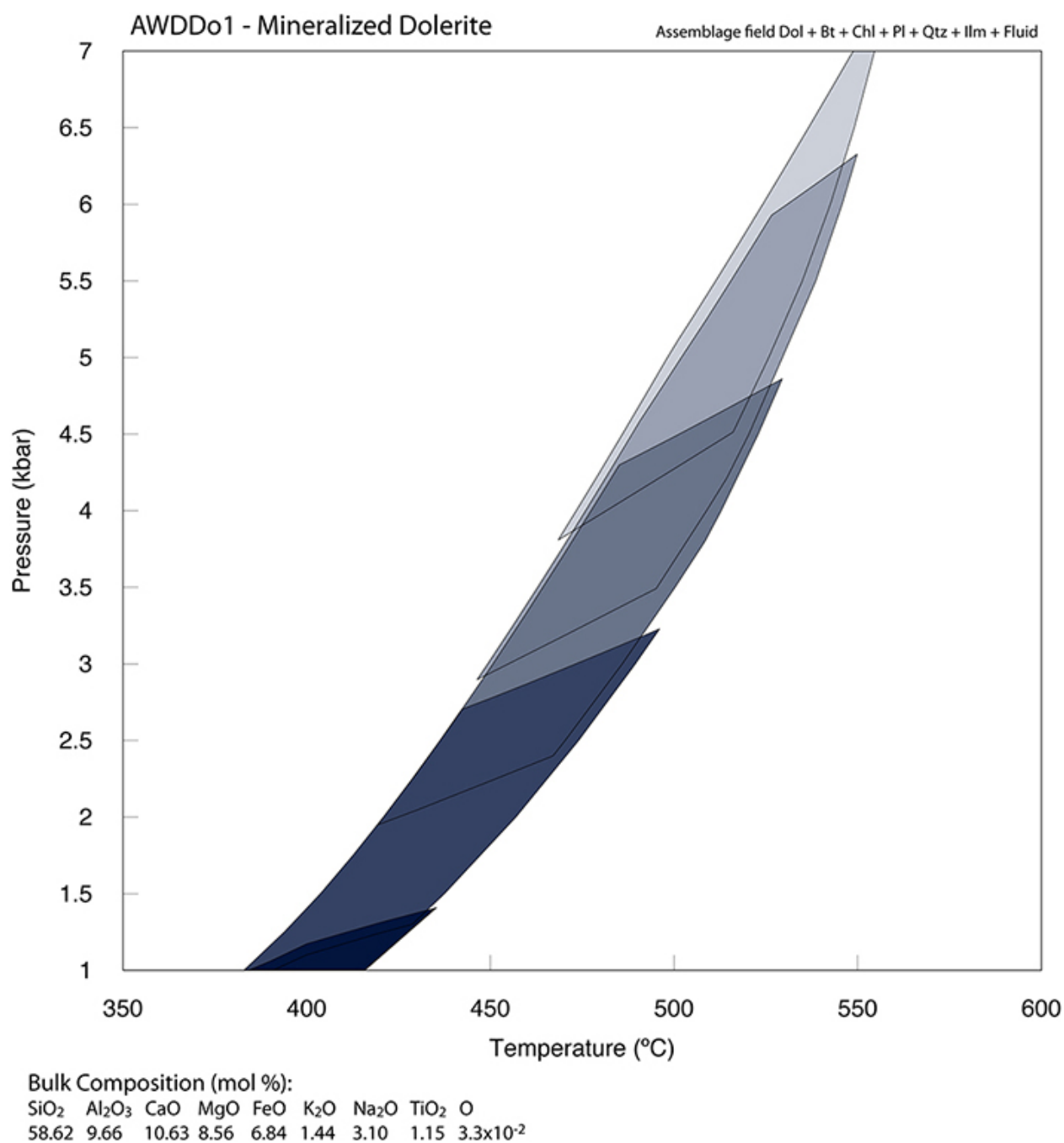
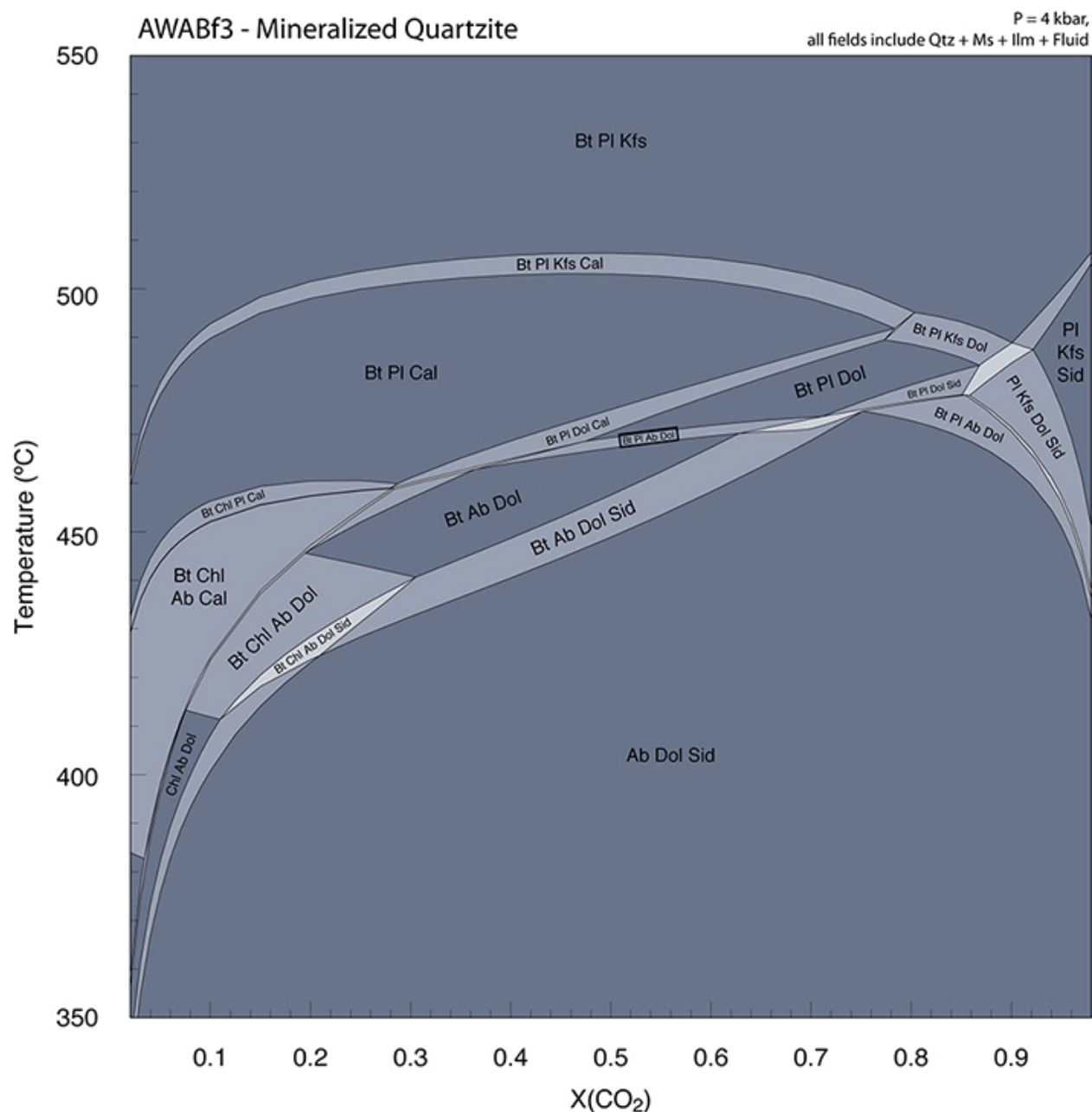


Figure 4.13: A diagram illustrating the change in position of assemblage field Dol-Bt-Chl-Pl-Ilm-Qtz as a function of $X(\text{CO}_2)$, in P-T space, for sample AWDDo1. Increasing $X(\text{CO}_2)$ shifts the calculated field towards lower temperatures and pressures.

AWABf3 - Mineralized Quartzite

Sample AWABf3 is a mineralized Banket Footwall quartzite with the mineral assemblage Chl-Dol-Pl-Ab-Ms-Ilm-Qtz(-Py). Pseudosection results are shown in figures 4.14 and 4.15. The coexistence of albite with a more calcic plagioclase places a very useful temperature constraint on this sample as the stability range of two feldspars, separated by the peristerite gap, defines a very narrow band running sub-horizontally across figure 4.14. Biotite is the predominant ferromagnesian silicate across much of the diagram, with chlorite coexisting at low values of $X(\text{CO}_2)$. Chlorite only occurs in the absence of biotite at very low temperatures and $X(\text{CO}_2)$. The presence of chlorite in the observed assemblage therefore, is inferred to be a retrograde product formed through the chloritization of biotite, with a H_2O -rich fluid, similar to that seen in unmineralized Tarkwa Phyllite (section 4.3.2). The stability range of dolomite (ankerite), in the absence of calcite and siderite, again defines a relatively narrow band with positive T - $X(\text{CO}_2)$ and P - T slopes, bounded to lower and higher values of both $X(\text{CO}_2)$ and pressure by the stability ranges of calcite and siderite respectively.

The most likely assemblage field is therefore Bt-Dol-Pl-Ab-Ms-Ilm-Qtz. This assemblage field is extremely well constrained in both P - T (Fig. 4.14) and T - $X(\text{CO}_2)$ (Fig. 4.15) space. As with sample AWDDo1, this field defines a continuous array that trends towards higher T and P with decreasing $X(\text{CO}_2)$, as shown in figures 4.16 and 4.17. However, the presence of a feldspar peristerite gap places additional constraints on conditions. When two plagioclase feldspars coexist in a rock, as is the case here, the difference in composition (i.e. the size of the miscibility gap) is sensitive to temperature, such that at lower temperatures the compositional gap is larger and it decreases with increasing temperature (Crawford, 1966). This relationship has been calculated for sample AWABf3 (presented in figure 4.18) and the observed plagioclase composition indicates likely temperatures are around 435°C , which correlates with a pressure of approximately 2.5 kbar and a fluid $X(\text{CO}_2)$ of around 0.7.



Bulk Composition (mol %):

SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O	TiO ₂	O
83.94	7.16	1.87	1.67	1.82	1.84	1.30	0.39	5.6x10 ⁻³

Figure 4.14: A T-X(CO₂) pseudosection for sample AWABf3, calculated at 4 kbar. The alteration assemblage field (boxed) is Dol-Bt-Pl-Ms-Pl-Ab-Ilm-Qtz, which is stable at conditions of approximately 465°C and X(CO₂) of 0.35–0.7.

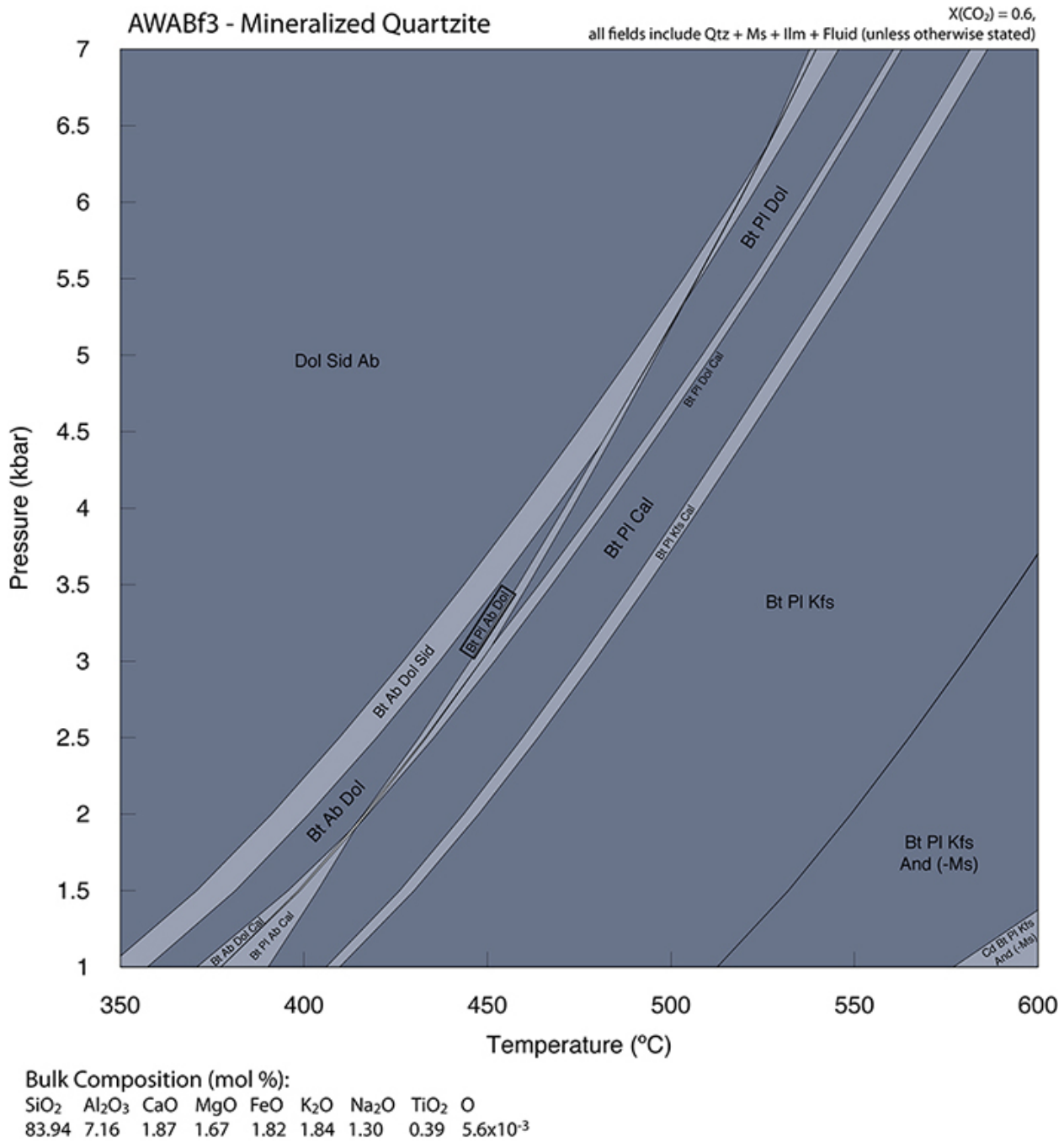


Figure 4.15: A pseudosection for sample AWABf3, calculated at an $X(\text{CO}_2)$ value of 0.6. The alteration assemblage field (boxed) is Dol-Btl-Ms-Pl-Ab-Ilm-Qtz, which is stable at conditions of 420–470°C and 2–5 kbar.

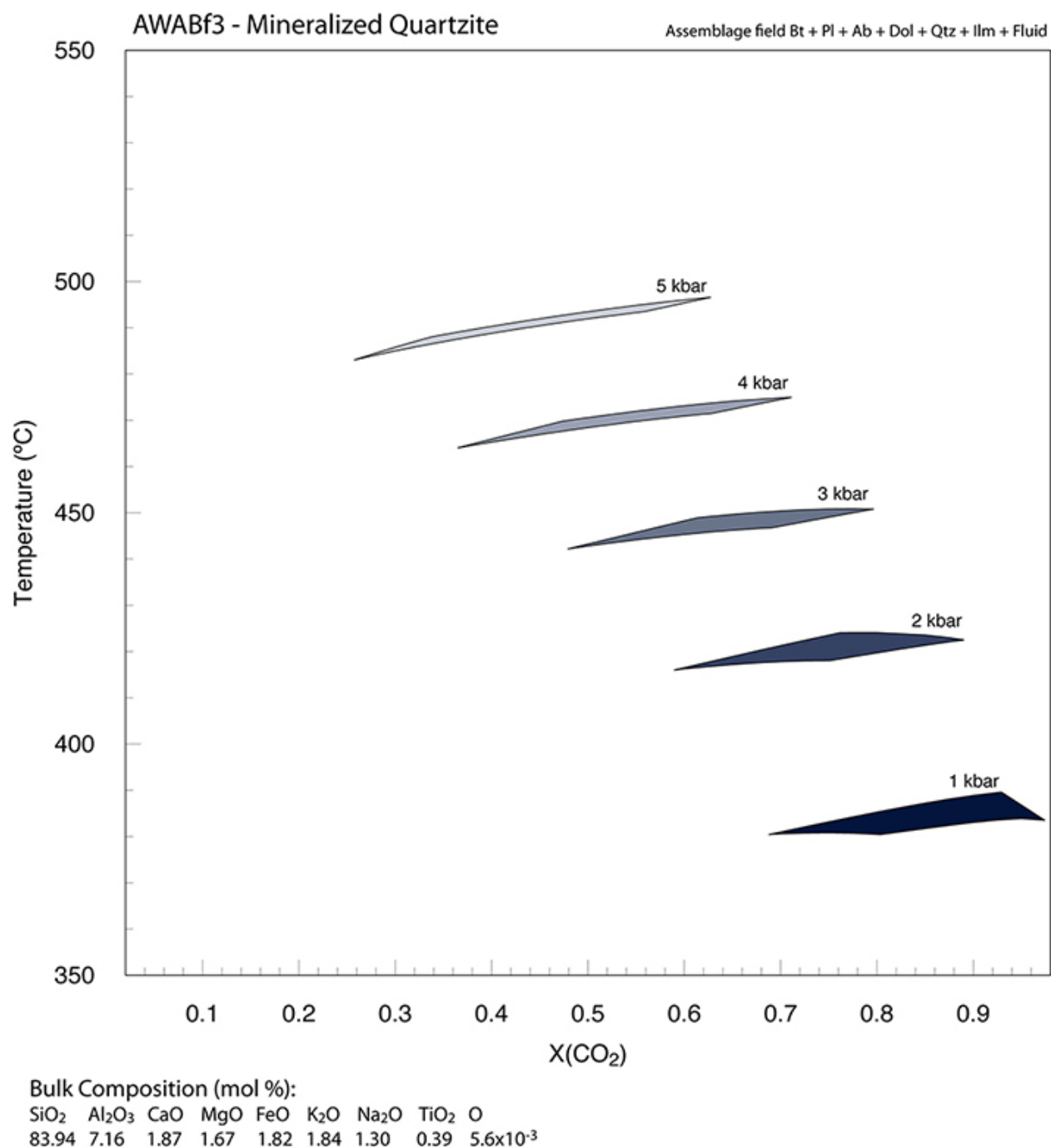


Figure 4.16: A diagram illustrating the change in position of assemblage field Dol-Btl-Ms-Pl-Ab-Ilm-Qtz as a function of pressure, in T-X(CO₂) space, for sample AWABf3. Increasing pressure shifts the calculated field towards higher temperatures and lower X(CO₂) values.

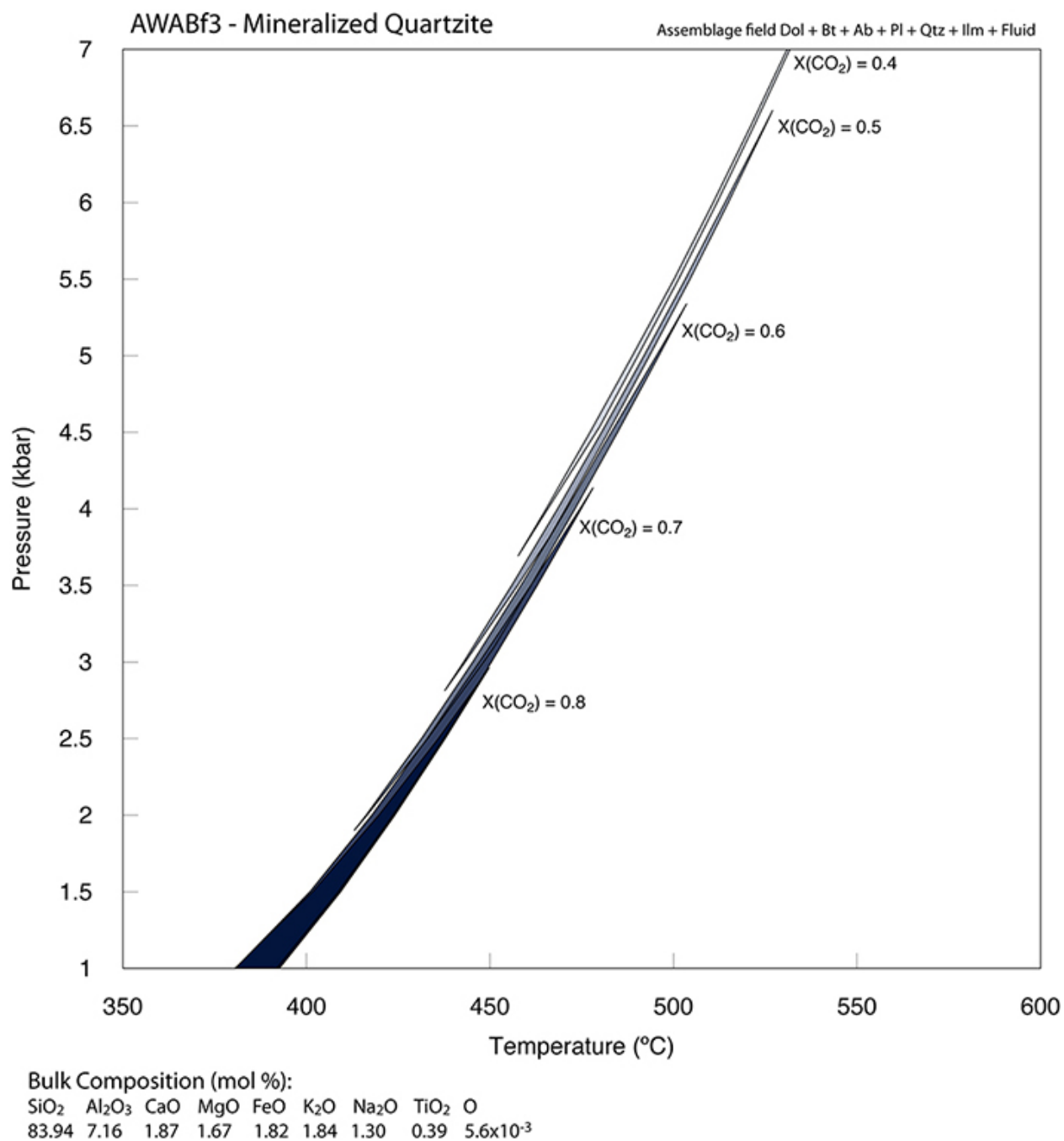


Figure 4.17: A diagram illustrating the change in position of assemblage field Dol-Btl-Ms-Pl-Ab-Ilm-Qtz as a function of $X(\text{CO}_2)$, in P-T space, for sample AWABf3. Increasing $X(\text{CO}_2)$ shifts the calculated field towards lower temperatures and pressures.

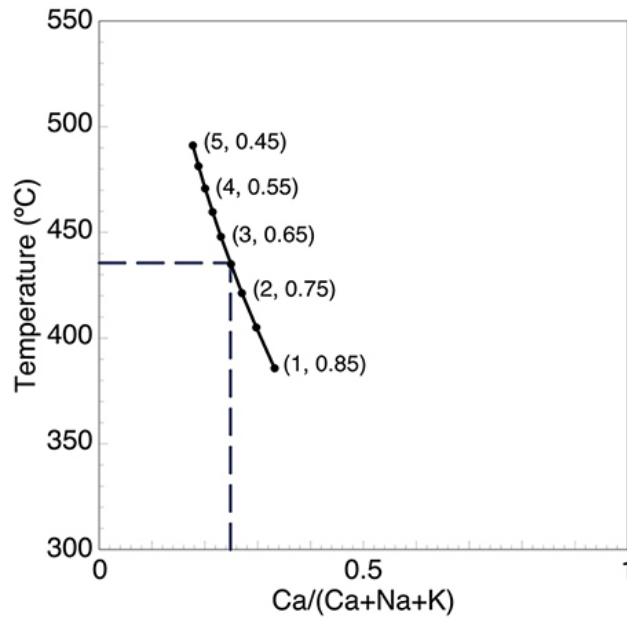


Figure 4.18: A diagram illustrating the change in modelled composition of plagioclase feldspar in assemblage field Ank-Btl-Ms-Pl-Ab-Ilm-Qtz, in sample AWABf3. Points are calculated as a function of temperature, pressure and $X(\text{CO}_2)$ that track with the assemblage field as per figure 4.16. The plotted line represents the Ca-rich limb of the peristerite miscibility gap as determined for sample AWABf3 with THERMOCALC using the [Holland & Powell \(1998\)](#) data set. Observed compositions suggest equilibrium temperatures of approximately 435°C at a pressure of 2.5 kbar and a fluid $X(\text{CO}_2)$ of 0.7.

AWDDo2 - Altered Dolerite

Sample AWDDo2 is an altered, but not mineralized, dolerite dyke from the west wall of the Damang pit, comprising the alteration assemblage Hbl-Cum-Bt-Chl-Dol-Pl-Ilm. Pseudosection results are shown in figures 4.19 and 4.20. A key feature of this assemblage is the coexistence of two amphiboles (hornblende and cummingtonite), which occurs across very high temperature portions of the diagrams. This mineral association does, however, show a temperature minimum of $\sim 475^{\circ}\text{C}$ at $X(\text{CO}_2)$ values of 0.6–0.7, at 3 kbar. At lower $X(\text{CO}_2)$, actinolite is stable in place of cummingtonite and there is a narrow zone of calculated three-amphibole stability. At higher $X(\text{CO}_2)$, Fe and Mg partition into siderite and dolomite (ankerite) in place of amphiboles. The stability range of quartz is restricted to temperatures below $\sim 480^{\circ}\text{C}$, at 3 kbar, as it is consumed in the production of amphiboles. As with previous samples, dolomite (ankerite) occurs in the absence of calcite and siderite in a relatively narrow band with positive T- $X(\text{CO}_2)$ and P-T slopes, bounded to lower and higher values of both $X(\text{CO}_2)$ and pressure by the stability ranges of calcite and siderite respectively.

The large number of phases and corresponding low variance of the observed assemblage, results in a well constrained field in both T- $X(\text{CO}_2)$ (Fig. 4.19) and P-T (Fig. 4.20) space. Once again, this field defines an array through P-T- $X(\text{CO}_2)$ space with a similar trajectory to samples AWDDo1 and AWABf3. However, AWDDo2 is significantly more constrained as the field disappears entirely above 3.6 kbar, and at $X(\text{CO}_2)$ values below 0.66. This is due to a relative change in the upper temperature limits of quartz and chlorite with increasing pressure, such that at lower pressures quartz is exhausted at lower temperatures than chlorite, while the converse is true at higher pressures. As such, above 3.6 kbar all Chl + Hbl + Cum assemblages also contain Qtz. The calculated assemblage field, however, indicates temperatures up to 50°C greater than those for AWABf3 (at 2.5 kbar). There are a number of possible explanations for this, which will be discussed in section 4.4.1.

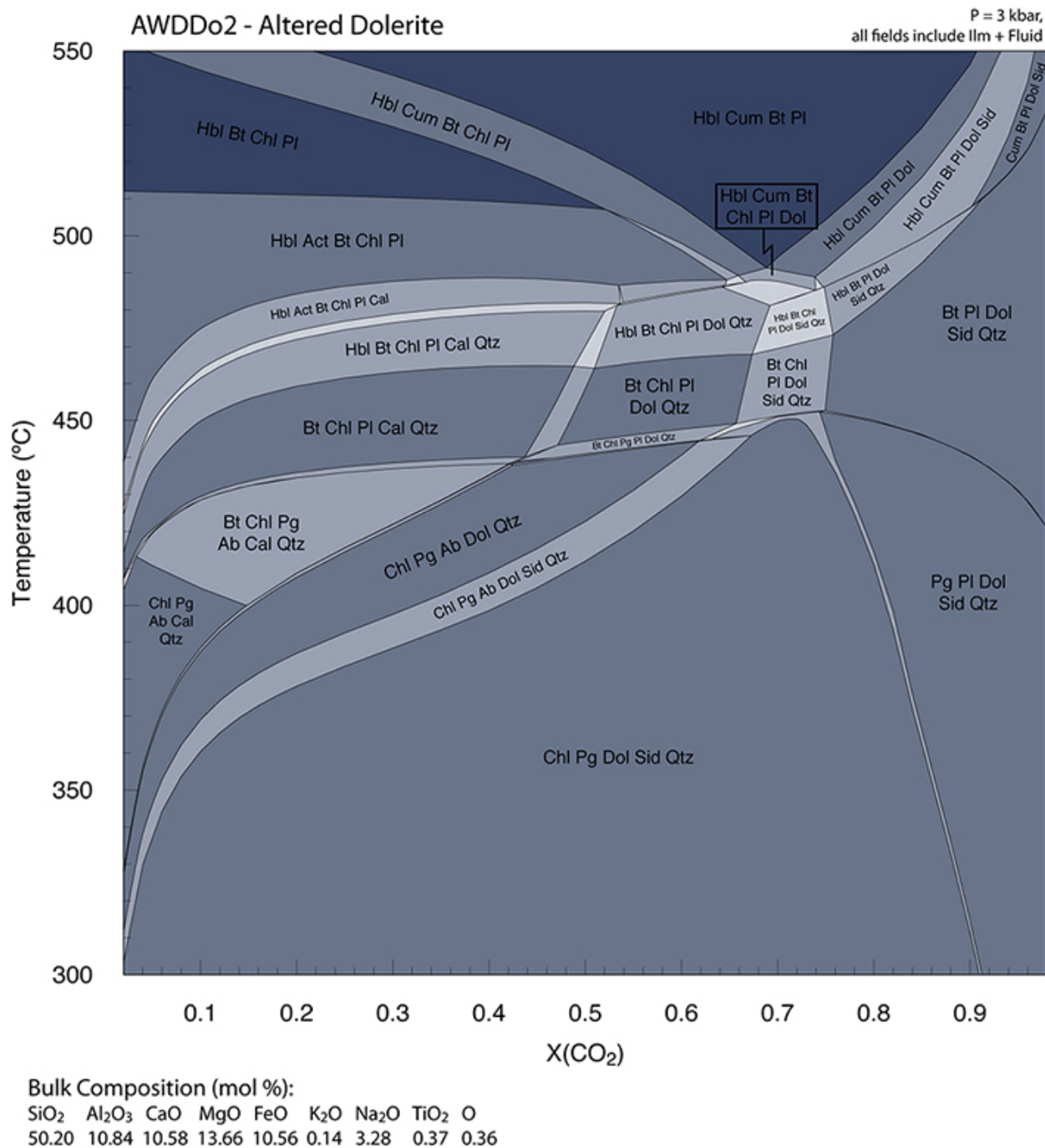


Figure 4.19: A T-X(CO₂) pseudosection for sample AWDDo2, calculated at 3 kbar. The alteration assemblage field (boxed) is Hbl-Cum-Bt-Chl-Dol-Pl-Ilm, which is stable at conditions of 485–490°C and X(CO₂) of 0.68–0.74.

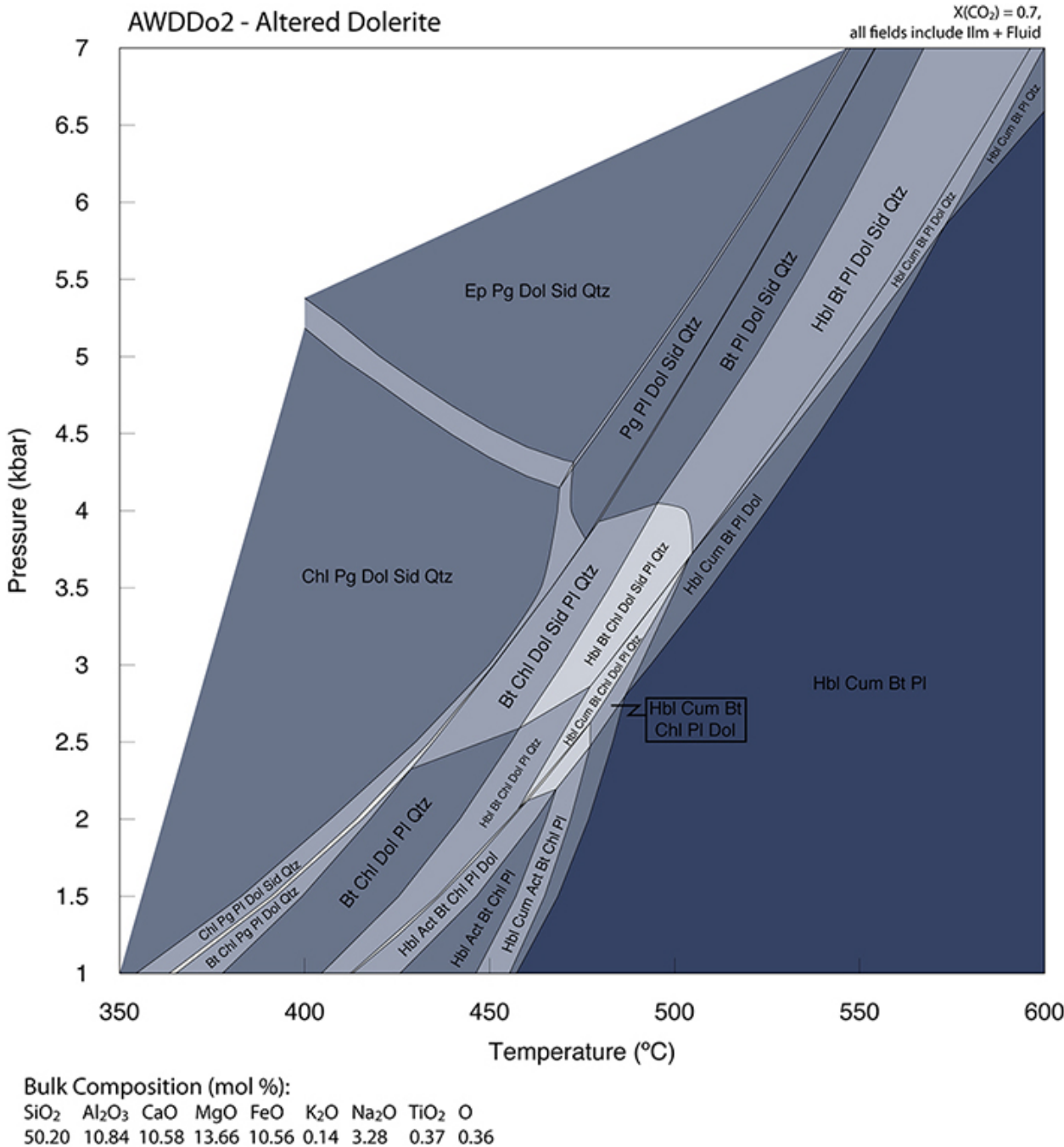


Figure 4.20: A pseudosection for sample AWDDo2, calculated at an $X(\text{CO}_2)$ value of 0.7. The alteration assemblage field (boxed) is Hbl-Cum-Bt-Chl-Dol-Pl-Ilm, which is stable at conditions of 475–500°C and 2.5–3.5 kbar.

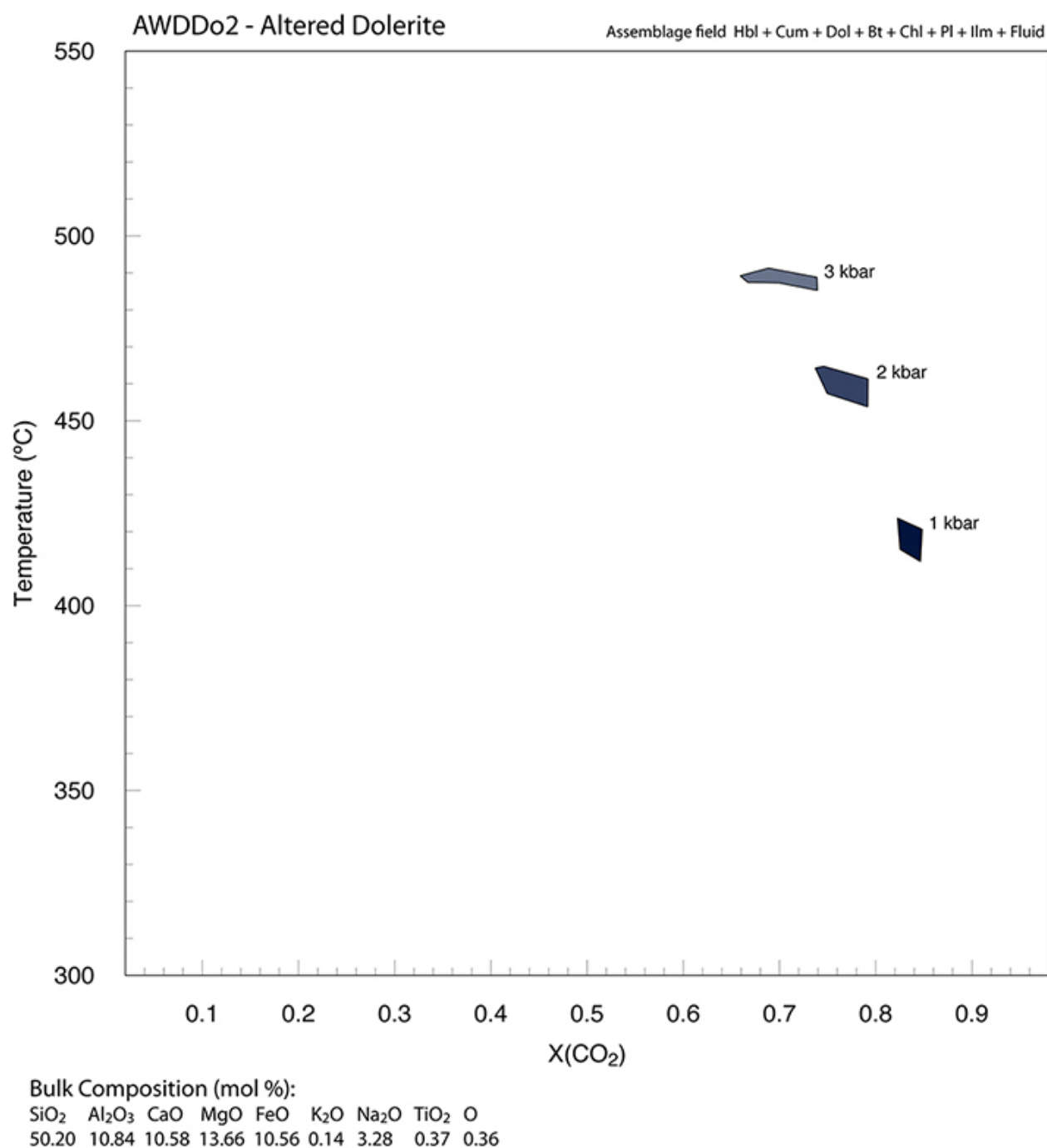


Figure 4.21: A diagram illustrating the change in position of assemblage field Hbl-Cum-Bt-Chl-Dol-Pl-Ilm as a function of pressure, in T-X(CO₂) space, for sample AWDDo2. Increasing pressure shifts the calculated field towards higher temperatures and lower X(CO₂) values before disappearing at pressures above 3.6 kbar.

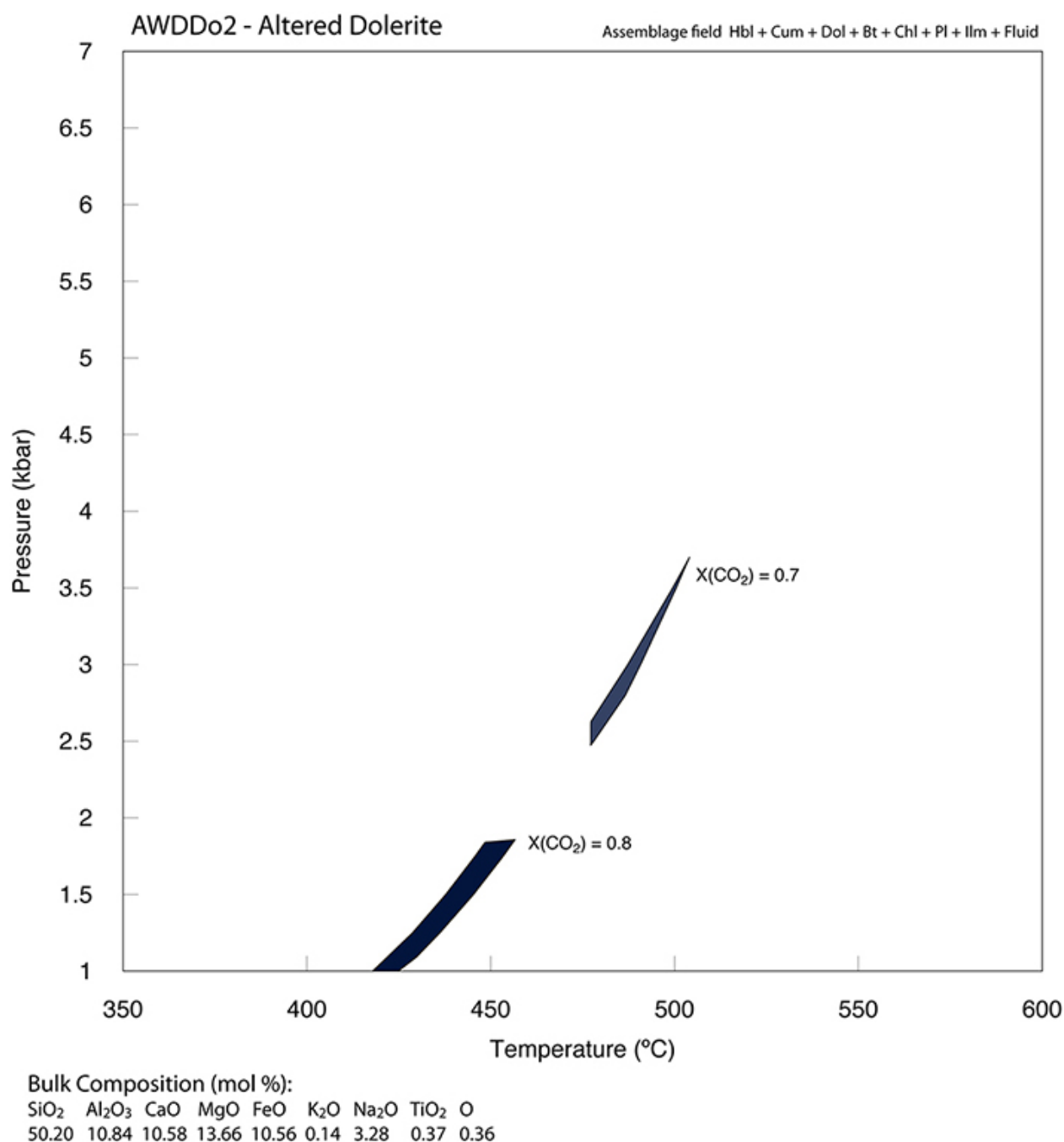


Figure 4.22: A diagram illustrating the change in position of assemblage field Hbl-Cum-Bt-Chl-Dol-Pl-Ilm as a function of $X(\text{CO}_2)$, in P-T space, for sample AWDDo2. Increasing $X(\text{CO}_2)$ shifts the calculated field towards lower temperatures and pressures. The field is not present at all below an $X(\text{CO}_2)$ of 0.66.

AWDP4 - Mineralized Tarkwa Phyllite

Sample AWDP4 is a mineralized Tarkwa Phyllite with the full mineral assemblage Bt-Chl-St-Ky-Pl-Tur-Ms-Ilm-Qtz(-Py-Po). Early attempts at modelling, however, indicated a strong disequilibrium between a number of the phases. In particular, the appearance of staurolite and kyanite did not occur until pressures and temperatures similar to regional metamorphic conditions. These phases are therefore interpreted to be relics of the metamorphic assemblage that were not removed during alteration. Conversely, the compositions of the micas and feldspars differ from those in the unmineralized Tarkwa Phyllite samples and suggested conditions much more in line with those determined for the other altered samples discussed above. As a result, modelling was completed using a bulk composition determined from chlorite, the micas, feldspars and quartz only, thereby removing the refractory metamorphic minerals from consideration. The observed assemblage to be modelled is therefore Bt-Chl-Pl-Ms-Qtz.

Pseudosection results are shown in figures 4.23 and 4.24. The observed assemblage is completely devoid of carbonate minerals and is therefore constrained to lie above the band of decarbonation reactions that cross figure 4.23 at $\sim 440\text{--}450^\circ\text{C}$. A key feature of the calculated pseudosections is the stability of paragonite, in addition to muscovite, over a very wide range of P-T-X(CO₂). The absence of paragonite in the observed assemblage constrains conditions to the low X(CO₂), high T portion of figure 4.23 and the low P part of figure 4.24. This paragonite-absent region also overlaps the stability range of coexisting biotite and chlorite. Quartz-absent assemblages occur in a similar geometry to the paragonite-absent fields but are displaced to lower temperatures. This therefore provides a lower temperature constraint.

The observed assemblage occurs in a fairly broad region of P-T-X(CO₂) space. As with previous samples, this field defines an array in P-T-X(CO₂) space, which again trends towards high P and T with decreasing X(CO₂) (Figs. 4.25 and 4.26) and is unbounded, in these diagrams, to lower pressures across the investigated X(CO₂) range. This array, however, exists only at pressures below about 4.4 kbar and at X(CO₂) values less than 0.75. This is due to an increase in the upper temperature to the stability range of quartz, such that at

higher pressures and $X(\text{CO}_2)$ all paragonite-absent assemblages are also quartz-free.

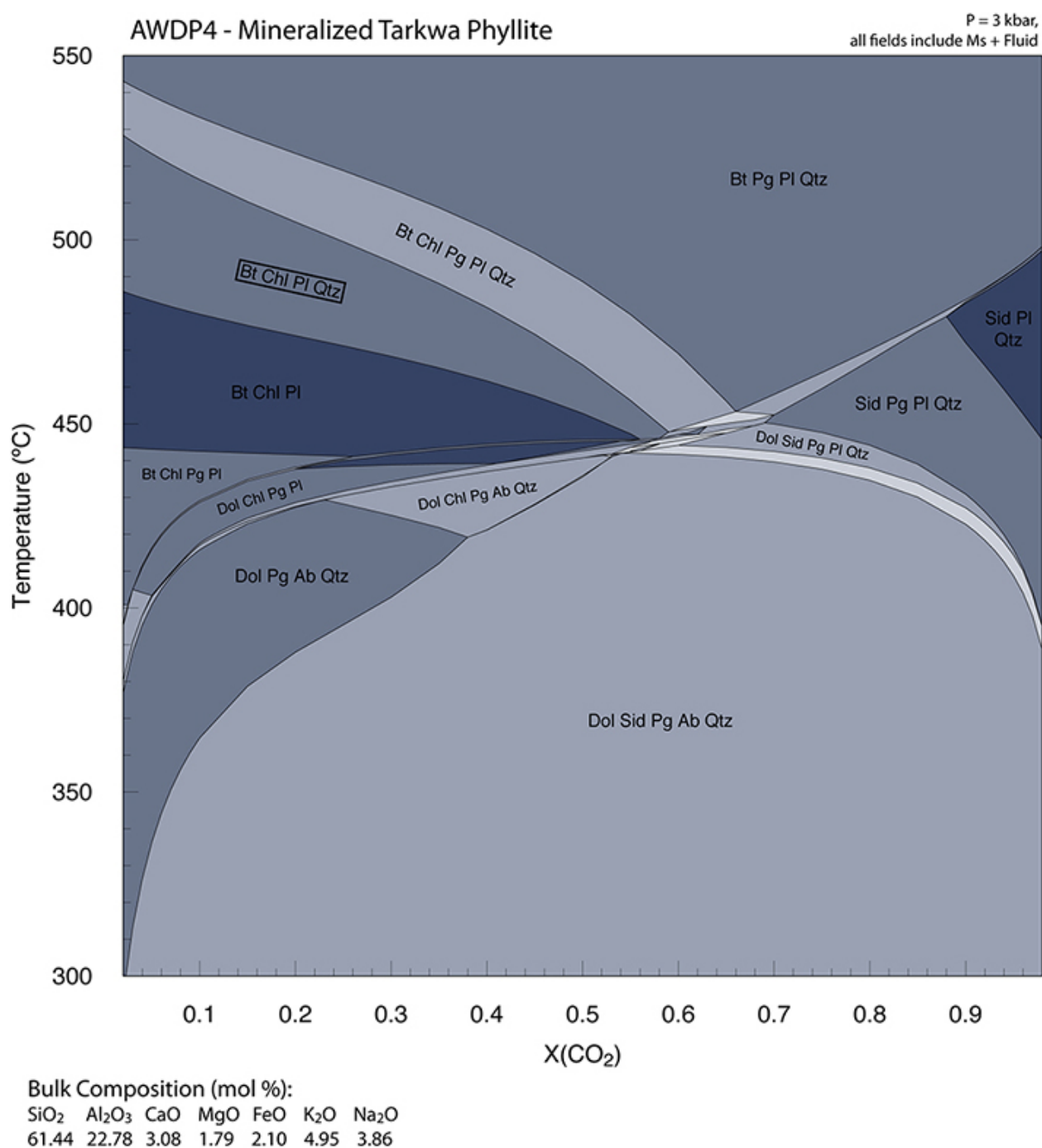


Figure 4.23: A T- $X(\text{CO}_2)$ pseudosection for sample AWDP4, calculated at 3 kbar. The alteration assemblage field (boxed) is Bt-Chl-Pl-Ms-Qtz, which is stable at conditions of 450–530°C and $X(\text{CO}_2)$ of 0–0.6.

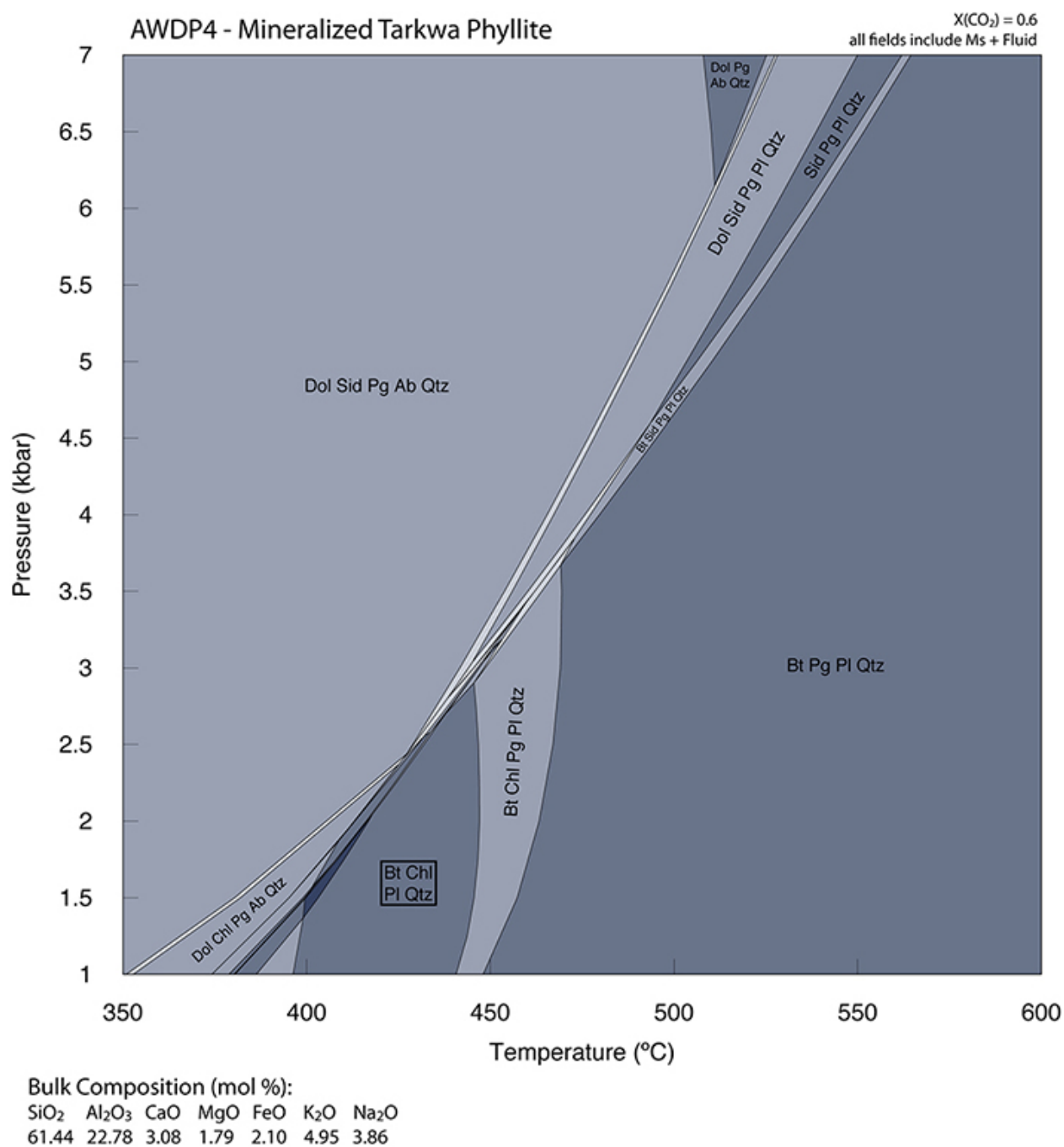


Figure 4.24: A pseudosection for sample AWDP4, calculated at an $X(\text{CO}_2)$ value of 0.6. The alteration assemblage field (boxed) is Bt-Chl-Pl-Qtz, which is stable at conditions of 395–445°C and <1–3 kbar.

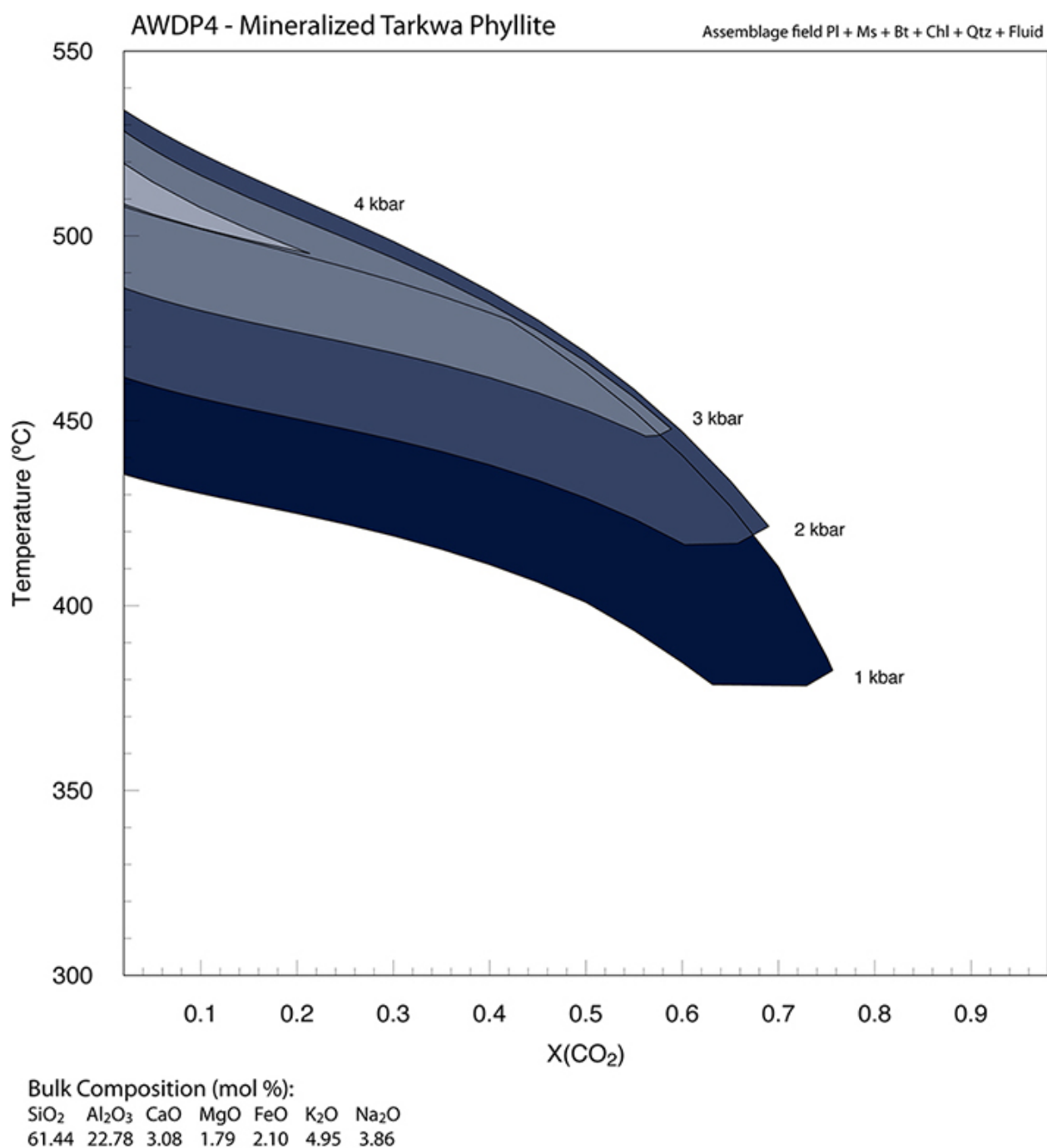


Figure 4.25: A diagram illustrating the change in position of assemblage field Bt-Chl-Pl-Ms-Qtz as a function of pressure, in T-X(CO₂) space, for sample AWDP4. Increasing pressure shifts restricts the field towards higher average temperatures (but lower maximum) and lower X(CO₂) values before disappearing at pressures above 3.4 kbar.

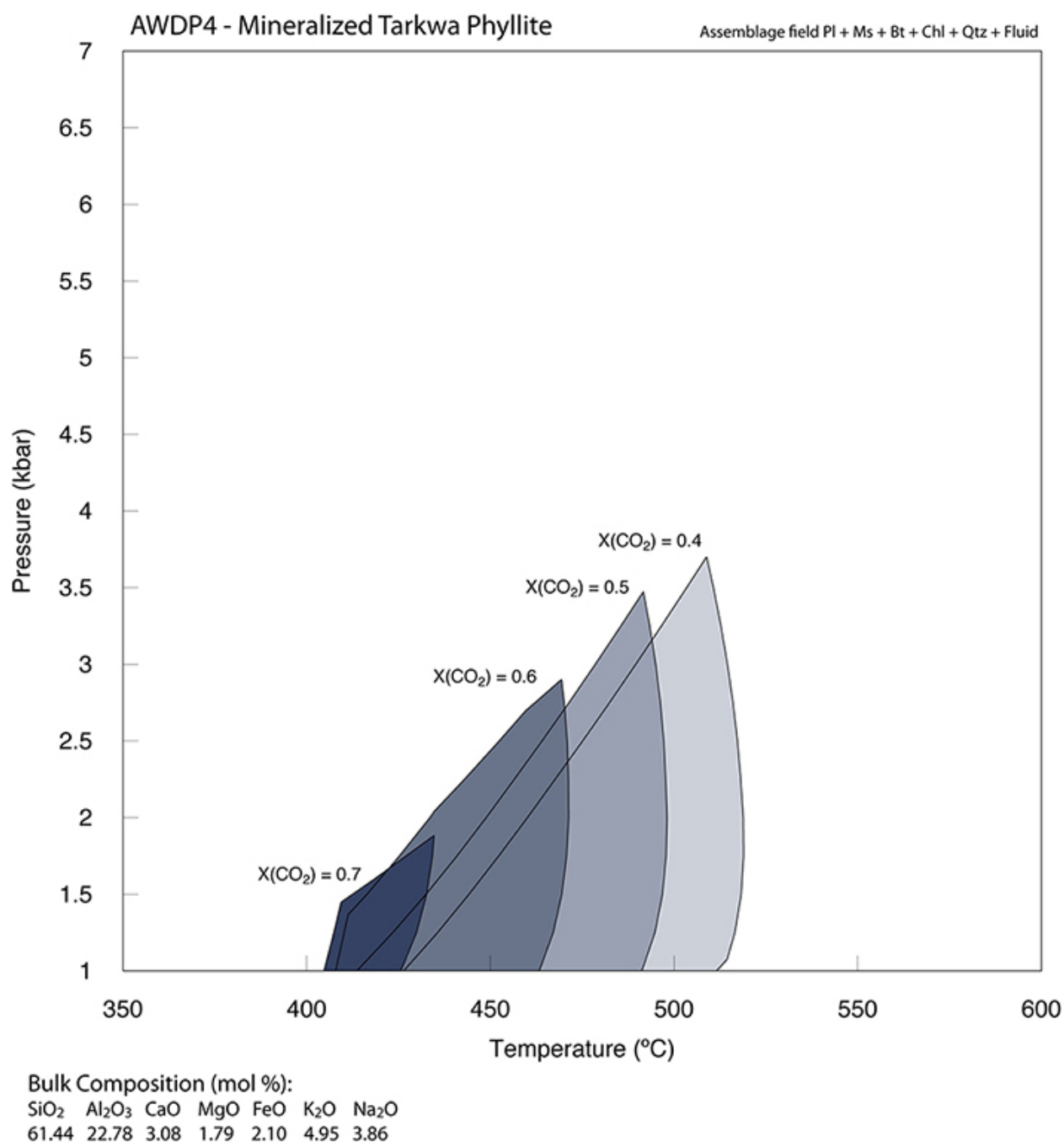


Figure 4.26: A diagram illustrating the change in position of assemblage field Bt-Chl-Pl-Ms-Qtz as a function of $X(\text{CO}_2)$, in P-T space, for sample AWDP4. Increasing $X(\text{CO}_2)$ shifts the calculated field towards lower averages temperatures and pressures. The field is not present at all above an $X(\text{CO}_2)$ of 0.75.

4.4.1 Summary and Discussion of Gold Mineralization Conditions

Modelling results for each of the altered/mineralized samples define arrays through P-T-X(CO₂) space, within which calculated phase compositions and modal proportions match well with observed parameters (Figures 4.27 and 4.28). Conditions for the gold mineralization event are estimated to occur where these arrays overlap, such that all samples agree. This assumes that the mineralizing fluid composition is entirely externally controlled. It is possible, however, that, since the mineralizing event is strongly rock controlled (see chapter 3), mineral equilibria will drive the fluid composition in different directions, which could prevent a perfect overlap of all samples.

The P-T-X(CO₂) arrays for samples AWDDo1, AWABf3, AWDDo2 and AWDP4 are plotted together in figure 4.29. It can be seen that there is not a single region of overlap for all four arrays. There is, however, a clear convergence towards lower pressure and temperature and higher X(CO₂) conditions. Samples AWDDo1 and AWDP4 show significant overlap around 1.5 kbar, 425°C and an X(CO₂) of 0.7. For this same X(CO₂) value, AWABf3 suggests higher pressures (around 2.5 kbar), while AWDDo2 indicates higher pressures and temperatures (around 3 kbar and 480°C). There are a number of possible explanations for these discrepancies:

1. Sample AWABf3 is from the Amoanda pit, while the other samples are from the main Damang pit. These two locations are approximately 5 km apart at present, while the minimum separation of calculated pressures is 0.7 kbar (1.8–2.5), which represents around 1.5 km of crustal depth. It is possible therefore, that during the gold mineralization event, Amoanda was at a slightly deeper, and therefore hotter, crustal level than Damang.
2. As was seen with sample AWDP1, the chloritization of biotite during retrogression resulted in the plagioclase feldspar composition becoming more sodic, due to the conversion of paragonite to muscovite (section 4.3.2). As discussed in the preceding section, a similar chloritization process is inferred to have affected sample AWABf3. This implies that the plagioclase composition may originally have been more calcic, which in turn would pro-

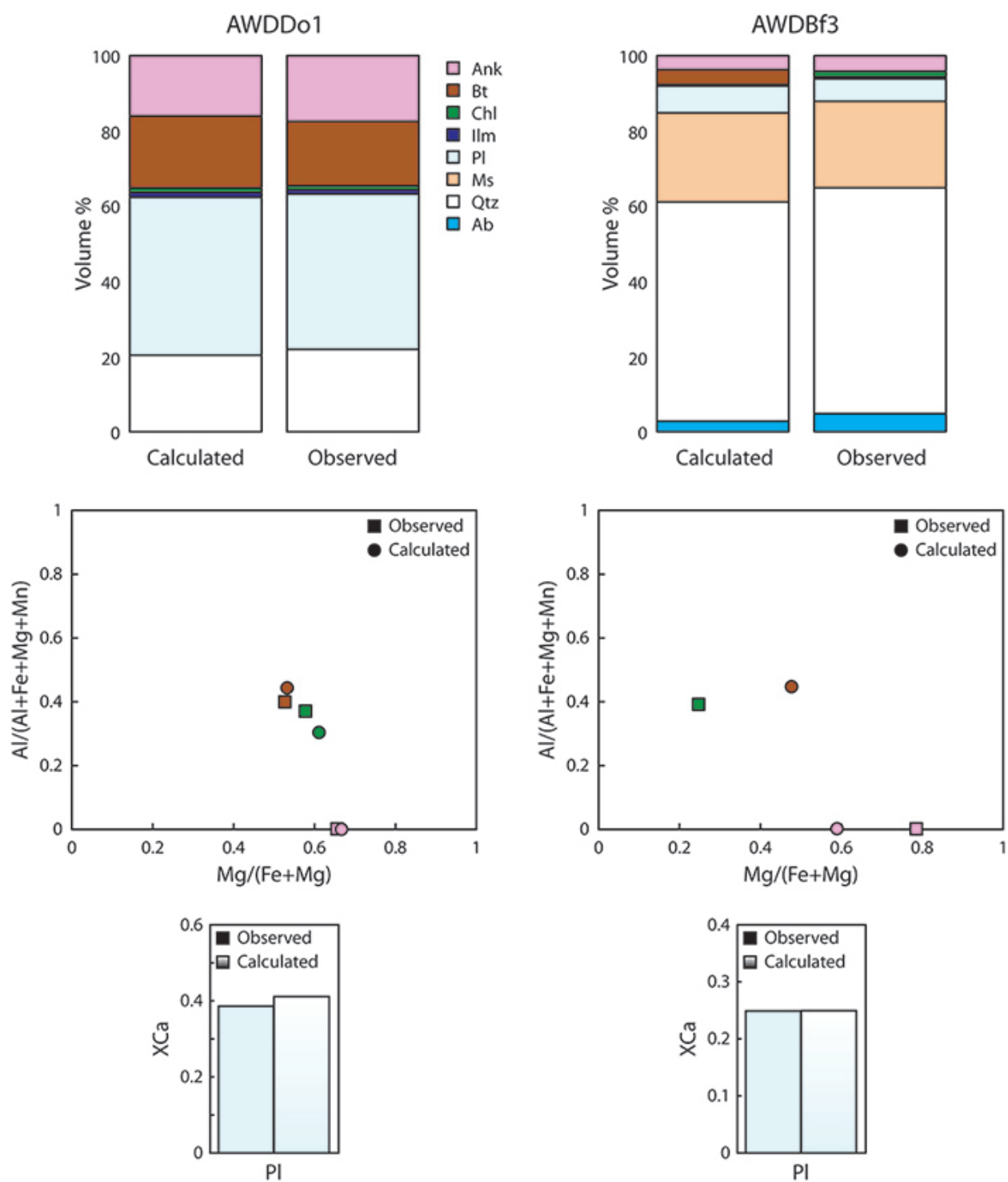


Figure 4.27: Comparisons of observed and calculated modes and phase compositions for AWDDo1 (left) and AWABf3 (right). In the lower panels, X_{Ca} for plagioclase = $Ca/(Ca+Na+K)$.

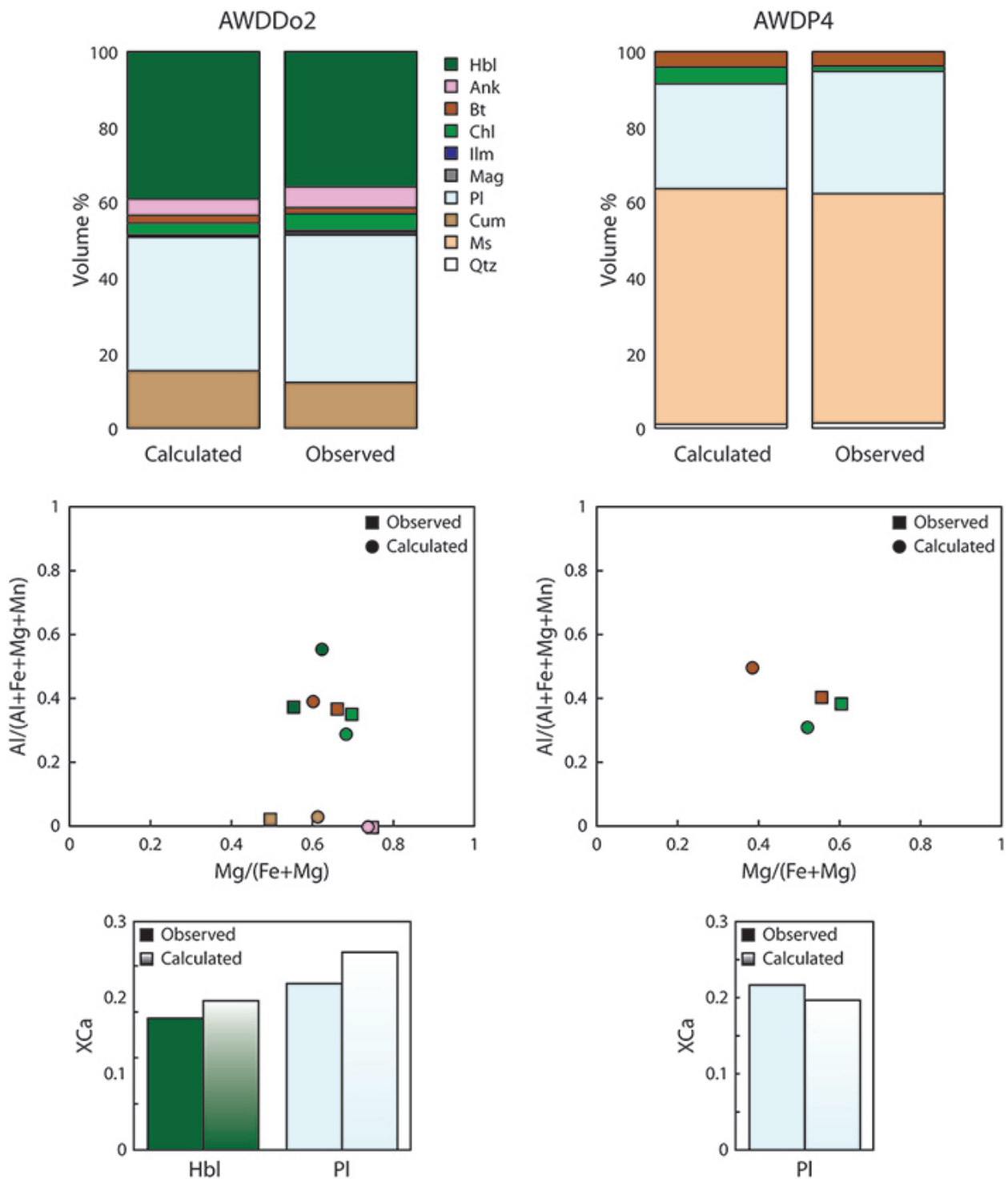


Figure 4.28: Comparisons of observed and calculated modes and phase compositions for AWDDo2 (left) and AWDP4 (right). In the lower panels, XCa for hornblende = $Ca/(Ca+Fe+Mg)$ and for plagioclase = $Ca/(Ca+Na+K)$.

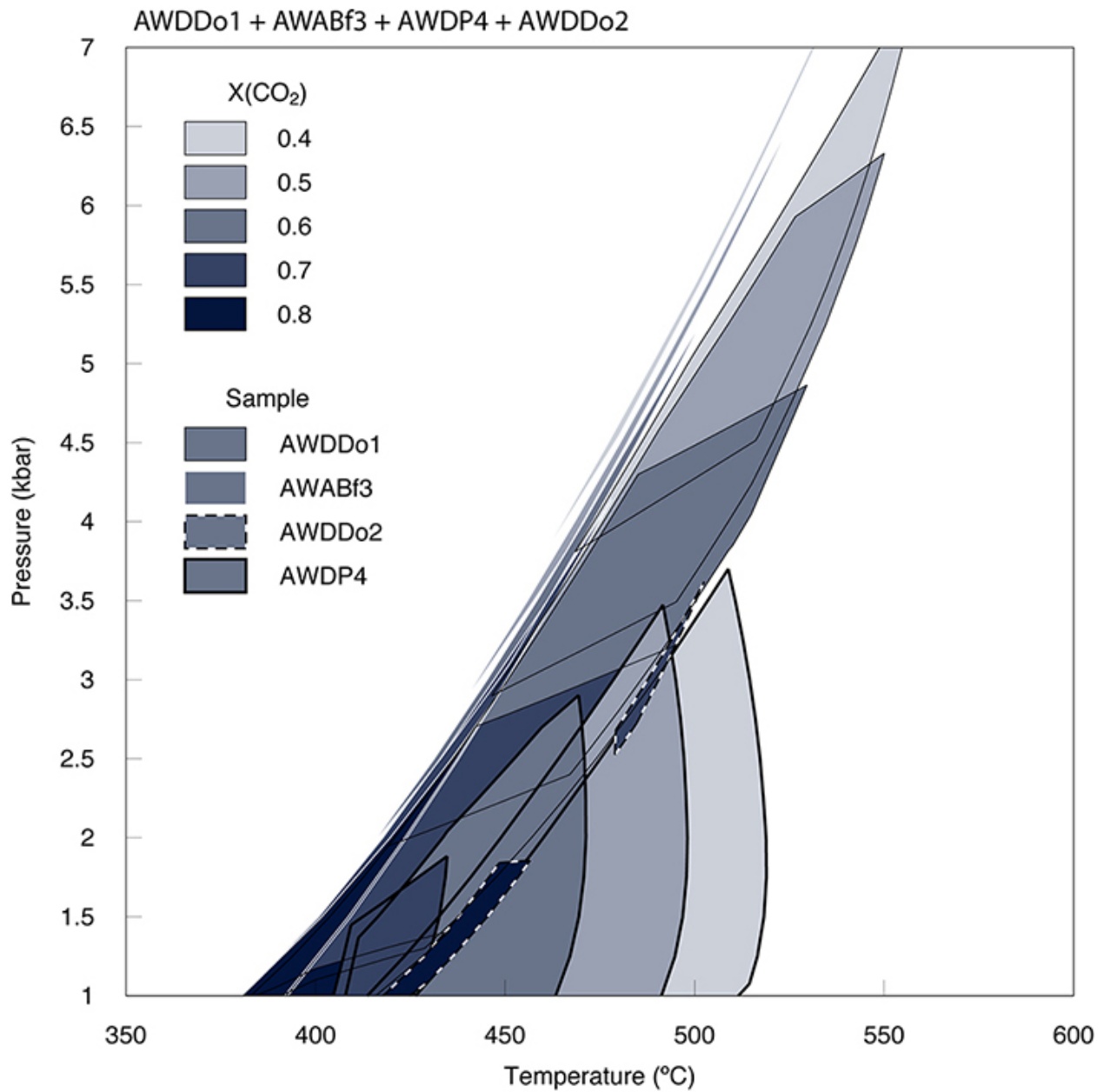


Figure 4.29: A diagram illustrating the P-T-X(CO₂) arrays of stable alteration assemblages, as shown in figures 4.13, 4.17, 4.22 and 4.26, plotted in P-T space. All arrays converge with decreasing temperature and pressure and increasing X(CO₂).

duce lower mineralization temperatures, and consequently pressures, than those suggested in figure 4.18.

3. Sample AWDDo2 exhibits only patchy alteration with incomplete replacement of metamorphic phases, rather than the more pervasive, vein-associated alteration seen in the other samples. It may therefore represent a product of a lower fluid/rock ratio such that alteration stopped at an earlier stage in the hydrothermal event, which, if mineralization was prolonged during post-peak-metamorphic cooling, could correspond to higher P-T conditions.

Overall, all modelled samples suggest conditions for hydrothermal alteration, and therefore gold mineralization, of around 400–450°C, 1–2 kbar and with a fluid composition of around 0.7 X(CO₂). This is broadly in line (although slightly higher temperature) with estimates from fluid inclusion studies at Damang (Schmidt Mumm *et al.*, 1997; Pigois, 2003) and other Ghanaian orogenic gold deposits (Schmidt Mumm *et al.*, 1997; Yao *et al.*, 2001; Wille & Klemm, 2004) (see also section 2.7.2).

4.5 Discussion of the P-T-X(CO₂) History of Damang

The results of THERMOCALC modelling discussed in this chapter successfully constrain both the P-T conditions of peak regional metamorphism and the P-T-X(CO₂) conditions of the gold mineralization event. The difference between these two sets of conditions therefore constrains a portion of the P-T path followed by the Damang region.

Figure 4.30 shows the P-T ranges of stable alteration assemblages calculated for the four hydrothermally altered samples, at a fluid X(CO₂) of 0.7, as per figure 4.29. Also shown are peak regional metamorphic conditions, as calculated from sample AWDP1 (section 4.3.2, figure 4.2), at 590°C and 5.25 kbar. From this point, a number of geothermal gradients are drawn. A simple linear geotherm from this point to the surface ($\sim 0, 0$) has a gradient of around 38°C/km (solid red line). It can clearly be seen that the region of likely conditions for mineralization falls significantly below this geotherm. Instead, it suggests a geothermal gradient of around 14°C/km (from peak conditions) (dashed red line). Importantly, it can

also be seen that sample AWDDo2, at an $X(\text{CO}_2)$ of 0.7, falls directly on this geotherm. Ultimately, this indicates that the Damang region experienced erosion and/or exhumation over and above a steady-state pressure and temperature decrease, consistent with a typical clockwise P-T path. This will be addressed further in chapter 7.

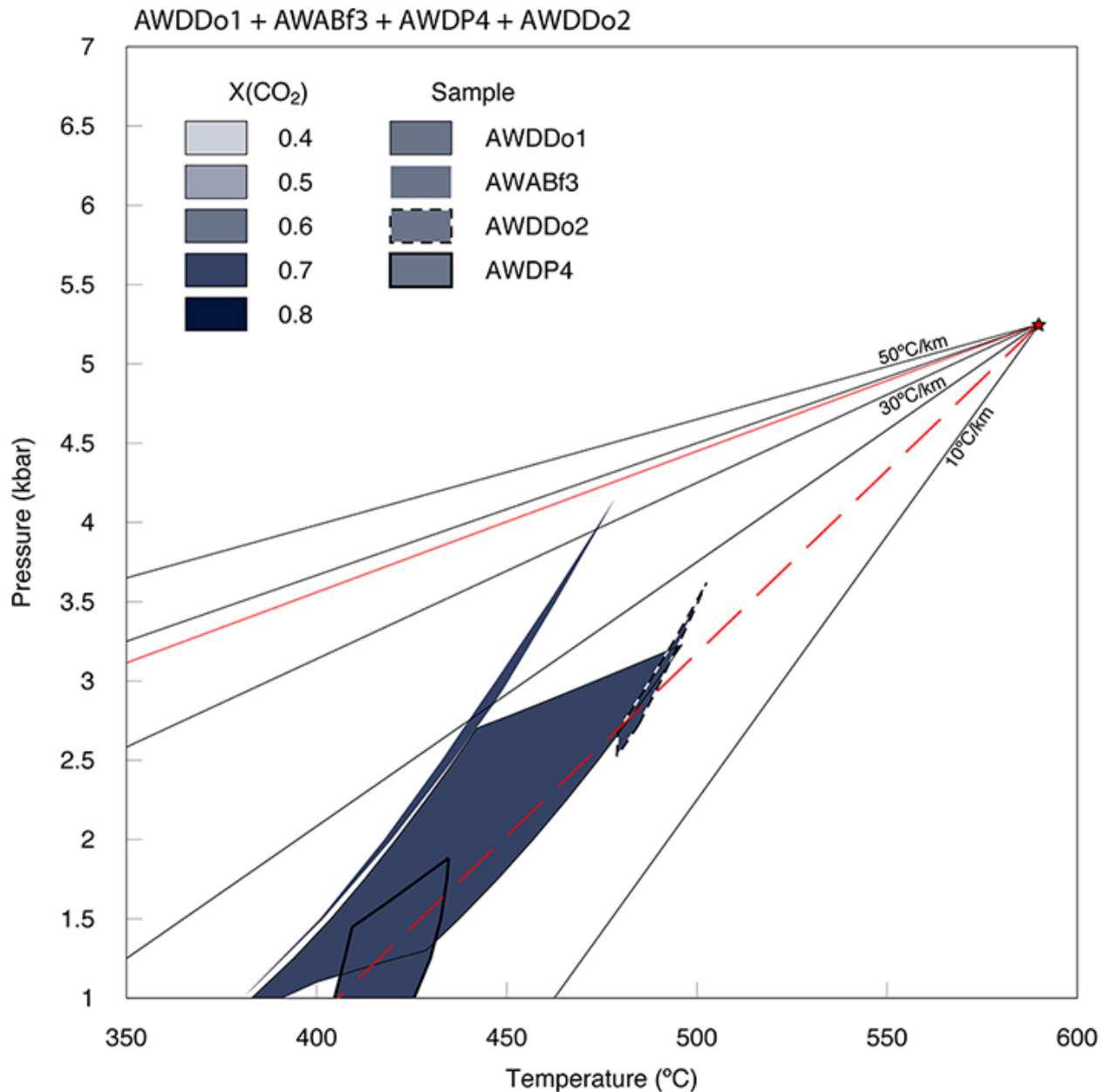


Figure 4.30: A diagram illustrating the P-T ranges of stable alteration assemblages, at a fluid $X(\text{CO}_2)$ of 0.7, as per figure 4.29. The red star represents peak regional metamorphic conditions, as per figure 4.2, with a range of geothermal gradients emanating from it. The solid red line represents a simple linear geotherm to the surface, with a gradient of $38^\circ/\text{km}$. The dashed red line represents a geotherm through the broad zone of overlap in modelled alteration conditions, with a gradient of $14^\circ/\text{km}$.

Chapter 5

Geochronology



A dump truck in the Damang pit.

It was the work of Ernest Rutherford ([Rutherford, 1900](#)) at the start of the twentieth century that began the revolution of isotope geochronology, which has since grown into a large and diverse field of research. The development of mass spectrometry since the 1970s significantly advanced the science and it is now used to date material as old as the solar system itself or as young as the Holocene. Isotope geochronology is especially useful in determining both the absolute and relative timings of metamorphic, magmatic and metasomatic episodes in a range of geologic settings.

Many ore systems exhibit a prolonged formational history and this is especially true of hydrothermal deposits in metamorphic terranes where various tectonic, magmatic, metamorphic and hydrothermal events may have occurred. In order to fully understand an ore system such as Damang, the age of these episodes must be resolved, not only in absolute terms, but also relative to one another such that rates and timescales of processes can be determined. Consecutive geologic events may be recorded by the growth of different minerals, each with distinct isotopic characteristics, and so a multi-system approach is required.

This chapter presents the results and interpretation of the geochronology conducted at Damang. This includes U-Pb isotopic analysis of zircons from basement igneous rocks, which provide both a framework and a maximum limit for all other processes. U-Th-total Pb chemical analysis of monazites from the Tarkwa Phyllite constrains the timing of regional metamorphism prior to gold mineralization. Re-Os isotope analysis of pyrite and pyrrhotite from hydrothermally altered rocks associated with gold mineralization can directly determine the age of mineralization. Finally, $^{40}\text{Ar}/^{39}\text{Ar}$ analysis of biotite from both hydrothermally altered zones and unmineralized host rocks gives the age at which the region cooled below $300\pm 50^\circ\text{C}$.

5.1 Principles of Geochronology

The principles of radiogenic isotope geochronology rely upon the spontaneous breakdown of unstable isotopes of certain elements in order to achieve a stable atomic configuration. This breakdown process is radioactivity and the newly formed isotopes are termed ‘radiogenic’.

The time taken for the amount of a radioactive isotope to reduce by a factor of two is known as the half-life.

The number of radiogenic daughter atoms (D^*) present after time t is proportional to the number of radioactive parent atoms initially present (P_0) through the decay constant (λ):

$$D^* = P_0(1 - e^{-\lambda t}) \quad (5.1)$$

or in terms of the number of radioactive parent atoms (P) present at time t :

$$D^* = P(e^{\lambda t} - 1) \quad (5.2)$$

The decay constant (λ) is related to the half-life ($T_{\frac{1}{2}}$):

$$T_{\frac{1}{2}} = \frac{\ln 2}{\lambda} \quad (5.3)$$

and is assumed to be constant through time. Equation 5.2 can be rearranged to give the ratio of daughter and parent atoms at time t :

$$\frac{(D - D_0)}{P} = e^{\lambda t} - 1 \quad (5.4)$$

where D is the total number of radiogenic daughter atoms, *i.e.* $D = D^* + D_0$ and D_0 is the number of non-radiogenic daughter atoms. D_0 is a constant that can either be assumed or calculated from a cogenetic suite of samples.

An age can therefore be calculated by rearranging equation 5.4 in to the form:

$$t_{(age)} = \frac{1}{\lambda} \ln \left(1 + \frac{(D - D_0)}{P} \right) \quad (5.5)$$

The age calculated from this ratio should be accurate, assuming certain conditions are met:

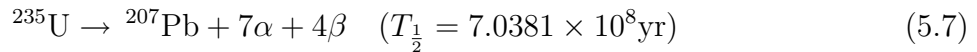
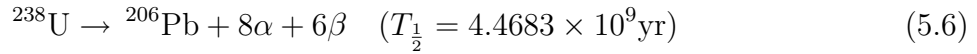
1. The system has remained closed throughout its history and has not gained or lost any parent or daughter atoms or any other intermediate daughter products.

2. The initial number of daughter atoms, prior to any addition through radioactive decay within the system, is estimated correctly.
3. λ is known accurately.
4. The values of D and P can be measured accurately with all analytical errors free from systematic errors.

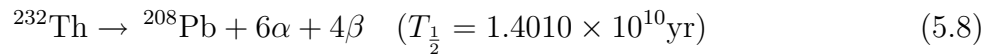
For the effective use of isotope geochronology, a decay system is needed with a $T_{\frac{1}{2}}$ on the order of the age of the material to be dated, which itself must contain sufficient and measurable D and P.

5.1.1 U-Th-Pb Isotope Systematics

Uranium has three naturally occurring nuclides, ^{234}U , ^{235}U and ^{238}U , which are all radioactive, although ^{234}U only occurs as an intermediate decay product of ^{238}U (figure 5.1). The decay schemes for ^{235}U and ^{238}U are as follows:



Thorium has six naturally occurring nuclides, although it exists predominantly as ^{232}Th , with the other five being intermediate daughters of the decay of ^{235}U , ^{238}U and ^{232}Th (figure 5.1). The decay scheme for ^{232}Th is as follows:



As the half-lives of the parent U and Th isotopes are orders of magnitude larger than the intermediate radioactive isotopes, the production rate of stable radiogenic Pb is effectively equal to the decay rate of the parents. This relationship is termed ‘secular equilibrium’ and implies that U and Th can be considered to directly decay to Pb.

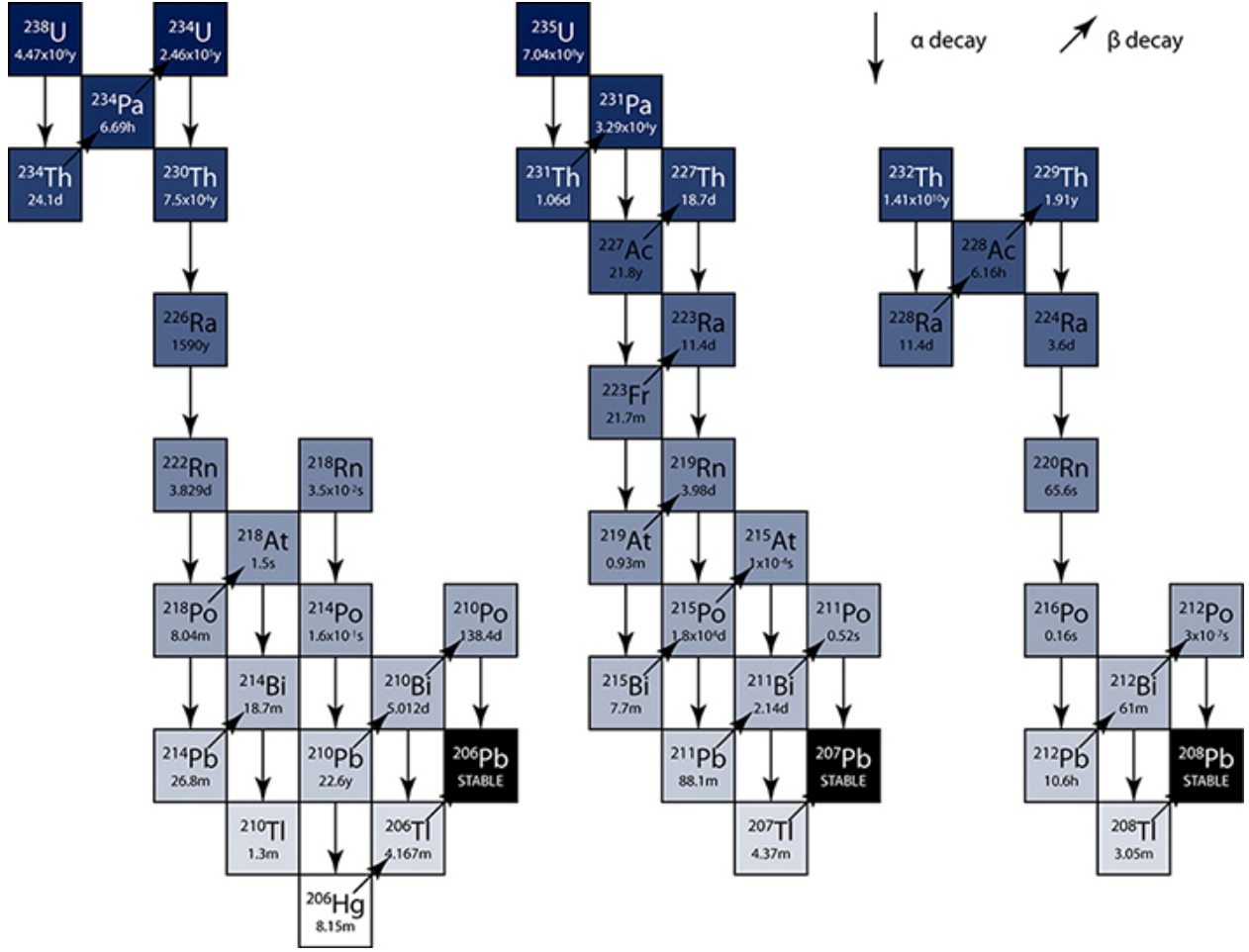


Figure 5.1: Radioactive decay schemes for uranium and thorium, with the half-lives of intermediate daughter products shown. Modified after [Faure \(1986\)](#).

Lead has four naturally occurring isotopes, non-radiogenic ^{204}Pb , and radiogenic ^{206}Pb , ^{207}Pb and ^{208}Pb , produced through the radioactive decay of ^{238}U , ^{235}U and ^{232}Th respectively. This allows equation 5.4 to be expressed for all of ^{238}U , ^{235}U and ^{232}Th :

$$(a) \frac{^{206}\text{Pb}^*}{^{238}\text{U}} = e^{\lambda^{238}t} - 1 \quad (b) \frac{^{207}\text{Pb}^*}{^{235}\text{U}} = e^{\lambda^{235}t} - 1 \quad (c) \frac{^{208}\text{Pb}^*}{^{232}\text{Th}} = e^{\lambda^{232}t} - 1 \quad (5.9)$$

where λ^{238} is the decay constant for ^{238}U , etc. Therefore, measuring the ratios of parent to daughter atoms can allow for the calculation of an age by rearranging equations 5.9a-c into the form of equation 5.5. Here, D_0 is assumed to be essentially zero, as very little Pb is incorporated into the zircon as it grows. The majority of measured Pb is therefore interpreted to be radiogenic.

The ages calculated from these independent ratios should be in mutual agreement, assuming

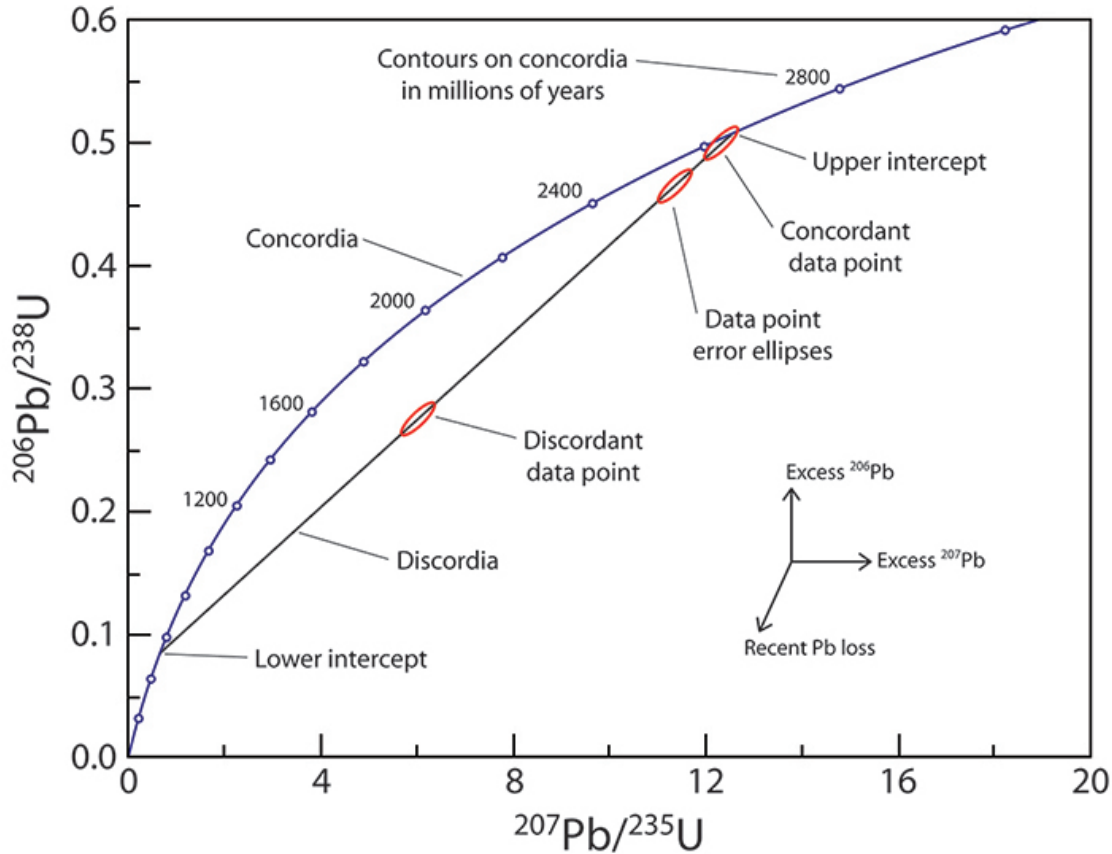


Figure 5.2: Example Wetherill concordia plot illustrating the effect of excess and/or loss of Pb. A discordia defines an upper intercept, representing the age of formation, and a lower intercept that may represent a later episode resulting in the partial ‘resetting’ of the system.

the conditions listed above are met. By comparing the two ratios, departures from the ideal can be assessed.

A two-dimensional plot of $^{207}\text{Pb}^*/^{235}\text{U}$ vs. $^{206}\text{Pb}^*/^{238}\text{U}$ defines a curve through time, termed concordia (Wetherill (1956) and Figure 5.2). If the above conditions are met then the measured ages will agree and points will plot on concordia (*i.e.* they will be ‘concordant’). Departures from the assumptions result in ‘discordant’ points, which plot above or below concordia, due to the two calculated ages not being in agreement. This can, however, be used to an advantage in cases where Pb loss has been extensive and an array of points below concordia define a ‘discordia’, the lower intercept of which provides a good indication for the timing of the Pb loss event.

5.1.2 Chemical U-Th-Pb Systematics

U-Th-Total Pb ‘chemical dating’ utilizes the same decay scheme outlined in Section 5.1.1 but involves measuring total UO_2 , ThO_2 and PbO concentrations rather than isotopic ratios. Since only total elemental (or oxide) concentrations are required, chemical dating can be carried out routinely on an electron microprobe, which is both quicker, in terms of sample preparation time or access to equipment, and cheaper than using isotopic methods. An age for each analysis can be determined by solving the following equation.

$$0 = \text{PbO} - 223.2 \left(\frac{\text{ThO}_2}{264} (e^{\lambda_{232}t} - 1) + \frac{0.9928 \times \text{UO}_2}{270} (e^{\lambda_{238}t} - 1) + \frac{0.0072 \times \text{UO}_2}{270} (e^{\lambda_{235}t} - 1) \right) \quad (5.10)$$

This method is suitable only for minerals that contain significant amounts of U and/or Th, where the amount of common Pb is negligible, or for samples of sufficient age that it becomes negligible, and in which variation of U/Th/Pb ratios are controlled only by radioactive decay (i.e. the standard geochronological assumption that the system is closed). One such appropriate mineral is monazite. Monazite commonly contains several weight percent ThO_2 and hundreds of ppm to a few weight percent UO_2 , leading to rapid accumulation of radiogenic Pb, while rarely containing common PbO exceeding 1 ppm (Parrish, 1990). Unlike isotopic methods, chemical dating is unable to detect discordant monazites, which would produce geologically meaningless ages. However, monazites are typically concordant, which reduces this concern (Cocherie *et al.*, 1998; Scherrer *et al.*, 2000). Additionally, a number of studies have looked into the behaviour of Pb in monazites and while diffusion and loss can occur, it is generally uncommon and not thought to be a major problem (Suzuki *et al.*, 1994; Montel *et al.*, 1996; Cocherie *et al.*, 1998).

Calculation of a population age can be achieved through a number of methods. Early studies used an isochron method in which measured PbO is plotted against ‘ ThO_2^* ’, a calculated value that represents the concentration of ThO_2 that would have been required to produce the entire concentration of radiogenic Pb derived from both Th and U (Suzuki & Adachi, 1991; Suzuki *et al.*, 1991, 1994). The slope of the regression line through these data points

is then proportional to the age of the sample. This approach assumes all analysed domains have the same age regardless of chemical composition. Later studies have tended towards the use of statistical treatment of data, using either poisson statistics directly from the electron probe analysis (Montel *et al.*, 1996), or through repeated measurements of a ‘compositional domain’ (i.e. population statistics) (Williams & Jercinovic, 2002).

5.1.3 Re-Os Isotope Systematics

Rhenium has two naturally occurring nuclides, ^{185}Re , which is stable, and ^{187}Re , which is radioactive. The decay scheme for ^{187}Re is as follows:

$$^{187}\text{Re} \rightarrow ^{187}\text{Os} + \beta^- + \bar{\nu} \quad (T_{\frac{1}{2}} = 4.56 \times 10^{10} \text{yr}) \quad (5.11)$$

Osmium has seven naturally occurring isotopes, all of which are stable on practical timescales, due to extremely long half lives, and of which only ^{187}Os is radiogenic. Modifying equation 5.4 allows for the production of ^{187}Os in a Re-bearing system to be described as follows:

$$\frac{^{187}\text{Os}}{^{186}\text{Os}} = \left(\frac{^{187}\text{Os}}{^{186}\text{Os}} \right)_i + \frac{^{187}\text{Re}}{^{186}\text{Os}} (e^{\lambda t} - 1) \quad (5.12)$$

where λ is the decay constant for ^{187}Re , and subscript ‘*i*’ denotes the initial ratio prior to closure of the system. Since this initial ratio is not known and cannot necessarily be assumed, the above equation must be applied to a suite of samples of the same age and initial $^{187}\text{Os}/^{186}\text{Os}$ ratio. Individually these samples should have different values for $^{187}\text{Os}/^{186}\text{Os}$ and $^{187}\text{Re}/^{186}\text{Os}$, thereby defining an isochron (Fig. 5.3). The slope of this isochron, with a gradient of $(e^{\lambda t} - 1)$, is proportional to the age of the suite of samples and the y-axis intercept is equal to the initial $^{187}\text{Os}/^{186}\text{Os}$ ratio.

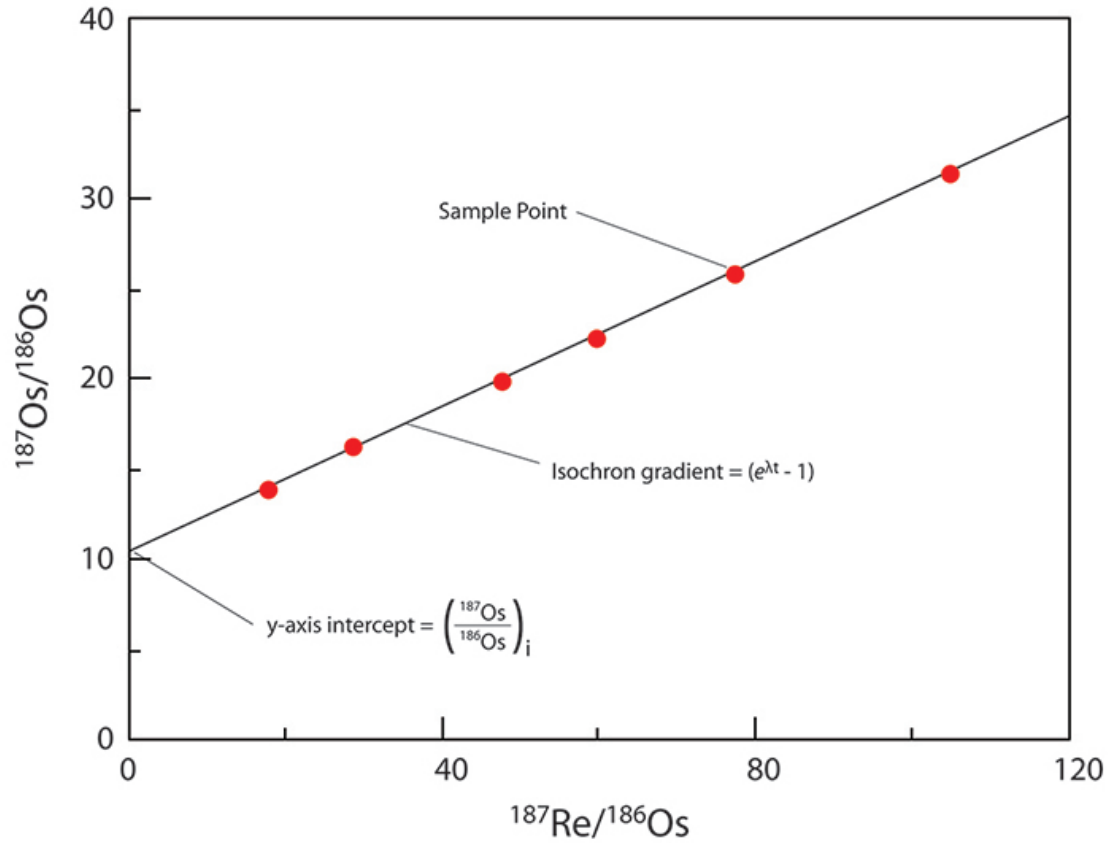
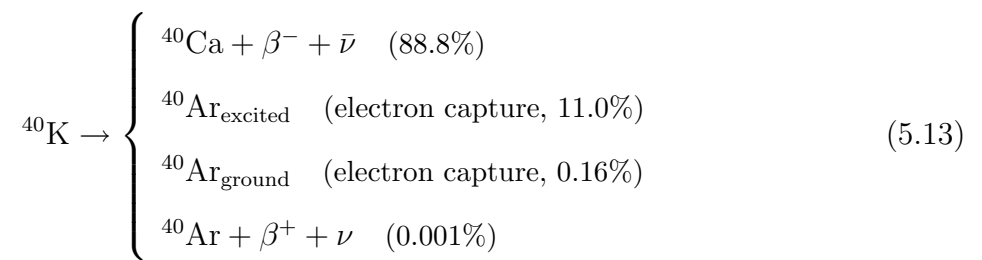


Figure 5.3: Schematic isochron plot for the Re-Os system in which a suite of samples define a line whose slope is proportional to the age of the samples and whose y-intercept equals the system's initial $^{187}\text{Os}/^{186}\text{Os}$ ratio

5.1.4 $^{40}\text{Ar}/^{39}\text{Ar}$ Isotope Systematics

The $^{40}\text{Ar}/^{39}\text{Ar}$ method of dating is a development of the previously more conventional K-Ar dating method. Potassium has three naturally occurring nuclides ^{39}K , ^{40}K and ^{41}K of which only ^{40}K is radioactive. The decay scheme for ^{40}K is as follows:



Although 88.8% of ^{40}K decays to ^{40}Ca , this is not a useful decay scheme to use since ^{40}Ca is the most abundant (96.9%) Ca isotope, and Ca is a common element in many rock

forming minerals. Therefore, accurately measuring radiogenic ^{40}Ca is difficult. By contrast, since Ar is gaseous, it is not readily incorporated into minerals and is rapidly lost when at temperatures above the mineral's closure temperature (the temperature below which Ar is assumed to remain trapped within the crystal).

The two branches of this decay chain (one producing ^{40}Ca , the other ^{40}Ar) have different decay constants:

$$(a) \lambda_e = 0.581 \times 10^{-10} \text{ y}^{-1} \quad (b) \lambda_\beta = 4.962 \times 10^{-10} \text{ y}^{-1} \quad (5.14)$$

where λ_e refers to the decay of ^{40}K to ^{40}Ar and λ_β the decay to ^{40}Ca . This produces a total decay constant (λ) of $5.543 \times 10^{-10} \text{ y}^{-1}$ and a total half-life for ^{40}K of $1.250 \times 10^9 \text{ y}$. As per equation 5.5, a K-Ar age can therefore be calculated:

$$t_{(age)} = \frac{1}{\lambda} \ln \left(1 + \left(\frac{\lambda}{\lambda_e} \right) \frac{{}^{40}\text{Ar}^*}{{}^{40}\text{K}} \right) \quad (5.15)$$

This technique was developed further by [Merrihue & Turner \(1966\)](#) into the $^{40}\text{Ar}/^{39}\text{Ar}$ dating method, which has several advantages over traditional K-Ar dating. In particular, only isotope ratios of Ar are measured, which can be done from the same sample, rather than separately determining the absolute concentrations of K and Ar.

$^{40}\text{Ar}/^{39}\text{Ar}$ dating is based upon the irradiation of K with thermal and fast neutrons in a nuclear reactor, which leads to the production of ^{39}Ar from ^{39}K through the reaction $^{39}\text{K}(n,p)^{39}\text{Ar}$. Although ^{39}Ar does decay over time to ^{39}K , its half-life of 269 years makes it essentially stable on the short time scales of analysis. The number of ^{39}Ar atoms formed from ^{39}K can be described as follows:

$${}^{39}\text{Ar} = {}^{39}\text{K} \Delta T \int \varphi(\epsilon) \sigma(\epsilon) d\epsilon \quad (5.16)$$

where ${}^{39}\text{K}$ is the number of such atoms in the sample, ΔT is the length of the irradiation, $\varphi(\epsilon)$ is the neutron flux density at energy ϵ , $\sigma(\epsilon)$ is the capture cross section of ^{39}K for neutrons having energy ϵ , and the integration is calculated for the entire energy spectrum

of the neutrons.

From equation 5.2, the number of radiogenic ^{40}Ar atoms present in the sample is given by:

$$^{40}\text{Ar}^* = \frac{\lambda_e}{\lambda} ^{40}\text{K}(e^{\lambda t} - 1) \quad (5.17)$$

Combining equations 5.16 and 5.17 gives the $^{40}\text{Ar}^*/^{39}\text{Ar}$ ratio after neutron irradiation:

$$\frac{^{40}\text{Ar}^*}{^{39}\text{Ar}} = \frac{\lambda_e}{\lambda} \frac{^{40}\text{K}}{^{39}\text{K}} \frac{1}{\Delta T} \int \frac{e^{\lambda t} - 1}{\varphi(\epsilon)\sigma(\epsilon)} d\epsilon \quad (5.18)$$

A parameter J can now be defined as:

$$J = \frac{\lambda}{\lambda_e} \frac{^{39}\text{K}}{^{40}\text{K}} \Delta T \int \varphi(\epsilon)\sigma(\epsilon) d\epsilon \quad (5.19)$$

which is substituted in to equation 5.18 to give:

$$\frac{^{40}\text{Ar}^*}{^{39}\text{Ar}} = \frac{e^{\lambda t} - 1}{J} \quad (5.20)$$

However, since $\varphi(\epsilon)$ and $\sigma(\epsilon)$ are not well known, parameter J must be determined empirically by irradiating a second sample, termed the flux monitor, which has a known age that can be inserted in to equation 5.20. This finally allows for a $^{40}\text{Ar}/^{39}\text{Ar}$ age to be calculated:

$$t = \frac{1}{\lambda} \ln \left(\frac{^{40}\text{Ar}^*}{^{39}\text{Ar}} J + 1 \right) \quad (5.21)$$

During sample irradiation additional Ar isotopes are produced other than just ^{39}Ar from ^{39}K . The most important reactions (although many others exist) are summarised in table 5.1. Each of these interferences must be corrected for, by monitoring reactions directly through including K_2SO_4 and CaF_2 in the irradiation and cross-referencing with other interferences.

Following irradiation, measurement of Ar isotopic ratios is made through step heating analysis. Step heating involves progressively heating the sample, in order to degas the contained Ar, and making measurements after each temperature interval. Ar may occupy a number of

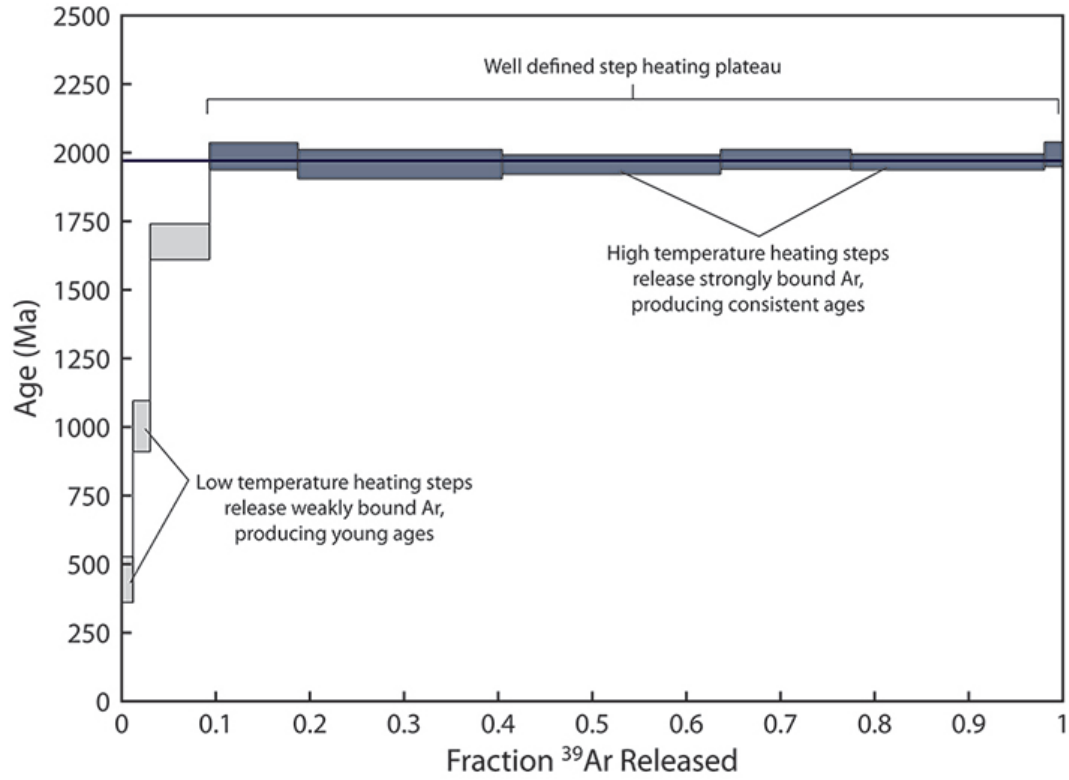


Figure 5.4: Example of an $^{40}\text{Ar}/^{39}\text{Ar}$ step heating plateau

different sites within a mineral lattice, some of which are more strongly bound than others. Relatively low temperatures heating steps will liberate the most weakly bound Ar in the sample first. These sites may well have lost radiogenic ^{40}Ar earlier in the mineral’s history resulting in ages that are too young. In contrast, the later high temperature heating steps release Ar that is strongly bound within the mineral and is assumed to have remained closed since the mineral cooled below its closure temperature. These heating steps should therefore produce ‘real’ ages. A step heating plateau is constructed by plotting apparent age for each heating step against the cumulative amount of Ar released, with a geologically meaningful age being defined by presence of a statistically coherent plateau, as illustrated in figure 5.4.

Table 5.1: Important interfering nuclear reactions produced during sample irradiation, after [McDougall & Harrison \(1999\)](#)

Ar Isotope Produced	K	Source element	
		Ca	Cl
^{36}Ar		$^{40}\text{Ca}(n, n\alpha)$	$^{35}\text{Cl}(n, \gamma, \beta)$
^{37}Ar		$^{40}\text{Ca}(n, \alpha)$	
^{38}Ar			$^{37}\text{Cl}(n, \gamma, \beta)$
^{39}Ar		$^{42}\text{Ca}(n, \alpha)$	
^{40}Ar	$^{40}\text{K}(n, p)$		

5.2 Analytical Methods

In this study, isotopic analyses were obtained from minerals separated from bulk rock samples. Chemical analyses of monazites were carried out on in situ grains in thin section.

5.2.1 Mineral Separation Procedures

Zircon Separation for U-Pb Geochronology

Separation of minerals from bulk rock samples was undertaken at NORDSIM at the Naturhistoriska Riksmuseet, Stockholm, Sweden. The separation process involved crushing the samples in a jaw crusher before grinding them to an average diameter of 0.45 mm in a disk mill. The samples were then passed over a Wilfley table, set at an angle of 17°, to separate the heavy mineral fraction. A second phase of separation, with the Wilfley table set at an angle of 5°, allowed for collection of the dominantly zircon-bearing fraction. Magnetic separation removed any remaining magnetite before treatment with oxalic acid in a sonic bath removed any iron in solution coating the grains. The separates were finally washed in ethanol and placed in an oven overnight to dry. Individual zircons were picked from the heavy separate using tweezers under a binocular microscope and set in a 25 mm epoxy resin mount. The mounts were then polished to give approximately equatorial cross-sections of the grains before being coated with *c.* 30nm of gold. Backscattered electron (BSE) and cathodoluminescence (CL) images were captured for all grains to give information on physical condition and nature of any growth zoning respectively.

Monazite identification for U-Th-total Pb Geochronology

Monazite crystals were not separated from their host rock but instead analysed in situ. Prior to this however, they were identified using an SEM at the University of Oxford, based on their high backscatter coefficient and EDS spectrum. Their coordinates within the sample were recorded for transfer to the electron microprobe. During this stage the crystals were

also assessed for compositional zonation, particularly with regards to Th content.

Sulphide Separation for Re-Os Geochronology

Separation of sulphide minerals from bulk rock samples was completed at the University of Oxford and involved cutting samples into centimetre-scale chips that were then lightly crushed with a percussion mortar. Crushing was kept to a minimum in order to preserve whole sulphide crystals as much as possible. Individual sulphide crystals were subsequently picked manually using tweezers before being cleaned in acetone in a sonic bath.

Mica Separation for $^{40}\text{Ar}/^{39}\text{Ar}$ Geochronology

Biotite crystals were separated from bulk rock samples at the University of Oxford and University of Manchester. The process initially involved cutting samples into centimetre-scale chips before lightly crushing them with a percussion mortar. The majority of samples containing fine-grained biotite, were subsequently sieved to a size fraction of 63–250 μm . Highly magnetic minerals, such as magnetite and pyrrhotite were manually separated using a hand held magnet. The remaining sample material was then treated with heavy liquids in order to concentrate the required micas. Initially bromoform (CHBr_3 , $\rho = 2.8\text{g}/\text{cm}^3$) was used to remove the heaviest (Fe-oxides, etc.) and lightest (quartz, feldspar) mineral fractions. A second treatment with di-iodomethane (CHI_2 , $\rho = 3.3\text{g}/\text{cm}^3$) further concentrated the micas, and other minerals of similar density, by removing any remaining light minerals. These mineral separates were finally passed through a Frantz isodynamic separator to further concentrate biotite. Finally, manual picking with tweezers under a binocular microscope was used to obtain pure biotite fractions. Samples with significantly coarser biotite were not subjected to this process. Instead biotite crystals were manually picked with tweezers immediately following crushing.

5.2.2 Analytical Procedures

U-Pb Zircon SIMS Analysis

Analysis of separated zircons was completed using the NORDSIM Cameca IMS 1280 large-format ion microprobe secondary ionization mass spectrometer (SIMS) at the Naturhistoriska Riksmuseet in Stockholm, Sweden. Analytical procedures were similar to those described by [Whitehouse *et al.* \(1997, 1999\)](#) and [Whitehouse & Kamber \(2005\)](#) and briefly summarized here.

A *c.* 8 nA defocused O_2^- primary beam was used to project the image of a 150 μm aperture on to the sample, producing an elliptical, flat-bottomed crater with an approximately 20 μm long axis. An energy window of 60 eV was used in the secondary ion beam optics with energy adjustments made using the $^{90}\text{Zr}^{16}\text{O}$ peak. U-Pb analyses with a mass resolution ($M/\Delta M$) of *c.* 5000 were performed using a peak switching routine, with a single ion-counting electron multiplier (EM) as the detection device. Mass calibration was maintained using the automatic routine in the Cameca CIPS software. Pb/U calibration was performed relative to Geostandards zircon 91500 with an accepted age of 1065.4 ± 0.3 Ma (1σ) and Pb and U concentrations of *c.* 15 and 80 ppm respectively ([Wiedenbeck *et al.*, 1995](#)). Data reduction was performed using Isoplot/Ex v.3 ([Ludwig, 2003](#)).

U-Th-Total Pb EPMA Analysis

Quantitative analysis of identified monazites was completed at the University of Oxford using a JEOL JXA-8800R electron probe microanalyser (EPMA) operating at 15 kV and 60 nA, to allow for optimal spatial resolution of approximately 0.5 μm . The EPMA is equipped with four wavelength-dispersive spectrometers. X-ray line, spectrometer crystal, concentration standard and counting time details for all analysed elements are summarised in table [5.2](#). Corrections for significant peak overlaps were made using the factors listed in table [5.3](#).

In addition to the concentration standards listed in table [5.2](#), an internal age standard,

Table 5.2: Elemental analytical conditions for EPMA analysis

Element	Standard	Oxide (wt %)	X-Ray Line	Crystal	Counting Time (s)	
					Peak	Background
P	CePO ₄	30.19	K α	TAP	30	15
Si	CaSiO ₃	50.91	K α	TAP	30	15
Al	Al ₂ O ₃	100	K α	TAP	30	15
Ti	Ti metal	166.81	K α	PETJ-1	30	15
Ca	Apatite-32	55.83	K α	PETJ-2	30	15
Mn	Mn metal	129.12	K α	LIF	30	15
Fe	Fe metal	128.65	K α	LIF	30	15
Pb	PbCrO ₄	69.06	M β	PETJ-2	600	300
U	UO ₂	100	M β	PETJ-1	300	150
Th	ThP ₂ O ₇	65.04	M α	PETJ-1	300	150
Y	Y metal	126.99	L α	TAP	30	15
La	LaPO ₄	69.66	L α	LIF	30	15
Ce	CePO ₄	69.81	L α	LIF	30	15
Pr	PrPO ₄	69.91	L β	LIF	30	15
Nd	NdPO ₄	70.33	L β	LIF	30	15
Sm	SmPO ₄	71.07	L β	LIF	30	15
Eu	EuPO ₄	71.26	L α	LIF	100	50
Gd	GdPO ₄	71.86	L α	LIF	30	15
Tb	TbPO ₄	72.05	L α	LIF	100	50
Dy	DyPO ₄	72.44	L α	LIF	100	50
Ho	HoPO ₄	72.69	L β	LIF	100	50
Er	ErPO ₄	72.94	L α	LIF	100	50
Tm	TmPO ₄	73.11	L α	LIF	100	50
Yb	YbPO ₄	73.52	L α	LIF	100	50
Lu	LuPO ₄	73.71	L α	LIF	100	50

sample DLB-22A, was used to standardize the age distribution by applying a correction factor to ThO₂ concentrations. DLB-22A is garnet-cordierite pelitic hornfels from the inner aureole of the eastern Bushveld complex near the Steelpoort pericline and contains numerous large monazites. Since the monazites grew during thermal metamorphism resulting from the intrusion of the Bushveld complex, its intrusion age of 2057.5 ± 4.2 Ma (Harmer & Armstrong, 2000) is a reliable measure of the timing of monazite growth.

Concentration errors and detection limits were calculated using the Poisson (counting) statistics approach of Ancey *et al.* (1978) with individual age errors calculated by propagating concentration errors through equation 5.10 as per Montel *et al.* (1996). Since all monazites are assumed to belong to a single population, measurements of multiple monazites can be

Table 5.3: Peak overlap correction factors in EPMA analysis

Interference	Factor
Y on P	0.04792
U on Pb	0.00081
Th on U	0.00537
Sm on Tm	0.18961
Mn on Dy	0.13014
Ce on Gd	0.11206
Pr on Eu	0.26179
Tb on Er	0.05504
Ho on Lu	0.15632

treated as making multiple measurements of a single homogeneous domain. A final age and associated uncertainty can then be obtained using population statistics, as used by [Williams & Jercinovic \(2002\)](#). Errors associated with the final age are given as at a 95% confidence interval, as recommended by [Lisowiec \(2006\)](#), unlike [Williams & Jercinovic \(2002\)](#) who simply used the standard error on the mean.

Re-Os Sulphide TIMS Analysis

Preparation and analysis of separated sulphide crystals was completed by Dr Dave Selby at Durham University and followed the procedure described by [Selby *et al.* \(2009\)](#), and references therein) and summarized here.

Sample material was accurately weighed and loaded into a carius tube ([Shirey & Walker, 1995](#)) together with a known amount of ^{185}Re - ^{190}Os tracer solution. The sample and tracer solution were digested using the carius tube method and equilibrated using a mix of 11N HCL (3 ml) and 15.5N HNO₃ (8 ml) at 220°C for 48 hours. Solvent extraction with CHCl₃, microdistillation and anion column and single-bead chromatography methods were used to isolate and purify Os and Re ([Selby & Creaser, 2001, 2003](#)). The pure Re and Os samples were loaded onto Ni and Pt filaments respectively ([Selby *et al.*, 2007](#)). Isotopic measurements were made using a Thermo Electron TRITON negative thermal ionization mass spectrometer (TIMS) via ion counting and using a secondary electron multiplier in peak-hopping mode for Os, and static Faraday collection for Re ([Creaser *et al.*, 1991](#)). Data reduction and age

calculation was completed using Isoplot/Ex v.3 (Ludwig, 2003)

$^{40}\text{Ar}/^{39}\text{Ar}$ Analysis

Separated biotite mineral fractions were weighed, wrapped in Al foil and vacuum sealed in quartz ampules (8 mm $\times$ 65 mm) along with the neutron fluence monitors Hb3gr ($t = 1073.6 \pm 5.3$ Ma; Jourdan *et al.* (2006)) placed at the top and bottom of the samples. Nuclear irradiation was carried out at the RODEO facility in Petten, Netherlands, on 10 May 2011, without Cd-shielding, for 26 hours in the high flux reactor. The J values determined from Hb3gr were 0.007266 ± 0.000052 (1σ) for vial 1 (all samples except DiArBt1-2) and 0.007287 ± 0.000032 (1σ) for vial 2 (sample DiArBt1-2 only).

Laboratory analytical conditions were similar to those described by Burgess & Turner (1998); Fernandes *et al.* (2000); Burgess *et al.* (2004); Flude (2005); Hennig *et al.* (2008) and Chalapathi Rao *et al.* (2011). The irradiated samples were step-heated in a Ta-resistance furnace over the temperature interval 500–1400°C, with steps of between 25–200 °C and a step duration of 30 minutes. During extraction, Ar gas was purified using a Zr-Al getter (SAES NP10) operating at 250°C. Prior to inlet to the mass spectrometer, the Ar gas is further purified for an additional five minutes by another Zr-Al getter (SAES ST 172), also operating at 250°C. The Ar gas is then sublimed on to a charcoal cold finger, held at liquid nitrogen temperature (-196°C) for a further five minutes. The charcoal finger is then heated to approximately 100°C, over five minutes, thereby releasing the gas.

Ar isotopic measurements were made using the 15cm, 90° sector, MS1 mass spectrometer at the University of Manchester, equipped with a Baur-Signer ion source and Faraday detector. Isotopic measurements were made for ^iAr (where $i = 36, 37, 38, 39, 40$), two contaminant peaks at masses 35 (for chlorine) and 41 (for hydrocarbons), and six zero masses over eleven cycles, lasting approximately 30 minutes in total. Final abundances of individual isotopes and isotopic ratios were determined by regressing the absolute abundance of ^{40}Ar and $^i\text{Ar}/^{40}\text{Ar}$ ratios back to the mass spectrometer inlet time, with the possibility of excluding any outlying data points. Plateaux were constructed and final weighted average ages and uncertainties

were calculated using Isoplot/Ex v.3 (Ludwig, 2003).

Blanks were not routinely measured during analysis but instead a daily air calibration was performed. This process also checks mass discrimination and ^{40}Ar sensitivity. The calibration process involves taking a known volume of air (a 1.071cm^3 aliquot from a 294.1cm^3 bottle containing air at approximately 7.5×10^{-3} torr) and analysing it as described above.

Interfering Ar isotopes produced during irradiation were monitored and corrected from K_2SO_4 ($(^{40}\text{Ar}/^{39}\text{Ar})\text{K}$, $(^{38}\text{Ar}/^{39}\text{Ar})\text{K}$ and $(^{37}\text{Ar}/^{39}\text{Ar})\text{K}$) and CaF_2 ($(^{39}\text{Ar}/^{37}\text{Ar})\text{Ca}$, $(^{38}\text{Ar}/^{37}\text{Ar})\text{Ca}$ and $(^{36}\text{Ar}/^{39}\text{Ar})\text{Ca}$). Furthermore, Ca-derived ^{36}Ar was corrected for using ^{37}Ar , which is entirely produced from ^{40}Ca and is not abundant in the atmosphere. Cl-derived ^{36}Ar was corrected for using Cl-derived ^{38}Ar , although in practice this is often unnecessary since the decay of ^{36}Cl (which is produced during irradiation) to ^{36}Ar has a half life of 3×10^5 yrs, and Cl-contents of many minerals is low. Therefore, if analysis is carried out within a year of irradiation (as is the case here) the amount of Cl-derived radiogenic ^{36}Ar formed is negligible. Ca-derived ^{39}Ar was also corrected for using Ca-derived ^{37}Ar , although in minerals with a K/Ca ratio greater than around 0.1 (as in biotite), the contribution from ^{42}Ca is very low. Finally, ^{40}Ar produced from ^{40}K was corrected for using the air calibration process described above.

5.3 Geochronology of Damang Host Rocks

The age of deposition of the Tarkwaian System is difficult to determine directly due to the the lack of any datable authigenic minerals. The detrital zircon study of Pigois *et al.* (2003) constrained the maximum age of deposition to 2133 ± 4 Ma. In contrast, the volcanic rocks of the Birimian basement and the intrusive Diorite Porphyry are ideally suited to a geochronological study.

As mentioned in section 3.1.3, the Diorite Porphyry is not well constrained with the rest of the geological history at Damang. Specifically, it is not known whether it is of Birimian age or is in fact post-Tarkwaian. If the latter were true, it could have profound implications for

the source of mineralizing fluids and/or heat generation driving their circulation, especially given the increasing recognition of mineralized granitoids elsewhere in Ghana (Yao *et al.*, 2001; Brownscombe, 2009). In order to put the age calculated for the Diorite Porphyry in context, an age for a Birimian Volcaniclastic unit has also been constrained.

5.3.1 AWADi - Diorite Porphyry

The Diorite Porphyry, sample AWADi, contained zircons that were highly cracked and exhibited numerous large inclusions. However, of those that were suitable for analysis the zircons were largely homogeneous with little compositional zoning (Fig. 5.5A). A total of 14 zircon grains were selected for analysis. Of these, four were strongly discordant along a discordia running within error of the origin. This Pb-loss can be attributed to loss during recent uplift and near-surface processes, and these grains were therefore discounted. The remaining 10 concordant zircon grains indicate an age of formation of 2164.6 ± 8.0 Ma (2σ , MSWD=1.6) (Fig. 5.6A). Tabulated results are presented in Table 5.4.

5.3.2 AWABv - Birimian Volcaniclastic

The Birimian Volcaniclastic sample AWABv contained euhedral zircons that showed extensive zoning, with many exhibiting distinct core and rim domains (Fig. 5.5B). 18 individual zircon grains were selected for analysis including 3 grains with core and rim zones, giving a total of 21 analysis points. 8 of these were discordant while the remaining 13 concordant zircon grains indicate a maximum formation age of 2178 ± 9.3 Ma (2σ , MSWD=1.8) (Fig. 5.6B). The core domains produce no discernible difference in age and are therefore not xenocrystic, but are instead interpreted to represent a short break in growth conditions and/or a change in magma system dynamics. Tabulated results are presented in Table 5.4.

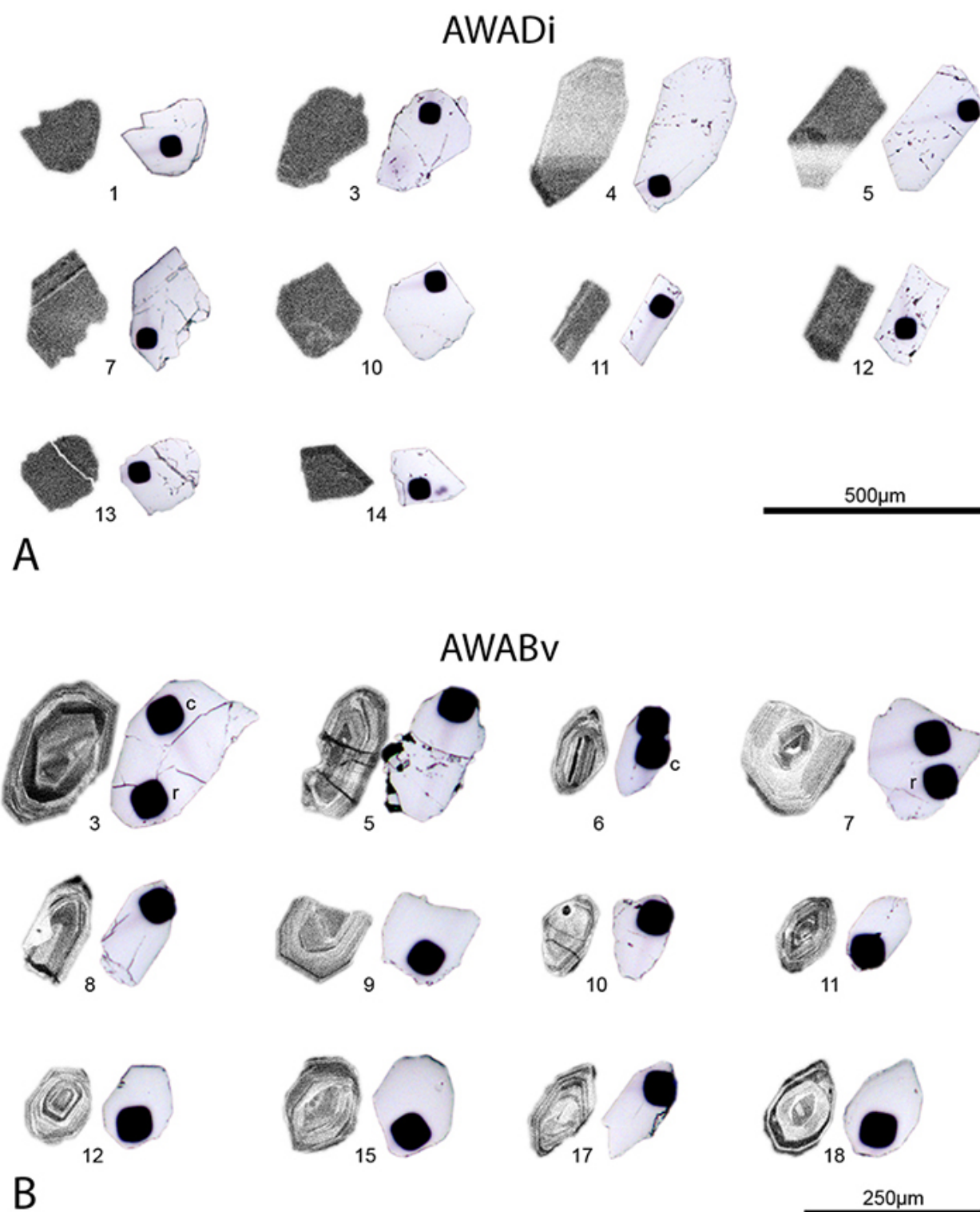


Figure 5.5: Cathodoluminescence (CL) (left) and optical (right) images of analysed zircons from for the Diorite Porphyry (A), and the Birimian Volcaniclastics (B). Optical images are taken post-analysis and the sputtering spot can be seen. The actual analysis spot is $\frac{1}{4}$ of the size and located in the centre of the sputtering spot. The CL images show that the Diorite Porphyry zircons are compositionally uniform, while the in the Birimian Volcaniclastic they are zoned, often with distinct core and rim domains.

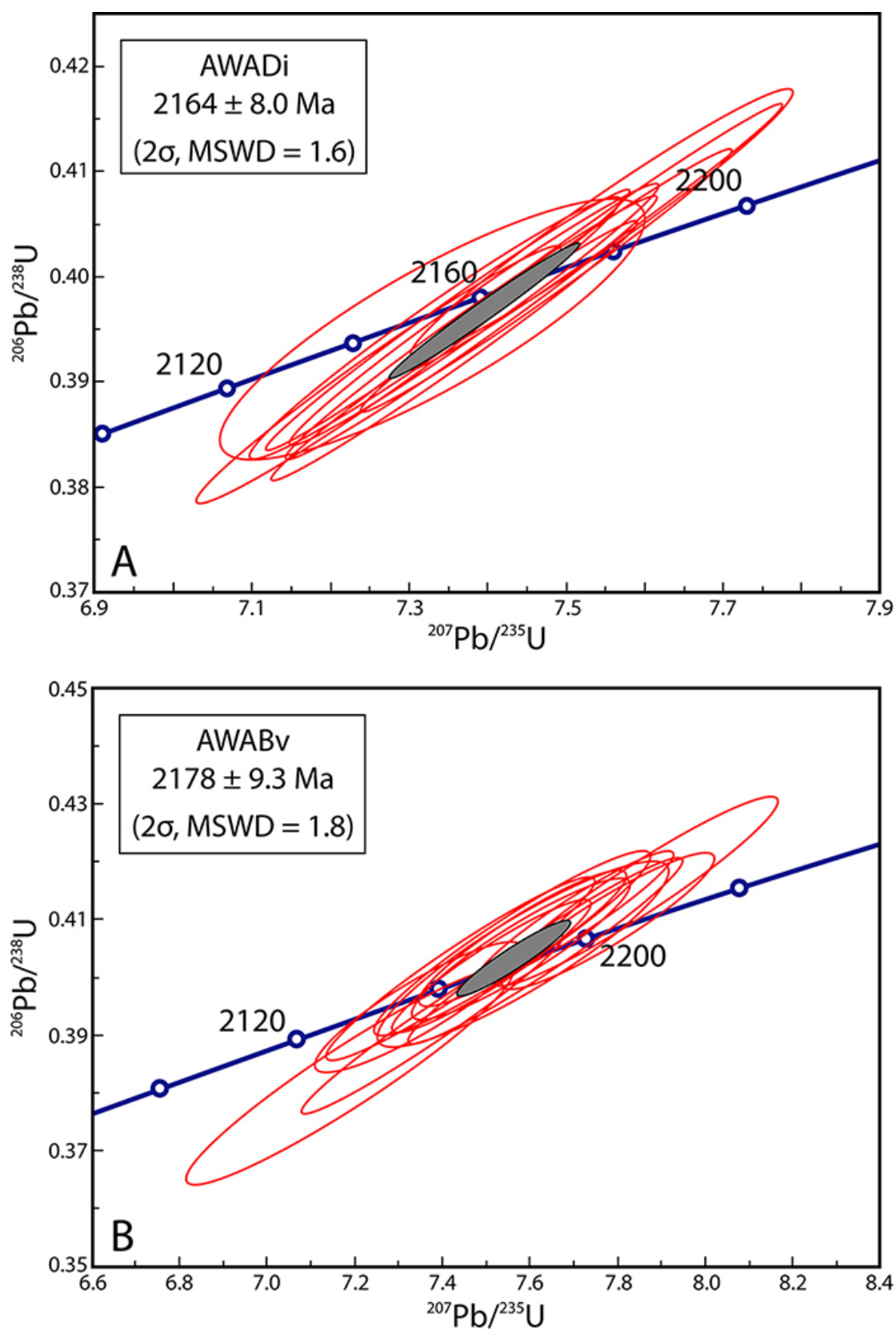


Figure 5.6: U/Pb concordia plots for the Diorite Porphyry (A), and the Birimian Volcaniclastics (B).

Table 5.4: Results of SIMS U-Pb zircon analysis

Spot I.D. ^a	[U] ppm	[Pb] ppm	Th/U	$f_{206}\%$	$^{207}\text{Pb}/^{206}\text{Pb}$	$\sigma\%$	$^{207}\text{Pb}/^{235}\text{U}$	$\sigma\%$	$^{206}\text{Pb}/^{238}\text{U}$	$\sigma\%$	ρ^b
Sample AWADi											
n3262-1	539.3	278.4	0.60	0.02	0.1347	0.28	7.3310	1.31	0.3951	1.28	0.98
n3262-3	619.3	330.8	0.65	0.03	0.1353	0.30	7.5471	1.31	0.4051	1.28	0.97
n3262-4	972.2	523.9	0.81	0.05	0.1349	0.18	7.3492	1.29	0.3961	1.28	0.99
n3262-5	458.6	230.3	0.55	0.05	0.1353	0.26	7.2585	1.31	0.3906	1.28	0.98
n3262-7	891.5	465.4	0.69	0.06	0.1362	0.19	7.3587	1.30	0.3929	1.28	0.99
n3262-10	803.4	420.6	0.68	0.02	0.1355	0.20	7.3798	1.31	0.3953	1.29	0.99
n3262-11	809.8	427.7	0.73	1.02	0.1417	0.47	7.3313	1.51	0.3950	1.28	0.85
n3262-12	1006.8	561.2	0.93	0.03	0.1358	0.19	7.4743	1.29	0.3997	1.28	0.99
n3262-13	660.1	343.3	0.64	0.03	0.1352	0.23	7.3830	1.30	0.3964	1.28	0.98
n3262-14	1078.5	593.0	0.82	0.02	0.1355	0.17	7.5391	1.29	0.4038	1.27	0.99
Sample AWABv											
n3267-3r	496.5	259.7	0.46	0.37	0.1387	0.29	7.7053	1.19	0.4106	1.13	0.95
n3267-3c	165.8	83.1	0.38	0.29	0.1369	0.40	7.5142	1.23	0.4021	1.16	0.94
n3267-5	107.5	52.9	0.28	0.06	0.1362	0.55	7.5806	1.35	0.4050	1.23	0.91
n3267-6c	106.0	53.0	0.34	0.15	0.1362	0.48	7.5849	1.28	0.4068	1.17	0.92
n3267-7r	114.9	58.9	0.39	0.08	0.1382	0.49	7.7800	1.28	0.4098	1.17	0.92
n3267-8	143.7	71.7	0.41	0.58	0.1389	0.54	7.3432	1.31	0.3950	1.16	0.88
n3267-9	98.7	48.6	0.30	0.11	0.1369	0.78	7.5847	1.80	0.4040	1.61	0.90
n3267-10	96.5	45.6	0.36	0.15	0.1371	0.66	7.1895	2.12	0.3831	2.00	0.95
n3267-11	214.9	105.5	0.40	0.07	0.1370	0.37	7.3783	1.65	0.3920	1.61	0.97
n3267-12	177.3	89.1	0.35	0.07	0.1374	0.42	7.6379	1.68	0.4046	1.62	0.97
n3267-15	121.4	58.7	0.25	0.08	0.1350	0.48	7.4437	1.68	0.4016	1.61	0.96
n3267-17	224.9	120.4	0.39	1.05	0.1444	0.39	7.8426	1.69	0.4148	1.61	0.95
n3267-18	111.2	55.3	0.31	0.12	0.1360	0.54	7.5613	1.70	0.4059	1.61	0.94

a: 'r' and 'c' indicate rim and core spots respectively, obtained on the same grain. b: error correlation between $^{207}\text{Pb}/^{235}\text{U}$ and $^{206}\text{Pb}/^{238}\text{U}$.

5.3.3 Discussion of U-Pb zircon results

The results of U-Pb zircon analysis clearly show that the Diorite Porphyry is of Birimian age and was intruded into the Birimian Supergroup prior to deposition of the Tarkwaian System. This implies that the Diorite Porphyry is akin to the Belt-type granitoids. Additionally, it precludes the possibility of it being either a direct source or modifying influence on the later gold mineralization.

5.4 Geochronology of the Damang Regional Metamorphic Event

5.4.1 Monazite Petrography

The Tarkwa Phyllite contains an abundance of very small monazite grains, typically 10-20 μ m in size. This size range means that in practice, one analysis can be performed per monazite grain. Grains are generally subhedral (Fig. 5.7A,C-D), but some are partially euhedral (Fig. 5.7B) and all occur as matrix phases interstitial amongst quartz, plagioclase and muscovite (Fig. 5.7A-B). This is important since monazites that occur in different microstructural settings, such as inclusions within porphyroblast phases, may be open to a different equilibration volume and therefore record a different part of the rock's metamorphic history (Montel *et al.*, 2000). Backscattered electron imaging indicates that grains are homogeneous and do not contain distinct core or overgrowth zones (Fig. 5.7C-D). Petrography indicates that they have the same growth history, which is likely due to a single, common growth event.

Within mineralized Tarkwa Phyllite samples, the monazites are highly altered, becoming anhedral in shape and surrounded by an irregular, broadly concentrically zoned domain of Apatite–Allanite–Epidote (Fig. 5.7E-F). This phenomenon was studied in detail by Finger *et al.* (1998) and further documented by Fraser (2000), with similar reaction textures noted by Dini *et al.* (2004) and Rasmussen & Muhling (2009). Finger *et al.* (1998) suggested that the inner apatite zone is a direct replacement of monazite, while the displaced REEs diffuse

outwards to form the surrounding allanite corona. They also described a chemical mass balance that suggests the breakdown reaction is initiated by an influx of hydrothermal Ca, Fe, Si and Al. Figure 3.47 shows that of these elements, there is only evidence for Fe addition. However, Ca, Si and Al, while not introduced by a fluid phase, are potentially remobilized during the alteration process. Furthermore, this agrees in principle with Spear (2010), whose thermodynamic calculations suggest the monazite–allanite transition is a function of the host rock’s CaO (and Al₂O₃) content, with higher Ca-contents favouring allanite stability. This in turn agrees with the summary of observations for the first appearance of metamorphic monazite reported by Spear & Pyle (2002).

Ultimately, these reaction textures support the assertion that the monazites are metamorphic in origin and not related to a hydrothermal event since the mineralization process is clearly monazite-destructive. Of equal importance is that Finger *et al.* (1998) also suggested that the relic monazite grains preserve their U-Th-Pb characteristics and are therefore still viable for chemical dating of the pre-mineralization metamorphic history.

5.4.2 Monazite Chemistry

The final calculation of a monazite U-Th-total Pb chemical age is best conducted using repeated measurements of a single, homogeneous domain, as described in section 5.1.2. Since the sizes of monazite grains in the Tarkwa Phyllite are such that only one analysis spot can be placed on each grain, this translates to making measurements of a homogeneous population. U-Th-total Pb analysis has been carried out on monazites from three different samples of the Tarkwa Phyllite and in order to assess whether all of these monazites belong to a single population, both within each sample as well as between samples, requires a detailed investigation of the monazite compositions.

The general formula of monazite is (LREE)PO₄, where the dominant LREE is typically Ce, although it can undergo several cation substitutions:

- Th₁Si₁REE₋₁P₋₁ – huttonite (ThSiO₄),

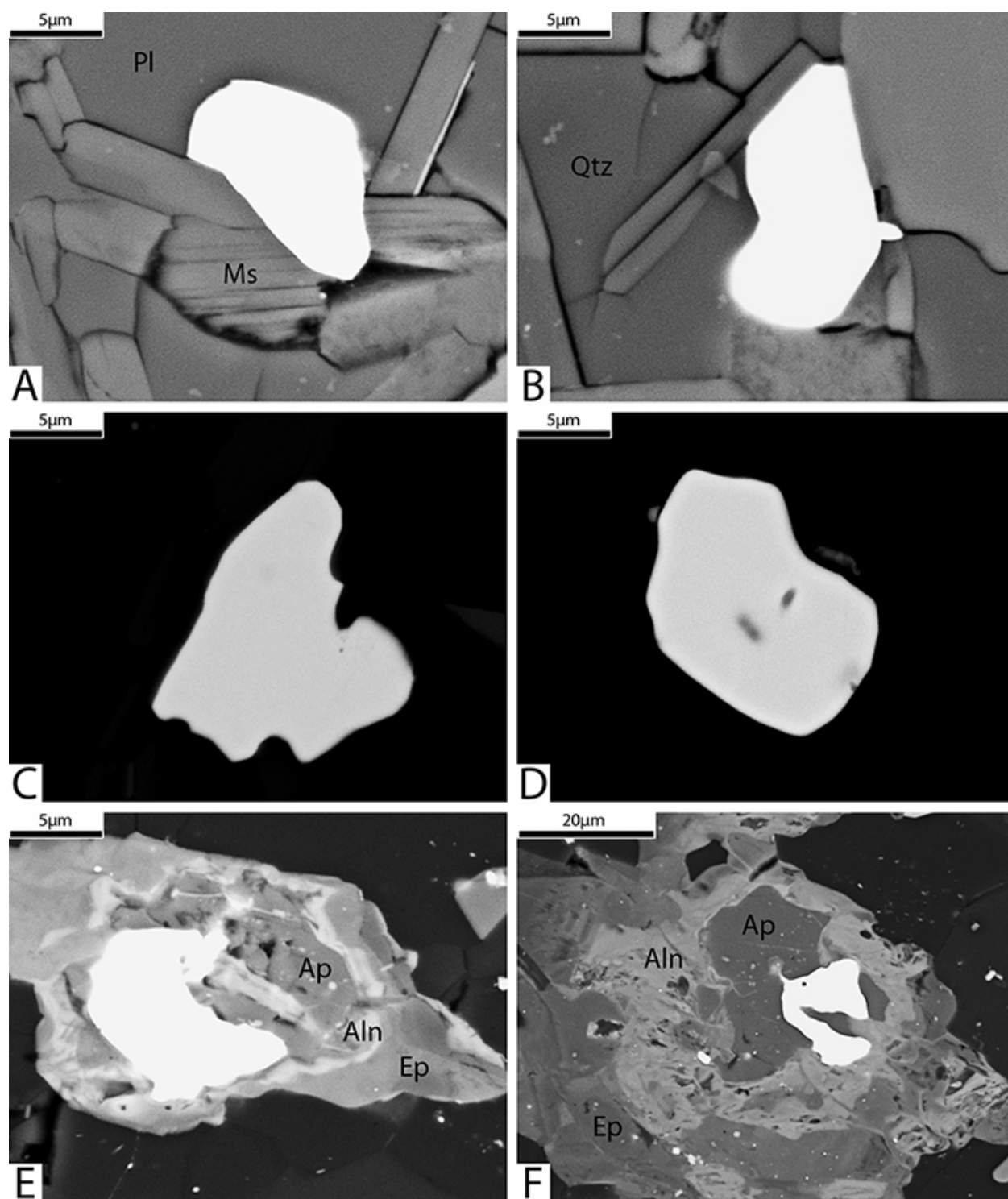


Figure 5.7: Backscattered electron images of monazites from the Tarkwa Phyllite. A–D, unaltered monazites in sample AWDP1. A–B have the contrast set such that their relationship to major matrix phases can be seen. C–D are contrasted to show the lack of internal compositional variation. E–F, hydrothermally altered monazites in sample AWDP2. Relic monazite is still visible surrounded by zones of apatite, allanite and epidote.

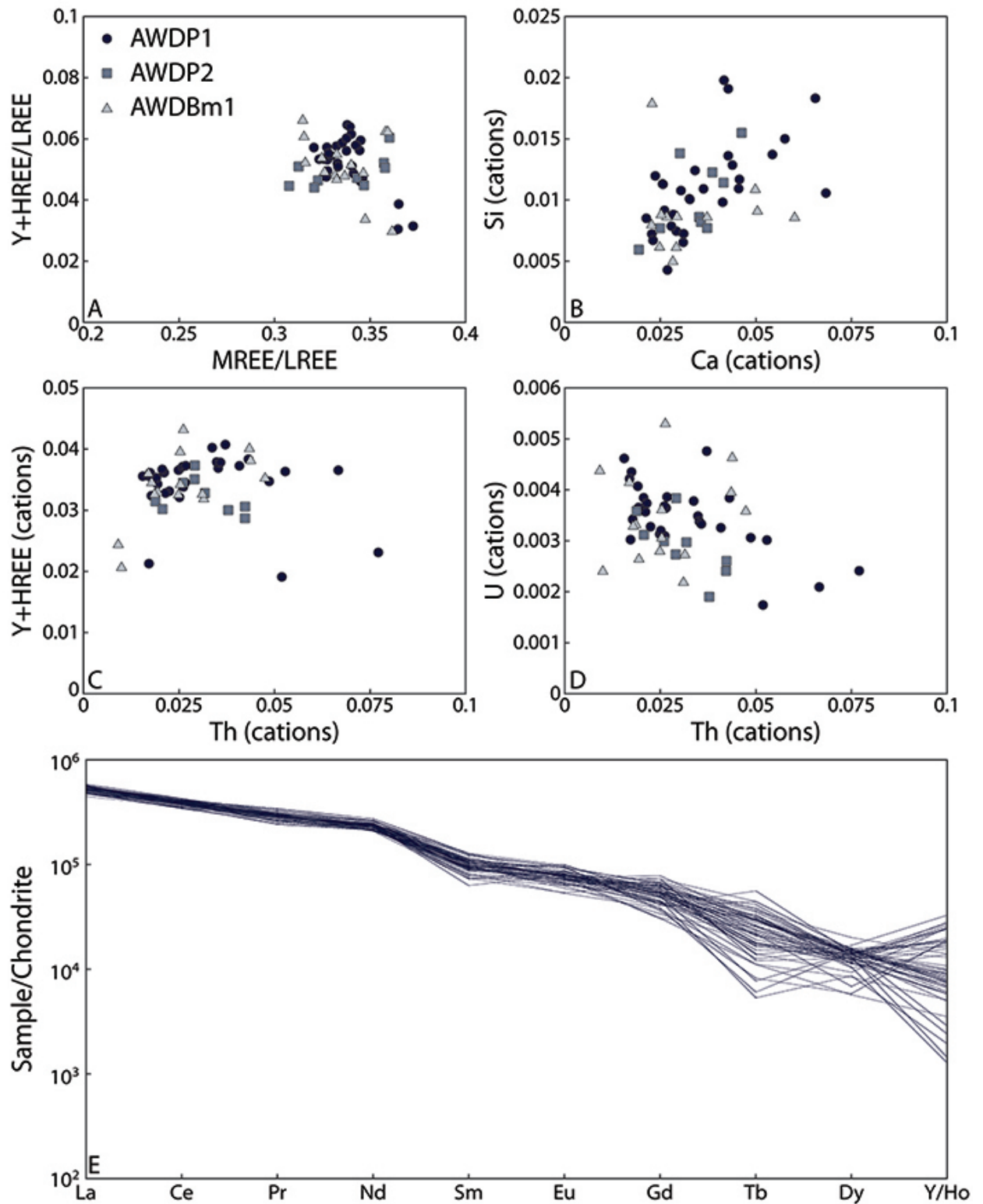


Figure 5.8: Plots of common compositional substitutions (A–D) and REE patterns (E) for Damang monazites, showing that all grains belong to a single, homogeneous population. In (E), the absent element Pm, is omitted from between Nd and Sm. Y is plotted in place of Ho where concentrations are below detection limits.

- $\text{Th}_1\text{Ca}_1\text{REE}_{-2}$ – cheralite ($\text{ThCa}(\text{PO}_4)_2$),
- U_1Th_{-1} ,
- $\text{YMHREE}_1\text{LREE}_{-1}$ (YMHREE = Yttrium and/or medium- and /or heavy REE).

These substitutions, as measured for the Tarkwa Phyllite monazites, are represented in figure 5.8A-D. These figures clearly show that, despite a small degree of scatter, all three samples are compositionally uniform and indistinguishable. Similarly, since REEs can substitute for each other, total REE trends are useful for constraining monazite populations. Figure 5.8E shows that all measured monazites exhibit identical REE trends. Overall, the compositions of monazites from all three samples of the Tarkwa Phyllite are within the same range, and all monazites can therefore be treated as a single homogeneous population.

5.4.3 Results of Monazite U-Th-Total Pb Analysis

U-Th-total Pb analysis was carried out on a total of 53 monazite grains (excluding grains discounted for having bad analysis totals or stoichiometry) from three samples of Tarkwa Phyllite, two of which are unmineralized (AWDP1 and AWDBm1) and one that is mineralized (AWDP2). Statistical treatment, as described in section 5.2.2, results in a final age and uncertainty of 2005 ± 26 Ma (95% C.I., MSWD = 2.1), as shown using the histogram approach of Montel *et al.* (1996) in figure 5.9A. The total probability histogram (the thick line in figure 5.9A) defines a function that it may be possible to fit two sub-populations to. However, as discussed in the preceding section, there is no petrographic or geochemical basis on which to define two separate populations. Therefore, there would be no statistical significance to defining two age groups in this case.

Results are also presented according to the isochron method, after Suzuki & Adachi (1991) (Fig. 5.9B), although this technique was not used to calculate the final age. Finally, a weighted average approach calculated in Isoplot/Ex v.3 (Ludwig, 2003) is shown in figure 5.9C. Tabulated results are presented in table 5.5.

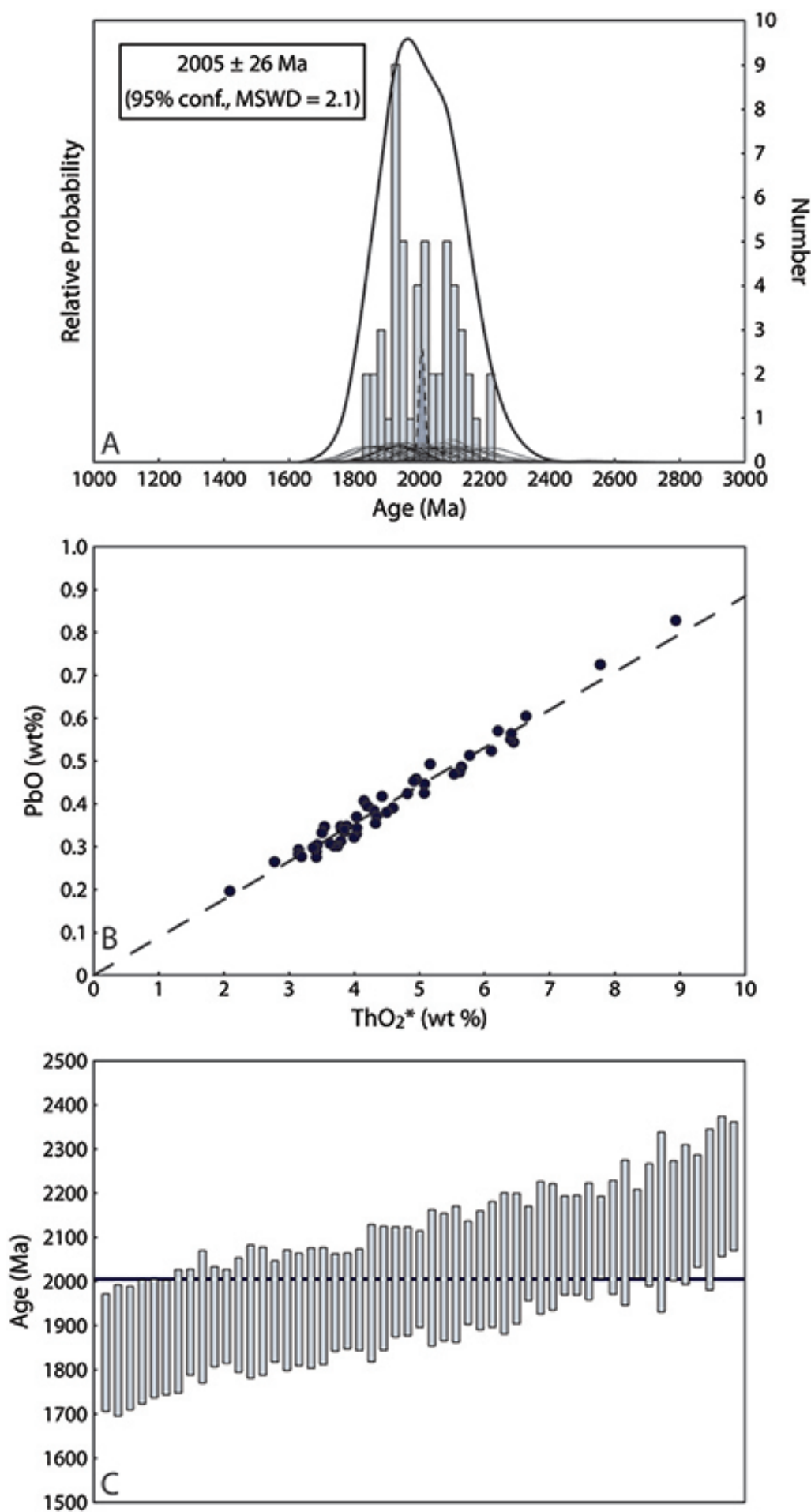


Figure 5.9: Results of U-Th-total Pb chemical dating of monazites from the Tarkwa Phyllite. A shows a histogram approach similar to [Montel *et al.* \(1996\)](#) in which the small bell-curves are the probability density functions for each analysed grain, the larger thick line is the sum of all of these functions, the smaller dashed line is the weighted average, and the block histogram gives the number of monazites within each age range. B presents the results in an isochron form, after [Suzuki & Adachi \(1991\)](#). C is a weighted average plot calculated in Isoplot/Ex v.3 ([Ludwig, 2003](#)). Samples are plotted in order of increasing age.

Table 5.5: Results of EPMA U-Th-Total Pb monazite analysis. Errors are 1σ . ThO₂ values are corrected using internal standard DLB-22A.

Grain	PbO (ppm)	±	ThO ₂ (ppm)	±	UO ₂ (ppm)	±	Age (Ma)	±
Sample AWDP1								
4	4860	134	42464	295	3680	116	1958	57
6	6040	149	53584	331	3320	110	2063	54
7	4250	125	35965	271	3910	119	1908	60
8	4470	128	36333	273	3780	117	2001	61
9	3430	112	27146	236	3490	113	1935	68
10	4930	135	37068	275	3750	117	2160	63
11	4070	122	26193	232	3920	120	2215	72
12	4580	130	33626	262	4100	122	2100	64
13	5700	145	49076	317	3360	111	2082	56
14	5700	145	49076	317	3360	111	2082	56
15	3400	112	22317	214	4230	124	2010	72
17	8280	175	78910	402	2680	99	2099	46
25	2930	104	18091	192	3470	112	2110	81
27	5130	137	37663	278	5240	138	2020	58
2-1	3220	109	26297	232	3650	115	1839	66
2-2	5240	139	44643	302	4320	125	1953	55
2-3	3430	112	20100	203	4660	130	2052	73
2-4	3470	113	18553	195	5040	136	2077	74
2-5	3330	111	17091	187	4630	130	2153	79
2-6	2760	101	15450	178	5010	135	1844	73
2-9	3020	105	21005	207	4260	125	1863	70
2-10	3080	106	20421	204	4200	124	1933	72
2-11	3310	110	26627	233	3650	115	1871	66
2-12	3020	105	22071	213	4060	122	1849	69
2-14	2880	103	19081	198	4000	121	1920	74
2-15	3140	108	23741	220	3770	117	1888	69
2-16	3550	114	27184	236	4270	125	1873	64
2-17	3710	117	27457	237	4220	124	1944	66
2-18	7240	163	68629	375	2340	92	2109	50
Sample AWDP2								
2-2	4740	132	44803	303	3000	105	1921	56
2-3	4240	125	39813	285	2180	89	2000	62
2-4	3840	119	30909	252	3170	107	2026	67
2-5	3450	113	27457	237	3440	112	1940	67
2-6	3050	106	19185	198	3980	120	2017	76
2-7	3800	118	29183	244	4180	123	1924	64
2-8	3470	113	21571	210	3540	114	2215	78
2-9	3910	120	33182	261	3370	111	1937	63
2-11	4700	132	44784	303	2780	101	1932	57

5.4.4 Discussion of Monazite U-Th-Total Pb Analysis

The monazite petrography, chemistry and U-Th-total Pb discussed indicate that all monazites in the Tarkwa Phyllite represent a single population grown together during a common metamorphic event. The remaining question, however, is during which part of the regional metamorphic cycle did the monazites grow?

The presence of monazite in metamorphic rocks has received a great deal of attention, with many different proposals for pressure-temperature conditions of monazite producing reactions. These include formation during garnet-grade (Catlos *et al.*, 2001), staurolite-in (Smith & Barreiro, 1990; Kohn & Malloy, 2004) and aluminosilicate-in (Wing *et al.*, 2003) prograde reactions, in addition to during hydrothermal processes (Townsend *et al.*, 2000). Additionally, there are many recorded occurrences of detrital monazites stable through low-grade metamorphism and even higher grade conditions (Parrish, 1990; Suzuki *et al.*, 1994).

Recent studies suggest that REEs in metamorphic rocks are partitioned between monazite at very low-metamorphic grades, often as detrital grains, followed by allanite at moderate grades and finally metamorphic monazite at the highest grades (Janots *et al.*, 2008; Rasmussen & Muhling, 2009; Spear, 2010). The formation of metamorphic monazite is therefore intimately associated with the breakdown of allanite, which typically occurs close to, but ultimately independent from, the staurolite-in isograd (Tomkins & Pattison, 2007; Corrie & Kohn, 2008; Janots *et al.*, 2008). Although this is also a function of CaO (and Al₂O₃) content (Rasmussen *et al.*, 2006; Spear, 2010).

Peak metamorphic mineral assemblages in the Tarkwa Phyllite were described and discussed in sections 3.4 and 4.3.2 respectively, both of which clearly show that the Damang region reached staurolitegrade metamorphic conditions. This implies that monazite growth was at or very close to peak metamorphism and the U-Th-total Pb age they provide is therefore representative of these conditions.

5.5 Geochronology of the Damang Gold Mineralization Event

Accurately constraining the timing of gold mineralization at Damang requires the dating of a mineral phase that is not only produced during the mineralization process, but whose daughter isotope blocking temperature is as high or higher than the temperature of mineralization, such that any calculated age represents the true formation age. Re-Os analysis of sulphide minerals, whose formation is directly associated with gold mineralization, is a suitable technique and has been used extensively for this purpose (Cardon *et al.*, 2008; Selby *et al.*, 2009; Morelli *et al.*, 2010). Despite this, this approach has not been applied to gold mineralization in Ghana. Instead, the majority of mineralization timing constraints are determined indirectly, relative to structural or magmatic events (see section 2.5).

5.5.1 Results of Re-Os Sulphide Analysis

Pyrite and pyrrhotite crystals were separated from a total of 15 individual samples, covering both sediments and dolerites and all observed parageneses; disseminated pyrite and pyrrhotite, and pyrite replacing magnetite along bedding planes. 5 of these samples contained insufficient Re (<0.5 ppb) for the analysis. Three of the remaining samples produced associated errors that are too large (i.e. greater than the absolute determined value) and are therefore of no use for constructing an isochron. On assessment of the remaining combined data set, two samples clearly fall below the trend produced by the rest of the data when plotted on a $^{187}\text{Re}/^{188}\text{Os}$ vs. $^{187}\text{Os}/^{188}\text{Os}$ diagram (Fig. 5.10). This suggests that these two samples (DoROpr1 and DoROpy3) have undergone significant Os loss since formation.

The five useable samples still do not produce a reliable isochron, as the scatter about any regression line is greater than can be accounted for by analytical error alone. Instead, an ‘errorchron’ can be drawn using a robust regression algorithm (ignoring analytical error) inside of Isoplot/Ex v.3 (Ludwig, 2003). The age given by this errorchron is $2191 \pm 370/-1600$ Ma (Fig. 5.10), which is very poorly constrained, particularly at the young end of the age range. Tabulated results are presented in Table 5.6.

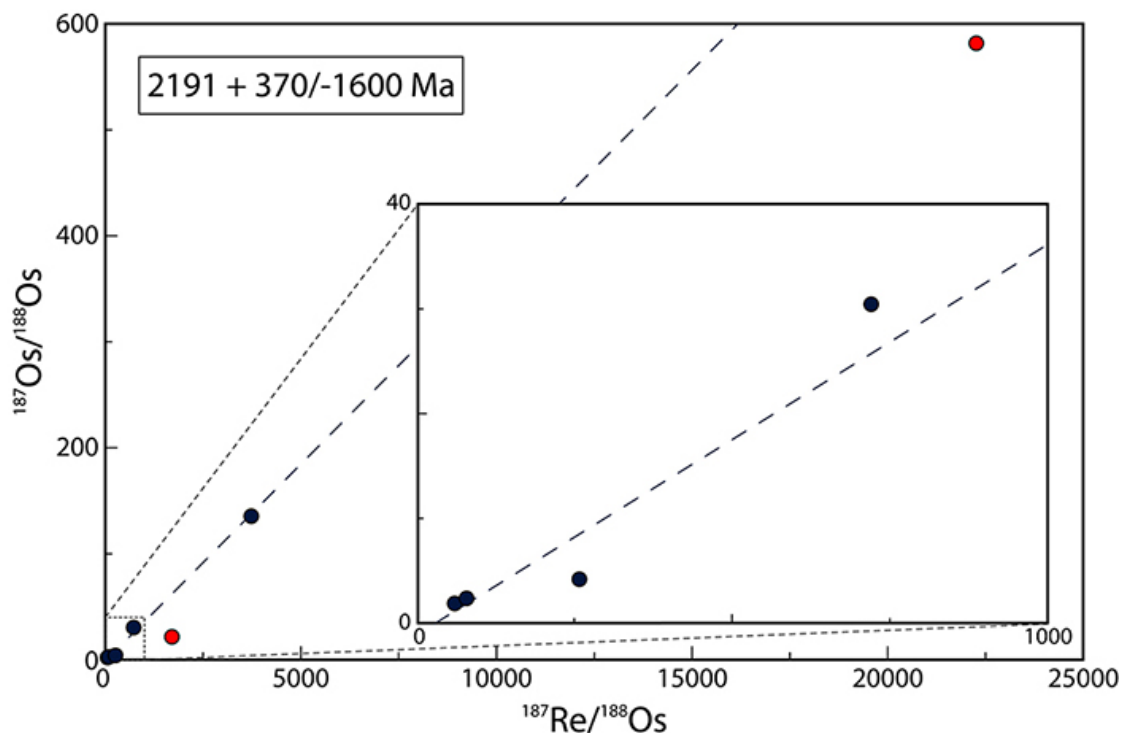


Figure 5.10: ‘Errorchron’ plot of Re/Os analysis of sulphide minerals from Damang. Red points are those discounted due to falling significantly below the trend of the other points. Inset is an enlargement of the lower left hand corner. Initial $^{187}\text{Os}/^{188}\text{Os}$ ratio is -1.1 ± 5.8 .

5.5.2 Discussion of Re-Os Sulphide Analysis

The poor age constraint and large degree of scatter in the data can be attributed to one or more processes. Loss of the daughter isotopes (in this case Os) is a common problem in many sulphide systems and certainly seems to have seriously affected some of the samples from Damang. In particular, the diffusion of Os in pyrrhotite is significantly greater than for pyrite resulting in pyrrhotite crystals commonly being isotopically reset (Brenan *et al.*, 2000; Morelli, 2008). Similarly, sulphide minerals, particularly pyrrhotite, have been shown to gain or lose the parent isotope (Re), which would also move data points off an isochron. Finally, sulphides can develop internal heterogeneity of isotope ratios, without any actual loss of either parent or daughter atoms (Barra *et al.*, 2003; Cardon *et al.*, 2008). The effect of this is that very large crystals, which are common at Damang, become broken up and not completely sampled during the sample preparation (exacerbated by the highly brittle nature of the sulphide minerals) and isotope separation stages. This means that a true isotope ratio is not obtained. This last process would affect each crystal differently, producing a wide

scatter of data and could explain the results seen at Damang.

Although the age calculated here is not well constrained, additional constraints can be placed on this range by other dating methods discussed in this chapter. In particular, U-Th-total Pb dating of monazites (section 5.4) of the pre-mineralization metamorphic event provides a maximum age constraint on the mineralization. Similarly, $^{40}\text{Ar}/^{39}\text{Ar}$ dating (section 5.6) of biotite from hydrothermal alteration assemblages associated with sulphide (and gold) mineralization can provide a minimum age constraint.

5.6 Thermal History of the Damang Region

The principle of assessing thermal histories through geochronology, termed thermochronology, relies on the fact that different radiogenic atoms diffuse through their host minerals at different rates. For a mineral such as zircon, diffusion is so slow that its ‘closure temperature’ (the temperature below which radiogenic daughter atoms are assumed to be retained by the mineral) is essentially the same as its formation temperature. By contrast, gaseous Ar is much more easily lost from its host minerals and closure temperatures are therefore considerably lower, and often lower than formation temperatures. Commonly quoted closure temperatures for K-bearing minerals are around 500°C for hornblende, 350°C for muscovite, 300°C for biotite and 150°C for K-feldspar (McDougall & Harrison, 1999).

In this study therefore, $^{40}\text{Ar}/^{39}\text{Ar}$ dating of biotite is a particularly useful tool since it allows not only for investigation of the regional cooling history, by recording the time at which the region cooled through around 300°C, but also places a minimum age constraint on mineralization, which occurred above the closure temperature for biotite (see chapter 4).

Biotite was selected for use over other K-bearing minerals because of its relative ease of separation and analysis. Both hydrothermal and metamorphic muscovite in the Tarkwaian sediments is extremely fine-grained, which make successfully obtaining a pure mineral separate very difficult. The abundant hornblende found in unmineralized dolerites was not used since hornblende has been known to produce rather irregular age spectra, despite apparently

Table 5.6: Results of TIMS Re-Os sulphide analysis

Sample	Bat	Lith	Para	Re	±	Os	±	^{187}Re	±	^{187}Os	±	Re/Os	±	Os/Os	±	rho	Model Age	% ^{187}Os
DoROpy2	2	Dol	diss py	0.66	0.01	21.50	2.65	0.41	0.02	16.77	0.18	727.50	78.91	30.30	3.28	0.99	2388.6	97.33
AWDDo5py	2	Dol	diss po	2.70	0.02	65.03	30.78	1.70	0.01	61.23	1.77	3730.64	1314.94	135.4	47.88	1.00	2127.8	99.39
AWDP4py	1	TP	diss py	1.06	0.01	91.38	0.39	0.67	0.01	7.71	1.61	66.58	0.93	1.57	0.012	0.67	688.69	48.82
AWDP4pr	1	TP	diss po	1.18	0.01	33.28	0.65	0.74	0.01	9.44	0.47	258.43	9.07	4.10	0.14	0.96	762.15	80.12
AWDBf5	1	BQ	py bed	0.60	0.01	45.34	0.44	0.38	0.01	6.57	0.75	81.35	2.18	2.21	0.05	0.76	1033.12	63.49
Data discounted due to 'over-large' errors.																		
AWDDo5pr	2	Dol	diss py	4.93	0.02	131.23	160.26	3.10	0.01	129.82	3.02	18.27E3	15.93E3	765.6	667.65	1.00	2461.8	99.89
AWDDo10py	2	Dol	diss py	3.07	0.02	83.08	157.39	1.93	0.00	82.23	1.18	18.94E3	25.60E3	807.3	1091.40	1.00	2503.2	99.89
DoROpy1	1	Dol	diss py	4.32	0.02	55.71	1015.77	2.71	0.01	55.65	0.36	36.36E4	46.93E5	7.46E3	9.63E4	1.00	1218.58	99.92
Data discounted for lying significantly below the trend of the other data points, representing significant Os loss.																		
DoROpy3	1	Dol	diss py	8.04	0.03	133.42	93.74	5.06	0.02	131.53	0.49	22.29E3	11.21E2	5.81E2	2.92E2	1.00	1541.46	99.83
DoROpr1	1	Dol	diss po	4.91	0.02	52.44	2.99	3.09	0.01	37.50	0.35	1722.44	92.79	21.72	1.17	0.99	724.78	96.22
Samples with insufficient Re for analysis.																		
BfROpy1	2	BQ	py bed	<0.5														
BfROpy2	2	BQ	diss py	<0.5														
BfROpy3	2	BQ	diss py	<0.5														
BqROpy1	1	BQ	diss py	<0.5														
TpROpy1	1	TP	diss py	<0.5														

N.B. Bat = Batch Number, Lith = Lithology, Para = Paragenesis, TP = Tarkwa Phyllite, Dol = Metadolomite, BQ = Banket Quartzite, 'diss py' = disseminated pyrite, 'diss po' = disseminated pyrrhotite, 'py bed' = pyrite replacing magnetite along bedding planes. Re and Os values given in ppb, Model Age given in Ma, $\text{Re/Os} = {}^{187}\text{Re}/{}^{186}\text{Os}$, $\text{Os/Os} = {}^{187}\text{Os}/{}^{186}\text{Os}$

Table 5.7: Details of samples used for $^{40}\text{Ar}/^{39}\text{Ar}$ analysis

Sample	Biotite generation	mass (mg)	J value
DoArBt4	Hydrothermal - within vein	10.54	0.007266 $\pm$ 0.000052
DoArBt4-2	Hydrothermal - within vein	12.33	
DoArBt4-3	Hydrothermal - within vein	12.30	
DoArBt2	Hydrothermal - wall rock alteration	9.87	
AWDDo6	Hydrothermal - wall rock alteration	10.41	
TpArBt1	Metamorphic - Tarkwa Phyllite	9.21	0.007287 $\pm$ 0.000032
AWABv1	Metamorphic - Birimian Volcaniclastic	8.53	
DiArBt1	Metamorphic - Diorite Porphyry	12.66	
DiArBt1-2	Metamorphic - Diorite Porphyry	8.48	

simple cooling histories (see [McDougall & Harrison \(1999\)](#) for a discussion). Additionally, since K-contents are much lower than in biotite (especially true for the Damang hornblendes) significantly more material would be required to achieve a well-constrained age.

5.6.1 Results of $^{40}\text{Ar}/^{39}\text{Ar}$ Analysis

$^{40}\text{Ar}/^{39}\text{Ar}$ analysis was carried out on nine samples of biotite from a number of petrographic settings: hydrothermal biotite within a quartz vein, hydrothermal biotite from altered dolerites, and metamorphic biotite from a range of unmineralized lithologies. Sample details are listed in table 5.7. Step-heating plateau results are presented in figure 5.11, along with weighted average age calculations, summarized in table 5.9, and tabulated in table 5.8.

DoArBt4

Sample DoArBt4 contains very coarse-grained biotite flakes within a quartz vein. It was divided into three aliquots for analysis: samples DoArBt4, DoArBt4-2 and DoArBt4-3. The step-heating plateaux for all are extremely well defined, producing consistent ages. Sample DoArBt4, with 97.13% of released ^{39}Ar , over 13 steps, produced an age of 1978.8 $\pm$ 6.2 Ma (2σ , MSWD = 0.48). Sample DoArBt4-2, with with 92.08% of released ^{39}Ar , over 15 steps, produced an age of 1975.0 $\pm$ 6.6 Ma (2σ , MSWD = 0.44). Sample DoArBt4-3, with 98.36% of released ^{39}Ar , over 13 steps, produced an age of 1974.0 $\pm$ 5.8 Ma (2σ , MSWD = 1.7). These three aliquots are therefore statistically identical.

DoArBt2

The step-heating plateau for sample DoArBt2 is well defined with 91.0% of released ^{39}Ar , over 11 steps, producing an age of 1926.4 ± 8.3 Ma (2σ , MSWD = 0.53). This age is significantly lower than for the three aliquots from sample DoArBt4. Inspection of Ca/K ratios calculated for each step shows that for the low temperature steps, particularly at 700°C, the proportion of Ca is significantly higher than for K, indicating contamination of the sample by a mineral other than biotite. Given the relatively low temperature release of Ca-derived Ar, this is interpreted to be carbonate, which is abundant in hydrothermal alteration zones and often intimately associated with biotite (c.f. section 3.6). The higher temperature steps however, typically have much lower Ca/K ratios with a high proportion of radiogenic ^{40}Ar and are therefore deemed to reliably represent Ar release from biotite.

AWDDo6

The step-heating plateau for sample AWDDo6 is reasonably well defined with 88.94% of released ^{39}Ar , over 5 steps, producing an age of 1927 ± 21 Ma (2σ , MSWD = 1.06). This age agrees well with sample DoArBt2. The low number of heating steps compared to other samples is due to the sample containing a very much lower proportion of Ar overall, which also results in the relatively large age uncertainty. As with sample DoArBt2, Ca/K ratios are high for the low temperature steps, which again is interpreted to represent carbonate contamination. Once again, the higher temperature steps have much lower Ca/K ratios with a high proportion of radiogenic ^{40}Ar and are therefore also deemed to reliably represent Ar release from biotite.

TpArBt1

Sample TpArBt1 produces a step-heating plateau that is again well defined with 79.14% of released ^{39}Ar , over 9 steps, producing an age of 1950.2 ± 8.3 Ma (2σ , MSWD = 0.67). This age falls between those defined by the aliquots of sample DoArBt4 and samples DoArBt2

and AWDDo6. The 850°C heating step produced an anomalously low age that has been excluded from the final age calculation. This anomaly can be ascribed directly to a problem during the analysis that resulted in the mass spectrometer ‘falling off’ the ion peak during analysis, rather than it being a function of the sample itself.

AWABv1

Sample AWABv1 produces a step-heating plateau that is reasonably well defined, but increases in age slightly as gas is released. Despite this, a plateau defined by 61.58% of released ^{39}Ar , over 9 steps, produces an age of 1898 ± 11 Ma (2σ , MSWD = 0.76), which is distinctly younger than all other samples.

DiArBt1

Sample DiArBt1 is the first aliquot from sample DiArBt1. The resulting step-heating plateau is very well defined with 80.67% of released ^{39}Ar , over 10 steps, producing an age of 1941.6 ± 6.6 Ma (2σ , MSWD = 0.56), which is similar to that for sample TpArBt1. As with sample TpArBt1, an anomalously low age heating step occurred at 1050°C, which again is due to an analysis problem and is not related to the sample.

DiArBt1-2

Sample DiArBt1-2 is the second aliquot from sample DiArBt1. The resulting step-heating plateau is extremely well defined with 90.59% of released ^{39}Ar , over 11 steps, producing an age of 1929.8 ± 4.4 Ma (2σ , MSWD = 0.72). This age is similar to, but just outside of 2σ error with sample DiArBt1. A likely explanation is that the most proximal monitor to this sample within the irradiation vial produced an anomalously high J value (0.007671 ± 0.000053), which is higher than for other monitors that were subjected to the same irradiation process (R. Burgess, *pers. comm.*). Instead, the second nearest monitor was used to define the J value. Although this provides a best estimate, the final age cannot be verified and this sample will therefore be excluded from further discussion.

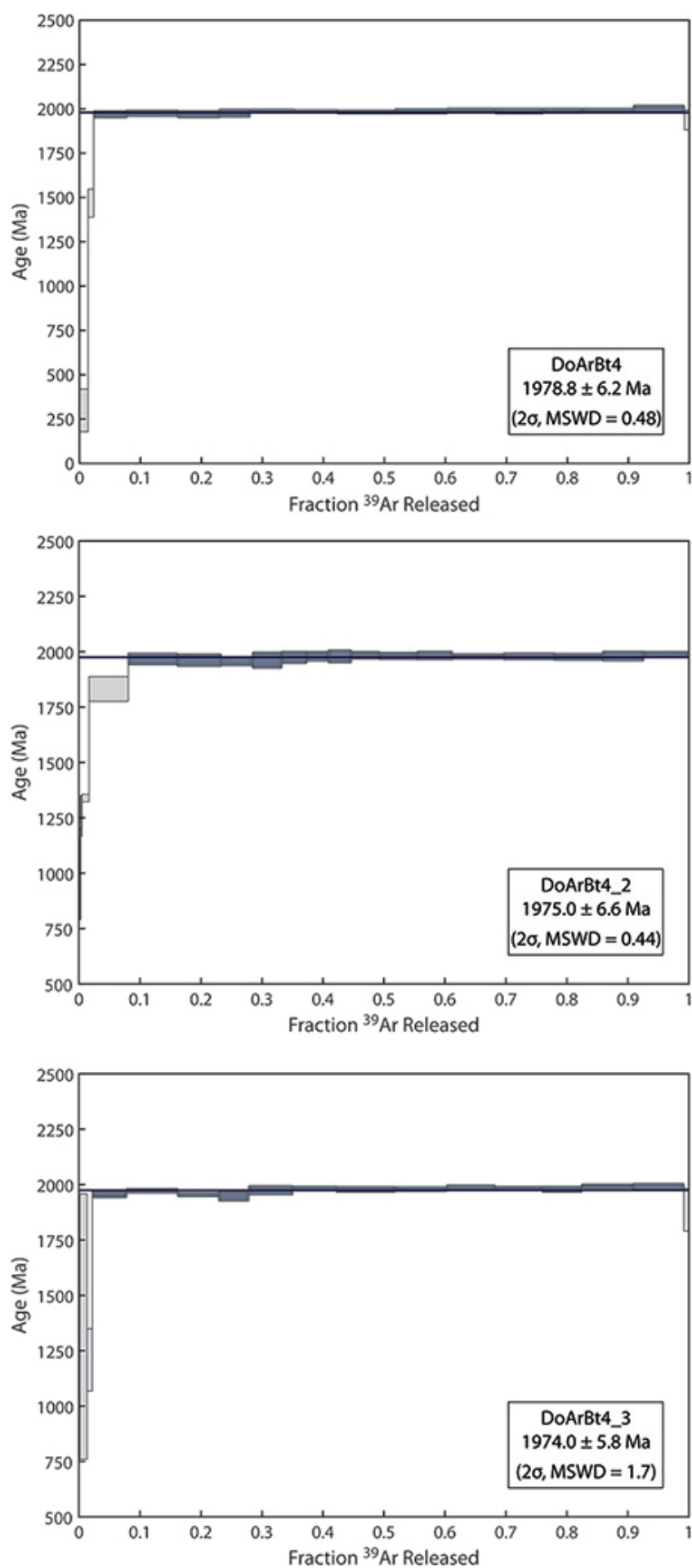
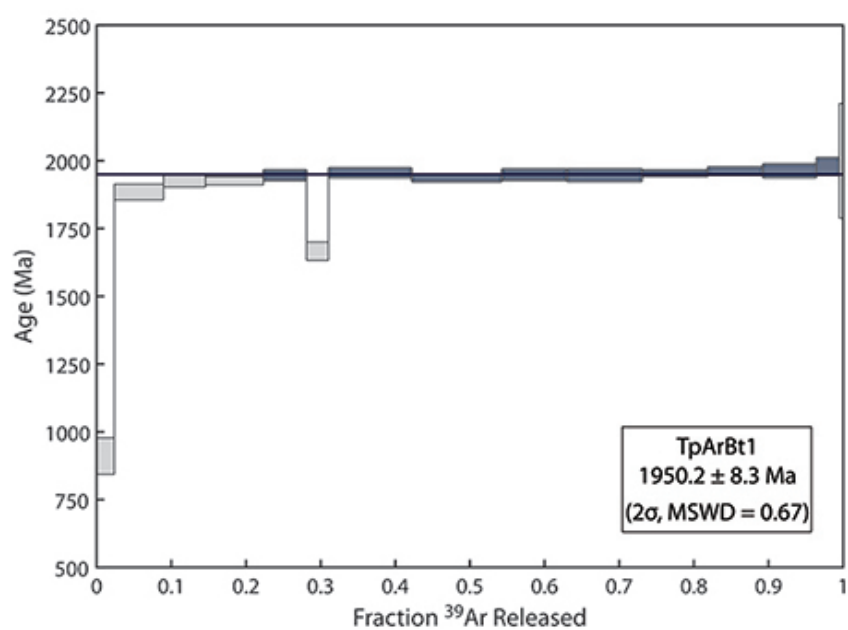
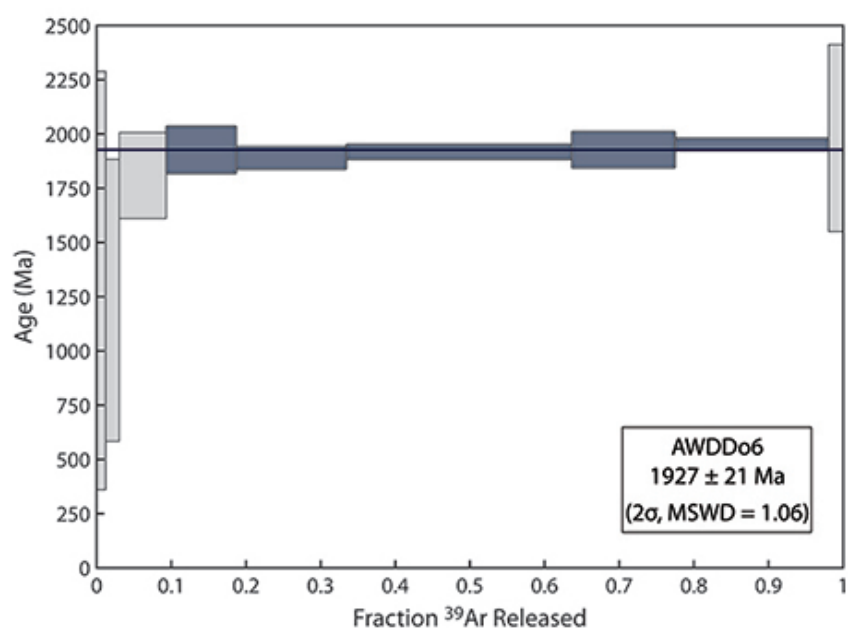
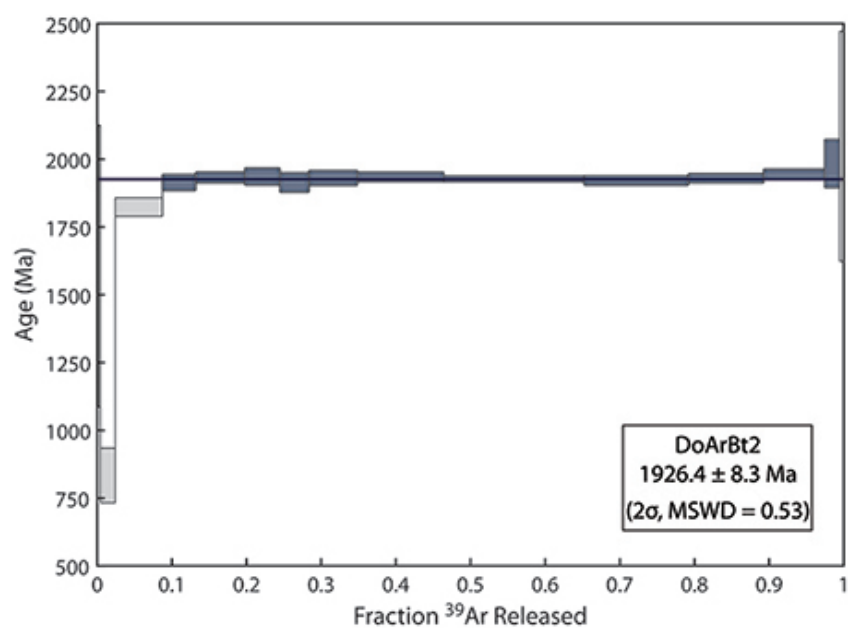


Figure 5.11: (Continued on following two pages) $^{40}\text{Ar}/^{39}\text{Ar}$ plateaux results.



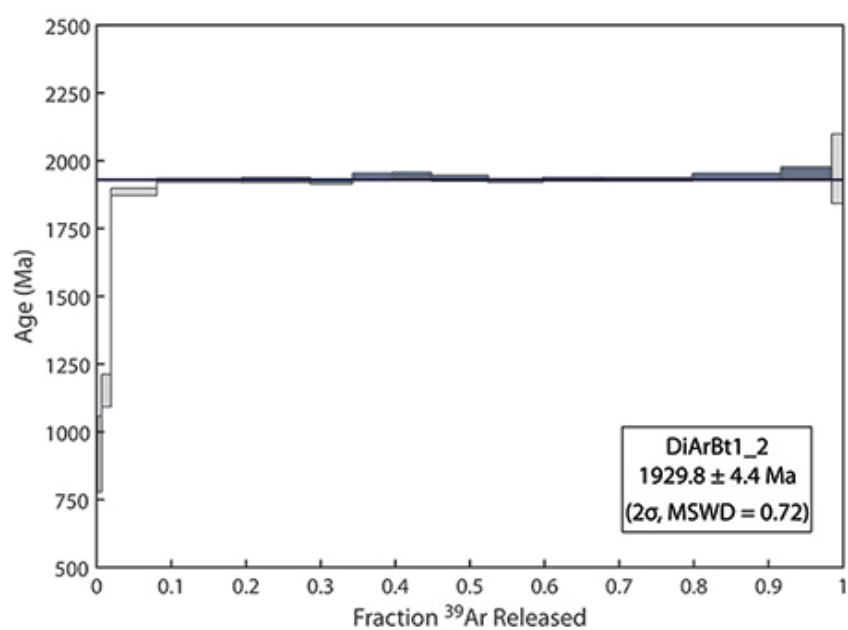
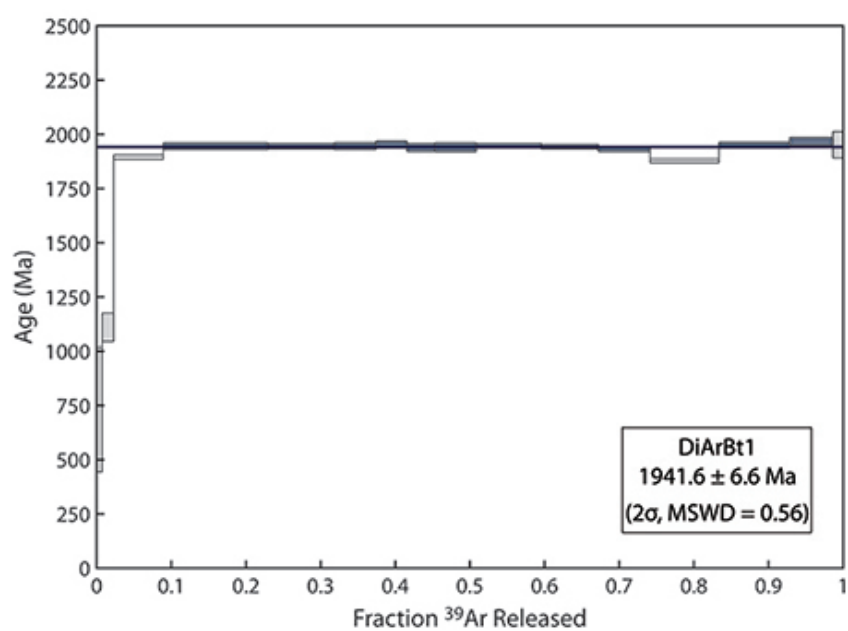
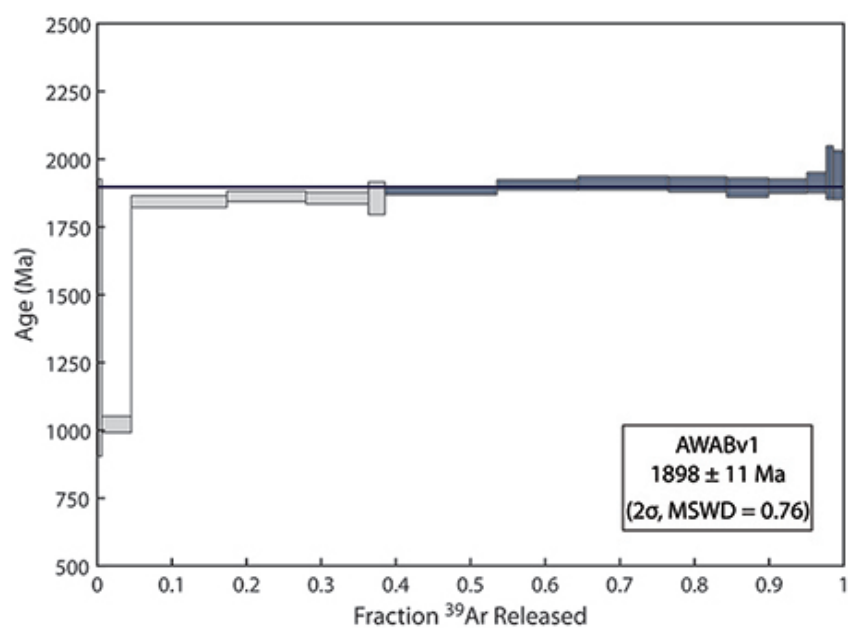


Table 5.8: Results of step-heating $^{40}\text{Ar}/^{39}\text{Ar}$ analysis

Temp. Step.	^{36}Ar	^{37}Ar	^{38}Ar	^{39}Ar	^{40}Ar	$^{36}\text{Ar}/^{40}\text{Ar}_{\text{air}}$	Ca/K	Cum. ^{39}Ar	$^{40}\text{Ar}^*/^{39}\text{Ar}$	$\%^{40}\text{Ar}^*$	Age (Ma)
Sample DoArBt4-2											
400	9.1 ± 0.1	1.2 ± 0.1	4.7 ± 0.2	17.4 ± 0.3	4403.3 ± 17.7	292.85	0.8 ± 0.2	0.1	100.7 ± 13.6	39.8	990.50 ± 103.39
500	3.7 ± 0.1	0.6 ± 0.1	3.1 ± 0.2	32.3 ± 0.2	5532.0 ± 2.4	292.85	0.2 ± 0.1	0.3	138.2 ± 7.0	80.4	1254.29 ± 46.49
600	7.7 ± 0.1	12.8 ± 0.3	9.8 ± 0.2	229.9 ± 0.6	36817.0 ± 22.9	292.85	0.6 ± 0.0	1.6	150.8 ± 1.3	93.9	1334.51 ± 10.44
700	4.5 ± 1.1	4.9 ± 0.3	7.0 ± 1.2	113.9 ± 0.4	28799.3 ± 15.9	292.85	0.5 ± 0.1	7.9	241.9 ± 5.9	95.4	1829.79 ± 29.13
750	1.1 ± 0.4	1.2 ± 0.2	2.9 ± 0.3	145.3 ± 0.2	39641.1 ± 46.9	292.85	0.1 ± 0.0	16.0	271.2 ± 2.8	99.2	1963.96 ± 15.08
775	0.7 ± 0.3	0.6 ± 0.3	2.3 ± 0.3	127.1 ± 0.4	34466.5 ± 31.4	292.85	0.1 ± 0.1	23.1	270.2 ± 3.2	99.4	1959.57 ± 16.50
800	0.8 ± 0.1	0.5 ± 0.0	1.8 ± 0.1	94.5 ± 0.3	25632.0 ± 19.9	292.85	0.1 ± 0.0	28.4	269.3 ± 2.2	99.1	1955.60 ± 12.86
825	0.5 ± 0.2	0.7 ± 0.2	1.9 ± 0.2	85.1 ± 0.2	23098.2 ± 96.1	292.85	0.1 ± 0.0	33.1	270.2 ± 4.0	99.3	1959.70 ± 19.77
850	0.4 ± 0.1	0.5 ± 0.1	1.5 ± 0.1	74.0 ± 0.1	20267.0 ± 57.5	292.85	0.1 ± 0.0	37.2	273.0 ± 2.9	99.5	1972.03 ± 15.50
875	0.6 ± 0.0	0.3 ± 0.0	1.2 ± 0.1	65.4 ± 0.2	18067.2 ± 7.9	292.85	0.0 ± 0.0	40.9	274.0 ± 2.4	99.0	1976.55 ± 13.52
900	0.4 ± 0.1	0.3 ± 0.0	1.1 ± 0.1	64.8 ± 0.1	17843.3 ± 8.3	292.85	0.1 ± 0.0	44.5	274.1 ± 3.4	99.3	1976.69 ± 17.20
925	0.5 ± 0.0	0.4 ± 0.1	1.1 ± 0.1	75.1 ± 0.1	20702.9 ± 6.8	288.41	0.1 ± 0.0	49.3	275.4 ± 1.8	99.3	1982.39 ± 11.77
950	0.6 ± 0.1	0.6 ± 0.1	1.4 ± 0.1	97.8 ± 0.1	26876.7 ± 12.1	288.41	0.1 ± 0.0	55.5	274.6 ± 1.6	99.3	1978.99 ± 11.17
975	0.7 ± 0.1	0.5 ± 0.1	1.3 ± 0.1	89.5 ± 0.1	24642.7 ± 11.4	288.41	0.1 ± 0.0	61.1	275.0 ± 2.1	99.2	1980.79 ± 12.65
1000	0.9 ± 0.0	0.7 ± 0.1	2.2 ± 0.1	134.6 ± 0.2	36906.2 ± 16.4	288.41	0.1 ± 0.0	69.6	273.9 ± 1.2	99.3	1975.95 ± 10.19
1025	0.8 ± 0.1	0.7 ± 0.1	2.2 ± 0.1	131.6 ± 0.2	36091.4 ± 15.7	288.41	0.1 ± 0.0	78.0	274.2 ± 1.7	99.4	1977.24 ± 11.44
1050	0.7 ± 0.0	0.6 ± 0.1	2.1 ± 0.1	124.6 ± 0.3	34091.7 ± 17.9	288.41	0.1 ± 0.0	85.9	273.7 ± 1.7	99.4	1975.08 ± 11.46
1100	0.7 ± 0.1	0.6 ± 0.2	1.7 ± 0.1	107.9 ± 0.1	29640.2 ± 83.2	288.41	0.1 ± 0.0	92.7	274.4 ± 2.4	99.3	1978.32 ± 13.63
1200	1.1 ± 0.0	0.8 ± 0.1	1.9 ± 0.1	115.2 ± 0.2	31918.9 ± 16.9	288.41	0.1 ± 0.0	100.0	276.1 ± 1.3	99.0	1985.39 ± 10.36
1400	0.0 ± 0.0	0.0 ± 0.0	0.0 ± 0.0	0.2 ± 0.0	23.7 ± 3.3	288.41	-1.8 ± -4.0	100.0	90.5 ± 316.7	81.7	911.36 ± 2504.58

Temp. Step.	^{36}Ar	^{37}Ar	^{38}Ar	^{39}Ar	^{40}Ar	$^{36}\text{Ar}/^{40}\text{Ar}_{\text{air}}$	C/Ka	Cum. ^{39}Ar	$^{40}\text{Ar}^*/^{39}\text{Ar}$	$\%^{40}\text{Ar}^*$	Age (Ma)
Sample DoArBt4											
500	0.9 ± 0.1	0.4 ± 0.0	0.7 ± 0.0	2.8 ± 0.0	818.4 ± 0.6	299.02	1.7 ± 0.3	0.3	196.6 ± 48.2	67.2	1600.59 ± 260.55
600	0.2 ± 0.0	2.1 ± 0.0	0.9 ± 0.1	17.4 ± 0.1	1459.9 ± 2.7	299.02	1.5 ± 0.1	2.3	80.0 ± 6.1	95.6	826.63 ± 50.75
700	1.5 ± 0.1	69.1 ± 0.1	1.3 ± 0.1	55.0 ± 0.1	13556.5 ± 31.2	299.02	15.5 ± 0.4	8.6	239.7 ± 3.6	96.9	1819.40 ± 19.21
750	0.4 ± 0.0	5.8 ± 0.0	0.8 ± 0.0	39.7 ± 0.1	10452.4 ± 31.6	299.02	1.8 ± 0.0	13.1	259.1 ± 3.2	98.8	1909.97 ± 16.76
800	0.5 ± 0.0	3.8 ± 0.0	1.0 ± 0.0	57.7 ± 0.1	15289.5 ± 4.9	293.82	0.8 ± 0.0	19.7	263.1 ± 2.1	99.1	1928.03 ± 12.56
825	0.0 ± 0.0	3.0 ± 0.1	0.7 ± 0.1	41.1 ± 0.1	10856.0 ± 24.9	293.82	0.9 ± 0.0	24.4	264.3 ± 3.5	100.0	1933.42 ± 17.81
850	0.4 ± 0.0	3.5 ± 0.0	0.7 ± 0.0	34.4 ± 0.1	8989.1 ± 4.9	293.82	1.3 ± 0.0	28.3	259.1 ± 3.8	98.8	1909.77 ± 19.41
900	0.2 ± 0.1	4.9 ± 0.0	0.9 ± 0.0	56.5 ± 0.1	14899.6 ± 12.6	293.82	1.1 ± 0.0	34.8	262.8 ± 2.9	99.5	1926.68 ± 15.65
950	0.5 ± 0.1	4.4 ± 0.0	1.6 ± 0.1	101.3 ± 0.1	26830.5 ± 25.3	293.82	0.5 ± 0.0	46.4	263.7 ± 1.9	99.4	1930.63 ± 11.92
1000	0.8 ± 0.1	3.3 ± 0.0	2.5 ± 0.3	164.7 ± 0.2	43337.0 ± 11.0	293.82	0.2 ± 0.0	65.2	262.1 ± 1.2	99.5	1923.62 ± 9.94
1025	0.9 ± 0.1	1.8 ± 0.1	2.2 ± 0.1	122.5 ± 0.3	32189.5 ± 55.0	293.82	0.2 ± 0.0	79.2	261.1 ± 1.9	99.2	1919.04 ± 12.00
1050	0.7 ± 0.1	1.3 ± 0.1	1.5 ± 0.0	88.7 ± 0.1	23456.9 ± 20.7	293.82	0.2 ± 0.0	89.3	262.6 ± 1.8	99.2	1925.79 ± 11.85
1100	0.5 ± 0.0	1.9 ± 0.0	1.4 ± 0.1	71.0 ± 0.1	19010.0 ± 16.0	293.82	0.3 ± 0.0	97.5	266.1 ± 2.0	99.3	1941.67 ± 12.46
1200	0.6 ± 0.1	3.6 ± 0.1	1.0 ± 0.0	18.5 ± 0.1	5234.7 ± 16.2	293.82	2.4 ± 0.2	99.6	274.8 ± 10.2	96.9	1979.71 ± 45.46
1400	0.8 ± 0.1	2.9 ± 0.1	0.4 ± 0.0	3.7 ± 0.1	1317.4 ± 1.3	293.82	9.7 ± 0.7	100.0	289.8 ± 50.3	81.4	2044.30 ± 212.40
Sample DoArBt4-3											
500	0.7 ± 0.1	0.1 ± 0.1	0.3 ± 0.0	3.0 ± 0.1	659.9 ± 0.4	290.55	0.5 ± 0.8	0.2	153.7 ± 48.6	70.4	1352.98 ± 301.10
600	0.4 ± 0.1	1.6 ± 0.1	0.5 ± 0.0	16.1 ± 0.1	2211.3 ± 2.7	290.55	1.3 ± 0.1	1.0	130.4 ± 10.4	94.7	1202.58 ± 70.31
700	1.5 ± 0.1	1.7 ± 0.1	2.1 ± 0.1	119.0 ± 0.2	32240.2 ± 19.2	290.55	0.2 ± 0.0	6.9	268.6 ± 1.4	98.7	1952.40 ± 10.66
750	1.1 ± 0.1	0.8 ± 0.1	2.7 ± 0.3	188.0 ± 0.2	51366.0 ± 49.8	290.55	0.1 ± 0.0	16.3	272.7 ± 1.1	99.4	1970.62 ± 9.86
775	0.9 ± 0.0	0.7 ± 0.0	2.6 ± 0.1	147.6 ± 0.1	39752.1 ± 33.1	290.55	0.1 ± 0.0	23.7	268.8 ± 1.0	99.4	1953.49 ± 9.56
800	0.6 ± 0.0	0.5 ± 0.1	1.8 ± 0.1	94.9 ± 0.3	25381.8 ± 34.1	290.55	0.1 ± 0.0	28.4	266.7 ± 2.4	99.3	1944.32 ± 13.52
850	0.8 ± 0.1	0.6 ± 0.1	2.2 ± 0.1	135.0 ± 0.4	36917.0 ± 26.1	290.55	0.1 ± 0.0	35.2	272.9 ± 2.2	99.4	1971.38 ± 12.88
900	0.9 ± 0.0	0.8 ± 0.1	2.2 ± 0.1	129.9 ± 0.1	35778.3 ± 31.0	290.55	0.1 ± 0.0	41.7	274.6 ± 1.1	99.3	1979.23 ± 9.96
950	0.9 ± 0.1	0.7 ± 0.0	2.3 ± 0.1	164.7 ± 0.1	45227.7 ± 60.0	290.55	0.1 ± 0.0	49.9	274.1 ± 1.2	99.4	1977.02 ± 10.17
975	1.0 ± 0.1	0.5 ± 0.0	2.5 ± 0.1	154.2 ± 0.1	42362.2 ± 31.3	290.55	0.0 ± 0.0	57.6	274.1 ± 1.1	99.3	1976.89 ± 9.95
1000	0.8 ± 0.0	0.6 ± 0.1	2.6 ± 0.2	163.8 ± 0.2	45278.8 ± 33.0	290.55	0.0 ± 0.0	65.8	276.1 ± 1.1	99.5	1985.40 ± 9.76
1025	0.8 ± 0.0	0.6 ± 0.0	2.7 ± 0.1	165.1 ± 0.2	45423.0 ± 28.5	290.55	0.1 ± 0.0	74.1	274.8 ± 0.9	99.5	1979.96 ± 9.55
1050	0.7 ± 0.0	0.6 ± 0.0	2.7 ± 0.1	164.4 ± 0.4	45076.3 ± 45.8	290.55	0.1 ± 0.0	82.3	274.0 ± 1.5	99.5	1976.62 ± 10.89
1100	0.8 ± 0.0	0.8 ± 0.1	3.2 ± 0.1	213.1 ± 0.3	59072.2 ± 39.6	290.55	0.0 ± 0.0	92.9	277.3 ± 1.1	99.6	1990.69 ± 9.81
1200	0.8 ± 0.0	0.4 ± 0.0	2.0 ± 0.1	127.5 ± 0.3	35398.6 ± 25.2	290.55	0.0 ± 0.0	99.3	277.0 ± 1.5	99.4	1989.71 ± 10.83
1400	0.9 ± 0.0	0.2 ± 0.0	0.7 ± 0.0	13.5 ± 0.1	3662.2 ± 2.1	290.55	0.2 ± 0.0	100.0	253.4 ± 10.5	93.1	1883.67 ± 49.01

Temp. Step.	^{36}Ar	^{37}Ar	^{38}Ar	^{39}Ar	^{40}Ar	$^{36}\text{Ar}/^{40}\text{Ar}_{\text{air}}$	Ca/K	Cum. ^{39}Ar	$^{40}\text{Ar}^*/^{39}\text{Ar}$	$\%^{40}\text{Ar}^*$	Age (Ma)
Sample AWDDo6											
500	0.4 ± 0.0	0.0 ± 0.0	0.3 ± 0.0	1.5 ± 0.1	342.0 ± 0.7	284.92	-0.1 ± -0.7	1.1	148.8 ± 76.6	65.2	1321.94 ± 482.39
600	0.0 ± 0.0	1.4 ± 0.0	-0.1 ± 0.0	2.6 ± 0.0	332.9 ± 0.5	299.02	6.9 ± 0.5	2.9	134.6 ± 49.0	104.6	1230.26 ± 325.15
700	0.2 ± 0.1	6.3 ± 0.1	0.0 ± 0.1	8.8 ± 0.1	2141.9 ± 1.3	299.02	8.9 ± 0.4	9.2	236.8 ± 20.4	96.7	1805.47 ± 98.65
800	0.6 ± 0.1	1.5 ± 0.1	0.2 ± 0.1	13.1 ± 0.0	3613.4 ± 2.0	299.02	1.4 ± 0.1	18.6	262.3 ± 12.1	95.2	1924.33 ± 55.15
900	0.5 ± 0.0	2.7 ± 0.1	0.4 ± 0.1	20.6 ± 0.1	5378.4 ± 15.9	299.02	1.6 ± 0.1	33.4	254.3 ± 5.8	97.4	1887.71 ± 27.81
1000	0.3 ± 0.0	2.3 ± 0.1	1.0 ± 0.0	42.2 ± 0.1	11103.1 ± 36.9	299.02	0.7 ± 0.0	63.7	260.3 ± 3.8	99.2	1915.28 ± 19.11
1100	0.3 ± 0.1	0.6 ± 0.1	0.4 ± 0.0	19.3 ± 0.1	5167.9 ± 16.3	299.02	0.4 ± 0.1	77.6	262.3 ± 9.1	98.2	1924.24 ± 41.87
1200	0.5 ± 0.0	1.6 ± 0.0	0.6 ± 0.0	28.6 ± 0.0	7820.7 ± 15.2	299.02	0.7 ± 0.0	98.2	267.6 ± 2.9	98.0	1948.35 ± 15.60
1400	0.9 ± 0.0	0.8 ± 0.0	0.2 ± 0.1	2.6 ± 0.1	976.4 ± 0.7	299.02	4.1 ± 0.4	100.0	275.0 ± 49.1	72.5	1980.57 ± 214.81
Sample DoArBt2											
500	0.9 ± 0.1	0.4 ± 0.0	0.7 ± 0.0	2.8 ± 0.0	818.4 ± 0.6	299.02	1.7 ± 0.3	0.3	196.6 ± 48.2	67.2	1600.59 ± 260.55
600	0.2 ± 0.0	2.1 ± 0.0	0.9 ± 0.1	17.4 ± 0.1	1459.9 ± 2.7	299.02	1.5 ± 0.1	2.3	80.0 ± 6.1	95.6	826.63 ± 50.75
700	1.5 ± 0.1	69.1 ± 0.1	1.3 ± 0.1	55.0 ± 0.1	13556.5 ± 31.2	299.02	15.5 ± 0.4	8.6	239.7 ± 3.6	96.9	1819.40 ± 19.21
750	0.4 ± 0.0	5.8 ± 0.0	0.8 ± 0.0	39.7 ± 0.1	10452.4 ± 31.6	299.02	1.8 ± 0.0	13.1	259.1 ± 3.2	98.8	1909.97 ± 16.76
800	0.5 ± 0.0	3.8 ± 0.0	1.0 ± 0.0	57.7 ± 0.1	15289.5 ± 4.9	293.82	0.8 ± 0.0	19.7	263.1 ± 2.1	99.1	1928.03 ± 12.56
825	0.0 ± 0.0	3.0 ± 0.1	0.7 ± 0.1	41.1 ± 0.1	10856.0 ± 24.9	293.82	0.9 ± 0.0	24.4	264.3 ± 3.5	100.0	1933.42 ± 17.81
850	0.4 ± 0.0	3.5 ± 0.0	0.7 ± 0.0	34.4 ± 0.1	8989.1 ± 4.9	293.82	1.3 ± 0.0	28.3	259.1 ± 3.8	98.8	1909.77 ± 19.41
900	0.2 ± 0.1	4.9 ± 0.0	0.9 ± 0.0	56.5 ± 0.1	14899.6 ± 12.6	293.82	1.1 ± 0.0	34.8	262.8 ± 2.9	99.5	1926.68 ± 15.65
950	0.5 ± 0.1	4.4 ± 0.0	1.6 ± 0.1	101.3 ± 0.1	26830.5 ± 25.3	293.82	0.5 ± 0.0	46.4	263.7 ± 1.9	99.4	1930.63 ± 11.92
1000	0.8 ± 0.1	3.3 ± 0.0	2.5 ± 0.3	164.7 ± 0.2	43337.0 ± 11.0	293.82	0.2 ± 0.0	65.2	262.1 ± 1.2	99.5	1923.62 ± 9.94
1025	0.9 ± 0.1	1.8 ± 0.1	2.2 ± 0.1	122.5 ± 0.3	32189.5 ± 55.0	293.82	0.2 ± 0.0	79.2	261.1 ± 1.9	99.2	1919.04 ± 12.00
1050	0.7 ± 0.1	1.3 ± 0.1	1.5 ± 0.0	88.7 ± 0.1	23456.9 ± 20.7	293.82	0.2 ± 0.0	89.3	262.6 ± 1.8	99.2	1925.79 ± 11.85
1100	0.5 ± 0.0	1.9 ± 0.0	1.4 ± 0.1	71.0 ± 0.1	19010.0 ± 16.0	293.82	0.3 ± 0.0	97.5	266.1 ± 2.0	99.3	1941.67 ± 12.46
1200	0.6 ± 0.1	3.6 ± 0.1	1.0 ± 0.0	18.5 ± 0.1	5234.7 ± 16.2	293.82	2.4 ± 0.2	99.6	274.8 ± 10.2	96.9	1979.71 ± 45.46
1400	0.8 ± 0.1	2.9 ± 0.1	0.4 ± 0.0	3.7 ± 0.1	1317.4 ± 1.3	293.82	9.7 ± 0.7	100.0	289.8 ± 50.3	81.4	2044.30 ± 212.40

Temp. Step.	^{36}Ar	^{37}Ar	^{38}Ar	^{39}Ar	^{40}Ar	$^{36}\text{Ar}/^{40}\text{Ar}_{\text{air}}$	Ca/K	Cum. ^{39}Ar	$^{40}\text{Ar}^*/^{39}\text{Ar}$	$\%^{40}\text{Ar}^*$	Age (Ma)
Sample TpArBt1											
500	0.8 ± 0.1	0.0 ± 0.0	0.2 ± 0.0	1.4 ± 0.2	182.2 ± 0.9	288.41	0.2 ± 0.6	0.1	-27.6 ± -155.2	-20.3	nd $\pm$
600	0.8 ± 0.0	0.2 ± 0.1	1.5 ± 0.1	28.1 ± 0.1	2744.4 ± 4.5	288.41	0.1 ± 0.0	2.3	90.0 ± 4.1	91.5	907.67 ± 33.16
700	1.2 ± 0.1	0.8 ± 0.1	0.0 ± 0.2	82.5 ± 0.1	21113.3 ± 64.1	288.41	0.1 ± 0.0	8.9	253.3 ± 2.9	98.4	1883.21 ± 15.93
750	0.7 ± 0.0	0.6 ± 0.1	0.0 ± 0.1	71.1 ± 0.1	18777.5 ± 60.9	288.41	0.1 ± 0.0	14.6	262.8 ± 2.5	99.0	1926.67 ± 14.12
800	0.8 ± 0.0	0.7 ± 0.1	0.1 ± 0.1	97.9 ± 0.2	25733.9 ± 19.7	288.41	0.1 ± 0.0	22.4	262.2 ± 1.5	99.1	1923.77 ± 10.71
825	0.6 ± 0.1	0.4 ± 0.1	1.1 ± 0.1	74.2 ± 0.1	19789.0 ± 8.7	284.92	0.1 ± 0.0	28.2	266.8 ± 2.2	99.1	1944.67 ± 12.85
850	2.0 ± 0.0	0.4 ± 0.1	3.0 ± 0.0	37.0 ± 0.0	8224.6 ± 3.9	284.92	0.1 ± 0.1	31.1	208.6 ± 3.2	93.1	1664.21 ± 18.58
900	0.5 ± 0.1	1.0 ± 0.1	2.0 ± 0.1	144.6 ± 0.2	38615.7 ± 100.0	284.92	0.1 ± 0.0	42.3	268.4 ± 1.9	99.6	1951.65 ± 12.10
950	1.0 ± 0.1	1.0 ± 0.1	2.6 ± 0.1	155.3 ± 0.1	41062.0 ± 20.1	284.92	0.1 ± 0.0	54.4	264.8 ± 1.5	99.3	1935.56 ± 10.72
975	0.6 ± 0.0	0.7 ± 0.1	1.8 ± 0.1	111.9 ± 0.3	29849.9 ± 81.5	284.92	0.1 ± 0.0	63.1	267.6 ± 2.2	99.5	1948.17 ± 13.14
1000	1.3 ± 0.1	0.9 ± 0.2	2.5 ± 0.1	129.2 ± 0.2	34552.1 ± 112.3	284.92	0.1 ± 0.0	73.1	266.9 ± 2.5	99.0	1945.11 ± 13.94
1025	0.6 ± 0.0	0.8 ± 0.1	1.9 ± 0.1	114.2 ± 0.2	30511.6 ± 11.7	284.92	0.1 ± 0.0	82.0	268.1 ± 1.2	99.4	1950.49 ± 10.06
1050	0.3 ± 0.1	0.8 ± 0.1	1.7 ± 0.1	94.6 ± 0.2	25373.9 ± 13.3	284.92	0.1 ± 0.0	89.4	269.9 ± 1.7	99.7	1958.39 ± 11.37
1100	1.0 ± 0.1	0.9 ± 0.1	1.6 ± 0.1	92.5 ± 0.1	25032.3 ± 75.8	284.92	0.1 ± 0.0	96.6	270.1 ± 2.7	98.9	1959.40 ± 14.68
1200	0.7 ± 0.0	0.8 ± 0.1	0.7 ± 0.1	38.2 ± 0.0	10565.1 ± 31.5	284.92	0.2 ± 0.1	99.5	274.3 ± 3.8	98.2	1977.74 ± 18.58
1400	1.1 ± 0.1	0.1 ± 0.1	0.3 ± 0.0	6.2 ± 0.1	2019.7 ± 1.0	284.92	0.2 ± 0.3	100.0	278.5 ± 24.2	84.2	1995.90 ± 105.26
Sample AWABv1											
500	0.7 ± 0.1	0.2 ± 0.1	1.3 ± 0.0	4.2 ± 0.2	889.8 ± 1.3	292.73	0.6 ± 0.4	0.5	163.4 ± 42.8	77.6	1411.79 ± 256.37
600	0.2 ± 0.0	0.9 ± 0.0	1.0 ± 0.0	34.6 ± 0.1	3648.5 ± 2.9	292.73	0.3 ± 0.0	4.4	104.1 ± 2.1	98.6	1016.00 ± 16.46
700	0.7 ± 0.0	2.4 ± 0.1	1.4 ± 0.1	112.8 ± 0.2	27694.7 ± 98.7	292.73	0.3 ± 0.0	17.1	244.1 ± 2.2	99.2	1840.60 ± 13.16
750	0.5 ± 0.1	1.1 ± 0.1	1.3 ± 0.0	95.2 ± 0.1	23761.1 ± 48.7	292.73	0.1 ± 0.0	27.8	248.6 ± 2.0	99.3	1861.40 ± 12.57
775	0.4 ± 0.1	1.0 ± 0.1	1.2 ± 0.1	74.9 ± 0.1	18558.9 ± 10.8	292.73	0.2 ± 0.0	36.3	246.7 ± 2.3	99.3	1852.71 ± 13.45
800	0.1 ± 0.0	0.2 ± 0.0	0.2 ± 0.0	19.0 ± 0.1	4714.5 ± 3.3	292.73	0.2 ± 0.1	38.4	246.9 ± 6.4	99.5	1853.56 ± 30.99
850	0.7 ± 0.1	2.0 ± 0.0	1.9 ± 0.1	133.4 ± 0.2	33812.5 ± 33.2	292.73	0.2 ± 0.0	53.4	252.5 ± 1.4	99.4	1879.84 ± 10.58
900	0.4 ± 0.1	0.7 ± 0.3	1.5 ± 0.1	96.1 ± 0.1	24789.0 ± 14.2	292.73	0.1 ± 0.1	64.3	257.3 ± 2.1	99.5	1901.79 ± 12.69
950	0.3 ± 0.1	0.9 ± 0.0	1.7 ± 0.1	107.4 ± 0.2	27832.0 ± 94.1	292.73	0.1 ± 0.0	76.4	258.8 ± 2.7	99.7	1908.57 ± 15.00
975	0.3 ± 0.1	0.5 ± 0.1	1.1 ± 0.0	69.8 ± 0.2	18043.5 ± 36.1	292.73	0.1 ± 0.1	84.3	258.1 ± 3.0	99.6	1905.12 ± 15.88
1000	0.4 ± 0.1	0.5 ± 0.1	0.8 ± 0.0	49.6 ± 0.2	12731.6 ± 7.3	292.73	0.1 ± 0.0	89.8	255.4 ± 3.8	99.2	1892.90 ± 19.52
1050	0.5 ± 0.0	0.5 ± 0.1	0.8 ± 0.0	45.5 ± 0.2	11779.7 ± 11.2	292.73	0.2 ± 0.1	95.0	256.4 ± 2.9	98.8	1897.74 ± 15.61
1100	0.3 ± 0.0	0.5 ± 0.0	0.5 ± 0.0	23.1 ± 0.1	6090.4 ± 3.2	292.73	0.3 ± 0.0	97.6	259.4 ± 4.1	98.3	1911.05 ± 20.32
1200	0.3 ± 0.0	0.6 ± 0.0	0.3 ± 0.0	8.5 ± 0.1	2360.5 ± 0.3	292.73	1.0 ± 0.1	98.5	267.5 ± 10.9	96.3	1947.71 ± 49.35
1400	0.7 ± 0.0	0.5 ± 0.0	0.3 ± 0.0	13.0 ± 0.0	3646.4 ± 3.3	292.73	0.5 ± 0.1	100.0	265.8 ± 9.9	94.3	1940.09 ± 45.27

Temp. Step.	^{36}Ar	^{37}Ar	^{38}Ar	^{39}Ar	^{40}Ar	$^{36}\text{Ar}/^{40}\text{Ar}_{\text{air}}$	Cum. ^{39}Ar	$^{40}\text{Ar}^*/^{39}\text{Ar}$	$\%^{40}\text{Ar}^*$	Age (Ma)
Sample DiArBt1										
500	0.4 ± 0.1	0.1 ± 0.1	0.7 ± 0.1	10.0 ± 0.2	798.4 ± 0.9	293.82	0.5	67.6 ± 16.3	84.8	720.89 ± 143.69
600	0.3 ± 0.1	1.3 ± 0.1	0.8 ± 0.0	30.6 ± 0.0	3637.4 ± 3.0	297.58	2.0	116.0 ± 4.4	97.7	1103.16 ± 31.98
700	0.7 ± 0.0	20.1 ± 0.0	2.0 ± 0.1	133.5 ± 0.2	34232.4 ± 22.3	297.58	8.8	254.6 ± 1.1	99.4	1889.30 ± 9.80
750	0.9 ± 0.1	2.2 ± 0.0	3.5 ± 0.1	277.6 ± 0.7	74230.6 ± 60.5	297.58	22.8	265.9 ± 1.6	99.6	1940.73 ± 11.03
775	0.9 ± 0.0	1.1 ± 0.0	2.7 ± 0.1	177.8 ± 0.4	47672.1 ± 37.3	297.58	31.8	266.1 ± 1.3	99.5	1941.57 ± 10.37
800	0.6 ± 0.1	0.7 ± 0.0	1.7 ± 0.1	109.7 ± 0.2	29439.6 ± 27.9	297.58	37.3	266.0 ± 1.7	99.4	1941.20 ± 11.33
825	0.5 ± 0.0	0.7 ± 0.0	1.3 ± 0.1	84.5 ± 0.1	22773.0 ± 19.0	297.58	41.6	267.5 ± 1.8	99.4	1947.50 ± 11.57
850	0.5 ± 0.0	0.6 ± 0.0	1.2 ± 0.0	72.1 ± 0.2	19274.8 ± 7.8	297.58	45.2	264.7 ± 2.0	99.2	1935.20 ± 12.23
900	0.4 ± 0.0	1.0 ± 0.0	1.8 ± 0.1	110.6 ± 0.2	29461.4 ± 84.2	297.58	50.8	264.8 ± 2.1	99.6	1935.85 ± 12.54
950	0.8 ± 0.0	1.5 ± 0.1	2.8 ± 0.1	175.4 ± 0.2	47037.3 ± 37.0	297.58	59.7	266.3 ± 1.1	99.5	1942.26 ± 9.75
975	0.8 ± 0.0	1.1 ± 0.1	2.5 ± 0.1	152.0 ± 0.2	40674.3 ± 24.4	297.58	67.3	265.5 ± 1.1	99.4	1938.64 ± 9.72
1000	1.2 ± 0.1	0.9 ± 0.1	2.5 ± 0.2	136.2 ± 0.2	36304.3 ± 16.5	297.58	74.2	263.5 ± 1.3	99.0	1929.61 ± 10.19
1050	2.9 ± 0.0	2.0 ± 0.1	5.5 ± 0.1	185.4 ± 0.2	47430.1 ± 26.9	297.58	83.6	250.7 ± 0.9	98.2	1871.45 ± 9.24
1100	0.9 ± 0.1	2.4 ± 0.1	3.1 ± 0.1	186.8 ± 0.3	50234.1 ± 33.2	297.58	93.0	267.0 ± 1.2	99.5	1945.50 ± 10.18
1200	0.8 ± 0.1	5.7 ± 0.0	2.2 ± 0.1	114.6 ± 0.3	31363.3 ± 28.0	297.58	98.8	271.0 ± 2.0	99.2	1963.41 ± 12.16
1400	0.9 ± 0.1	1.8 ± 0.0	0.7 ± 0.1	23.7 ± 0.1	6632.9 ± 6.3	297.58	100.0	267.6 ± 6.7	95.9	1948.19 ± 31.08
Sample DiArBt1-2										
550	0.6 ± 0.0	0.2 ± 0.1	0.7 ± 0.1	7.1 ± 0.1	854.2 ± 0.6	300.53	0.6	93.7 ± 17.3	78.2	938.76 ± 135.07
600	0.2 ± 0.1	0.4 ± 0.1	0.4 ± 0.1	16.9 ± 0.0	2167.1 ± 2.2	300.53	1.9	124.7 ± 8.3	97.7	1166.15 ± 57.42
700	0.6 ± 0.1	16.8 ± 0.1	1.1 ± 0.1	77.2 ± 0.1	19742.5 ± 22.9	300.53	8.0	252.8 ± 2.8	99.2	1884.51 ± 13.04
750	0.6 ± 0.1	1.3 ± 0.1	1.5 ± 0.1	145.4 ± 0.2	38416.6 ± 40.7	300.53	19.4	261.9 ± 1.3	99.5	1925.76 ± 6.01
775	0.6 ± 0.1	0.5 ± 0.1	1.6 ± 0.0	115.5 ± 0.2	30579.6 ± 19.2	300.53	28.5	262.1 ± 1.8	99.4	1926.84 ± 8.31
800	0.4 ± 0.0	0.4 ± 0.0	1.2 ± 0.1	72.4 ± 0.2	19093.5 ± 6.6	300.53	34.2	261.0 ± 1.7	99.4	1921.78 ± 7.74
850	0.3 ± 0.0	0.4 ± 0.1	1.2 ± 0.1	67.2 ± 0.2	18007.5 ± 16.0	300.53	39.5	265.4 ± 2.5	99.4	1941.70 ± 11.25
900	0.4 ± 0.0	0.6 ± 0.0	1.0 ± 0.1	67.1 ± 0.1	18026.5 ± 34.4	300.53	44.8	265.8 ± 2.3	99.4	1943.41 ± 11.25
950	0.5 ± 0.1	0.7 ± 0.1	1.4 ± 0.1	97.1 ± 0.1	25857.0 ± 12.1	300.53	52.4	263.6 ± 2.2	99.4	1933.83 ± 9.91
975	0.6 ± 0.1	0.6 ± 0.0	1.4 ± 0.1	92.9 ± 0.1	24595.6 ± 19.7	300.53	59.8	261.8 ± 1.6	99.3	1925.33 ± 9.91
1000	0.6 ± 0.1	0.7 ± 0.0	1.5 ± 0.0	100.4 ± 0.0	26682.8 ± 15.8	300.53	67.7	263.0 ± 1.3	99.4	1930.90 ± 5.87
1050	0.7 ± 0.0	1.2 ± 0.1	2.4 ± 0.1	156.0 ± 0.1	41337.8 ± 53.4	300.53	80.0	262.5 ± 1.0	99.5	1928.79 ± 4.55
1100	0.8 ± 0.0	1.6 ± 0.0	2.3 ± 0.1	151.2 ± 0.7	40432.5 ± 55.6	300.53	91.9	264.7 ± 2.5	99.4	1938.62 ± 11.06
1200	1.3 ± 0.3	2.7 ± 0.4	0.4 ± 0.4	84.9 ± 0.3	23241.2 ± 28.7	300.53	98.6	268.1 ± 4.5	98.3	1953.60 ± 20.01
1400	0.8 ± 0.7	1.3 ± 0.4	-0.6 ± 0.4	18.4 ± 0.3	5254.4 ± 12.2	300.53	100.0	271.5 ± 28.2	95.3	1968.85 ± 124.55

Table 5.9: Summary of $^{40}\text{Ar}/^{39}\text{Ar}$ results, in age order

Sample	Age (Ma, $\pm 2\sigma$)
DoArBt4	1978.8 $\pm$ 6.2
DoArBt4-2	1975.0 $\pm$ 6.6
DoArBt4-3	1974.0 $\pm$ 5.8
TpArBt1	1950.2 $\pm$ 8.3
DiArBt1	1941.6 $\pm$ 6.6
DiArBt1-2	1929.8 $\pm$ 4.4
AWDDo6	1927 $\pm$ 21
DoArBt2	1926.4 $\pm$ 8.3
AWABv1	1898 $\pm$ 11

5.6.2 Discussion of $^{40}\text{Ar}/^{39}\text{Ar}$ Results

The $^{40}\text{Ar}/^{39}\text{Ar}$ results presented above cover a wide range of ages (approximately 80 Ma) with no identifiable trend related to biotite petrography (c.f. table 5.7). Given the close spatial distribution of these samples, it is unrealistic to suggest that these ages record a true variation in cooling through the same temperature. It is interpreted therefore, that this spread of ages is related to different closure temperatures for each sample. As such, there is the potential to extract a detailed cooling history for the Damang region.

5.6.3 Diffusion and the Concept of Closure Temperature

Loss of radiogenic Ar from a K-bearing mineral is predominantly through diffusion, a process that follows Fick's Second Law:

$$\frac{\partial C}{\partial t} = D \frac{\partial^2 C}{\partial x^2} \quad (5.22)$$

where C is the concentration of the diffusing species and D is the diffusion coefficient. The diffusion coefficient is a function of temperature that is given by the Arrhenius relationship:

$$D = D_0 \exp(-E/RT) \quad (5.23)$$

where D_0 is a frequency factor, E is the activation energy, R is the gas constant and T is absolute temperature. Since, in a cooling setting, T is a function of time, then D must also be a function of time ([Dodson, 1973](#)), given by:

$$D = D(0) \exp(-t/\tau) \quad (5.24)$$

where $D(0)$ is the diffusion coefficient at time zero ($t = 0$) and τ is a constant representing the mean life of the change in D .

Combining equations [5.23](#) and [5.24](#) gives the relationship:

$$\frac{E}{RT} = \frac{t}{\tau} + \ln \left(\frac{D_0}{D(0)} \right) \quad (5.25)$$

Differentiating this with respect to time (t) and rearranging gives:

$$\tau = \frac{-RT^2}{E \, dT/dt} \quad (5.26)$$

which, for a negative dT/dt (i.e. cooling) becomes a positive term.

[McDougall & Harrison \(1999\)](#) describe how the calculation of fractional loss of ^{40}Ar from minerals define expressions in terms of a dimensionless parameter:

$$tD/r^2 (\equiv Fo) \quad (5.27)$$

where r is the radius of the diffusion volume. Substituting equation [5.23](#) in to equation [5.27](#) and rearranging gives:

$$\frac{E}{RT} = \ln(Fo^{-1}tD_0/r^2) \quad (5.28)$$

However in this expression, D is independent of time, which contradicts the necessary relationship given in equation 5.24. So, defining a time varying diffusion coefficient, given by:

$$\zeta = \int_0^1 \frac{D(T)}{r^2} dt \quad (5.29)$$

and substituting in to equation 5.28 gives a general relationship of the form:

$$\frac{E}{RT} = \ln(A\tau D_0/r^2) \quad (5.30)$$

where A is a constant related to the diffusion volume geometry (55 for a sphere, 27 for an infinite cylinder, 8.7 for a plane sheet) (Dodson, 1973). Finally, this can be rearranged to give Dodson's (1973) expression for closure temperature, which is iterative in T_c :

$$T_c = \frac{E / R}{\ln(A\tau D_0/r^2)} \quad (5.31)$$

Dodson (1973) also provides a more accurate description of what a mineral's closure temperature represents, as illustrated in figure 5.12.

Equation 5.31 shows that the closure temperature of a given mineral is a function of a number of parameters including grain size (r), cooling rate (in τ), diffusion model (A) as well as estimates for E and D_0 .

Closure temperature can be related to the idea of a diffusion volume. That is, the mineral volume over which Ar can diffuse. For small grains the diffusion volume and the grain size are the same, and there exists a linear relationship between closure temperature and grain size (Wright *et al.*, 1991; Markley *et al.*, 2002; Alexandre, 2011). However, for larger grains (above a radius of around 250 μm) these studies showed that this relationship breaks down, suggesting a limit to the diffusion volume and implying that larger grains can also lose Ar by other 'multipath' mechanisms (Lee, 1995). By contrast, other studies have shown the

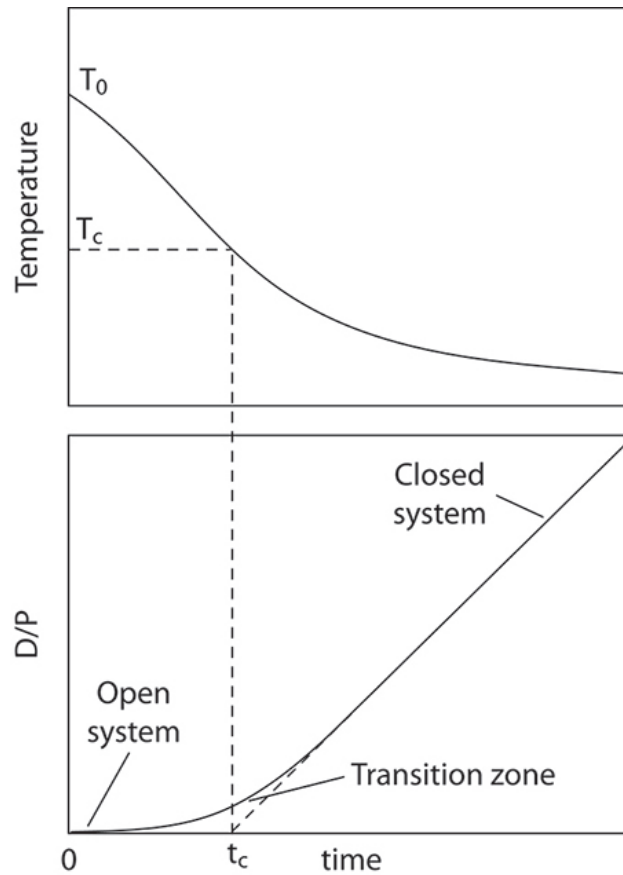


Figure 5.12: Diagrammatic definition of closure temperature in which T_c is closure temperature and t_c is mineral age. During slow cooling, the daughter to parent isotope ratio (D/P) transitions from a completely open state to a completely closed state, after which it increases linearly. T_c and t_c correspond to the extrapolation of this linear trend back to $D/P = 0$. After [Dodson \(1973\)](#).

age-grain size relationship continues to larger grain sizes of over $500\ \mu\text{m}$ up to macroscopic grain scale ([Onstott *et al.*, 1991](#); [Hess *et al.*, 1993](#); [Hodges *et al.*, 1994](#)). More recent ideas regarding this apparent discrepancy include the role of mechanical deformation of grains, which serves to reduce the effective diffusion volume while not affecting the macroscopic grain size ([Baxter, 2010](#)).

The biotite separates analysed in this study cover a broad range of grain sizes. The aliquots from sample DoArBt4 were entirely hand-picked and contain biotite grains that are up to several millimetres in size. Furthermore, these grains are enclosed within a competent quartz vein and as such have likely experienced little microstructural deformation, supporting the idea of a relatively large effective diffusion volume. Samples AWDDo6, DoArBt2 and AWABv1 contain grains up to a maximum diameter of $250\ \mu\text{m}$ due to the separation procedure (see section [5.2.1](#)), giving a maximum r value of $125\ \mu\text{m}$, although some samples

may well be finer than this on average. Samples DiArBt1 and TpArBt1 contain mechanically separated grains supplemented by coarser hand picked grains, such that their average grain size is intermediate to the other two sets. This general trend fits well with the principle of a grain size controlled closure temperature being responsible for the range of calculated ages.

Mineral composition has also been suggested as having a control on closure temperature, with Fe-, octahedral Al- and halogen-rich biotites being less retentive to Ar (Harrison *et al.*, 1985; Grove, 1993, cited in McDougall & Harrison (1999)). This influences equation 5.31 by affecting the values of E and D_0 . All biotites from Damang exhibit a narrow compositional range, particularly in terms of Fe/Mg ratio and octahedral Al contents (section 4.3.1, Fig. 4.1). Therefore, variation in E and D_0 is not thought to have a notable effect on closure temperature.

5.6.4 Calculation of Closure Temperatures

Following equation 5.31, closure temperatures have been calculated for a range of regional cooling rates and grain sizes, using an infinite cylinder diffusion model ($A=27$) (Harrison *et al.*, 1985) and values of E and D_0 after Grove & Harrison (1996) ($E = 47.1 \pm 1.5$ kcal/mol, $D_0 = 0.075^{+0.049}_{-0.021}$ cm²/s). These results are presented in table 5.10.

To a first order approximation, initial cooling rates are on the order of 10°C/Ma, assuming a linear cooling path between peak metamorphic conditions ($\sim 595^\circ\text{C}$ at ~ 2005 Ma) and closure of the oldest biotite samples ($\sim 300^\circ\text{C}$ at ~ 1975 Ma). As described above, these oldest samples are the coarsest and, using an r value of 500 μm , a cooling rate is 10°C/Ma, indicates a closure temperature of around 347°C.

If the range of calculated $^{40}\text{Ar}/^{39}\text{Ar}$ ages is the result of grain size variation, the youngest sample, at around 1900 Ma, would be inferred to have the smallest grain size. The smallest average grain size that could reasonably be expected for this, or any other, sample would be around 200 μm , giving an r value of around 100 μm . Table 5.10 shows that for any given cooling rate, the difference in closure temperature between a grain with $r = 500$ μm and

Table 5.10: Calculated closure temperatures in biotite as a function of cooling rate and grain size using the formula of [Dodson \(1973\)](#).

r (μm)	Cooling Rate $^{\circ}\text{C}/\text{Ma}$					
	0.1	0.5	1	5	10	20
50	231	248	255	274	283	292
100	245	263	272	292	301	310
200	261	280	289	310	320	329
250	266	286	294	316	326	339
500	283	304	313	336	347	357

one with $r = 100 \mu\text{m}$ is around $50\text{--}60^{\circ}\text{C}$. This implies a cooling rate between the oldest and youngest samples on the order of $0.75^{\circ}\text{C}/\text{Ma}$.

These simple calculations indicate that if the regional cooling history involves initial, relatively rapid cooling, followed by a period of much slower cooling, then the observed spread of ages can indeed be accounted for by a realistic variation in grain size.

5.6.5 DIFFARG Modelling and a Regional Cooling History

A more detailed investigation of the link between measured $^{40}\text{Ar}/^{39}\text{Ar}$ ages and regional cooling can be made by modelling the diffusion of argon in biotite over time with the program DIFFARG ([Wheeler, 1996](#)), which comprises a MATLAB[®] ([MATLAB[®], 2009](#)) script run, in this study, inside the freely available GNU Octave ([Eaton, 2002](#), <http://www.octave.org>). DIFFARG is a one-dimensional finite difference model that solves equation 5.22, combined with a source term that represents ^{40}Ar production over the run time of the model. Model input includes a cooling history as a sequence of temperature-time coordinates, a mineral type, shape and grain radius, a spatial grid mesh size, and the model time until the next output display (i.e. the length of time for that ‘run’).

The grid mesh defines the resolution of the grid used for the finite difference calculations, with a larger mesh size (i.e. smaller grid node separation) producing more ‘accurate’ results. For mesh sizes above around 20 nodes there exists a broadly linear relationship between grid size and apparent age (C. Warren, *pers. comm.*). So, if the model is run with a range of different mesh sizes, a regression of this mesh size against apparent age can be used to give

the apparent age at an effectively infinite mesh size.

DIFFARG also allows the use of either a fully explicit or a Crank-Nicholson algorithm (Crank, 1975). For the former, each run is divided into a non-arbitrary number of time steps, used in the calculations. With the latter however, a user-defined time step forms a scaling factor against the fully explicit algorithm. For example, inputting a time step of 10 results in each time step being 10 times larger than in the fully explicit case. The primary benefit of this is to reduce computational time without compromising excessively on result accuracy.

In this study, models were run using grain radii of 100 and 500 μm , mesh sizes of 20, 40 and 80, and using the Crank-Nicholson algorithm with a time step of 10. Cooling histories were then varied in order to match model and measured ages. All models were run with a fixed grain rim apparent age of 0 Ma, meaning that once Ar has diffused to the edge of the grain it is removed from the system and there does not exist an Ar reservoir in the pore fluid. Although a full cooling history from peak metamorphic conditions (approximately 595°C at 2005 Ma) is being considered, to save on computational time model runs were begun 5 Ma after peak metamorphism, at a temperature closer to 500 °C (depending on cooling rate chosen). In all cases biotites are still completely open to Ar at this temperature and so bulk age calculations are not affected. Models were run for a total of 400 Ma to ensure that the linear portion of the time vs. apparent age graph (lower panel of figure 5.12) has been reached.

Preliminary investigations showed that the cooling rates given in section 5.6.4 ultimately result in ages that are too young, particularly for the coarser (older) samples, as the lower cooling rate during the second stage of the model continues to drive Ar loss, rather than the system becoming closed precisely at around 347°C (c.f. figure 5.12). The implication is that a higher initial cooling rate is required, followed by a prolonged period of much slower cooling.

Results of the cooling history model that best fits the measured $^{40}\text{Ar}/^{39}\text{Ar}$ ages are presented in figure 5.13. Initial cooling at a rate of 17°C/Ma is followed by prolonged cooling at a rate of 0.15°C/Ma for the remainder of the model run. The modelled age for a 500 μm biotite is

1977.7 Ma, which is within error of all DoArBt4 aliquot samples. The calculated age for a 100 μm biotite is 1919.3 Ma, which is within error of both AWDDo6 and DoArBt2, and marginally older than sample AWABv1. One possible explanation for this slight discrepancy with sample AWABv1 is that being part of the Birimian basement, this sample would have been subjected to pre-Tarkwaian deformation. As such the biotite grains may be more microstructurally deformed than other samples, which would result in them producing a younger $^{40}\text{Ar}/^{39}\text{Ar}$ age.

Ultimately, this simple two stage cooling model fits well with the measured $^{40}\text{Ar}/^{39}\text{Ar}$ ages and highlights the general form of the regional cooling history, with initially rapid cooling followed by much more prolonged slow cooling. DIFFARG modelling was begun at 2005 Ma, although given the error on this value of ± 26 Ma, the first stage of the cooling could show some variation, for example, if cooling started earlier then a slightly lower cooling rate would be required. By contrast, the second stage of cooling is essentially fixed by the spread of measured ages which, if they are indeed a function of grain size, necessitates such a prolonged slow cooling rate.

5.7 Discussion of the Geochronology of the Damang Deposit

The results of the geochronological study presented here provide reliable time constraints on the magmatic, metamorphic and thermal history at Damang (Fig. 5.14).

U/Pb zircon ages from Birimian age igneous rocks are consistent with those that have been reported from across Ghana (2137–2200 Ma) (see section 2.5).

U-Th-total Pb monazite dating of the Tarkwa Phyllite places regional metamorphism at 2005 ± 26 Ma, which is notably younger than other estimates of regional metamorphism in Ghana, which is generally placed at 2086 ± 4 – 2106 ± 2 Ma (see section 2.5). Within any internally-heated metamorphic terrane, more deeply buried (and therefore hotter) rocks may reach their metamorphic peak at a later time (England & Thompson, 1984; Thompson & England, 1984; Stüwe *et al.*, 1993; Stüwe, 1998). Although a time difference of approximately

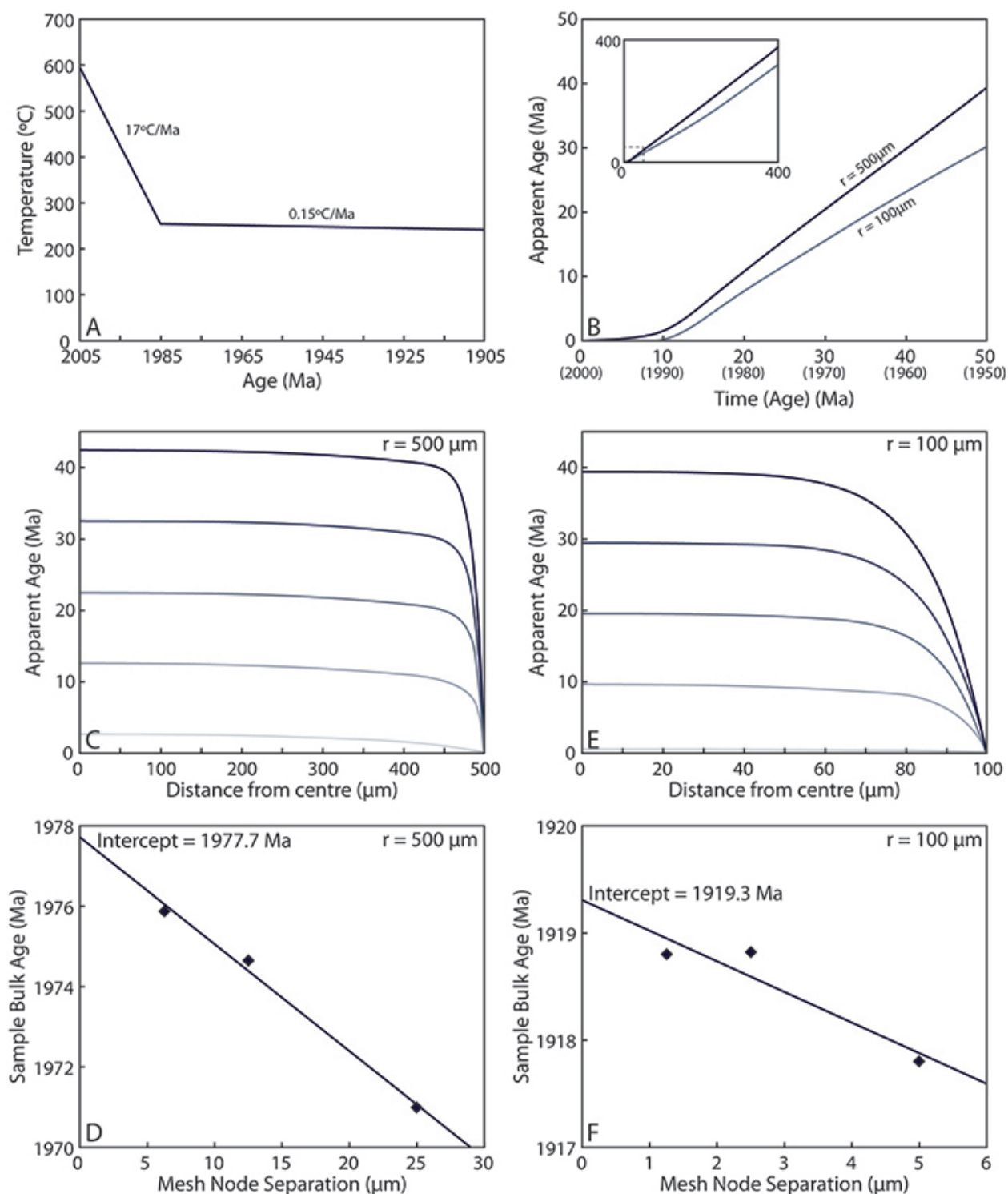


Figure 5.13: Results of DIFFARG modelling of $^{40}\text{Ar}/^{39}\text{Ar}$ ages. A is the modelled cooling history. B shows apparent sample age against model run time, similar to the lower panel of figure 5.12, for grain radii of 100 and $500\ \mu\text{m}$. The main panel shows the first 50 Ma of the model, while the inset shows the result after 400 Ma. This clearly shows how a smaller grain size results in a younger bulk age and results in the system staying open for a longer period. C shows the apparent age as a function of position within a $500\ \mu\text{m}$ radius biotite grain, with profiles drawn every 10 Ma for the first 50 Ma of the model. D gives apparent bulk age as a function of mesh size, for a $500\ \mu\text{m}$, radius biotite grain, with the intercept of the regression giving the most reliable estimate of a model age. Panels E and F are as for C and D respectively, but for a $100\ \mu\text{m}$ radius biotite grain.

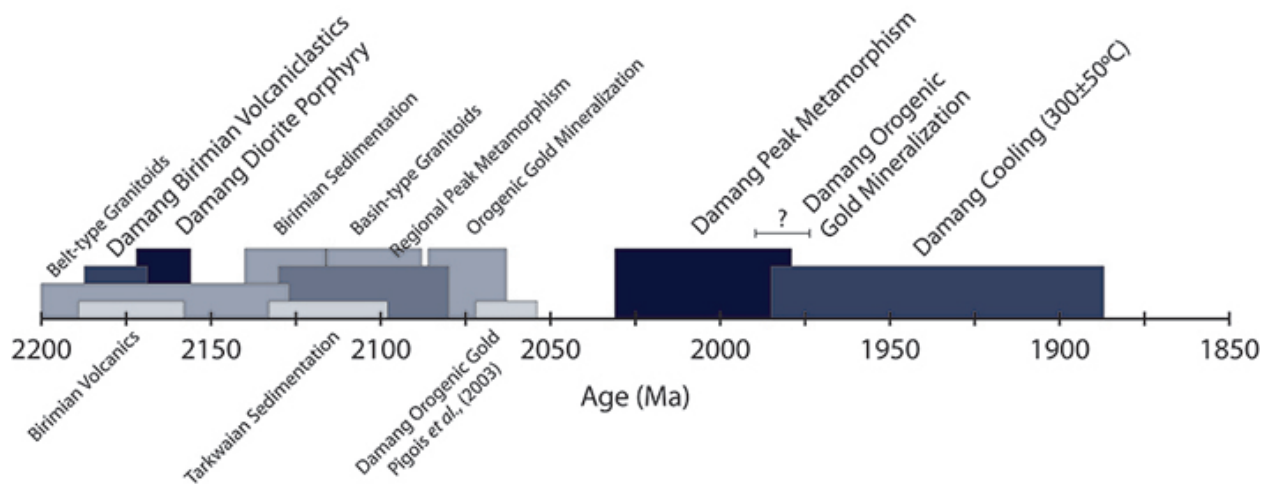


Figure 5.14: Summary of geochronology results in the context of regional Ghanaian chronology (see section 2.5). Darker blue boxes are results from this study, while paler blue boxes are from published data (Table 2.2). Birimian igneous rocks at Damang are consistent with Belt-type granitoids and typical Birimian volcanics. Regional metamorphism, and consequently gold mineralization, is significantly younger than determined elsewhere in Ghana due to Damang's differing (higher-grade) metamorphic history.

80 Ma may be too long to be explained by this process alone, especially considering there is no evidence for even higher grade metamorphism and partial melting later than 2005 Ma. This will be addressed further in chapter 7. The age of metamorphism determined here does correlate with the $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende ages of 1978 ± 37 and 2006 ± 40 Ma from Birimian volcanics in Ghana presented by Feybesse *et al.* (2006). These ages represent regional cooling through approximately $500\pm50^\circ\text{C}$, which despite being a slightly lower temperature than peak metamorphism at Damang, still provides a broadly comparable age estimate.

Re/Os isotopic dating of sulphide minerals did not produce a precise age and so the exact timing of the Damang mineralization event is not well constrained through this method. However, the timing of mineralization can be indirectly constrained to lie between regional metamorphism at 2005 ± 26 Ma and $^{40}\text{Ar}/^{39}\text{Ar}$ biotite ages beginning at 1978.8 ± 6.2 Ma. This estimate is again younger than previous estimates for the timing of orogenic gold mineralization in Ghana, which are typically around 2080–2110 Ma. However, given that these dates are also indirectly determined from estimates of regional metamorphism, this is to be expected.

More importantly however, this proposed age range for mineralization does not agree with the U/Pb hydrothermal xenotime dating of Pigois *et al.* (2003), which constrained miner-

alization at Damang to 2063 ± 9 Ma. The analysed xenotime grains were determined to be hydrothermal in origin since they were only found within mineralized samples. By contrast, in this study very little xenotime was found, precluding a geochronological study. Occasional occurrences of xenotime were, however, identified in unaltered samples of the Tarkwa Phyllite, suggesting that they may have an origin other than a purely hydrothermal one. In addition, a number of analyses lie above concordia (Fig. 13 of [Pigois *et al.* \(2003\)](#)), suggesting the influence of excess ^{206}Pb in at least some of the grains.

The $^{40}\text{Ar}/^{39}\text{Ar}$ biotite ages obtained in this study are also significantly younger than those presented by [Pigois *et al.* \(2003\)](#) at 2029 ± 4 and 2034 ± 4 Ma for Damang. A detailed explanation for this discrepancy cannot be provided. However, these two ages represent two aliquots from a single sample, and while they are, expectedly, consistent, they are not otherwise verified with any other samples. Similarly, the ages given by [Pigois *et al.* \(2003\)](#) are consistent with their estimate for the timing of mineralization (see above), but since that age is also uncertain in the context of this study, the entire history cannot be fully verified.

Overall, the ages determined for Birimian igneous activity, regional metamorphism and regional cooling in this study provide an internal consistency that broadly correlates with the regional framework. The younger estimates for regional metamorphism and mineralization are interpreted to be due to Damang's higher grade metamorphic history. Although the differences with the study of [Pigois *et al.* \(2003\)](#) cannot be so readily explained, this study is rather more comprehensive, particularly with regards to $^{40}\text{Ar}/^{39}\text{Ar}$ analysis.

Chapter 6

The Application of Portable Infrared Reflectance Spectroscopy



The 'Galamsey', local artisanal miners at Amoanda.

Infrared (IR) reflectance spectroscopy is an extremely useful tool for the quick identification of minerals as well as certain crystallographic and compositional properties. As is discussed below, IR spectroscopy is particularly useful for the identification of hydrous minerals, including micas, clays and amphiboles, as well as carbonates. This makes it ideal for studying hydrothermal systems, which are commonly characterized by alteration assemblages dominated by these mineral types. Identification of different mineral phases, assemblages and physical properties throughout hydrothermal systems can shed light on the nature of hydrothermal fluids, which in turn can be used to better understand the ore forming process. These mineralogical changes can ultimately be used as vectors to mineralization that are employed during exploration for new or larger deposits.

IR spectroscopy can be implemented through airborne hyperspectral surveys or through portable, field-based spectrometers. Both of these methods have proved invaluable to understanding ore systems as well as a number of other applications in Western Australia and the USA where the proportion of outcrop is relatively high (Prost, 1980; Buckingham & Sommer, 1983; Gladwell *et al.*, 1983; Whitney *et al.*, 1983; Crowley & Zimbelman, 1997; Thompson *et al.*, 1999; Rowan & Mars, 2003, Gold Fields Ltd., *pers. comm.*). However, in Ghana and other tropical – sub-tropical terrains, high levels of vegetation and a deep weathering profile have hindered its development, although some success has been documented (Swalf *et al.*, 2003). While airborne surveys are virtually impossible due to the vegetation cover, portable spectrometers that can be carried between the trees are still a viable option.

6.1 Principles of Infrared Reflectance Spectroscopy

Infrared light covers the portion of the electromagnetic spectrum (Fig. 6.1) between wavelengths of 700 nm and 100 μ m and can be further differentiated into four subdivisions:

- near infrared (700–1350 nm),
- short wavelength infrared (1350–2500 nm),
- mid infrared (2500–30,000 nm) and

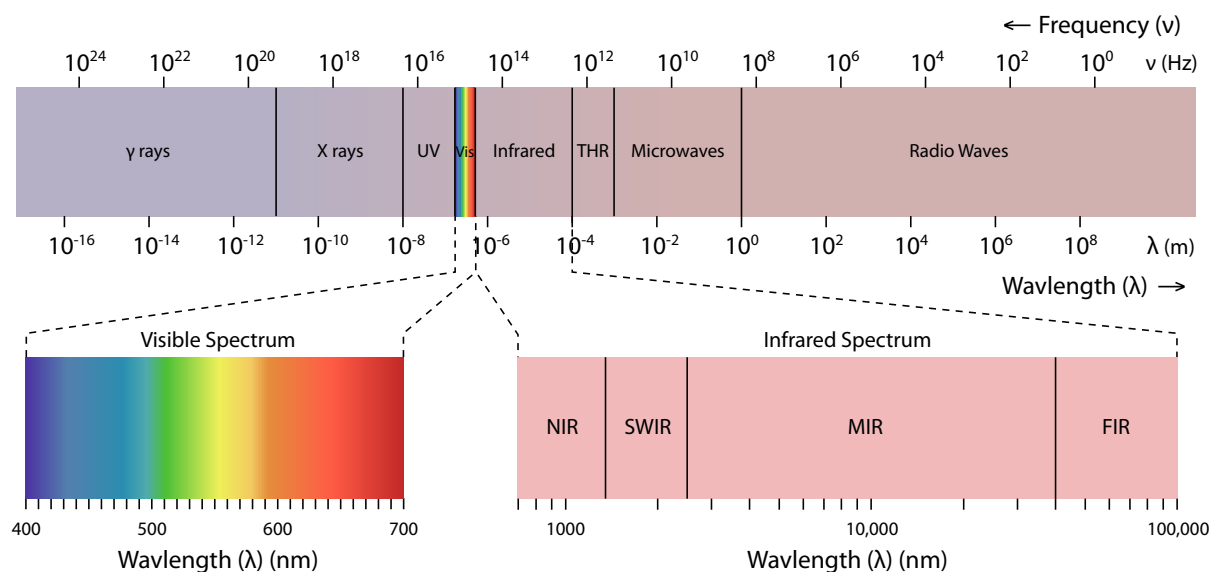


Figure 6.1: The electromagnetic (EM) spectrum showing the major subdivisions of the entire spectrum and the Infrared Region. Abbreviations: UV, Ultraviolet; Vis, Visible; THR, Terahertz Radiation; NIR, Near Infrared; SWIR, Short Wavelength Infrared; MIR, Mid Infrared; FIR, Far Infrared

- far infrared (30–100 μm).

Infrared reflectance spectroscopy measures the amount of infrared light reflected from a sample for a given wavelength. Atomic processes within the minerals in the sample cause a certain proportion of the incident light to be absorbed leaving ‘gaps’ beneath the reflected spectrum baseline. These absorption features occur at specific wavelengths depending up on the mineralogy of the sample allowing for quick and reliable mineral identification to be made through simple pattern matching.

6.1.1 The Processes of Absorption

When a sample is illuminated by an infrared light source the photons are either reflected or scattered from the surface of grains to be detected and measured by the spectrometer, or are absorbed by the constituent minerals. This absorption occurs at specific wavelengths, dependent upon which minerals are present as well as their chemistry and other physical properties.

Absorption of photons by a material follows Beers Law:

$$I = I_o \exp(-kx) \quad (6.1)$$

where I is the observed light intensity, I_o is the original light intensity, k is the absorption coefficient in units of length^{-1} and x is the distance travelled through the material. k is dependent upon the absorbing material and upon wavelength for each individual material. Photons are absorbed by minerals through a variety of different processes ([Clark, 1995](#)).

Electronic Processes

Electronic processes are those relating to the energy states or ‘energy bands’ of electrons shared between atoms within a mineral, or ‘crystal field’. These atoms or ions can absorb photons of a particular wavelength, resulting in a change from one energy state to another. This is predominantly encountered in minerals containing transition elements with unfilled electron shells, particularly iron, as it is the most common. The energy levels of unfilled orbitals around the atom of a transition element within a crystal are split into two levels. Electrons can be moved from the lower to the higher energy level by the absorption of a photon if it has same energy as the difference between the two levels. The transition of an electron from a higher to a lower energy level results in the emission of a photon at a different wavelength to that that was absorbed. These energy levels and their splitting are controlled by several factors: the valence state of the atom, its coordination number, the symmetry of its crystallographic site and the degree of distortion, and the type of ligands formed and the metal-ligand interatomic distance. Clearly, with such a range of complex factors, the energy levels for any particular atom will vary greatly from one mineral to another. Therefore, the wavelength of absorbed photons will also vary, for the same atom, between minerals.

Inter-element transitions, or charge transfers, occur by a similar process but involve the movement of an electron between ions and/or ligands, or between valence states in the same element. The absorption resulting from these processes are in fact two to three orders of magnitude stronger than the crystal field transitions and are also highly diagnostic of

mineralogy. Iron again is strongly affected by these processes and has its primary absorption bands centred in the ultraviolet, spreading up into the visible blue. This is what gives iron oxides their red colour in visible light.

Vibrational Processes

The bonds between atoms within a mineral are not static but are constantly vibrating. This vibration covers a range of movement including stretching, bending and twisting, known as modes, with the frequency (and therefore wavelength) of the vibration controlled by the strength of the bond and the mass of the atoms. It is the absorption of photons at these specific wavelengths that result in characteristic spectral absorption patterns. A molecule with N atoms has $3N - 6$ normal modes of vibration, termed ‘fundamentals’. Additional vibrations also occur at wavelengths that are approximately multiples of the fundamentals. These are termed overtones if they are multiples of a single fundamental, or combinations if they are due to different types of vibration. Each successive overtone or combination is 30 to 100 times weaker than the previous one. However, modern infrared reflectance spectrometers can easily measure even the weak second and third overtones and combinations, thereby extracting diagnostic information from the highly complex spectra of many minerals.

6.1.2 Short Wavelength Infrared Spectroscopy

Short wavelength infrared (SWIR) spectroscopy primarily utilizes vibrational absorption process, particularly those related to the bonds within certain anionic complexes, such as hydroxyl (OH), water (H₂O), carbonate (CO₃) and to a lesser extent ammonia (NH₄) (Fig. 6.2).

Water has 3 fundamental vibrations, which in the vapour phase occur at 2663 nm (asymmetric OH stretch), 2738 nm (symmetric OH stretch) and 6270 nm (H-O-H bend). These are shifted in liquid water due to hydrogen bonding. Clearly these wavelengths are within the mid-infrared and so the absorption features observed in the SWIR are in fact the first overtones and combinations of these. The first overtones of the OH stretches occur close to

1400 nm, while the combinations of the OH stretches and the H-O-H bend is at about 1900 nm (Fig. 6.2). Therefore, a mineral that produces a spectrum with a feature at 1400 nm, but not at 1900 nm contains OH but no water.

The hydroxyl ion has a single stretching mode but its wavelength is determined by the ion to which it is bonded. In addition, due to OH ions typically residing in a range of different crystallographic sites and bonded to many different cations, there can be several different 'OH' absorption features in a single spectrum. The most important of these involve metal-OH bonds, such as AlOH, FeOH and MgOH which have diagnostic combinations (metal-OH bend and OH stretch) at approximately 2160-2220 nm, 2230-2295 nm and 2300-2360 nm respectively (Fig. 6.2). Additional minor absorptions can commonly be observed in clays close to 1550 nm and between 1750-1850 nm.

The carbonate CO_3^{2-} anion exhibits four fundamentals, with another two being degenerate, that is they occur at the same wavelength and thus cannot be distinguished. They are at 7067 nm (asymmetric stretch), 9407 nm (symmetric stretch, which is not infrared active in minerals), 11,400 nm (out-of-plane bend) and 14,700 nm (in-plane bend). Again, these fundamentals are in the mid-infrared and higher so it is overtones and combinations of these that are measured in the SWIR. In particular, an overtone of the asymmetric stretch at 2300-2350 nm is highly diagnostic (Fig. 6.2). Additional, more minor, carbonate absorption features also occur at 1870 nm, 1990 nm and 2155 nm and these can be used to distinguish between carbonate and MgOH, which both have their largest SWIR absorption features in the same range of wavelengths. Care must also be taken when interpreting features in the MgOH/ CO_3^{2-} band as secondary AlOH absorptions occur in this range (*c.f.* kaolinite and illite in figure 6.2). If however, a feature in this band is significantly deeper than the AlOH absorption, it can be taken to indicate the presence of MgOH and/or carbonate.

Any mineral which incorporates any of the above molecules as major components are suitable for SWIR analysis: phyllosilicates and other hydrous silicates (e.g. epidote and amphibole), carbonates, and certain sulphates and ammonium-bearing minerals. Dominant mineral phases that are present in the samples but do not contain any of the above com-

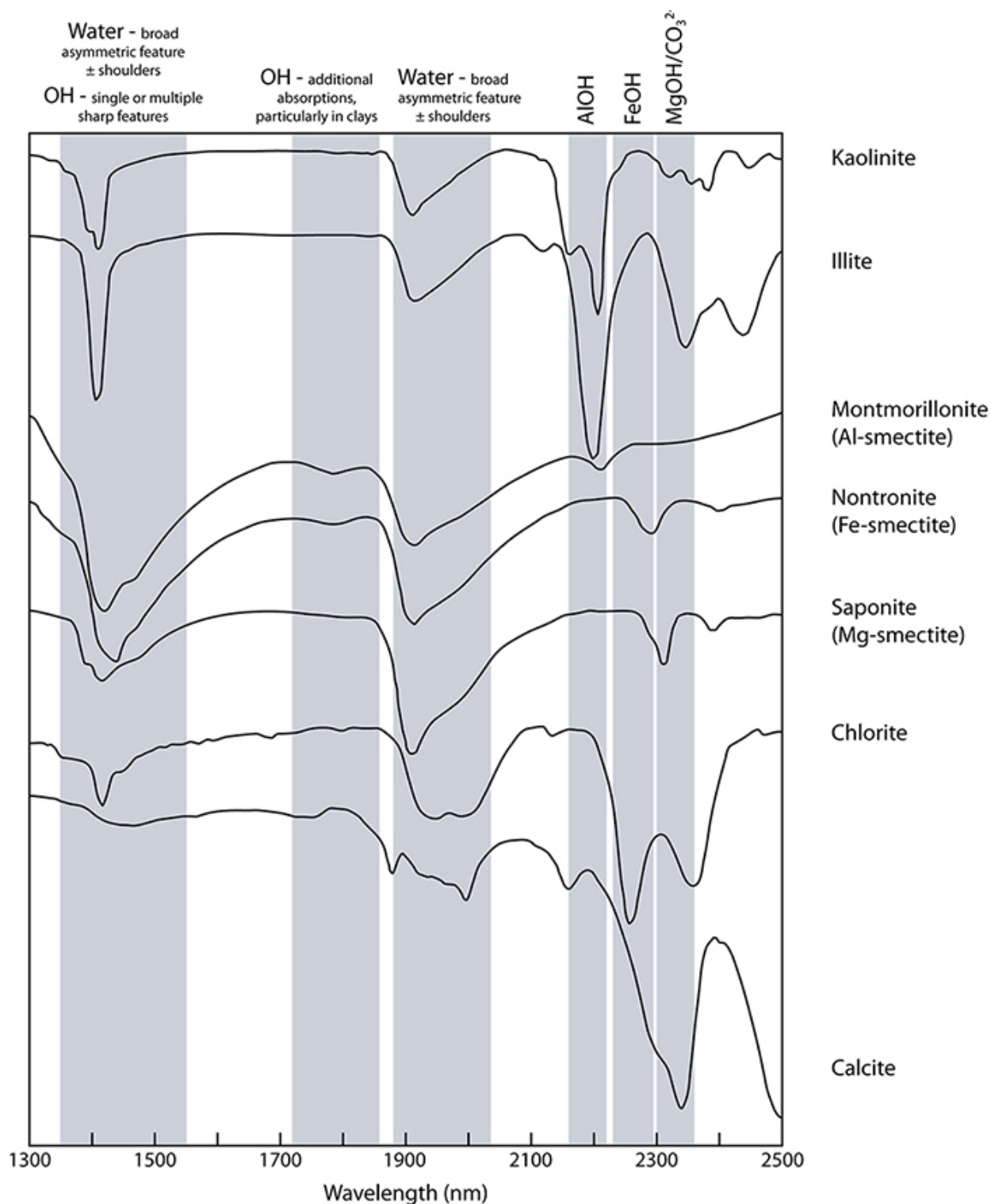


Figure 6.2: Example spectra for a range of minerals in the SWIR.

ponents (e.g. quartz and feldspar) will therefore not show up in SWIR analysis. However, any relatively minor mineral constituents may still be recognizable. This is extremely useful for the study of hydrothermal systems since the majority of the alteration occurs within these phases, as opposed to quartz and feldspar. There are a number of mineral groups that

are detrimental to SWIR analysis due to their very low reflectance (such as sulphides, magnetite and carbonaceous material) and can reduce the appearance of otherwise characteristic features.

6.1.3 Visible – Near Infrared Spectroscopy

The majority of modern infrared spectrometers measure reflected light down to wavelengths of 350 nm, covering the whole of the visible spectrum as well as the near infrared (NIR) and the SWIR. Absorption features observed in the visible – near infrared (VNIR) are primarily due to electronic processes in iron, and other transition metals such as copper and manganese, and are typically much broader than those in the SWIR. Any mineral that contains a significant amount of iron will exhibit absorption features in the VNIR: iron oxides, Fe-bearing silicates such as pyroxenes, chlorites and biotites, sulphates, Fe-, Cu- and Mn-carbonates, and other Mn-bearing silicates.

6.2 Analytical Methods

All infrared spectra in this study were collected using an Analytical Spectral Devices Inc. (ASD) TerraspecTM spectrometer owned by Gold Fields Ltd., with each spectra representing 60 integrations over a 20 second interval. Samples were taken from a large number of reverse-circulation (RC), diamond-drill (DD) and air-core (AC) drill cores from across the Damang region, covering the full range of lithologies, both palaeoplacer and hydrothermal mineralization styles as well as a diversity of gold grades. These downhole spectra were obtained at 1m intervals (for RC and AC cores), or less (for DD cores), from laboratory pulps.

Pulps were used in this study as they provided the largest sample database in terms of both the number of samples as well as the range of lithologies and alteration styles. The pulp making process can have adverse effects on the clay minerals in a sample, although this was not considered a problem in this case as clay minerals do not form a major component of the host

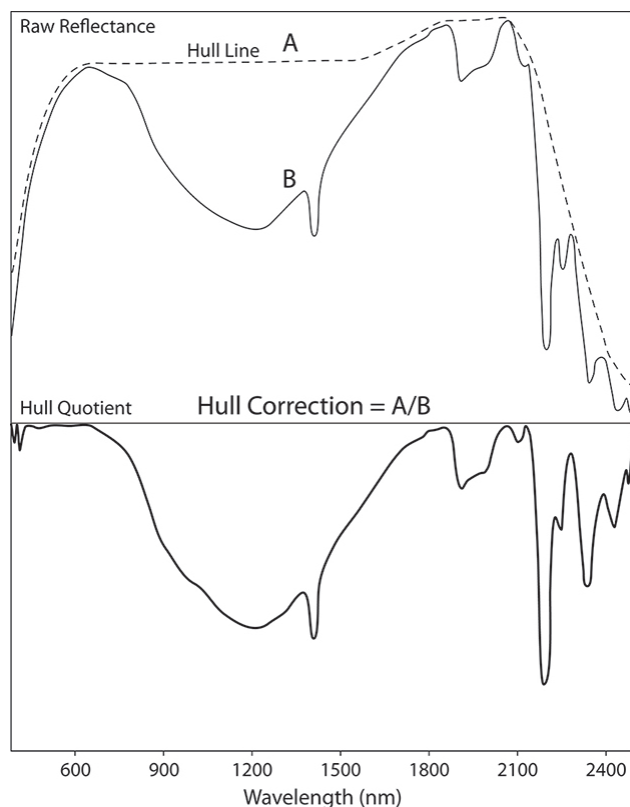


Figure 6.3: Calculation of the Hull quotient from an infrared reflectance spectrum. The upper panel represents the raw data collected by the spectrometer, while the lower panel is the adjusted Hull quotient.

rock or alteration assemblages. Pulps typically produce brighter (more reflective) spectra compared to fresh rock, but with more subdued absorption features, which reduces the ability to distinguish fine variation in mineralogy and makes it difficult to identify low reflectivity minerals. This is not an issue in the study, however, as the mineralogical changes associated with the targeted hydrothermal alteration are significant. The benefit of using pulps is evident in a homogenization effect, providing representative sampling mineral proportions, as well as the opportunity to compare spectra directly with gold assays. Spectral analysis and calculation of the hull quotient was completed using The Spectral GeologistTM software package. The Hull quotient is calculated as per figure 6.3 and serves to enhance absorption features, making them more readily identifiable, and acts as a normalization procedure.

6.3 Results of Infrared Spectral Analysis

6.3.1 Diagnostic Spectral Parameters

Several distinct spectral parameters have been identified that characterize particular aspects of the Damang gold deposit, as discussed below, and summarized in Table 6.1 and Figure 6.4. Of those, four parameters are the most useful when considering hydrothermal mineralization zones at Damang: the Ferrous-iron Response, AlOH/H₂O Absorption Depth Ratio, Water/OH Absorption Depth, and Automated Mineral Identification, all of which are obtained directly via The Spectral GeologistTM software package.

The Ferrous-iron Response spectral parameter measures the gradient of the reflectance spectrum (i.e. the raw data collected directly by the spectrometer) between the wavelengths of 1350 nm and 1650 nm, with a value greater than 1 indicating the presence of a positive slope (Fig. 6.4A). An increase in this value indicates an increase in the relative proportion of Fe²⁺-bearing phases, which result in a broad absorption outside of the short wavelength range, centred at approximately 1000 nm. This parameter becomes significantly reduced when sulphides and/or magnetite are present due to their ability to lower overall sample reflectance.

All other parameters discussed here are calculated using the hull quotient spectrum (Fig. 6.4B), which allows for more accurate measurements of absorption wavelengths and relative absorption depths. Relative absorption depth refers to the depth of an absorption feature from a horizontal baseline coincident with the maximum value for the individual spectrum, down to the minimum value within the selected range.

The spectral parameter AlOH/H₂O Absorption Depth Ratio is a measure of the relative absorption depths of the AlOH and H₂O absorption features (depth AlOH/depth H₂O), which occur at 2204 ± 24 nm and 1910 ± 30 nm respectively. This parameter is commonly used in studies of white micas as a measure of ‘crystallinity’, whereby increasing values, caused by a deepening and sharpening of the AlOH absorption feature, and representing

increased crystallographic ordering associated with the $2M$ versus $1M$ polytype (Kruse & Hauff, 1991). In lower temperature settings, an increase in ‘crystallinity’ is marked by a shallowing of the H_2O absorption feature due to a decrease in the number of water-bearing smectite layers mixed into the white mica crystal structure. The parameter therefore represents the mica/smectite ratio in a multi-layered crystal. The $AlOH/H_2O$ Absorption Depth Ratio can be applied to non-white mica bearing samples as an overall measure of the relative proportions of $AlOH$ and H_2O bonds, which is controlled by the relative proportions of minerals, each with a different $AlOH/H_2O$ Absorption Depth Ratio.

The spectral parameter Water/ OH Absorption Depth measures the relative absorption depth of the Water/ OH feature at 1400 ± 50 nm. The term ‘Water’ is used here to distinguish between this feature and the ‘ H_2O ’ feature discussed above. An increased Water/ OH Absorption Depth should indicate an increase in the proportion of overall water or O-H bonds in the sample. The broad, shallow absorption feature of water is distinguishable from the typically deeper and narrower OH absorption. This parameter is also useful for determining whether water in a sample is lattice bound water or whether it is simply ‘wet’ by assessing the overall shape of the water feature.

The Spectral AssistantTM is an automated mineral identification tool that matches collected spectra against a library of known mineral shortwave infrared spectra (Berman *et al.*, 1999). The spectral library is built from a large number of reference samples with each stored ‘typical’ spectra representing the average of measured natural variation.

The spectral parameters $AlOH$ Wavelength and $MgOH$ Wavelength, in addition to $AlOH/H_2O$ Absorption Depth Ratio, are particularly helpful in the identification and characterization of the weathering profile at Damang. These two ‘wavelength’ parameters record the wavelength at which the minima of the $AlOH$ and $MgOH$ absorption features occur, universally in the ranges 2204 ± 24 nm and 2348 ± 48 nm respectively.

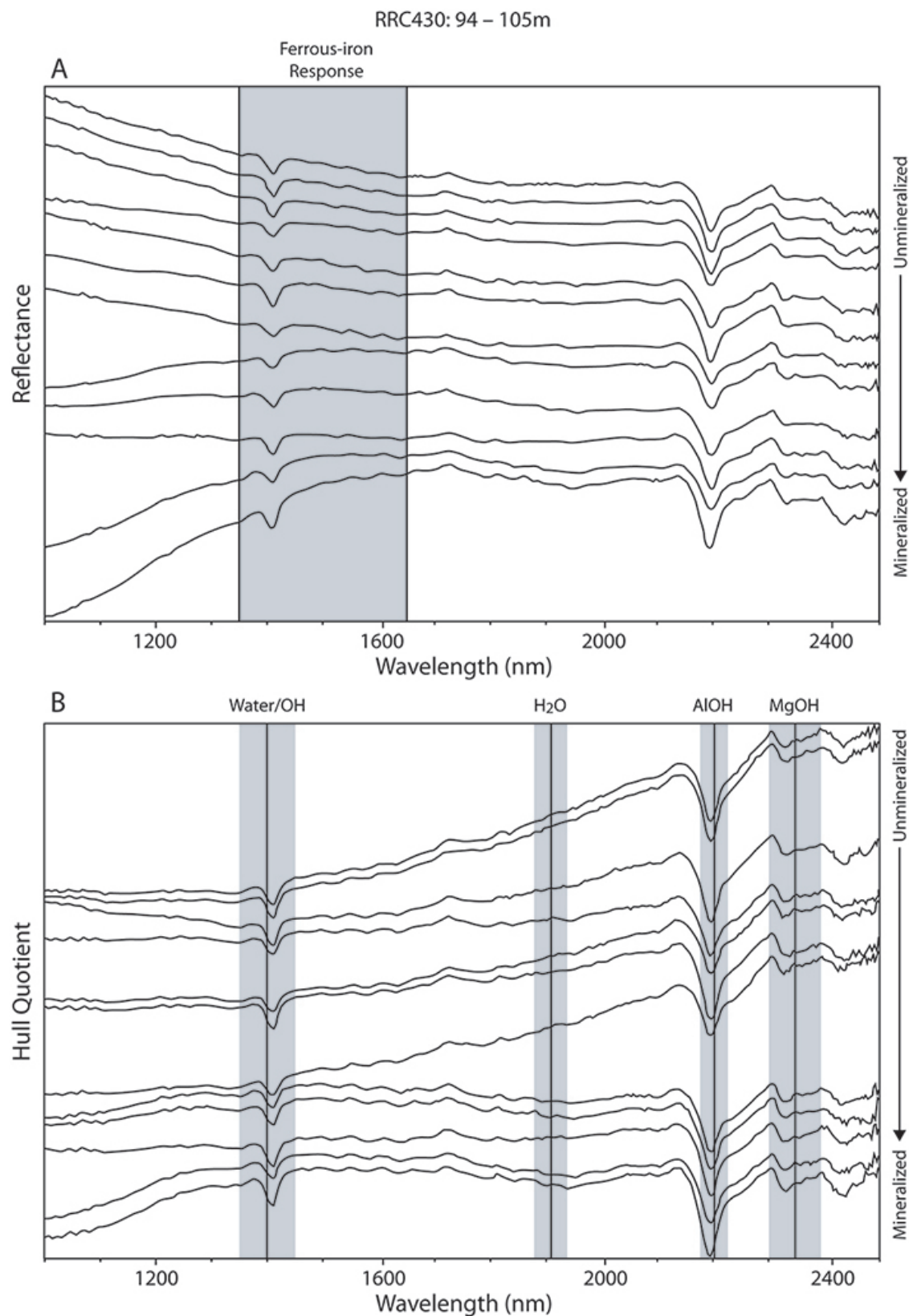


Figure 6.4: Diagram illustrating changes in spectral features as a function of progressive increasing hydrothermal alteration in the Banket Quartzite. Plots of spectra between 94–105m depth (top–bottom) in exploration drill core RRC430, at 1m intervals. The reflectance spectra (A) show the raw data collected by the spectrometer, while B presents the corrected hull quotient for the same spectra. Shaded zones represent the wavelength ranges of diagnostic spectral parameters discussed in the text.

Table 6.1: Summary table of diagnostic spectral parameters used in this study.

Parameter Name	Wavelength Range (nm)	Description
Ferrous-iron Response	1350–1650	Gradient of the reflectance spectra across the wavelength range. Values greater than 1 indicate the presence of Fe ²⁺ -bearing phases.
AlOH/H ₂ O Absorption Depth Ratio	1910 ± 30, 2204 ± 24	Relative absorption depth of AlOH divided by the relative absorption depth of H ₂ O.
Water/OH Absorption Depth	1400 ± 50	Relative absorption of the Water/OH feature.
AlOH Wavelength	2204 ± 24	Wavelength of the minimum for the AlOH feature.
MgOH Wavelength	2348 ± 48	Wavelength of the minimum for the MgOH feature.
TSA Mineral Identification	-	Spectral pattern matching by The Spectral Assistant™

6.3.2 Spectral Characterization of Unmineralized Lithologies

The Ferrous-iron Response and $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio spectral parameters are the most useful in identifying both unmineralized and mineralized (see below) lithologies. The widely differing mineral assemblages in the Damang sediments compared the dolerites results in measurable variations in these two spectral parameters. However, in hand sample, and especially in core chips, the dark colour and fine-grained nature of both lithologies can often make them difficult to identify, necessitating the need for an alternative technique of lithology identification. Unmineralized sediments typically exhibit lower values for both the Ferrous-iron Response and $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio than do the unmineralized dolerites (Figs. 6.5, 6.6 and 6.7). In addition, as expected, automated mineral identification effectively distinguishes the mineral assemblages of the two lithologies. While the sediments are dominated by the identification of muscovite and/or illite (although from petrographic observation, all white mica is muscovite), the dolerites are typically characterized by hornblende $\pm$ epidote. Despite observed variation in individual spectral parameters, the use of all three together to identify a particular lithology is dependable.

6.3.3 Spectral Correlation with Gold Mineralization

The spectral parameters Ferrous-iron Response, $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio, Water/OH Absorption Depth and Mineral Identification exhibit strong down hole (Figs. 6.5 and 6.6) and three-dimensional (Fig. 6.7) spatial correlation with gold mineralization in both the sediments (Banket Quartzite and Tarkwa Phyllite) and dolerites at Damang.

Within the sedimentary rocks calculated values of the Ferrous-iron Response and $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio both exhibit localized, roughly symmetrical, increases around zones of gold mineralization (Fig. 6.5). In contrast, Water/OH Absorption Depth shows a corresponding decrease associated with gold mineralization (Fig. 6.5). Automated mineral identification is not particularly useful in resolving gold mineralization in the Damang sediments because both altered and unaltered mineral assemblages are characterized by the

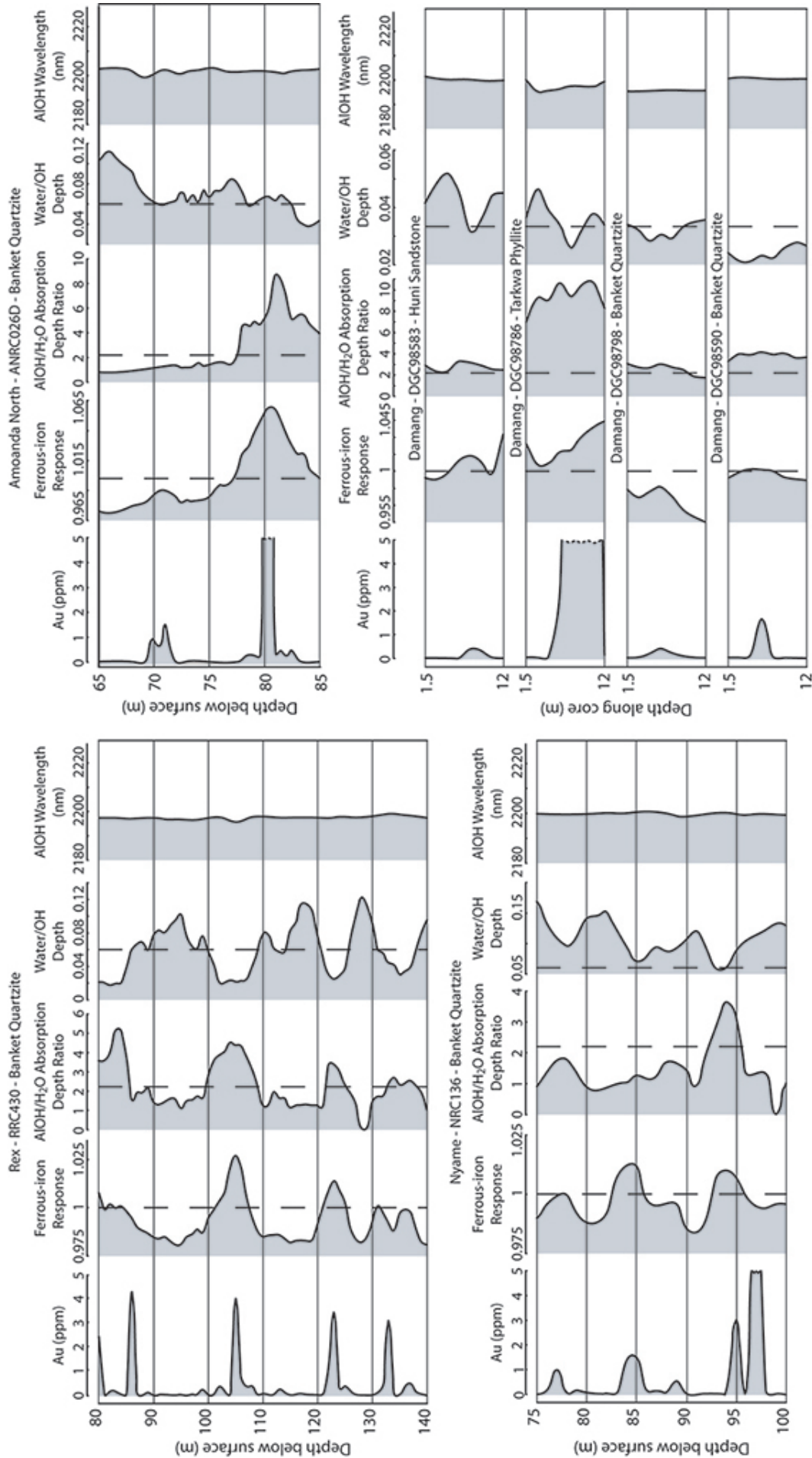
identification of muscovite/illite.

In contrast to the sediments, automated mineral identification is especially useful in characterizing zones of alteration in the dolerites. Unaltered dolerite is dominated by the identification of hornblende and lesser epidote, while the altered variety is primarily characterized by Mg-chlorite with intermediate (Fe-Mg) chlorite towards the margins of alteration zones. Lesser montmorillonite, tourmaline, biotite and ankerite are also occasionally identified. Montmorillonite may, however, be an artefact of the pulp making process. Chlorite is in fact present in the unaltered dolerite but is masked by the spectrally stronger hornblende. The lack of spectrally identifiable biotite in the altered dolerites, despite its widespread occurrence as an alteration phase, can be ascribed to shared features with chlorite spectra, its higher absorption coefficients and the fine grain size of the pulp. The dolerites exhibit a decrease in both the Ferrous-iron Response and $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio associated with zones of mineralization (Fig. 6.6). Water/OH Absorption Depth is highly variable within the Damang dolerites and shows no strong correlation with gold mineralization.

6.3.4 Spectral Characteristics and 3D Orebody Modelling

Grade shells derived from cut-offs utilized at the Damang Mine were defined using 3D interpolation LeapfrogTM software. Modelling was carried out on gold assays derived from 1.5 m composites obtained from 13 m deep, vertical grade control holes covering an area of 250 by 110m on level RL825 of the Damang open pit. The study area was located in the eastern cut back section of the pit where host lithologies are dominated by dolerite and

Figure 6.5: (Following page) Downhole plots of gold content and spectral parameters in sediments from selected exploration and grade control drill core from across the Damang region showing a strong positive correlation between gold mineralization, the Ferrous-iron Response and the $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio and a negative correlation with Water/OH Absorption Depth. By contrast, AlOH Wavelength values show negligible variation through zones of mineralization. Note that spectral parameters measure hydrothermal mineral assemblages and not gold content itself. Damang grade control cores are collared at pit level 825RL. Dashed lines for the Ferrous-iron Response, $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio and Water/OH Absorption Depth are at 1.0, 2.2 and 0.06 respectively and represent approximate boundaries between mineralized and unmineralized spectral signatures.



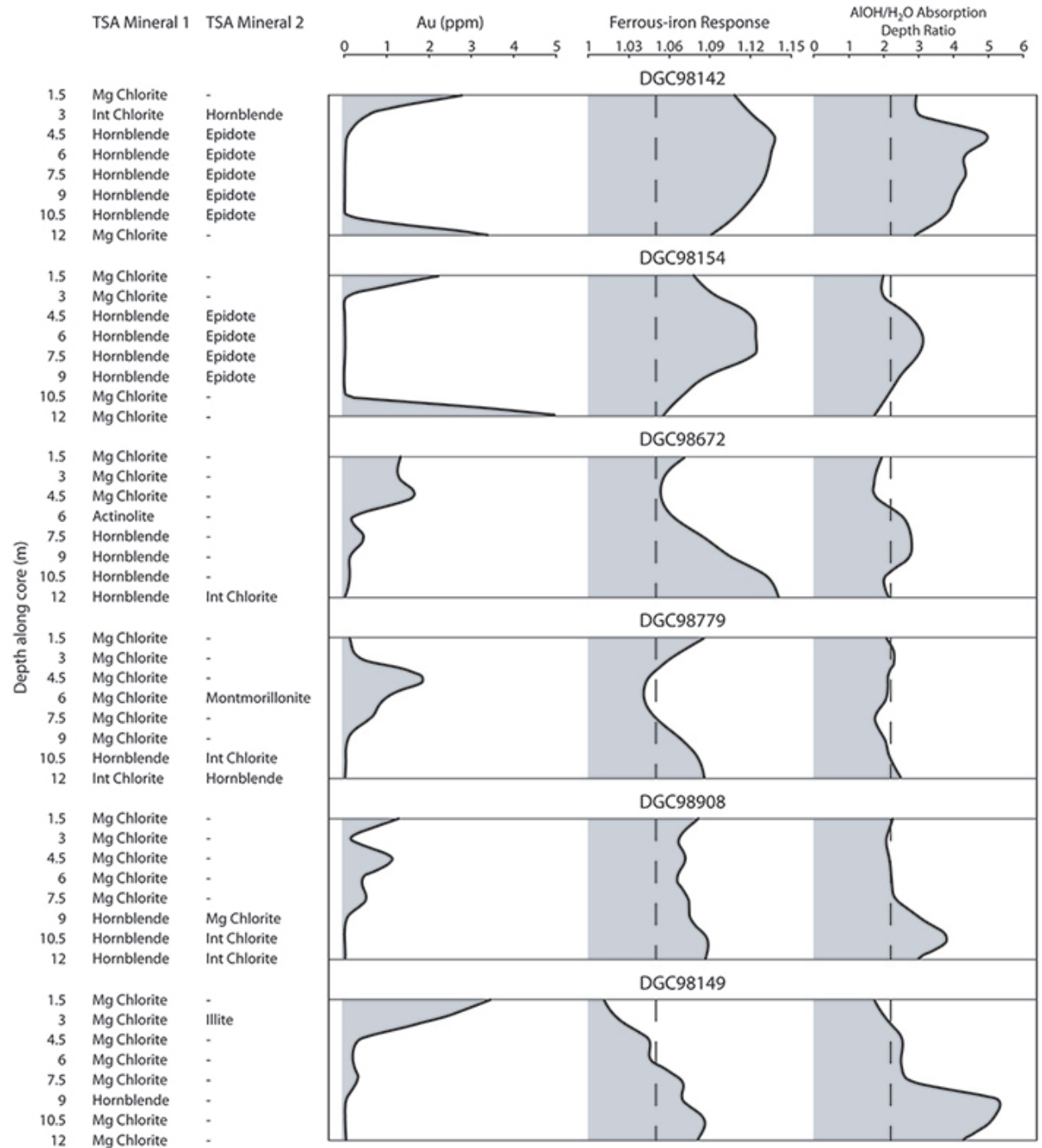


Figure 6.6: Downhole plots of gold content and spectral parameters in dolerites from selected grade control drill cores from the Damang pit showing a strong negative correlation between gold mineralization, the Ferrous-iron Response and the AlOH/H₂O Absorption Depth Ratio, in addition to a relationship with TSA Mineral Identification. All cores are collared at pit level 825RL. Dashed lines for the Ferrous-iron Response and AlOH/H₂O Absorption Depth Ratio are at 1.05 and 2.2 respectively and represent approximate boundaries between mineralized and unmineralized spectral signatures.

Tarkwa Phyllite. The ore shells encompass arrays of sub-horizontal quartz veins and associated alteration haloes which cut across steep, east-dipping lithologies on the eastern limb of the Damang anticline. Following gold analysis, spectra were obtained from each of the laboratory pulps. Spectrally determined minerals and parameters, as well as the interpolated grade shells, were imported into DataMineTM modelling software and examined in 3D together with logged lithologies (Fig. 6.7). Using this approach, it is possible to examine the relationships between spectral characteristics and lithology, alteration and gold distribution.

Figure 6.7 illustrates two 2D cross-sections through the 3D model construct in which a dolerite sill intrudes the Tarkwa Phyllite. As shown in Figures 6.5 and 6.6, unmineralized dolerite is characterized by high values of the Ferrous-iron Response and AlOH/H₂O Absorption Depth Ratio, while within the ore shells mineralized dolerite exhibits a marked decrease in both the Ferrous-iron Response values and the AlOH/H₂O Absorption Depth Ratio values. In the Tarkwa Phyllite, the Ferrous-iron Response is generally low with an accompanying low AlOH/H₂O Absorption Depth Ratio. Mineralized Tarkwa Phyllite within the ore shells however, is characterized by elevated values of both the Ferrous-iron Response and AlOH/H₂O Absorption Depth Ratio. It is important to emphasize that as interpolated ore shells are mathematical estimations of gold grade distribution, and because spectral parameters are derived from composited samples, there are instances where the spectral correlations are imperfect.

6.3.5 Spectral Correlation with the Weathered Profile

The results presented above apply directly to unweathered rocks at Damang. The weathered sediment and transported overburden intervals, however, exhibit distinctly different infrared spectral signatures due to the presence of numerous clay minerals, particularly kaolinite and minor halloysite, in the altered assemblage (Fig. 6.8). Throughout the weathered profile, AlOH Wavelength values are elevated relative to the fresh signature, from approximately 2198 nm (cf. Figure 6.5) to 2206 nm (Fig. 6.8). The AlOH/H₂O Absorption Depth Ratio is greatly increased in the upper portions of the weathered profile and steadily decrease

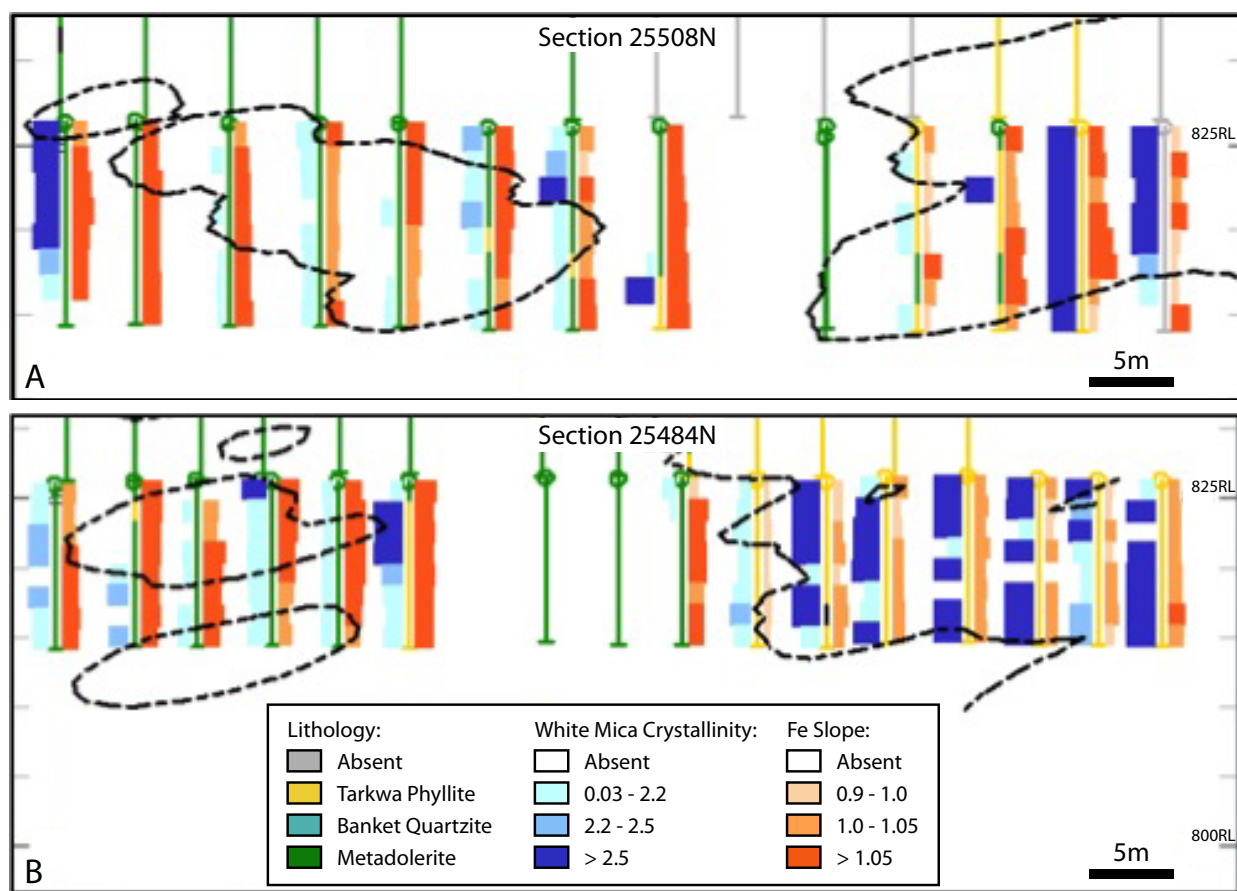


Figure 6.7: Vertical sections of grade control holes in the Damang pit illustrating the relationship between lithology, mineralization and two spectral parameters, the Ferrous-iron Response and the $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio, that best correlate with gold distribution. Lithologies are shown along drill hole traces, with the $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio and the Ferrous-iron Response plotted as histograms along either side of the drill hole (left and right respectively). Higher values of the two spectral parameters are reflected by increasing colour intensities. Dashed lines are projections of 3D ore shells on to the line of section.

with depth to a ‘fresh’ signature. In addition, a surficial transported overburden signature is characterized by a very high MgOH Wavelength and a lowered $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio. Of greater relevance to this study however, is a persistence in the correlation between gold mineralization and the Ferrous-iron Response throughout the weathered profile (Fig. 6.8). This association is particularly significant as the weathered profile is the most easily accessible portion of the Damang system and can be sampled quickly and cheaply through air-core drilling and manual augering. The rapid identification of a spectral signature associated with gold mineralization has important benefits for regional exploration at Damang.

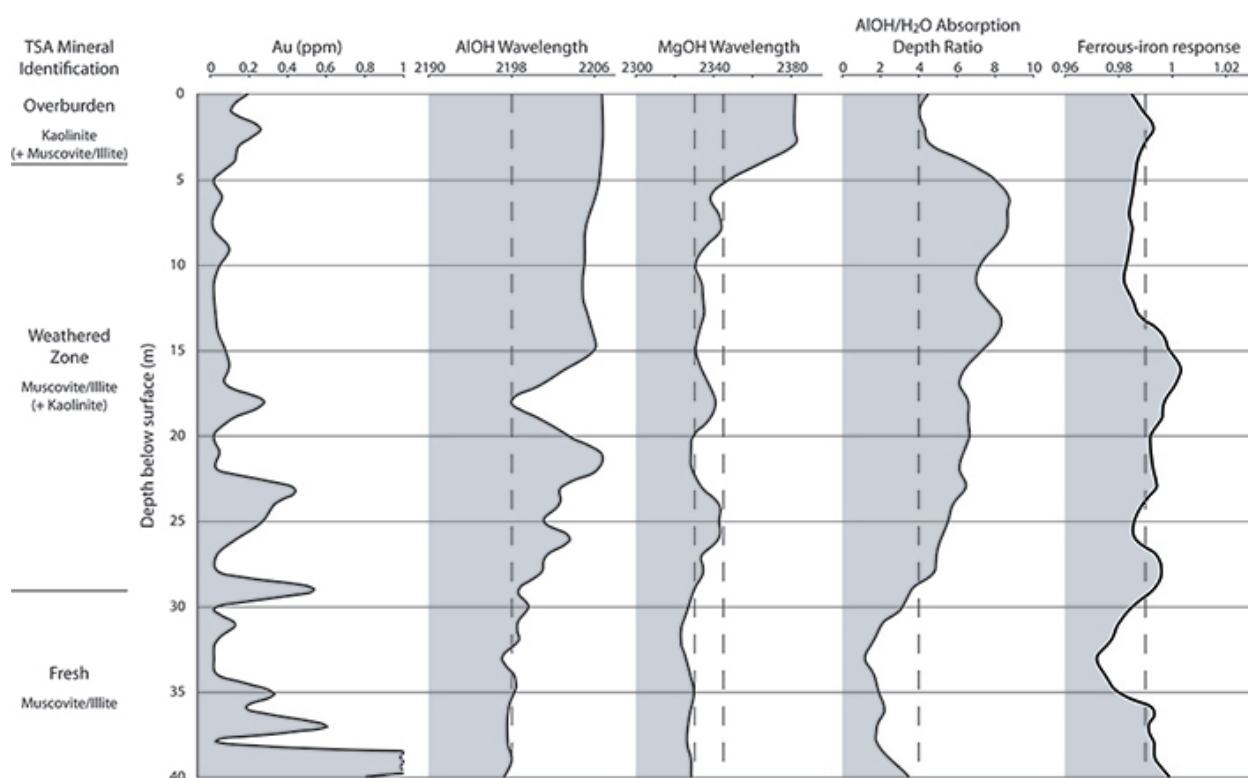


Figure 6.8: Downhole plot of gold content and spectral parameters in weathered Banket Quartzite from exploration hole RRC440 at Rex. The upper weathering profile is identifiable by elevated ALOH Wavelength, elevated and variable MgOH Wavelength and an elevated ALOH/H₂O Absorption Depth Ratio that decreases with depth. A transported overburden is characterized by very high MgOH Wavelength and a low ALOH/H₂O Absorption Depth Ratio. The Ferrous-iron Response still shows a good positive correlation with gold mineralization. Dashed lines for ALOH Wavelength, MgOH Wavelength, ALOH/H₂O Absorption Depth Ratio and the Ferrous-iron Response are at 2198, 2330 and 2345, 4.0 and 0.99 respectively and represent approximate boundaries between weathered and fresh spectral signatures.

6.3.6 Quantification of Spectral Parameters

The observed changes described here are relative to a local ‘background’ value in surrounding, unmineralized samples. The magnitude of changes and the absolute values associated with gold mineralization are therefore somewhat variable. Despite this, approximate boundaries between spectral signatures for mineralized and unmineralized zones can be suggested and are summarized in Table 6.2. A high ALOH/H₂O Absorption Depth Ratio (>2.2) is particularly effective as an identifier of mineralized Tarkwa Phyllite (Fig. 6.7B). A lower ALOH/H₂O Absorption Depth Ratio (<2.2), occurring at the margin of the ore shell, reflects Tarkwa Phyllite that has not been subjected to hydrothermal fluid ingress. With respect to dolerite, a Ferrous-iron Response value of >1.05 remains the most important spectral parameter

characterizing this lithology.

6.4 Discussion of Infrared Spectral Data

All of the observed changes in spectral parameters can be ascribed to specific petrographic features readily identified in the various lithologies examined. The raised Ferrous-iron Response recorded in altered sediments is interpreted to be due to the replacement of magnetite by pyrite as well as by the presence of ankerite, which is otherwise not identified, and Fe-bearing tourmaline. Although both magnetite and sulphide minerals typically suppress the Ferrous-iron Response, magnetite appears to have a stronger absorption effect than pyrite and so the replacement results in a net increase. The AlOH/H₂O Absorption Depth Ratio increases in the altered sediments due to the new growth of relatively high-temperature hydrothermal muscovite compared to metamorphic muscovite in the matrix of unaltered lithologies (Kruse & Hauff, 1991). The observed decrease in values of Water/OH Absorption Depth is due to the overall shape of the hull quotient spectrum. The downward curve of the reflectance spectra at wavelengths below 1700 nm, for mineralized samples (Fig. 6.4A), leads to the calculation of a hull quotient with a lower overall absorption across these wavelengths and consequently a significantly ‘flatter’ spectrum (Fig. 6.4B). This overall lower hull quotient across this range of wavelengths is represented by the decreased Water/OH Absorption Depth values.

Figure 6.5 also presents data for AlOH Wavelength, which is used to spectrally determine the composition of white micas. Elemental substitutions, such as Fe and Mg, into the muscovite crystal lattice cause a relative distortion of the Al-O-H bonds which results in absorption of infrared energy at a different wavelength. Specifically, higher values are indicative of increased Fe or Mg due to a phengitic substitution. All sediments at Damang show negligible variation in the AlOH Wavelength absorption feature indicating no compositional variation. This is supported by compositional data obtained by conventional energy dispersive spectrometry, which also shows no variation between altered and unaltered sediments (Fig. 6.9). Such invariant compositional data suggests homogenization postdating hydrothermal

Table 6.2: Summary table of spectral parameter values associated with unmineralized and gold-bearing samples of metasediment and dolerite at Damang. When considering dolerites, White Mica Crystallinity simply represents Depth AIOH/Depth H₂O; see text for discussion. ‘U’ and ‘M’ indicate unmineralized and mineralized lithologies respectively.

		Ferrous-iron Response	AIOH/H ₂ O Absorption Depth Ratio	Water/OH		AIOH Wavelength	MgOH Wavelength	TSA Mineral Identification
				Depth				
Metasediments	U	<1.0	<2.2	>0.06		~2198 nm	-	Muscovite/Illite
	M	>1.0	>2.2	<0.06		~2198 nm	-	Muscovite/Illite
Metadolerites	U	>1.05	>2.2	-		-	-	Hornblende/Epidote
	M	<1.05	<2.2	-		-	-	Mg-Chlorite
Weathered Profile	U	<0.99	>4.0	-		>2198 nm	>2330 nm	Kaolinite
	M	>0.99						Kaolinite
Overburden		<0.99	4.0	-		>2198 nm	>2345 nm	Kaolinite

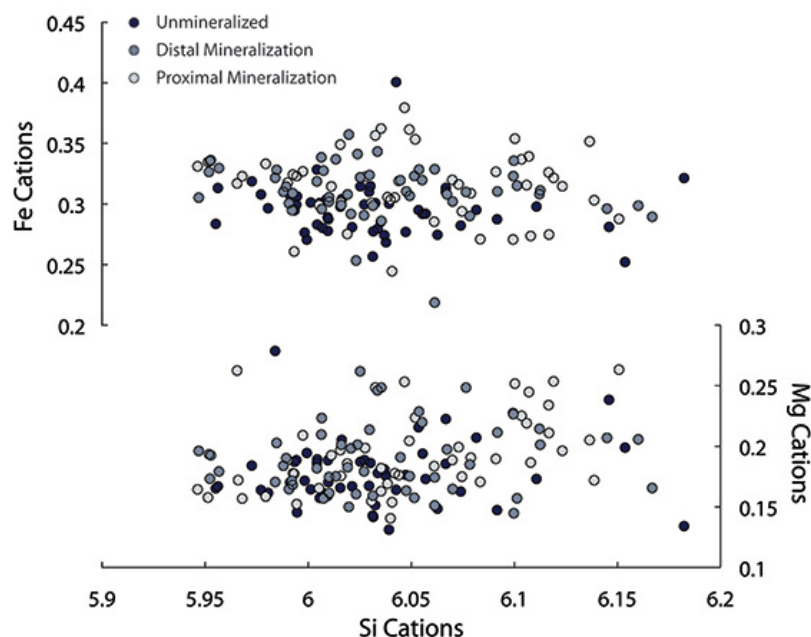


Figure 6.9: Plots of Fe and Mg cations contents of muscovites in Banket Quartzite at Damang as determined by energy dispersive spectroscopy. No compositional variation is evident between different alteration zones.

activity. The observed agreement between spectral and conventional geochemical techniques supports the use of infrared spectroscopy for this purpose at Damang.

The decrease in the Ferrous-iron Response values associated with mineralized dolerites can be attributed entirely to the formation of sulphide minerals, which contribute to suppress the Ferrous-iron Response. The observed decrease in the $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio in the altered dolerites is due to an overall shallowing of the AlOH absorption feature relative to the Water feature and can be ascribed to an increase in the relative proportion of chlorite, which has a weaker AlOH absorption than hornblende, which is destroyed. Hydrothermal processes that accompany gold mineralization in dolerite (i.e. replacement of amphibole by biotite and chlorite, formation of ankerite, and sulphidization) occasionally have a conflicted effect on the Ferrous-iron Response parameter and the result is subtler than with the sediments. For this reason, lower Ferrous-iron Response values alone are not necessarily the sole indicator of gold mineralization (Fig. 6.7B). However, as a reduced $\text{AlOH}/\text{H}_2\text{O}$ Absorption Depth Ratio also occurs in mineralized dolerite, it is the combination of these two features that best identifies gold mineralization in this lithology.

While automated mineral identification is a useful tool that facilitates the rapid and objective identification of mineral spectra, there are limitations that must be taken into account when using this software. Mineral identification is limited to a mixture of two minerals, even if,

as is typically the case, the sample comprises an assemblage containing several minerals. The two minerals identified are those with the strongest spectral response, but are not necessarily the two most abundant phases. It is for this reason that carbonate minerals, with their high absorption coefficients and poor response to the pulp making process, are rarely identified by automated mineral identification tools despite their ubiquity as alteration phases at Damang. The Al-O-H bonds, while exhibiting a dominant shortwave infrared absorption feature at 2204 ± 24 nm also produce minor absorption features that coincide directly with the carbonate absorption feature at 2322 ± 20 nm, which becomes masked and unidentifiable (Fig. 6.4B). This feature demonstrates the fact that absolute mineral identifications are not always reliable, although changes in the mineralogy of a sample do, nevertheless, faithfully reflect changes to the overall appearance of the infrared spectra. Moreover, the identification of a particular mineral may not represent an actual increase in the abundance of that mineral in one sample relative to another, but could in fact indicate a decrease in modal proportion of another mineral that had previously been masking it. While The Spectral AssistantTM is adequate, this library will ideally be superseded over time with deposit specific reference libraries, supported by detailed petrographic observations that mitigate the issues described above.

Within the weathered profile, elevated ALOH Wavelength values, up to approximately 2206 nm, and an increased ALOH/H₂O Absorption Depth Ratio are due to the presence of significant kaolinite, which has a very strong ALOH absorption feature. This feature occurs at a different wavelength in kaolinite compared to white micas. The observed decline in the ALOH/H₂O Absorption Depth Ratio with depth through the weathered profile is interpreted to be related to the overall decrease in the proportion of kaolinite. This reduces the relative absorption depth of the ALOH feature while leaving the wavelength relatively unaffected. The exact causes of the overburden signature are not fully understood but it may be related to unknown water-bearing clay minerals accompanying soil and organic matter, the latter causing unquantifiable interferences on mineral spectra. An increased Ferrous-iron Response is still associated with gold mineralization throughout the weathered profile, a feature that is presumably related to Fe-bearing minerals preserved from the unweathered protolith.

It is pertinent to recall that these parameters are not measuring gold directly but represent changes in the host rock mineral assemblages caused by fluid flow and hydrothermal alteration associated with the gold-bearing quartz veins. Since the degree of gold mineralization is not necessarily related to the intensity of hydrothermal alteration a quantitative association between spectral parameter intensity and gold mineralization should not be expected. Similarly, close to the margins of ore shells defined in Figure 6.7, spectral parameters tend towards unmineralized values as the effect of hydrothermal alteration is reduced.

Chapter 7

Discussion and Conclusions



The end product.

The preceding chapters have addressed specific aspects of the Damang system and certain portions of its history have been determined. In particular, thermodynamic modelling of mineral assemblages in chapter 4 provided a portion of the pressure-temperature path for the Damang region, while modelling of Ar diffusion in biotite in chapter 5 constrained a temperature-time path following peak metamorphism. This chapter aims to integrate these separate components into a single coherent tectonic (pressure-temperature-time) and mineralization history and synthesise a mineralization model for the Damang deposit. The implications of this model for exploration at both regional and local scales are discussed.

7.1 Thermal Modelling

An appropriate tool for constraining a full post-peak metamorphic pressure-temperature-time history is the computer program THERMODL, as used by Waters (1990). THERMODL is a numerical model that comprises a one-dimensional explicit finite-difference approximation to the heat equation, similar to that used by Oxburgh & Turcotte (1974); Bickle *et al.* (1975); England (1978) and more recently Smye *et al.* (2011).

The heat equation describes the movement of heat and in one dimension is given by:

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} - u \frac{\partial T}{\partial z} + \frac{A}{\rho \cdot C_p} \quad (7.1)$$

where T is absolute temperature, κ is thermal diffusivity ($31.56 \text{ km}^2/\text{Ma}$, which is approximately $1 \times 10^{-2} \text{ cm}^2/\text{s}$), z is distance, which here is depth in the crust, u is velocity, A is in situ heat production, ρ is density and C_p is the specific heat capacity. The first term on the right hand side of this equation describes the diffusion of heat and is identical in form to Fick's Second Law, discussed for the diffusion of Ar in minerals in chapter 5. The second term describes the advection of heat, while the third term describes, in this context, crustal radiogenic heat production.

Due to compositional variation in the crust, crustal heat production is itself a function of

depth, being greater in more evolved rocks nearer the surface. It is given by the relationship (Lachenbruch, 1968):

$$A_z = A_o \exp(-z/D) \quad (7.2)$$

where A_z is heat production at depth z , A_o is a constant representing a notional heat production at $z = 0$ (5×10^{-6} W/m³), and D is a constant with the dimensions of length (8 km).

κ is given by:

$$\kappa = \frac{k}{\rho.Cp} \quad (7.3)$$

where k is the thermal conductivity ($2.25 \text{ Wm}^{-1}\text{K}^{-1}$). The product $\rho.Cp$ is therefore kept constant at $2.25 \times 10^6 \text{ J/km.K}$.

The initial geothermal profile, prior to cooling and exhumation, is constrained from conditions for peak metamorphism calculated in chapter 4 and is calculated using the steady-state heat equation:

$$T_z = T_{zo} + \frac{D^2.A_o}{k} - \frac{D^2.A_o \exp(-z/D)}{k} + P.z \quad (7.4)$$

where T_{zo} is the temperature at the surface and $P = \partial T / \partial z$ at ∞ . P is parameterized by setting the heat flux across the base of the lithosphere (q), here kept at a constant depth of 100 km, according to the relationship:

$$q = -k \frac{\partial T}{\partial z} \quad (7.5)$$

In order to set a temperature of 597°C at a depth of 20 km (5.28 kbar), as constrained from modelled conditions of peak metamorphism (section 4.3.2), an initial basal flux of 52.5 mW/m^2 is required.

Finally, an exhumation rate and model run time are defined and applied to one or more ‘marker’ points whose depth-temperature coordinates are calculated and tracked through time as the model progresses. A simple two-stage exhumation history was considered and

Table 7.1: Summary of THERMODL variables

	Exhumation Rate (mm/yr)		Start
	Stage 1	Stage 2	Age (Ma)
Model 1	1.0	0.002	2005
Model 2	2.6	0.002	2005
Model 3	2.6	0.002	2025
Model 4	2.6	0.01	2005
Model 5	2.6	0	2005

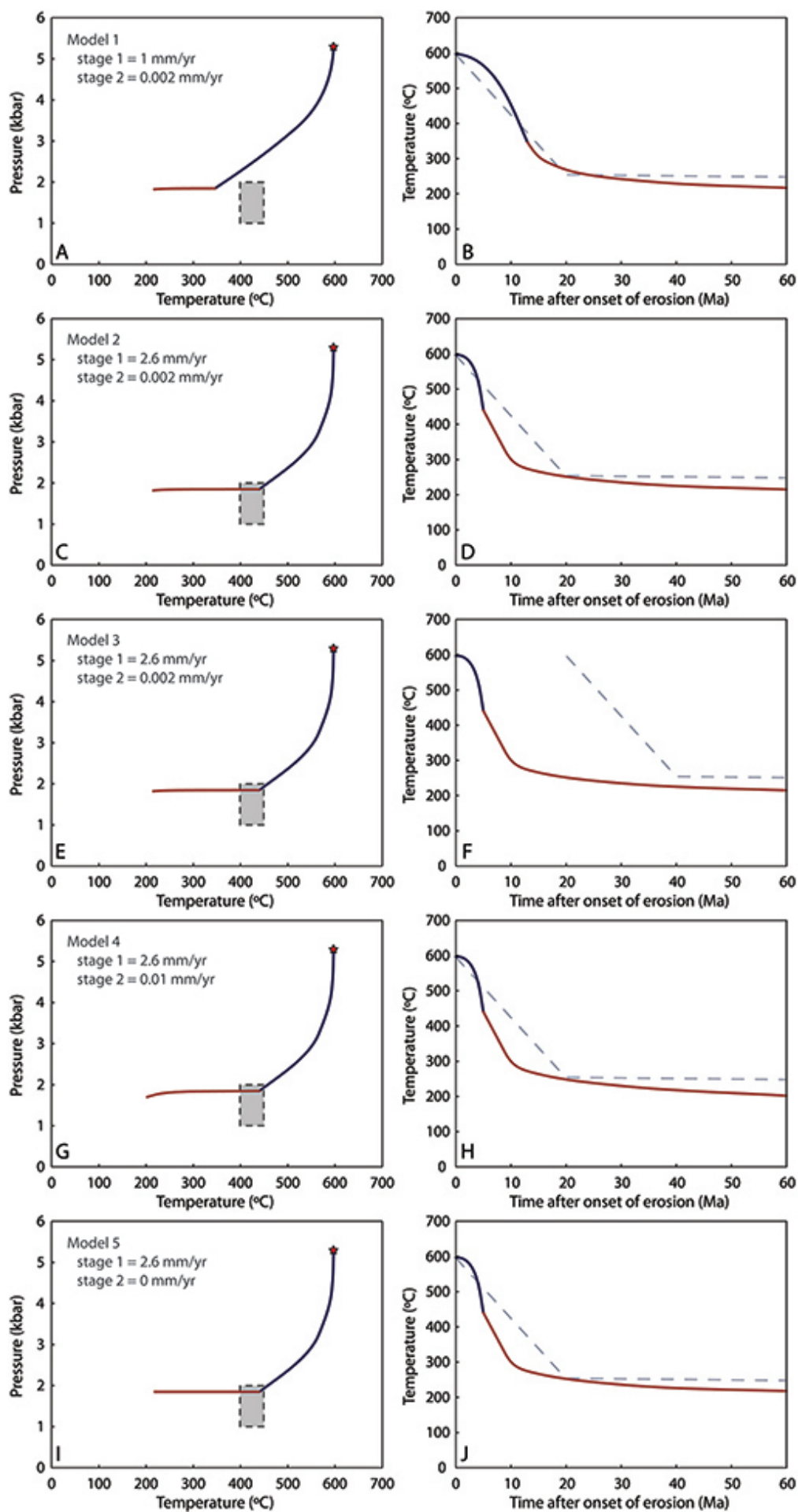
the exhumation rates of each stage were varied in order to obtain a best fit to the known pressure-temperature-time points. Exhumation is assumed, in the first instance, to commence immediately following peak metamorphism at 2005 Ma.

7.1.1 Thermal Modelling Results

The results of five separate model runs are given in figure 7.1, with model variables summarised in table 7.1.

Models 1 and 2 differ in the exhumation rate chosen during stage 1, at 1 mm/yr and 2.6 mm/yr respectively. Using a higher exhumation rate produces a better correlation with pressure-temperature conditions of gold mineralization, while both produce a good fit with the temperature-time constraints. Model 3 utilises the same exhumation rates as for Model 2, but commences 20 Ma earlier, i.e. peak metamorphism occurs at 2025 Ma. A good fit is still achieved with temperatures during stage 2 of the exhumation, as determined from Ar diffusion modelling. If exhumation were to commence later it is not possible to match the known constraints and would essentially require instantaneous cooling. Models 4 and 5 investigate the effect of changing exhumation rate during stage 2 from 0.01 mm/yr (model 4) to 0 mm/yr (model 5), although both rates produce a good fit with the temperature-time

Figure 7.1: (Following page) Results of thermal modelling with the program THERMODL. A single pressure-temperature-time path is presented, which is constrained from the results of chapters 4 and 5. Grey dashed line is the $^{40}\text{Ar}/^{39}\text{Ar}$ cooling history from chapter 5. In a two-stage exhumation history, initial exhumation must be at least 2.6 mm/yr, for 5 Ma, to intersect the likely conditions of hydrothermal alteration. The resulting thermal overstep then controls subsequent cooling and the exhumation rate during stage 2 is of lesser importance.



constraints. After a model run time of 60 Ma, the difference in temperature between these two models is just 16°C, which equates to approximately 4% of the entire cooling interval to that point.

7.1.2 Discussion of Thermal Modelling

The thermal modelling results show that, to a first order, following peak metamorphism the Damang region underwent a short period of rapid exhumation followed by a prolonged period of negligible exhumation. An exhumation rate during stage 1 of 2.6 mm/yr for 5 Ma is the minimum required to intersect the likely conditions of gold mineralization. A faster rate would require a shorter period of time, while a slower rate does not pass through the appropriate pressure-temperature space. Estimates of present day erosion rates in active orogens, such as in Taiwan, often exceed 3–5 mm/yr ([Burbank, 2002](#); [Dadson *et al.*, 2003](#); [Siame *et al.*, 2011](#)), which would imply a total exhumation rate of 2.6 mm/yr could be accounted for by simple surface erosion alone. However, erosion in present day tropical terrains may not be analogous to the Palaeoproterozoic and so such a high rate may indicate a component of tectonic exhumation.

The temperature overstep generated during rapid exhumation in stage 1 is so great that the exhumation rate chosen for stage 2, although necessarily low (less than 0.01 mm/yr), has only a second order effect, and any cooling imposed by exhumation here is negligible compared to that required to return to steady state following stage 1.

The decrease from high to very low rates of exhumation over a relatively short time period (<10 Ma) further supports the hypothesis of a tectonic component. Such tectonic exhumation would also explain the difference between peak metamorphic conditions determined at Damang and the lower grade conditions observed elsewhere, including at Tarkwa. The mechanism for this tectonically enhanced exhumation, however, is unclear and could range from enhanced erosion due to topography generation following overthrusting to extensional unroofing during orogenesis. Furthermore, there are no structures encountered within the Damang camp that could account for the required movement. However, returning to the

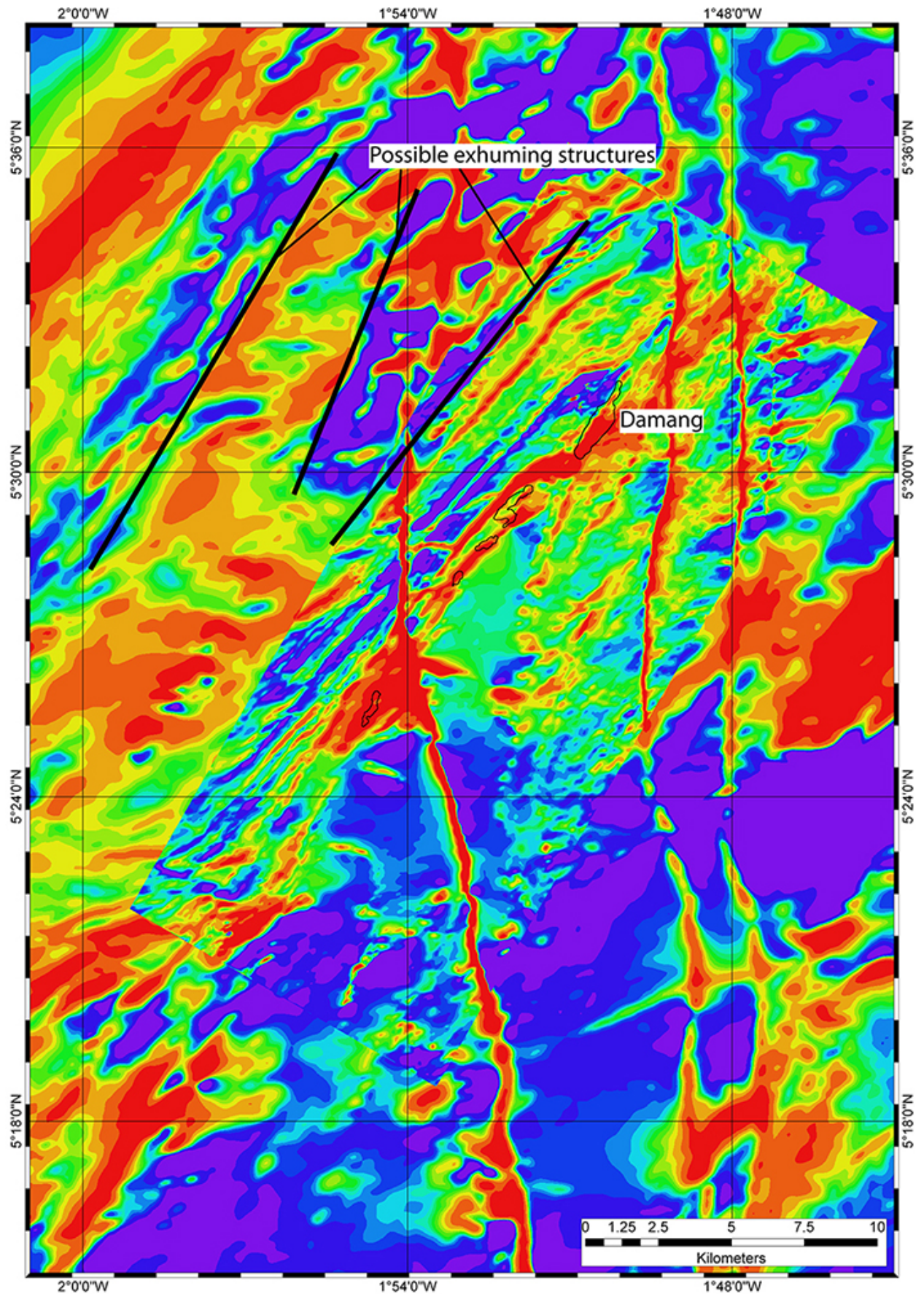


Figure 7.2: Reduced to pole (RTP) aeromagnetic data for the Tarkwa–Damang region, with a high resolution overlay of the Damang camp. The main operational pits at Damang are outlined and potential exhuming structures are delineated. Data provided by Gold Fields Ltd. Colour scale is qualitative from blue (low) to red (high).

the Damang regional aeromagnetic data shown in figure 2.19, a number of possible strike-discordant structures are visible immediately west of Damang (Fig. 7.2). Although the exact nature of these structures is not known, they may represent faults along which exhumation of the Damang region could have occurred. Interestingly, these potential structures are well defined to the west of Damang but become less apparent southwards towards Tarkwa. The implication is that Damang may have experienced a greater degree of exhumation than at Tarkwa, with the entire Tarkwa–Damang region tilting about an axis located close to Tarkwa. This accounts for the apparent lack of amphibolite facies metamorphism at Tarkwa as that area was not buried and heated to the same level as Damang.

7.2 Synthesis and a Mineralization Model

7.2.1 Generation of Damang's Quartz Vein Array

Tunks *et al.* (2004) have already shown that a Damang-style fault-fracture mesh can be generated in a convergent setting, with horizontal compression (σ_1) and vertical extension (σ_3) (Fig. 7.3). The exhumation history interpreted for the Damang region implies the involvement of a tectonic component, such that exhumation along compressional faults is appropriate. The fault-fracture mesh present at Damang is therefore the local manifestation of a regional-scale compressive event occurring during TD₃.

7.2.2 Generation of a Mineralizing Fluid

The specific origin of mineralizing fluids responsible for orogenic gold deposits is often difficult to constrain and is thus a contentious issue (see section 1.2.1). Prograde metamorphic devolatilization of hydrous minerals in a range of rock types is commonly cited as a generic source for fluid (Kerrick & Fyfe, 1981; Kerrich & Cassidy, 1994; Phillips & Powell, 2010). The timing of this fluid release in relation to the metamorphic cycle and timing of gold mineralization is also debated. Phillips & Powell (2010) argue for devolatilization and concomitant gold mineralization during prograde burial, while Pettke *et al.* (1999); Stüwe *et al.*

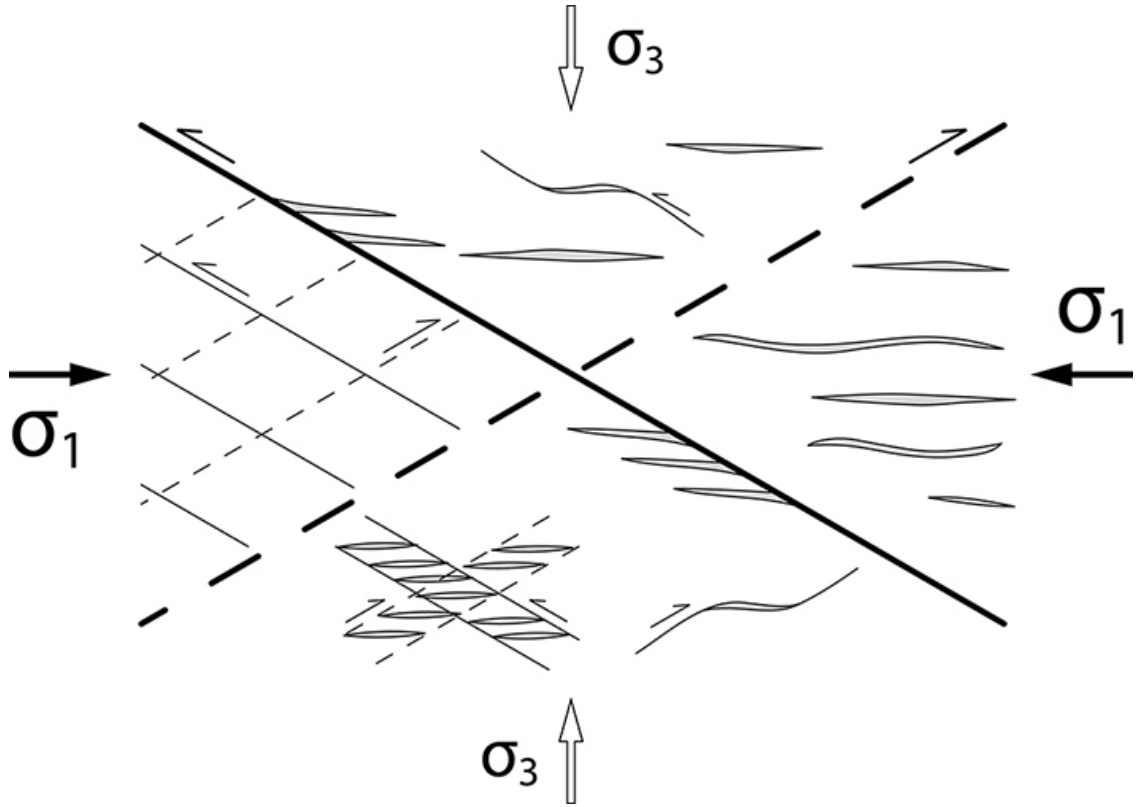


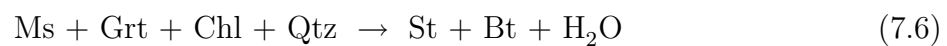
Figure 7.3: Possible stress controlled development of a fault-fracture mesh in a horizontal compressive field, as interpreted for Damang during mineralization. σ_1 represents the maximum compressive stress direction, which is horizontal, σ_3 is the minimum compressive stress direction, which is vertical and equates to extension in the opposite direction. After [Tunks *et al.* \(2004\)](#).

(1993) and [Stüwe \(1998\)](#) suggest that fluid may be generated in deep rocks that are still prograding, after shallower rocks have reached their metamorphic peak. The latter case accords with the models of [England & Thompson \(1984\)](#) and [Thompson & England \(1984\)](#) (see section 5.7). However, [Yardley & Cleverley \(2011\)](#) describe how many metamorphic dehydration reactions only produce small amounts of fluid very slowly. This contradicts the apparently rapid rate at which many orogenic gold deposits appear to form in association with high fluid flux events. By contrast, they propose that vast amounts of fluid may be generated through dehydration reactions during rapid exhumation and cite recent orogenic gold mineralization in Taiwan as an example of this (e.g. [Craw *et al.*, 2010](#)). This argument has clear implications for fluid generation at Damang, given a similarly rapid exhumation history. In addition, any fluid generation must be intimately coeval with the formation of the fault-fracture mesh, and therefore exhumation, since the presence of extensional fractures like those at Damang require fluid pressure to be equal to or greater than lithostatic pressure in order for the fractures to be held open. Indeed, this high fluid pressure, in combination

with the regional stress field shown in figure 7.3, is key to the generation of the fault-fracture mesh.

The assertion of slowly released water through prograde dehydration reactions given by [Yardley & Cleverley \(2011\)](#) assumes complete equilibrium throughout the prograde cycle. However, kinetic processes have a significant impact on the initiation of metamorphic reactions. In particular, the energy required to nucleate new metamorphic phases and the slow dissolution of reactant phases results in a temperature overstep being required, i.e. a reaction may commence at a higher temperature (and therefore later in time) than would be predicted from equilibrium calculations ([Waters & Lovegrove, 2002](#); [Pattison & Tinkham, 2009](#)). This delayed reaction results in the reactant assemblage becoming further and further out of equilibrium, such that when the reaction does commence it proceeds very rapidly. If this principle is applied to a dehydration reaction, then it is indeed possible for large volumes of water to be released quickly.

The peak mineral assemblage observed and calculated for the Tarkwa Phyllite at Damang contains abundant staurolite, produced through the reaction (c.f. Figs. 4.2 and 4.3):



[Pattison & Tinkham \(2009\)](#) calculated that the temperature overstep required for this reaction is as much as 30°C, while [Waters & Lovegrove \(2002\)](#) suggest that other chlorite-consuming reactions may require even greater temperature oversteps of 40°C. The effect of a 30°C overstep on reaction 7.6, with regard to the the Tarkwa Phyllite samples AWDP1 and AWDBm1 (section 4.3.2; Figs. 4.2 and 4.3) is shown in figure 7.4. It is seen that reaction 7.6 is now consistent with estimated conditions of peak metamorphism. This implies that the dehydration of chlorite could have occurred at or immediately before peak conditions and is therefore coeval with the onset of exhumation. This same principle can be applied to chlorite-dehydration in the other sedimentary lithologies at Damang, particularly the Banket Series, as well as the dolerite intrusions (c.f. Fig. 4.6).

Furthermore, the calculated peak assemblage field for samples AWDP1 and AWDBm1 are

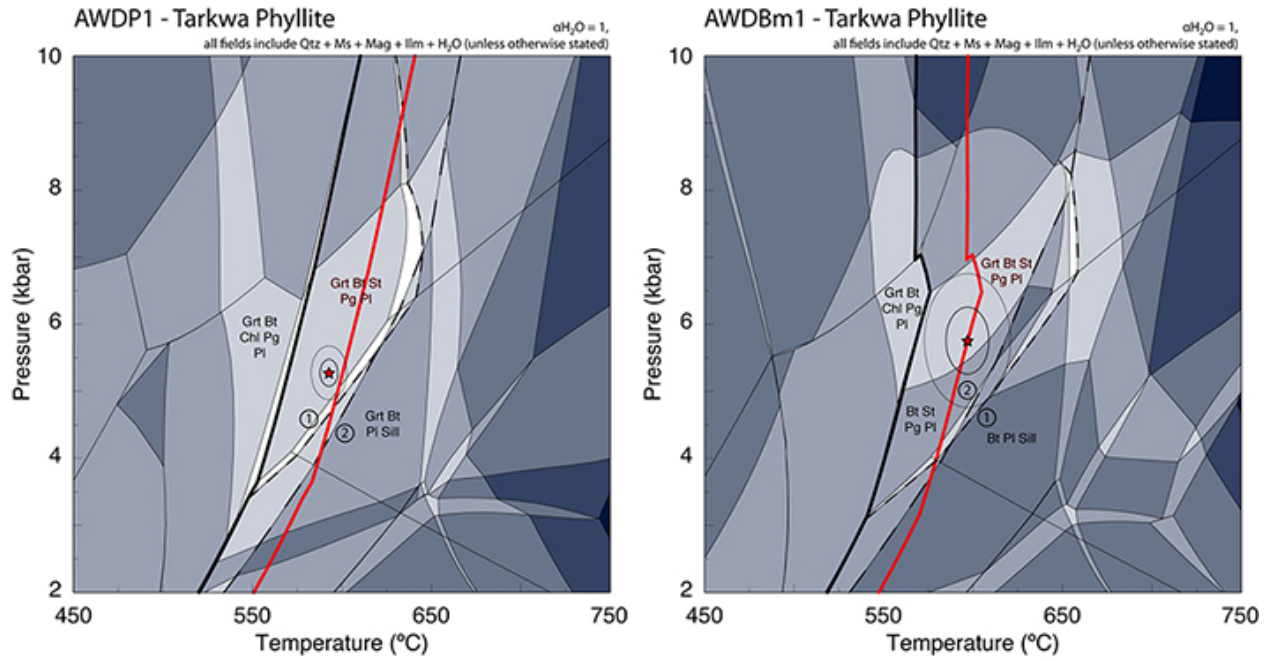
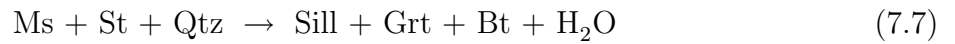


Figure 7.4: P-T pseudosections for samples AWDP1 and AWDBm1, as per figures 4.2 and 4.3. The heavy black line marks the extent of chlorite stability, while the heavy red line is this same reaction band with a 30°C overstep, which falls through likely peak conditions for both samples. Other key dehydration reactions are highlighted with heavy dashed lines: short dash = staurolite dehydration, occurring at point 1 through the reaction $\text{Ms} + \text{St} + \text{Qtz} \rightarrow \text{Sill} + \text{Bt} + \text{H}_2\text{O} \pm \text{Grt}$; long dash = paragonite dehydration, occurring at point 2 through the reaction $\text{Pg} + \text{Qtz} \rightarrow \text{Ab} + \text{Sill} + \text{H}_2\text{O}$.

bounded to their low pressure side by a number of other dehydration reactions, including:



with positive P/T slopes, such that decompression across them would result in the further production of fluid. Reactions such as 7.7 have clearly not occurred in the rocks at Damang however. This is likely due to the necessary temperature overstep not being achieved since Pattison & Tinkham (2009) suggest that reaction 7.7 may require an overstep of as much as 70°C. This is larger than that for reaction 7.6 since less water is released and therefore the entropy change (ΔS) across the reaction is lower. As a result, a larger overstep (ΔT) is required in order to produce a similar reaction affinity (i.e. the Gibbs free energy potential) (A), according to the relationship $A = \Delta T * \Delta S$ (Waters & Lovegrove, 2002; Pattison & Tinkham, 2009). The presence of chlorite in the Tarkwa Phyllite however, is not thought to be due to reaction 7.6 not going to completion since it is demonstrably associated with later retrogression (section 4.3.2).

The reactions discussed here and highlighted on figure 7.4 are dehydration reactions. However, since the mineralizing fluid at Damang had an $X(\text{CO}_2)$ of around 0.7, there must be a considerable contribution of CO_2 . The exact source of all of this CO_2 is uncertain, although there are a number of possibilities. Firstly, the Huni Sandstone does contain a reasonably large proportion of carbonate, which if decomposed at depth could introduce a small amount of CO_2 . The Birimian at Damang primarily comprises volcanoclastic sediments (see section 2.8). Elsewhere in Ghana, volcanoclastic sediments are often associated with a chemical sediment facies, particularly along the margins of the Birimian belts (see section 2.2.1). This chemical sediment facies includes considerable carbonate horizons, which if similar rocks are present under Damang could produce large quantities of CO_2 . Furthermore, although the fluid interacting at Damang has an $X(\text{CO}_2)$ of around 0.7, the original fluid sourced at depth may have been more hydrous. If hydration reactions were more dominant than carbonation reactions during fluid ascent, then the fluid would become progressively more CO_2 rich by the time it reaches Damang's structural level.

It is of course pertinent to consider that the development of a hydrothermal alteration assemblage necessarily implies that the fluid was out of equilibrium with the host rock. Therefore it is unlikely that fluids generated from the rocks now hosting gold mineralization at Damang, e.g. through reaction 7.6, are themselves responsible for the mineralization. Instead, the fluids are likely to have a contribution from a deeper, and therefore hotter, source area beneath Damang, generated through similar reactions to those described here. Low density fluids, once generated through devolatilization reactions, will tend to rise through the crust out of their region of formation through the grain boundary network and along pressure gradients both into fault structures and parallel to stratigraphy. Fluid flow along fold limbs towards hinge zones is therefore an important mechanism in focussing mineralizing fluids.

As discussed above and in section 5.7, within any internally-heated metamorphic terrane, rocks that are buried deeper reach their metamorphic peak at a later time (England & Thompson, 1984; Thompson & England, 1984). This principle explains why the age of peak metamorphism at Damang (2005 ± 26 Ma, section 5.4) is later than that typically considered

for lower grade rocks from Ghana. The important implication however, is that at the time the Damang region reached its metamorphic peak, deeper rocks were still devolatilizing and releasing fluid that fed the fault-fracture network forming at Damang during exhumation. This is similar to the scenarios described by (Pettke *et al.*, 1999; Stüwe *et al.*, 1993; Stüwe, 1998). The difference with these examples is that, at Damang, devolatilization reactions were crossed during decompression, rather than through continued burial and heating of deeper rocks. The exhumation however, was relatively localized since extensive regional-scale exhumation, incorporating much hotter rocks may induce melting, for which there is currently no evidence.

7.3 Implications for Exploration

The mineralization model presented here has a number of implications for exploration both at Damang and regionally by providing a set of potential targeting criteria.

7.3.1 Exploration at Damang

All lithologies, with the exception of the Kawere Group, are potentially prospective for gold mineralization. The most important factor that determines the extent to which a veined rock is mineralized is the presence of magnetite, due to its role in wall rock sulphidation and consequent gold mineralization. This is particularly relevant when considering why one dolerite intrusion may be mineralized while another is essentially barren. This relationship also explains the high gold contents exhibited by the Banket Series quartzites, and to a lesser extent the Tarkwa Phyllite, compared to the Huni Sandstone. Continued exploration of all lithologies is therefore justified.

In zones of focussed fluid flow, such as the Damang pit, mineralization should be expected to continue to depth through the fault-bounded zone of fractured and reactive rocks. The lower limit to mineralization may correlate with either a structure controlling the exhumed domain is reached, below which rocks experienced less relative exhumation and are consequently

unveined, or a depth at which during the mineralization event the rocks were either too ductile to fracture and vein effectively or fluid pressure was insufficient to maintain open fractures. This is supported by recent (mid-2011) exploration drilling beneath the base of the Damang pit that has encountered gold mineralization for a further 250 m below the current pit floor (Fig. 7.5). These drill holes regularly ended in mineralized zones, implying that the mineralization continues to greater depths than this and represents part of a potentially much larger gold mineralized system.

While Damang has undergone significant exhumation relative to surrounding rocks, this is not true of Tarkwa. As such, Tarkwa mine has not developed as extensive a vein fracture network nor comparable amounts of fluid as at Damang. The fault-fracture array is therefore predicted to be more extensive, particularly vertically, at Damang, where the maximum differential exhumation exists. Since the vertical extent of the fault-fracture mesh controls the maximum depth of mineralization, this depth is predicted to decrease southwards towards Tarkwa. In contrast, the lack of significant economic gold mineralization north of the Damang pit is interpreted to be due to a second order effect whereby the less favourable Huni Sandstone occurs at the surface. It is suggested that an extension of the mineralization seen in the Damang pit plunges northwards, with the host rocks, and is present at depth beneath the Huni Sandstone.

7.3.2 Regional Exploration

Ultimately, this thesis proves that the Tarkwaian System is prospective for gold mineralization and exploration should not be restricted to the Birimian Supergroup. Furthermore, the starkly different structural setting at Damang compared to other Ghanaian orogenic gold deposits implies that exploration should also not be restricted to large-scale shear zones. Although regional TD₁ folding and faulting and the intersection with later TD₂ faults provides the first order structural controls on gold mineralization at Damang, the generation of the mineralized quartz vein array is spatially and genetically linked to Damang's exhumation history, relative to that of the surrounding rocks. As such, other regions of anomalously

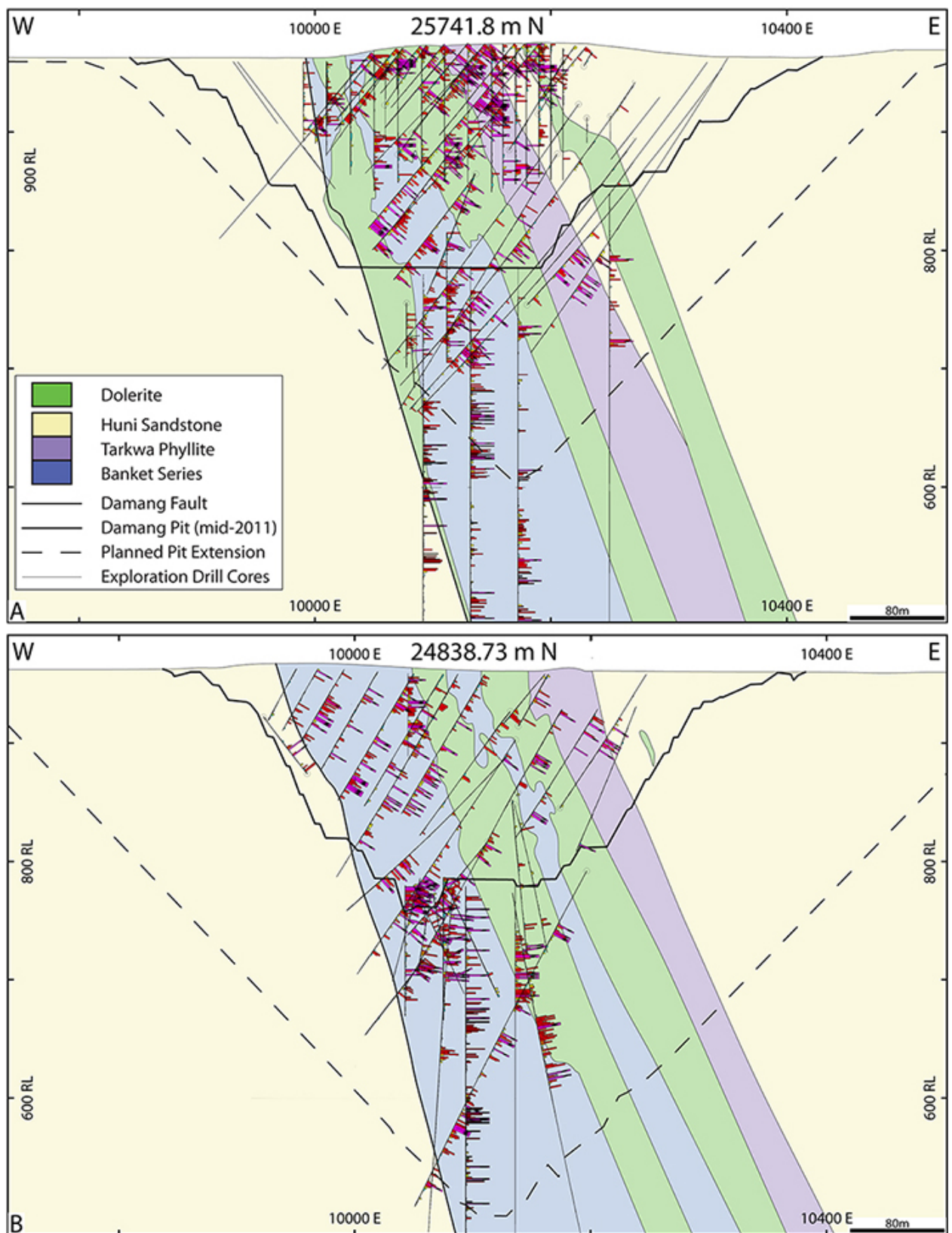


Figure 7.5: Geological cross-sections through the Damang Pit, as per figure 2.10 with the results of recent (mid-2011) exploration drilling showing extensive gold mineralization at depth beneath the floor of the current pit. Drill cores are projections from all drilling within 25 m of the line of section. A, 25741.8 m N; B, 24838.73 m N, according to the AGL grid (see appendix A).

higher metamorphic grade across the Man Shield that may have experienced similar exhumation histories are also prospective for gold mineralization. This may therefore discount thermal anomalies associated with contact aureoles around granitoid intrusions and instead focus should be directed at regions with a higher regional metamorphic peak

7.4 Thesis Conclusions

As with many ore deposits, Damang has a history comprising numerous distinct structural, igneous, metamorphic and hydrothermal events. Economic gold mineralization occurs at either end of this history with palaeoplacer style gold deposited during sedimentation of the Tarkwaian Group Banket Series conglomerates. U-Pb zircon dating of the Birimian Volcaniclastic and Diorite Porphyry give ages of 2178 ± 9.3 Ma and 2164.6 ± 8.0 Ma respectively. The Diorite Porphyry is therefore akin to the belt-type granitoids and was emplaced during construction of the Ashanti belt prior to deposition and deformation of the Tarkwaian System. This is overprinted by hydrothermal gold mineralization towards the end of Damang's history. Although orogenic gold mineralization is the last major identifiable geologic event to occur at Damang it was promoted by many of the preceding structural and hydrothermal episodes. The Damang deposit is ultimately the result of a confluence of favourable interactions between several regional- and local-scale processes. A diagrammatic summary of the history of the Damang deposit is presented in figure 7.6.

Early structural deformation at Damang, during the TD₁ event of [Tunks *et al.* \(2004\)](#), resulted in regional folding and faulting, including the Damang Fault, during burial. This deformation acted as a first order control on hydrothermal gold mineralization. Further faulting during TD₂ produced large-scale fluid pathways, particularly at intersections with TD₁ faults. TD₂ was accompanied by the intrusion of a suite of calc-alkaline – tholeiitic gabbros – diorites and a pervasive Cal+Chl+Hem+Ab+Qtz 'Pink Haematite' alteration interpreted to be associated with an oxidizing fluid of magmatic origin. Extensive silicification as part of the Pink Haematite alteration changed the rheology of the altered rocks and promoted brittle fracturing during the later gold mineralization. Where Pink Haematite alteration

was particularly intense, however, the resulting conversion of magnetite to haematite may, in fact, have suppressed later gold precipitation.

Basic petrography and thermodynamic modelling of mineral assemblages with the computer program THERMOCALC indicates that Damang experienced a higher grade metamorphic peak than is typically documented for the Tarkwaian System across Ghana. Modelled conditions from the Tarkwa Phyllite and dolerite intrusives are consistent at around 595°C and 5.5 kbar. The absence of amphibolite facies minerals in the Banket Series conglomerates from Tarkwa mine implies that this high grade peak is restricted to the immediate Damang region. Peak metamorphism occurred at 2005 ± 26 Ma, as determined from U-Th-total Pb dating of metamorphic monazites in the Tarkwa Phyllite. This is significantly later than is typically quoted for greenschist facies metamorphism elsewhere in Ghana and is a direct consequence of Damang's higher grade peak. These monazites are destroyed during the later gold-associated hydrothermal event so this age provides a maximum constraint on the timing of orogenic gold mineralization.

Hydrothermal gold mineralization is hosted in a vertically extensive, sub-horizontal fault-fracture quartz vein array formed in a compressive structural setting during TD₃. Gold mineralization is associated with localized potassic–carbonation–sulphidation alteration around quartz vein caused by a reducing, gold- and CO₂-bearing aqueous fluid. Gold precipitation is directly related to the sulphidation of magnetite. Overall, this fluid shares compositional characteristics with those responsible for orogenic gold mineralization elsewhere in Ghana, and globally. Therefore, a comparable fluid source can be invoked for mineralization at Damang.

The mineralogical changes produced during gold-associated hydrothermal alteration, particularly with regards to hydrous and Fe-oxide/Fe-sulphide phases impart a characteristic signature that is discernible with portable infrared spectroscopy. This spectral signature is consistent across the Damang region and can be also detected in heavily weathered rock. This technique is therefore a viable and useful exploration tool for use at Damang and in similar terrains.

$^{40}\text{Ar}/^{39}\text{Ar}$ dating of both hydrothermal and metamorphic biotite from Damang constrains the timing of cooling through approximately $300\pm 50^\circ\text{C}$ and produces a range of ages between 1978.8 ± 6.2 Ma and 1898 ± 11 Ma. This broad range of ages is interpreted to be a function of grain size and modelling of the diffusion of Ar through biotite using the computer program DIFFARG is consistent with this, implying extremely slow cooling through the interval $250\text{--}300^\circ\text{C}$. This places a minimum age constraint on the timing of orogenic gold mineralization, which is again significantly later than that quoted from elsewhere in Ghana and due to the later timing of peak metamorphism, which preceded gold mineralization, at Damang.

Thermodynamic modelling of gold-associated alteration assemblages with THERMOCALC constrains the likely conditions of gold mineralization to $400\text{--}450^\circ\text{C}$, 1–2 kbar, which agrees broadly with conditions typically considered for other Ghanaian deposits as well as fluid inclusion studies. These conditions, in combination with the new age constraints, imply that immediately following peak metamorphism the Damang region underwent rapid exhumation in the hanging wall of one or more controlling faults. Numerical modelling with the computer program THERMODL suggests an initial rate of around 2.6 mm/yr. This decompression is localized to the Damang region but related to regional compression. This TD_3 structural episode resulted in the formation of the fault-fracture quartz vein array. Subsequent exhumation and consequent cooling was extremely slow, at rates of less than 0.01 mm/yr, and likely nearer 0 mm/yr.

The source of the mineralizing fluid is still equivocal although dehydration reactions, such as the consumption of chlorite, occurring in all lithologies at Damang close to conditions of peak metamorphism are a possible source. These reactions are likely supplemented by major devolatilization occurring during rapid exhumation. This increased fluid pressure and fed the fault-fracture mesh, leading to quartz vein precipitation and associated gold-bearing alteration.

Orogenic gold mineralization at Damang clearly occurred significantly post-peak metamorphism and it is interpreted that much of the mineralizing fluid was generated during subsequent regional exhumation. Neither of these observations support the metamorphic model

of orogenic gold deposit formation, which would require fluid production during burial prior to peak metamorphism.

Ultimately, this thesis shows that the Tarkwaian is prospective for gold mineralization and although Damang is unique amongst currently known deposits in Ghana, its structural, metamorphic and hydrothermal history is not necessarily unique. Therefore the idea of further Damang-like deposits being discovered is not unreasonable.

7.4.1 Future Work

Certain aspects of the work presented in this thesis could be expanded on to further constrain the history of the Damang region.

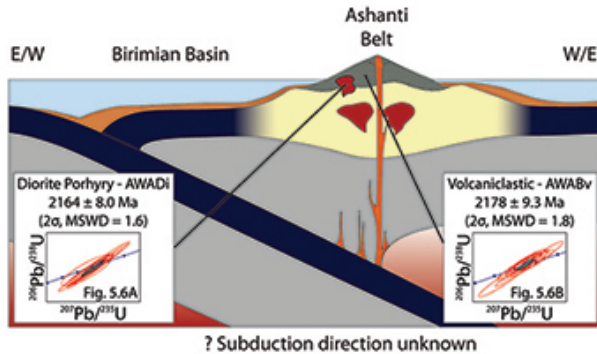
Thermodynamic Modelling

Recent updates to the THERMOCALC thermodynamic data set have incorporated S-bearing phases (Evans *et al.*, 2010; Holland & Powell, 2011). This is clearly relevant to the Damang system, although at present the data set is calibrated for rocks with very low (<0.1 mol %) sulphur contents and is therefore not applicable to the very sulphide-rich rocks at Damang. As the data set and calibrations are further improved however, the inclusion of sulphur into the thermodynamic models should be an obvious objective.

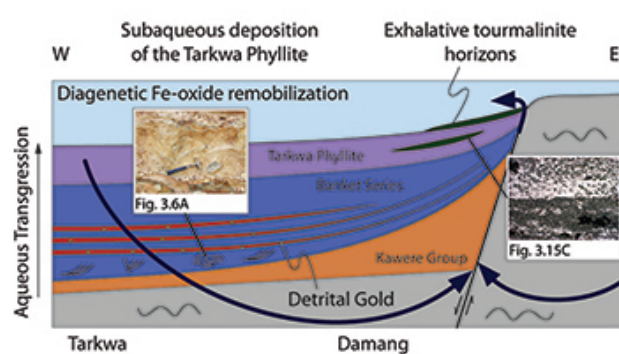
Geochronology

Since monazites are so abundant in the Tarkwa Phyllite, it is likely that some of them will be significantly larger than 10 μm in diameter. However, since these were not found in thin section, bulk sample crushing and mineral separation techniques, as applied to the zircons in this study, would be required. If sufficient coarse monazites were recovered, they could be analysed for U-Th-Pb isotopes, thereby providing a more precise estimate of the timing of peak regional metamorphism.

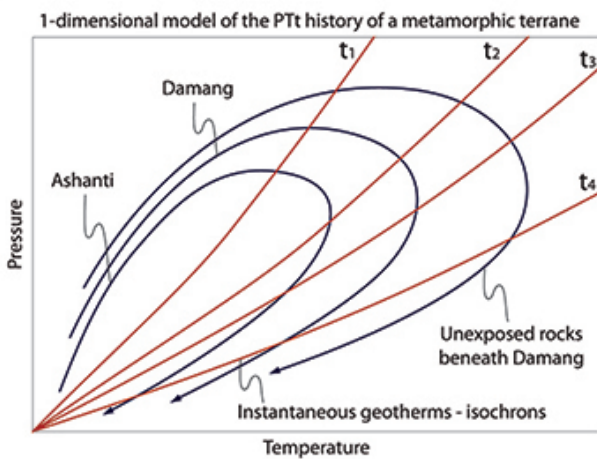
① Megascale – Growth of the Ashanti Belt – 2178 ± 9.3 - 2164.6 ± 8 Ma



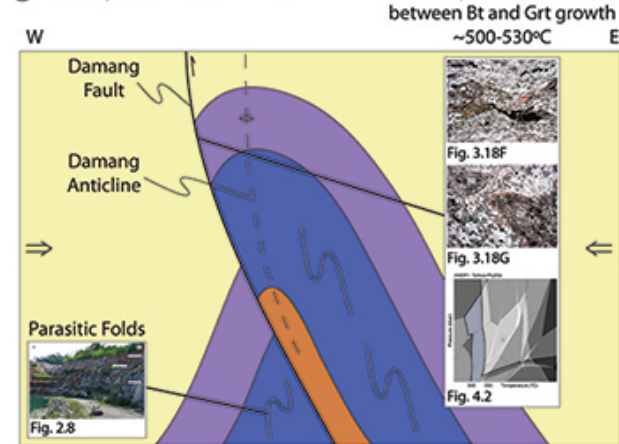
② Regional scale – Deposition of the Tarkwaian System – $<2133 \pm 4$ Ma



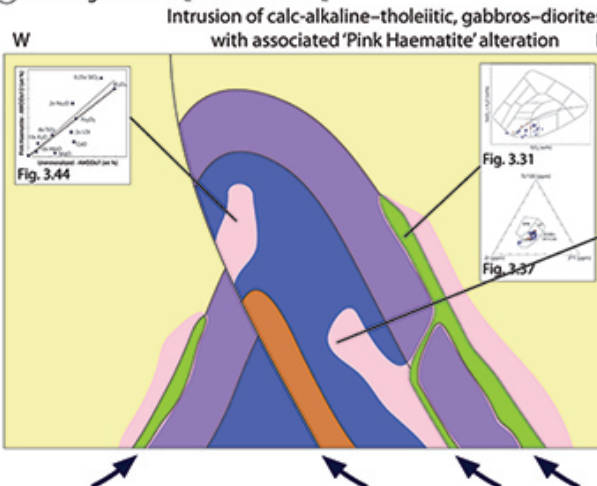
③ Regional scale – Onset of regional metamorphism



④ Damang scale – TD₁ deformation - t₁



⑤.1 Damang scale – TD₂ deformation - t₂



⑤.2 Damang scale – TD₂ deformation - t₂

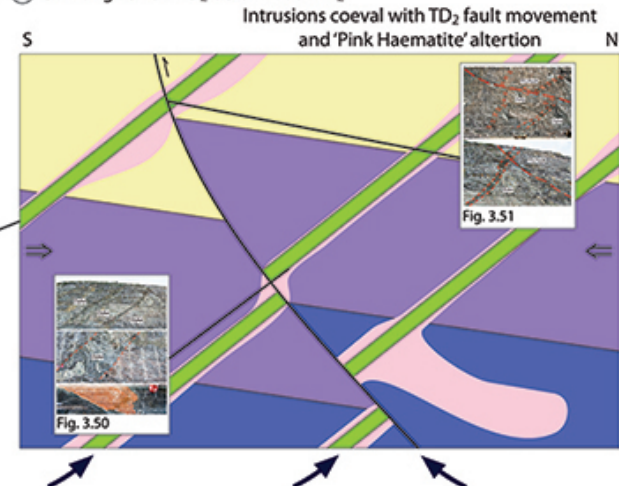
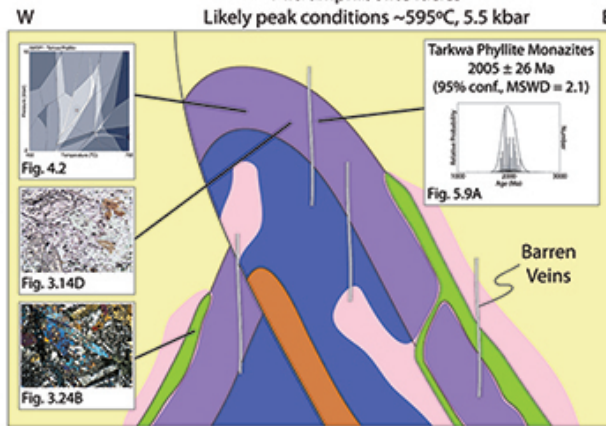
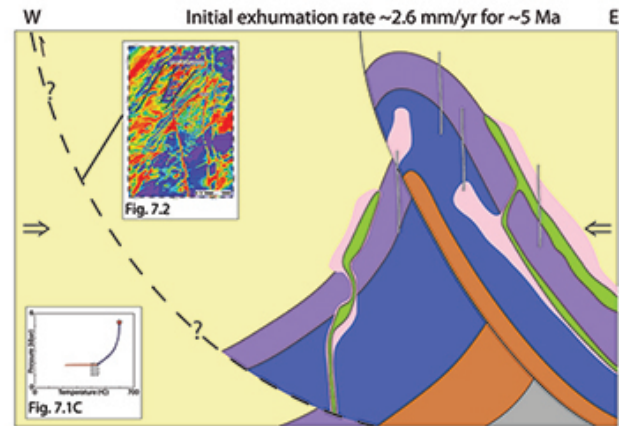


Figure 7.6: Schematic, sequential diagram illustrating the geological history and mineralization model for the Damang region, supported by figures presented in this thesis. See section 7.4 for discussion.

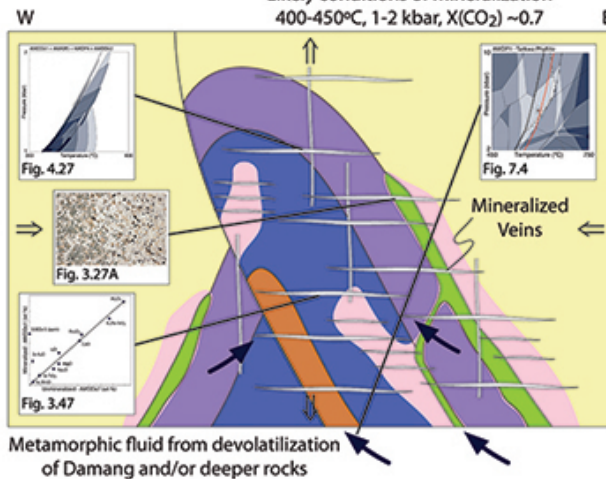
⑥ Damang scale – peak metamorphism - t_3 - 2005 ± 26 Ma
Mid amphibolite facies



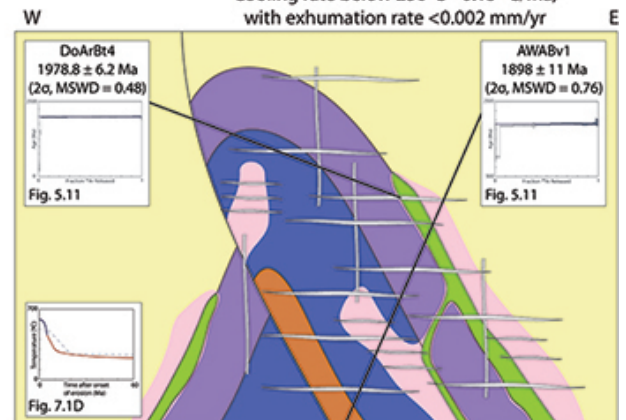
⑦ Regional scale – Commencement of regional exhumation - $t_3 < t < t_4$



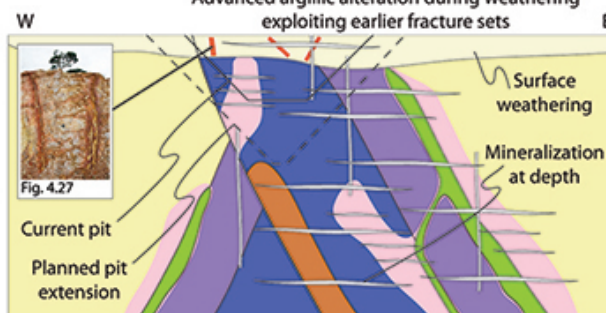
⑧ Damang scale – TD_3 deformation and gold mineralization - t_4
Likely conditions of mineralization
 $400\text{--}450^\circ\text{C}$, 1–2 kbar, $X(\text{CO}_2) \sim 0.7$



⑨ Regional scale – Post-exhumation cooling - after 1978.8 ± 6.2 Ma
Cooling rate below $250^\circ\text{C} \sim 0.15^\circ\text{C}/\text{Ma}$,
with exhumation rate < 0.002 mm/yr



⑩ Damang scale – Recent weathering and clay 'veining'
Advanced argillic alteration during weathering
exploiting earlier fracture sets



Similarly, directly constraining the timing of gold mineralization has yet to be achieved. Although, as shown here, Re/Os sulphide analysis was inconclusive, other techniques could be applied to the sulphide minerals, such as common Pb dating, that may be able to constrain their age of formation.

Additional $^{40}\text{Ar}/^{39}\text{Ar}$ analysis of other mineral phases, such as hornblende and muscovite, would provide further constraints on Damang's cooling history. Although these phases were not investigated in this study due to difficulties in analysis (hornblende) and mineral separation (muscovite) a more dedicated study may overcome these problems and produce a valuable addition to the entire Damang data set.

Geophysics

Detailed analysis and modelling of regional geophysical data, beyond the immediate Damang camp, may shed more light on the nature of the structures identified to the west of Damang. Thus, their role in the mineralization model presented above can be assessed. Additionally, seismic profiles across the Damang camp may serve to identify and locate structures potentially controlling exhumation, thereby placing a maximum depth limit on gold mineralization. Further detailed geomagnetic analysis across the Damang camp may help to elucidate zones of mineralization, particularly in the Banket Series quartzites where there is an accompanying destruction of magnetite. With regards to the dolerite intrusions however, the effect is much less marked. Furthermore, magnetite destruction is localized to vein margins and may not be detectable on larger scale surveys.

Gold Analysis

Detailed analysis of gold chemistry and physical morphology may be useful in determining a source, particularly with regards to the possible remobilization of early detrital gold during the later hydrothermal event.

References

- ABOUCAMI, W., BOHER, M., MICHARD, A., & ALBAREDE, F. 1990. A major 2.1 Ga event of mantle magmatism in West Africa: An early stage of crustal accretion. *Journal of Geophysical Research*, **95**, 17605–17629.
- ALEXANDRE, P. 2011. Comparison between grain size and multi-mineral $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology. *Geochimica et Cosmochimica Acta*, **75**(15), 4260–4272.
- ALLIBONE, A., TEASDALE, J., CAMERON, G., & ETHERIDGE, M. 2002. Timing and structural controls on gold mineralization at the Bogoso gold mine, Ghana, West Africa. *Economic Geology*, **97**(5), 949–969.
- ALLIBONE, A., HAYDEN, P., CAMERON, G., & DUKU, F. 2004. Paleoproterozoic Gold Deposits Hosted by Albite-and Carbonate-Altered Tonalite in the Chirano District, Ghana, West Africa. *Economic Geology*, **99**(3), 479–497.
- ANCEY, M., BASTENAIRE, F., & TIXIER, R. 1978. Application des méthodes statistiques en microanalyse. *Pages 323–347 of: Microanalyse, microscopie électronique à balayage*. Les Éditions du Physicien, Orsay.
- APPIAH, H. 1991. Geology and mine exploration trends of Prestea Goldfields, Ghana. *Journal of African Earth Sciences*, **13**(2), 235–241.
- ARCULUS, R.J. 2003. Use and abuse of the terms calcalkaline and calcalkalic. *Journal of Petrology*, **44**(5), 929–935.
- BARLEY, M.E., & GROVES, D.I. 1992. Supercontinent cycles and the distribution of metal deposits through time. *Geology*, **20**(4), 291–294.
- BARRA, F., RUIZ, J., MATHUR, R., & TITLEY, S. 2003. A Re-Os study of sulfide minerals from the Bagdad porphyry Cu-Mo deposit, northern Arizona, USA. *Mineralium Deposita*, **38**, 585–596.
- BASSOT, J.P., & CAEN-VACHETTE, M. 1984. Données géochronologiques et géochimiques nouvelles sur les granitoids de l'Est Sénégal: implications sur l'histoire du Birrimien de cette région. *Pages 191–209 of: Géologique Africaine*.
- BAXTER, E.F. 2010. Diffusion of Noble Gases. *In: ZHANG, Y., & CHERNIAK, D.J. (eds), Diffusion in Minerals and Melts*. Reviews in Mineralogy and Geochemistry, **72**.

- BEGG, G.C., GRIFFIN, W.L., NATAPOV, L.M., O'REILLY, S.Y., GRAND, S.P., O'NEILL, C.J., HRONSKY, J.M.A., POUDJOM DJOMANI, Y., SWAIN, C.J., DEEN, T., & P.BOWDEN. 2009. The lithospheric architecture of Africa: Seismic tomography, mantle petrology and tectonic evolution. *Geosphere*, **5**, 23–50.
- BERGE, J. 2011. Paleoproterozoic, turbidite-hosted, gold deposits of the Ashanti gold belt (Ghana, West Africa): Comparative analysis of turbidite-hosted gold deposits and an updated genetic model. *Ore Geology Reviews*, **39**(1–2), 91–100.
- BERGENE, J. 1944. West African Banket: some surveying, geological and other features. *Transactions of the Institute of Materials, Minerals and Mining*, **53**, 253–293.
- BERMAN, M., BISCHOF, L., & HUNTINGTON, J. 1999. Algorithms and software for the automated identification of minerals using field spectra or hyperspectral imagery. *Pages 222–232 of: Proceedings of the 13th International Conference on Applied Geologic Remote Sensing, Volume 1, Vancouver, Canada, 1–3 March 1999*.
- BERMAN, R.G. 1988. Internally consistent thermodynamic data for minerals in the system $\text{Na}_2\text{O}-\text{K}_2\text{O}-\text{CaO}-\text{MgO}-\text{FeO}-\text{Fe}_2\text{O}_3-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{TiO}_2-\text{H}_2\text{O}-\text{CO}_2$. *Journal of Petrology*, **29**, 445–522.
- BICKLE, M., HAWKSEWORTH, C., ENGLAND, P.C., & ATHEY, D. 1975. A preliminary thermal model for regional metamorphism in the Eastern Alps. *Earth and Planetary Science Letters*, **26**, 13–28.
- BIERLEIN, F.P., GROVES, D.I., & CAWOOD, P.A. 2009. Metallogeny of accretionary orogens – The connection between lithospheric processes and metal endowment. *Ore Geology Reviews*, **36**(4), 282–292.
- BLENKINSOP, T., SCHMIDT MUMM, A., KUMI, R., & SANGMOR, S. 1994. Structural geology of the Ashanti gold mine. *Geologisches Jahrbuch Reihe*, **D100**, 131–153.
- BOHER, M., ABOUCHAMI, W., MICHARD, A., ALBARDE, F., & ARNDT, N.T. 1992. Crustal growth in West Africa at 2.1 Ga. *Journal of Geophysical Research*, **97**, 345–369.
- BOSSIERE, G., BONKOUNGOU, I., PEUCAT, J., & PUPIN, J. 1996. Origin and age of Paleoproterozoic conglomerates and sandstones of the Tarkwaian Group in Burkina Faso, West Africa. *Precambrian Research*, **80**, 153–172.
- BRABHAM, G. 1998. *The regional geological setting and nature of the Damang stockwork gold deposit - a new type of gold deposit in Ghana?* MSc Thesis, University of Western Australia.
- BRAY, A. 1928. Notes on the Banket reefs of the gold coast colony. *Transactions of the Institute of Materials, Minerals and Mining*, **38**, 21–69.

- BRENAN, J.M., CHERNIAK, D.J., & ROSE, L.A. 2000. Diffusion of osmium in pyrrhotite and pyrite: implications for closure of the Re-Os isotopic system. *Earth and Planetary Science Letters*, **180**, 399–413.
- BROWN, T., HETHERINGTON, L., HANNIS, S., BIDE, T., BENHAM, A.J., IDOINE, N., & LUSTY, P. 2009. *World Mineral Production 2003-07*. Nottingham, UK: British Geological Survey.
- BROWNSCOMBE, W. 2009. *The Tinga anomaly: a new style of gold mineralization in Ghana?* MSc Thesis, University of Oxford.
- BUCKINGHAM, W.F., & SOMMER, S.E. 1983. Mineralogical characterization of rock surfaces formed by hydrothermal alteration and weathering - application to remote sensing. *ECONOMIC GEOLOGY*, **78**, 664–674.
- BURBANK, D.W. 2002. Rates of erosion and their implications for exhumation. *Mineralogical Magazine*, **66**(1), 25–52.
- BURGESS, R., & TURNER, G. 1998. Laser argon-40–argon-39 age determinations of Luna 24 mare basalts. *Meteoritics and Planetary Science*, **33**, 921–935.
- BURGESS, R., KIVIETS, G.B., & HARRIS, J.W. 2004. Ar-Ar age determinations of eclogitic clinopyroxene and garnet inclusions in diamonds from the Venetia and Orapa kimberlites. *Lithos*, **77**, 113–124.
- CANN, J. 1970. Rb, Sr, Y, Zr and Nb in some ocean floor basaltic rocks. *Earth and Planetary Science Letters*, **10**, 7–11.
- CARDON, O., REISBERG, L., ANDRE-MAYER, A.S., LEROY, J., MILU, V., & ZIMMERMANN, C. 2008. Re-Os Systematics of pyrite from the Bolcana porphyry copper deposit, Apuseni Mountains, Romania. *Economic Geology*, **103**(8), 1695–1702.
- CARRANZA, E.J.M., OWUSU, E.A., & HALE, M. 2009. Mapping of prospectivity and estimation of number of undiscovered prospects for lode gold, southwestern Ashanti Belt, Ghana. *Mineralium Deposita*, **44**(8), 915–938.
- CATLOS, E.J., HARRISON, T.M., KOHN, M.J., GROVE, M., RYERSON, F.J., MANNING, C.E., & UPRETI, B.N. 2001. Geochronologic and thermobarometric constraints on the evolution of the Main Central Thrust, central Nepal Himalaya. *Journal of Geophysical Research*, **106**(B8), 16177–16204.
- CHALAPATHI RAO, N.V., BURGESS, R., LEHMANN, B., MAINKAR, D., PANDE, S.K., HARI, K.R., & BODHANKAR, N. 2011. $^{40}\text{Ar}/^{39}\text{Ar}$ ages of mafic dykes from the Mesoproterozoic Chhattisgarh basin, Bastar craton, Central India: Implication for the origin and spatial extent of the Deccan Large Igneous Province. *Lithos*, **125**(3-4), 994–1005.

- CLARK, R.N. 1995. Reflectance spectra. *Pages 178–188 of: Rock physics & phase relations: a handbook of physical constants*. American Geophysical Union.
- CLINE, J.S., HOFSTRA, A., MUNTEAN, J.L., TOSDAL, R.M., & HICKEY, K.A. 2005. Carlin-type gold deposits in Nevada: Critical geologic characteristics and viable models. *Pages 451–484 of: HEDENQUIST, J.W., THOMPSON, J.F.H., GOLDFARB, R.J., & RICHARDS, J.P. (eds), ECONOMIC GEOLOGY 100TH ANNIVERSARY VOLUME*. Society of Economic Geologists.
- COCHERIE, A., LEGENDRE, O., PEUCAT, J.J., & KOUAMELAN, A.N. 1998. Geochronology of polygenetic monazites constrained by in situ electron microprobe Th-U-total lead determination: Implications for lead behaviour in monazite. *Geochimica et Cosmochimica Acta*, **62**(14), 2475–2497.
- COGGON, R., & HOLLAND, T. 2002. Mixing properties of phengitic micas and revised garnet-phengite thermobarometers. *Journal of Metamorphic Geology*, **20**(7), 683–696.
- CORRIE, S.L., & KOHN, J. 2008. Trace-element distributions in silicates during prograde metamorphic reactions: implications for monazite formation. *Journal of Metamorphic Geology*, **26**, 451–464.
- COUNCIL, THE WORLD GOLD. 2011 (February). *Gold Demand Trends. Full Year 2010*. [Online]. Available at: http://www.gold.org/investment/research/regular_reports/gold_demand_trends/. File 2010 Q4. [accessed 10 August 2011].
- COX, K.G., BELL, J.D., & PANKHURST, R.J. 1979. *The interpretation of igneous rocks*. London, UK: Allen & Unwin.
- CRANK, J. 1975. *The mathematics of diffusion*. 2 edn. Oxford, UK: Clarendon Press.
- CRAW, D., UPTON, P., YU, B.-S., HORTON, T., & CHEN, Y.-G. 2010. Young orogenic gold mineralisation in active collisional mountains, Taiwan. *Mineralium Deposita*, **45**(7), 631–646.
- CRAWFORD, M.L. 1966. Composition of plagioclase and associated minerals in some schists from Vermont, USA, and South Westland, New Zealand, with inferences about the peristerite solvus. *Contributions to mineralogy and petrology*, **13**(3), 269–294.
- CREASER, R.A., PAPANASTASSIOU, D.A., & WASSERBURG, G.J. 1991. Negative thermal ion mass spectrometry of osmium, rhenium and iridium. *Geochimica et Cosmochimica Acta*, **55**, 397–401.
- CROWLEY, J.K., & ZIMBELMAN, D.R. 1997. Mapping hydrothermally altered rocks on Mount Rainier, Washington, with Airborne Visible/Infrared Imaging Spectrometer [AVIRIS] data. *Geology*, **25**(6), 559–562.

- CROWSON, P. 2011. Demand, supply and prices: The horse, cart and driver of the mineral industries (Abstract). *Page 8 of: Fermor 2011: Ore deposits in an evolving Earth*.
- DADSON, S.J., HOVIUS, N., CHEN, H., DADE, W.B., HSIEH, M.-L., WILLET, S.D., HU, J.-C., HONG, M.-J., CHEN, M.-C., STARK, C.P., LAGUE, D., & JIN, J.-C. 2003. Links between erosion, runoff variability and seismicity in the Taiwan orogen. *Nature*, **426**(6967), 648–651.
- DAMPARE, S.B., SHIBATA, T., ASIEDU, D.K., OSAE, S., & BANOENG-YAKUBO, B. 2008. Geochemistry of Paleoproterozoic metavolcanic rocks from the southern Ashanti volcanic belt, Ghana: Petrogenetic and tectonic setting implications. *Precambrian Research*, **162**(3-4), 403–423.
- DAVIS, D., HIRDES, W., SCHALTEGGER, U., & NUNOO, E. 1994. U-Pb age constraints on deposition and provenance of Birimian and gold-bearing Tarkwaian sediments in Ghana, West Africa. *Precambrian Research*, **67**, 89–107.
- DE LA ROCHE, H., LETERRIER, J., GRANDCLAUDE, P., & MARCHAL, M. 1980. A classification of volcanic and plutonic rocks using R_1R_2 -diagram and major-element analyses – its relationships with current nomenclature. *Chemical Geology*, **29**, 183–210.
- DEER, W.A., HOWIE, R.A., & ZUSSMAN, J. 1992. *An introduction to the rock-forming minerals*. 2 edn. Harlow, UK: Longman.
- DIDIER, J. 1987. Contribution of enclave studies to the understanding of origin and evolution of granitic magmas. *Geologisches Rundschau*, **76**(1), 41–50.
- DIENER, J., POWELL, R., WHITE, R., & HOLLAND, T. 2007. A new thermodynamic model for clino- and orthoamphiboles in the system $\text{Na}_2\text{O}-\text{CaO}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{H}_2\text{O}-\text{O}$. *Journal of Metamorphic Geology*, **25**, 631–656.
- DIENER, J.F.A., & POWELL, R. 2011. Revised activity-composition models for clinopyroxene and amphibole. *Journal of Metamorphic Geology*.
- DINI, A., ROCCHI, S., & WESTERMAN, D.S. 2004. Reaction microtextures of REE-Y-Th-U accessory minerals in the Monte Capanne pluton (Elba Island, Italy): a possible indicator of hybridization processes. *Lithos*, **78**(1-2), 101–118.
- DODSON, M.H. 1973. Closure temperature in cooling geochronological and petrological systems. *Contributions to Mineralogy and Petrology*, **40**(3), 259–274.
- DORAIS, M.J., WHITNEY, J.A., & RODEN, M.F. 1990. Origin of mafic enclaves in the Dinkey Creek Pluton, Central Sierra Nevada Batholith, California. *Journal of Petrology*, **31**(4), 853–881.

- DROOP, W.G.T.R. 1987. A general equation for estimating Fe^{3+} concentrations in ferromagnesian silicates and oxides from microprobe analyses, using stoichiometric criteria. *Mineralogical Magazine*, **51**, 431–435.
- DUBÉ, B., & GOSSELIN, P. 2007. Greenstone-hosted quartz-carbonate vein deposits. *Pages 49–73 of: Mineral Deposits of Canada: A synthesis of major deposit-types, district metallogeny, the evolution of geological provinces and exploration methods*. Geological Association of Canada, Mineral Deposits Division, Special Publication No. 5.
- EATON, J.W. 2002. *GNU Octave Manual*. Network Theory Limited. <http://www.octave.org>.
- EISENLOHR, B.N. 1989. *The structural geology of the Birimian and Tarkwaian rocks of southwest Ghana*. Hannover, Germany: Bundesanstalt für Geowissenschaften und Rohstoffe.
- EISENLOHR, B.N. 1992. Conflicting evidence on the timing of mesothermal and paleoplacer gold mineralisation in early Proterozoic rocks from southwest Ghana, West Africa. *Mineralium Deposita*, **27**, 23–29.
- EISENLOHR, B.N., & HIRDES, W. 1992. The structural development of the early Proterozoic Birimian and Tarkwaian rocks of southwest Ghana, West Africa. *Journal of African Earth Sciences*, **14**(3), 313–325.
- ENGLAND, P.C. 1978. Some thermal considerations of the Alpine metamorphism – past, present and future. *Tectonophysics*, **46**, 21–40.
- ENGLAND, P.C., & THOMPSON, A.B. 1984. Pressure–Temperature–Time Paths of Regional Metamorphism I. Heat Transfer during the Evolution of Regions of Thickened Continental Crust. *Journal of Petrology*, **25**(4), 894–928.
- ENNIH, N., & LIÉGEOIS, J.-P. 2008. The boundaries of the West African craton, with special reference to the basement of the Moroccan metacratonic Anti-Atlas belt. *Pages 1–17 of: ENNIH, N., & LIÉGEOIS, J.-P. (eds), The boundaries of the West African craton*, vol. 297. Geological Society London Special Publications.
- EVANS, K.A., POWELL, R., & HOLLAND, T.J.B. 2010. Internally consistent data for sulphur-bearing phases and application to the construction of pseudosections for mafic greenschist facies rocks in $\text{Na}_2\text{O}-\text{CaO}-\text{K}_2\text{O}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{CO}_2-\text{O}-\text{S}-\text{H}_2\text{O}$. *Journal of Metamorphic Geology*, **28**(6), 667–687.
- FAURE, G. 1986. *Principles of Isotope Geology*. 2 edn. New York, USA: John Wiley and Sons.
- FERNANDES, V.A., BURGESS, R., & TURNER, G. 2000. Laser argon-40-argon-39 age studies of Dar al Gani 262 lunar meteorite. *Meteoritics and Planetary Science*, **35**, 1355–1364.

- FEYBESSE, J.L., & MILÉSI, J.-P. 1994. The Archean/Proterozoic contact zone in West Africa: a mountain belt of décollement thrusting and folding on a continental margin related to 2.1 Ga convergence of Archean cratons? *Precambrian Research*, **69**, 199–227.
- FEYBESSE, J.L., BILLA, M., GUERROT, C., DUGUEY, E., LESCUYER, J., MILÉSI, J.-P., & BOUCHOT, V. 2006. The paleoproterozoic Ghanaian province: Geodynamic model and ore controls, including regional stress modeling. *Precambrian Research*, **149**(3-4), 149–196.
- FINGER, F., BROSKA, I., ROBERTS, M.P., & SCHERMAIER, A. 1998. Replacement of primary monazite by apatite-allanite-epidote coronas in an amphibolite facies granite gneiss from the eastern Alps. *American Mineralogist*, **83**(3-4), 248–258.
- FLUDE, S. 2005. *Rhyolite volcanism in Iceland: Timing and timescales of eruption*. Ph.D. thesis, University of Manchester.
- FRASER, J. 2000. *The structural and metamorphic evolution of the deep crust in the Hunza Karakoram, Pakistan*. Ph.D. thesis, University of Oxford.
- FRIMMEL, H.E. 2008. Earth's continental crustal gold endowment. *Earth and Planetary Science Letters*, **267**(1-2), 45–55.
- GALIPP, K., KLEMD, R., & HIRDES, W. 2003. Metamorphism and geochemistry of the Paleoproterozoic Birimian Sefwi Volcanic Belt (Ghana, West Africa). *Geologisches Jahrbuch Reihe*, **D111**, 151–191.
- GLADWELL, D.R., LETT, R.E., & LAWRENCE, P. 1983. Application of reflectance spectrometry to mineral exploration using portable radiometers. *ECONOMIC GEOLOGY*, **78**, 699–710.
- GOLDFARB, R., GROVES, D., & GARDOLL, S. 2001. Orogenic gold and geologic time: a global synthesis. *Ore Geology Reviews*, **18**, 1–75.
- GOLDFARB, R., BAKER, T., DUBÉ, B., GROVES, D., HART, C., & GOSSELIN, P. 2005. Distribution, character, and genesis of gold deposits in Metamorphic Terranes. *Pages 407–450 of: HEDENQUIST, J.W., THOMPSON, J.F.H., GOLDFARB, R.J., & RICHARDS, J.P. (eds), ECONOMIC GEOLOGY 100TH ANNIVERSARY VOLUME*. Society of Economic Geologists.
- GOSSELIN, P., & DUBÉ, B. 2005. Gold deposits of the World: Distribution, geological parameters and gold content. *Geological Survey of Canada Open File*, **4895**, 1–271.
- GRANT, J.A. 1986. The isocon diagram - a simple solution to Gresens' equation for metasomatic alteration. *Economic Geology*, **81**, 1976–1982.

- GREEN, E., HOLLAND, T.J.B., & POWELL, R. 2007. An order-disorder model for omphacitic pyroxenes in the system jadeite-diopside-hedenbergite-acmite, with applications to eclogitic rocks. *American Mineralogist*, **92**(7), 1181–1189.
- GRESENS, R.L. 1967. Composition-volume relationships of metasomatism. *Chemical Geology*, **2**, 47–55.
- GROVE, M. 1993. *Thermal histories of southern California basement terranes*. Ph.D. thesis, University of California, Los Angeles.
- GROVE, M., & HARRISON, T.M. 1996. $^{40}\text{Ar}^*$ diffusion in Fe-rich biotite. *American Mineralogist*, **81**, 940–951.
- GROVES, D.I. 1993. The crustal continuum model for late-Archaeon lode-gold deposits of the Yilgarn Block, Western Australia. *Mineralium Deposita*, **28**(6), 366–374.
- GROVES, D.I., & BIERLEIN, F.P. 2007. Geodynamic settings of mineral deposit systems. *Journal of the Geological Society*, **164**, 19–30.
- GROVES, D.I., & GOLDFARB, R.J. 2007. The role of exhumation in the temporal distribution of ore deposits – A discussion. *Economic Geology*, **102**(1), 155–158.
- GROVES, D.I., PHILLIPS, G.N., HO, S.E., HOUSTOUN, S.M., & STANDING, C.A. 1987. Craton-scale distribution of Archean greenstone gold deposits; predictive capacity of the metamorphic model. *Economic Geology*, **82**(8), 2045–2058.
- GROVES, D.I., GOLDFARB, R.J., GEBRE-MARIAM, M., HAGEMANN, S.G., & ROBERT, F. 1998. Orogenic gold deposits: A proposed classification in the context of their crustal distribution and relationship to other gold deposit types. *Ore Geology Reviews*, **13**(1-5), 7–27.
- GROVES, D.I., GOLDFARB, R.J., ROBERT, F., & HART, C.J.R. 2003. Gold deposits in metamorphic belts: Overview of current understanding, outstanding problems, future research, and exploration significance. *Economic Geology*, **98**(1), 1–29.
- GROVES, D.I., CONDIE, K.C., GOLDFARB, R.J., HRONSKY, J.M.A., & VIELREICHER, R.M. 2005. Secular changes in global tectonic processes and their influence on the temporal distribution of gold-bearing mineral deposits. *Economic Geology*, **100**, 203–224.
- HAMMOND, N.Q., ROBB, L.J., FOYA, S., & ISHIYAMA, D. 2011. Mineralogical, fluid inclusion and stable isotope characteristics of Birimian orogenic gold mineralization at the Morila Mine, Mali, West Africa. *Ore Geology Reviews*, **39**, 218–229.

- HARCOUËT, V., GUILLOU-FROTTIER, L., BONNEVILLE, A., BOUCHOT, V., & MILÉSI, J.P. 2007. Geological and thermal conditions before the major Palaeoproterozoic gold-mineralization event at Ashanti, Ghana, as inferred from improved thermal modelling. *Precambrian Research*, **154**(1-2), 71–87.
- HARMER, R.E., & ARMSTRONG, R.A. 2000. Abstracts, Bushveld Complex Workshop, National Research Foundation, Gethlane Lodge, South Africa. *American Geophysical Union Reference*.
- HARRISON, T.M., DUNCAN, I., & MCDUGALL, I. 1985. Diffusion of ^{40}Ar in biotite: temperature, pressure and compositional effects. *Geochimica et Cosmochimica Acta*, **49**(11), 2461–2468.
- HART, C. 2007. Reduced intrusion-related gold systems. *Pages 95–112 of: Mineral Deposits of Canada: A synthesis of major deposit-types, district metallogeny, the evolution of geological provinces and exploration methods*. Geological Association of Canada, Mineral Deposits Division, Special Publication No. 5.
- HASTIE, A.R., KERR, A.C., PEARCE, J.A., & MITCHELL, S.F. 2007. Classification of altered volcanic island arc rocks using immobile trace elements: Development of the Th-Co discrimination diagram. *Journal of Petrology*, **48**(12), 2341–2357.
- HASTINGS, D.A. 1982. On the tectonics and metallogenesis of West Africa: a model incorporating new geophysical data. *Geoexploration*, **20**, 295–327.
- HEDENQUIST, J.W., ARRIBAS, R.A., & GONZALEZ, U.E. 2000. Exploration for epithermal gold deposits. *Pages 245–277 of: Gold in 2000*. Society of Economic Geologists, Reviews in Economic Geology, **13**.
- HEIN, K.A.A. 2010. Succession of structural events in the Goren greenstone belt (Burkina Faso). *Journal of African Earth Sciences*, **56**(2–3), 83–94.
- HELGESON, H.C., DELANY, J.M., NESBITT, H.W., & BIRD, D.K. 1978. Summary and critique of the thermodynamic properties of rock-forming minerals. *American Journal of Science*, **278A**, 229 pp.
- HENNIG, D., LEHMANN, B., BURGESS, R., & ALI NEKOUVAGHT TAK, M. 2008. Geology, geochemistry and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of the Cerro Millo epithermal high-sulfidation gold prospect, southern Peru. *Ore Geology Reviews*, **34**, 304–316.
- HENRY, D.J., GUIDOTTI, C.V., & THOMSON, J.A. 2005. The Ti-saturation surface for low-to-medium pressure metapelitic biotite: implications for geothermometry and Ti-substitution mechanisms. *American Mineralogist*, **90**, 316–328.

- HESS, J.C., LIPPOLT, H.J., GURBANOV, A.G., & MICHALSKI, I. 1993. The cooling history of the late Pliocene Eldzhurtinskiy granite (Caucasus, Russia) and the thermochronological potential of grain-size/age relationships. *Earth and Planetary Science Letters*, **117**(3-4), 393–406.
- HILL, R.I. 1988. San Jacinto Intrusive Complex 1. Geology and mineral chemistry, and a model for the intermittent recharge of tonalitic magma chambers. *Journal of Geophysical Research*, **93**(B9), 10325–10348.
- HILL, R.I., SILVER, L.T., CHAPPELL, B.W., & TAYLOR JR., H.P. 1985. Solidification and recharge of SiO₂-rich plutonic magma chambers. *Nature*, **313**, 643–646.
- HIRDES, W., & DAVIS, D. 1998. First U-Pb zircon age of extrusive volcanism in the Birimian Supergroup of Ghana, West Africa. *Journal of African Earth Sciences*, **27**(2), 291–294.
- HIRDES, W., & DAVIS, D. 2002. U-Pb Geochronology of Paleoproterozoic Rocks in the Southern Part of the Kedougou-Kéniéba Inlier, Senegal, West Africa: evidence for diachronous accretionary development of the Eburnean Province. *Precambrian Research*, **118**, 83–99.
- HIRDES, W., & NUNOO, B. 1994. The Proterozoic paleoplacers at Tarkwa Gold Mine, SW Ghana: Sedimentology, mineralogy, and precise age dating of the Main Reef and West Reef, and bearing on the investigations on the source area aspects. *Pages 247–311 of: OBERTHÜR, T. (ed), Metallogensis of selected gold deposits in Africa*. Geologisches Jahrbuch D100.
- HIRDES, W., SAAGER, R., & LEUBE, A. 1987. The Tarkwaian Group of Ghana: new aspects relating to its tectonic setting, structural evolution, gold mineralization and provenance area. *Abstr. International Conference on Metallogeny Related to Tectonics, Arusha*, 19–20.
- HIRDES, W., SAAGER, R., & LEUBE, A. 1988. New structural, radiometric and mineralogical aspects of the Au-bearing Tarkwaian group of Ghana. *Bicentennial Gold 88, Melbourne*, **146**, 148.
- HIRDES, W., DAVIS, D., & EISENLOHR, B.N. 1992. Reassessment of Proterozoic granitoid ages in Ghana on the basis of U/Pb zircon and monazite dating. *Precambrian Research*, **56**, 89–96.
- HIRDES, W., SENGER, R., ADJEL, J., EFA, E., LOH, G., & TETTEY, A. 1993. Explanatory notes for the geological map of southwest Ghana 1:100,000. Sheets Wiawso (0603C), Kukuom (0603B), Goaso (0603A), Sunyani (0703D), Berekum (0703C). *Geologisches Jahrbuch*, 1–139.

- HIRST, T. 1938. The geology of the Tarkwa goldfield and adjacent country. *Gold Coast Geological Survey Bulletin*, **10**, 24.
- HITZMAN, M., ORESKEN, N., & EINAUDI, M.T. 1992. Geological characteristics and tectonic setting of Proterozoic iron-oxide (Cu-U-Au-REE) deposits. *Precambrian Geology*, **58**, 241–287.
- HODGES, K.V., & MCKENNA, L.W. 1987. Realistic propagation of uncertainties in geologic thermobarometry. *American Mineralogist*, **72**, 671–680.
- HODGES, K.V., HARRIES, W.E., & BOWRING, S.A. 1994. $^{40}\text{Ar}/^{39}\text{Ar}$ age gradients in micas from a high-temperature-low-pressure metamorphic terrain: Evidence for very slow cooling and implications for the interpretation of age spectra. *Geology*, **22**, 55–58.
- HOLDAWAY, M.J. 2000. Application of new experimental and garnet Margules data to the garnet-biotite geothermometer. *American Mineralogist*, **85**(7-8), 881–892.
- HOLLAND, T., & BLUNDY, J. 1994. Non-ideal interactions in calcic amphiboles and their bearing on amphibole-plagioclase thermometry. *Contributions to Mineralogy and Petrology*, **116**(4), 433–447.
- HOLLAND, T., & POWELL, R. 1998. An internally consistent thermodynamic data set for phases of petrological interest. *Journal of Metamorphic Geology*, **16**, 309–343.
- HOLLAND, T.J.B. 2009 (January). *Muscovite-Phengite*. [Online]. Available at: <http://www.esc.cam.ac.uk/research/research-groups/holland/thermocalc/muscovite> [accessed 10 February 2009].
- HOLLAND, T.J.B., & POWELL, R. 1985. An internally consistent thermodynamic dataset with uncertainties and correlations: 2. Data and results. *Journal of Metamorphic Geology*, **3**, 343–370.
- HOLLAND, T.J.B., & POWELL, R. 1990. An enlarged and updated internally consistent thermodynamic dataset with uncertainties and correlations: the system $\text{K}_2\text{O}-\text{Na}_2\text{O}-\text{CaO}-\text{MgO}-\text{MnO}-\text{FeO}-\text{Fe}_2\text{O}_3-\text{Al}_2\text{O}_3-\text{TiO}_2-\text{SiO}_2-\text{C}-\text{H}_2-\text{O}_2$. *Journal of Metamorphic Geology*, **8**(1), 89–124.
- HOLLAND, T.J.B., & POWELL, R. 2003. Activity-composition relations for phases in petrological calculations: an asymmetric multicomponent formulation. *Contributions to Mineralogy and Petrology*, **145**(4), 492–501.
- HOLLAND, T.J.B., & POWELL, R. 2011. An improved and extended internally consistent thermodynamic dataset for phases of petrological interest, involving a new equation of state for solids. *Journal of Metamorphic Petrology*, **29**, 333–383.

- HOLLAND, T.J.B., BAKER, J.M., & POWELL, R. 1998. Mixing properties and activity-composition relationships of chlorites in the system MgO-FeO-Al₂O₃-SiO₂-H₂O. *European Journal of Mineralogy*, **10**, 395–406.
- IRVINE, T.N., & BARAGAR, W.R.A. 1971. A guide to the chemical classification of the common volcanic rocks. *Canadian Journal of Earth Science*, **8**, 523–548.
- JANOTS, E., ENGI, M., BERGER, A., ALLAZ, J., SCHWARZ, J.-O., & SPANDLER, C. 2008. Prograde metamorphic sequence of REE minerals in pelitic rocks of the Central Alps: implications for allanite–monazite–xenotime phase relations from 250 to 610°C. *Journal of Metamorphic Geology*, **26**, 509–526.
- JENSEN, L.S. 1976. A new cation plot for classifying subalkalic volcanic rocks. *Ontario Geological Survey Miscellaneous Paper*, **66**, 22 p.
- JOHN, T., KLEMD, R., HIRDES, W., & LOH, G. 1999. The metamorphic evolution of the Paleoproterozoic (Birimian) volcanic Ashanti belt (Ghana, West Africa). *Precambrian Research*, **98**, 11–30.
- JOURDAN, F., VERATI, C., & FÉRAUD, G. 2006. Intercalibration of the Hb3gr ⁴⁰Ar/³⁹Ar dating standard. *Chemical Geology*, **231**, 77–189.
- JUNNER, N.R. 1935. *Gold in the Gold Coast*. Ghana Geological Survey Memoir 4.
- JUNNER, N.R. 1940. *Geology of the Gold Coast and Western Togoland with Revised Geological Map*. Ghana Geological Survey Bulletin 11.
- JUNNER, N.R., HIRST, T., & SERVICE, H. 1942. *The Tarkwa Goldfield*. Gold Coast Geological Survey Memoir 6.
- KERN, R., & WEISBROD, A. 1967. *Thermodynamics for Geologists*. San Francisco, USA: Freeman, Cooper and Company.
- KERRICH, R. 1993. Perspectives on genetic models for lode gold deposits. *Mineralium Deposita*, **28**, 362–365.
- KERRICH, R., & CASSIDY, K. 1994. Temporal relationships of lode gold mineralization to accretion, magmatism, metamorphism and deformation - Archean to present: a review. *Ore Geology Reviews*, **9**, 263–310.
- KERRICH, R., & FYFE, W.S. 1981. The gold–carbonate association: Source of CO₂, and CO₂ fixation reactions in Archaean lode deposits. *Chemical geology*, **33**(1-4), 265–294.
- KESLER, S.E., & WILKINSON, B.H. 2006. The role of exhumation in the temporal distribution of ore deposits. *Economic Geology*, **101**, 919–922.

- KESLER, S.E., & WILKINSON, B.H. 2010. Global hydrothermal gold resources for the next millenium. *Pages 8–18 of: GOLDFARB, R.J., MARSH, E.E., & MONECKE, T. (eds), The challenge of finding new mineral resources: Global metallogeny, innovative exploration, and new discoveries.* Society of Economic Geologists Special Publication 15.
- KESSE, G.O. 1984. Resources of Gold in Phanerozoic Epithermal Deposits. *Pages 645–659 of: FOSTER, R.P. (ed), Gold '82: The geology, geochemistry and genesis of gold deposits.* Rotterdam, The Netherlands: Balkema.
- KITCO. 2011 (October). *Kitco - Gold precious metals.* [Online]. Available at: <http://www.kitco.com> [accessed 3 October 2011].
- KLEMD, R., HIRDES, W., OLESCH, M., & OBERTHÜR, T. 1993. Fluid inclusions in quartz-pebbles of the gold-bearing Tarkwaian conglomerates of Ghana as guides to their provenance area. *Mineralium Deposita*, **28**, 334–343.
- KLEMD, R., HUNKEN, U., & OLESCH, M. 2002. Metamorphism of the country rocks hosting gold-sulfide bearing quartz-veins in the Paleoproterozoic southern Kibi-Winneba belt (SE-Ghana). *Journal of African Earth Sciences*, **35**, 199–211.
- KOHN, M.J., & MALLOY, M.A. 2004. Formation of monazite via prograde metamorphic reactions among common silicates: Implications for age determination. *Geochimica et Cosmochimica Acta*, **68**(1), 101–113.
- KORZHINSKII, D.S. 1959. *Physiochemical basis of the analysis of the paragenesis of minerals.* New York, USA: Consultants Bureau. Translated from the Russian edition.
- KRETZ, R. 1983. Symbols for rock-forming minerals. *American Mineralogist*, **68**(1-2), 277–279.
- KRUSE, F.A., & HAUFF, P.L. 1991. Identification of illite polytype zoning in disseminated gold deposits using reflectance spectroscopy and X-ray diffraction - potential for mapping with imaging spectrometers. *IEEE Transactions on Geoscience and Remote Sensing*, **29**(1), 101–104.
- KUNO, H. 1968. Differentiation of basaltic magmas. *Pages 623–688 of: HESS, H.H., & POLDERVAART, A. (eds), Basalts*, vol. 2. New York: Wiley Interscience.
- LACHENBRUCH, A.H. 1968. Preliminary geothermal model of the Sierra Nevada. *Journal of Geophysical Research*, **73**(22), 6977–6989.
- LAFFITTE, P. 1957. *Introduction à l'étude des roches métamorphiques et des gîtes métallifères: physio-chimie et thermodynamique.* Paris, France: Masson.

- LARGE, R.R., BULL, S.W., & MASLENNIKOV, V.V. 2011. A carbonaceous sedimentary source-rock model for Carlin-type and orogenic gold deposits. *Economic Geology*, **106**, 331–358.
- LEDRU, P., JOHAN, V., MILÉSI, J.-P., & TEGYEY, M. 1994. Markers of the last stages of the Palaeoproterozoic collision: evidence for a 2 Ga continent involving circum-South Atlantic provinces. *Precambrian Research*, **69**, 169–191.
- LEE, J.K.W. 1995. Multipath diffusion in geochronology. *Contributions to Mineralogy and Petrology*, **120**(1), 60–82.
- LEUBE, A., HIRDES, W., MAUER, R., & KESSE, G. 1990. The early Proterozoic Birimian supergroup of Ghana and some aspects of its associated gold mineralization. *Precambrian Research*, **46**, 139–165.
- LISOWIEC, N. 2006. Precision estimation in electron microprobe monazite dating: Repeated measurements versus statistical (Poisson) based calculations. *Chemical geology*, **234**(3–4), 223–235.
- LUDWIG, K.R. 2003. User's Manual for Isoplot 3.00. A geochronological toolkit for Microsoft Excel. *Berkeley Geochronology Center Special Publication*, **4a**, 1–71.
- MARKLEY, M.J., TEYSSIER, C., & COSCA, M. 2002. The relation between grain size and $^{40}\text{Ar}/^{39}\text{Ar}$ date for Alpine white mica from the Siviez-Mischabel Nappe, Switzerland. *Journal of structural geology*, **24**(12), 1937–1955.
- MATLAB®. 2009. *version 7.9.0.259 (R2009b)*. Natick, Massachusetts: The MathWorks Inc.
- MCCUAIG, T.C., & KERRICH, R. 1998. P-T-t-deformation-fluid characteristics of lode gold deposits: evidence from alteration systematics. *Ore Geology Reviews*, **12**, 381–453.
- MCCUAIG, T.C., KERRICH, R., GROVES, D.I., & ARCHER, N. 1993. The nature and dimensions of regional and local gold-related hydrothermal alteration in tholeiitic metabasalts in the Norseman goldfields: the missing link in a crustal continuum of gold deposits? *Mineralium Deposita*, **28**(6), 420–435.
- MCDUGALL, I., & HARRISON, T.M. 1999. *Geochronology and thermochronology by the $^{40}\text{Ar}/^{39}\text{Ar}$ method*. 2 edn. Oxford, UK: Oxford University Press.
- MCKEITH, T.D., SCHODDE, R.C., & BALTIS, E.J. 2010. Gold discovery trends. *SEG Newsletter*, 1, 20–26.
- MERRIHUE, C., & TURNER, G. 1966. Potassium-argon dating by activation with fast neutrons. *Journal of Geophysical Research*, **71**(11), 2852–2857.

- MIKUCKI, E.J., & RIDLEY, J.R. 1993. The hydrothermal fluid of Archæan lode-gold deposits at different metamorphic grades: compositional constraints from ore and wallrock alteration assemblages. *Mineralium Deposita*, **28**, 469–481.
- MILÉSI, J.-P., LEDRU, P., ANKRAH, P., JOHAN, V., & MARCOUX, E. 1991. The metallogenic relationship between Birimian and Tarkwaian gold deposits in Ghana. *Mineralium Deposita*, **26**, 228–238.
- MILÉSI, J.-P., LEDRU, P., FEYBESSE, J.L., DOMMANGET, A., & MARCOUX, E. 1992. Early proterozoic ore deposits and tectonics of the Birimian orogenic belt, West Africa. *Precambrian Research*, **58**, 305–344.
- MIYASHIRO, A. 1978. Nature of alkalic volcanic series. *Contributions to Mineralogy and Petrology*, **66**, 91–104.
- MONTEL, J.M., FORET, S., VESCHAMBRE, M., NICOLLET, C., & PROVOST, A. 1996. Electron microprobe dating of monazite. *Chemical Geology*, **131**(1-4), 37–53.
- MONTEL, J.M., KORNPORST, J., & VIELZEUF, D. 2000. Preservation of old U-Th-Pb ages in shielded monazite: example from the Beni Bousera Hercynian kinzigites (Morocco). *Journal of Metamorphic Geology*, **18**(3), 335–342.
- MORELLI, J.M. 2008. *Rhenium-osmium geochronology of low-level sulfide minerals from hydrothermal ore deposits – Applications, limitations, and implications*. Ph.D. thesis, University of Alberta.
- MORELLI, R.M., BELL, C.C., CREASER, R.A., & SIMONETTI, A. 2010. Constraints on the genesis of gold mineralization at the Homestake Gold Deposit, Black Hills, South Dakota from rhenium–osmium sulfide geochronology. *Mineralium Deposita*, **45**(5), 461–480.
- MULLEN, E.D. 1983. MnO/TiO₂/P₂O₅: a minor element discriminant for basaltic rocks of oceanic environments and its implications for petrogenesis. *Earth and Planetary Science Letters*, **62**, 53–62.
- MUMIN, A.H., & FLEET, M.E. 1995. Evolution of gold mineralization in the Ashanti Gold Belt, Ghana: Evidence from carbonate compositions and parageneses. *Mineralogy and Petrology*, **55**, 265–280.
- MUMIN, A.H., FLEET, M.E., & CHRYSOULIS, S.L. 1994. Gold mineralization in As-rich mesothermal gold ores of the Bogosu-Prestea mining district of the Ashanti Gold Belt, Ghana: remobilization of "invisible" gold. *Mineralium Deposita*, **29**, 445–460.
- MUMIN, A.H., FLEET, M.E., & LONGSTAFF, F.J. 1996. Evolution of hydrothermal fluids in the Ashanti Gold Belt, Ghana: Stable isotope geochemistry of carbonates, graphite, and quartz. *Economic Geology*, **91**, 135–148.

- OBERTHÜR, T., VETTER, U., SCHMIDT MUMM, A., WEISER, T., AMANOR, J.A., GYAPONG, W.A., KUMI, R., & BLENKINSOP, T.G. 1994. The Ashanti gold mine at Obuasi, Ghana: Mineralogical, geochemical, stable isotope and fluid inclusion studies on the metallogenesis of the deposit. *Pages 31–129 of: OBERTHÜR, T. (ed), Metallogenesis of selected gold deposits in Africa*. Geologisches Jahrbuch D100.
- OBERTHÜR, T., SCHMIDT MUMM, A., VETTER, U., SIMON, K., & AMANOR, J.A. 1996. Gold mineralization in the Ashanti Belt of Ghana: Genetic constraints of the stable isotope geochemistry. *Economic Geology*, **91**, 289–301.
- OBERTHÜR, T., WEISER, T., AMANOR, J.A., & CHRYSOULIS, S.L. 1997. Mineralogical siting and distribution of gold in quartz veins and sulfide ores of the Ashanti mine and other deposits in the Ashanti belt of Ghana: genetic implications. *Mineralium Deposita*, **32**, 2–15.
- OBERTHÜR, T., VETTER, U., DAVIS, D., & AMANOR, J.A. 1998. Age constraints on gold mineralization and Paleoproterozoic crustal evolution in the Ashanti belt of southern Ghana. *Precambrian Research*, **89**, 129–143.
- ONSTOTT, T.C., PHILLIPS, D., & PRINGLE-GODELL, L. 1991. Laser microprobe measurement of chlorine and argon zonation in biotite. *Chemical geology*, **90**(1-2), 145–168.
- OXBURGH, E., & TURCOTTE, D. 1974. Thermal gradients and regional metamorphism in overthrust terrains with special reference to the Eastern Alps. *Schweizerische Mineralogische und Petrographische Mitteilungen*, **54**, 642–662.
- PARRISH, R.R. 1990. U-Pb dating of monazite and its application to geological problems. *Canadian Journal of Earth Sciences*, **27**(11), 1431–1450.
- PATTISON, D.R.M., & TINKHAM, D.K. 2009. Interplay between equilibrium and kinetics in prograde metamorphism of pelites: an example from the Nelson aureole, British Columbia. *Journal of Metamorphic Geology*, **27**(4), 249–279.
- PEARCE, J.A., & CANN, J.R. 1973. Tectonic setting of basic volcanic rocks determined using trace element analyses. *Earth and Planetary Science Letters*, **19**, 290–300.
- PECCERILLO, A., & TAYLOR, S.R. 1976. Geochemistry of Eocene calc-alkaline volcanic rocks from the Kastamonu Area, Northern Turkey. *Contributions to Mineralogy and Petrology*, **58**, 63–81.
- PETTKE, T., DIAMOND, L.W., & VILLA, I.M. 1999. Mesothermal gold veins and metamorphic devolatilization in the northwestern Alps: The temporal link. *Geology*, **27**(7), 641–644.

- PHILLIPS, G.N., & POWELL, R. 1993. Link between gold provinces. *Economic Geology*, **88**(5), 1084–1098.
- PHILLIPS, G.N., & POWELL, R. 2009. Formation of gold deposits: Review and evaluation of the continuum model. *Earth Science Reviews*, **94**(1-4), 1–21.
- PHILLIPS, G.N., & POWELL, R. 2010. Formation of gold deposits: a metamorphic devolatilization model. *Journal of Metamorphic Geology*, **28**(6), 689–718.
- PIGOIS, J.-P. 2003. *The Damang gold deposit, Ghana: A Palaeoproterozoic orogenic gold deposit overprinting weak Tarkwa-type palaeoplacer mineralization*. Ph.D. thesis, University of Western Australia.
- PIGOIS, J.-P., GROVES, D., FLETCHER, I., MCNAUGHTON, N., & SNEE, L. 2003. Age constraints on Tarkwaian palaeoplacer and lode-gold formation in the Tarkwa-Damang district, SW Ghana. *Mineralium Deposita*, **38**(6), 695–714.
- PITCAIRN, I.K., TEAGLE, D.A.H., CRAW, D., OLIVO, G.R., KERRICH, R., & BREWER, T.S. 2006. Sources of metals and fluids in orogenic gold deposits: insights from the Otago and Alpine Schists, New Zealand. *Economic Geology*, **101**(8), 1525–1546.
- PLIMER, I.R. 1984. The role of fluorine in submarine exhalative systems with special reference to Broken Hill, Australia. *Mineralium Deposita*, **19**, 19–25.
- PLIMER, I.R. 1987. The association of tourmalinite with stratiform scheelite deposits. *Mineralium Deposita*, **22**, 282–291.
- POWELL, R. 1978. *Equilibrium thermodynamics in petrology: An introduction*. London, UK: Harper & Row.
- POWELL, R. 1985. Geothermometry and geobarometry: a discussion. *Journal of the Geological Society, London*, **142**, 29–38.
- POWELL, R. 1987. Darken's quadratic formalism and the thermodynamics of minerals. *American Mineralogist*, **72**(1-2), 1–11.
- POWELL, R., & HOLLAND, T. 2008. On thermobarometry. *Journal of Metamorphic Geology*, **26**(2), 155–179.
- POWELL, R., & HOLLAND, T.J.B. 1988. An internally consistent dataset with uncertainties and correlations: 3. Applications to geobarometry, worked examples and a computer program. *Journal of Metamorphic Geology*, **6**, 173–204.
- POWELL, R., & HOLLAND, T.J.B. 1994. Optimal geothermometry and geobarometry. *American Mineralogist*, **79**, 120–133.

- POWELL, R., HOLLAND, T., & WORLEY, B. 1998. Calculating phase diagrams involving solid solutions via non-linear equations, with examples using THERMOCALC. *Journal of Metamorphic Geology*, **16**, 577–588.
- PRIGOGINE, I., & DEFAY, R. 1954. *Chemical Thermodynamics*. London, UK: Longmans, Green & Co.
- PROST, G. 1980. Alteration mapping with airborne multispectral scanners. *ECONOMIC GEOLOGY*, **75**, 894–906.
- RAMBERG, H. 1952. *The origin of metamorphic and metasomatic rocks*. Chicago, USA: University of Chicago Press.
- RASMUSSEN, B., & MUHLING, J.R. 2009. Reactions destroying detrital monazite in greenschist-facies sandstones from the Witwatersrand basin, South Africa. *Chemical geology*, **264**(1-4), 311–327.
- RASMUSSEN, B., MUHLING, J.R., FLETCHER, I.R., & WINGATE, M.T.D. 2006. In situ SHRIMP U-Pb dating of monazite integrated with petrology and textures: Does bulk composition control whether monazite forms in low-Ca pelitic rocks during amphibolite facies metamorphism? *Geochimica et Cosmochimica Acta*, **70**, 3040–3058.
- REID, J.B, EVANS, O.C., & FATES, D.G. 1983. Magma mixing in granitic rocks of the Central Sierra Nevada, California. *Earth and Planetary Science Letters*, **66**, 243–261.
- RIDLEY, J.R., & DIAMOND, L.W. 2000. Fluid chemistry of orogenic lode gold deposits and implications for genetic models. *SEG Reviews in Economic Geology*, **13**, 141–162.
- ROBB, L.J. 2005. *Introduction to Ore-Forming Processes*. Oxford, UK: Blackwell Publishing.
- ROBIE, R.A., & WALDBAUM, R. 1968. Thermodynamic properties of minerals and related substances at 298.15 K (25°C) and one atmosphere (1.013 bars) pressure and at higher temperatures. *U.S. Geological Survey Bulletin*, **1259**, 256 pp.
- ROBIE, R.A., HEMINGWAY, B.S., & FISHER, J.R. 1979. Thermodynamic properties of minerals and related substances at 298.15 K (25°C) and 1 bar (10⁵ Pascals) pressure and at higher temperatures. *U.S. Geological Survey Bulletin*, **1452**, 456 pp.
- ROCK, N.M.S., & GROVES, D.I. 1988. Do lamprophyres carry gold as well as diamonds? *Nature*, **332**, 253–255.
- ROWAN, L.C., & MARS, J.C. 2003. Lithologic mapping in the Mountain Pass, California area using advanced spaceborne thermal emission and reflection radiometer (ASTER) data. *Remote Sensing of the Environment*, **84**, 350–366.

- RUTHERFORD, E. 1900. A radioactive substance emitted from thorium compounds. *Philosophical Magazine*, **49**, 1–14.
- SAWKINS, F.J. 1990. *Metal Deposits in Relation to Plate Tectonics*. 2 edn. New York, USA: Springer-Verlag.
- SCHERRER, N.C., ENGI, M., GNOS, E., JAKOB, V., & LIECHTI, A. 2000. Monazite analysis; from sample preparation to microprobe age dating and REE quantification. *Schweizerische Mineralogische und Petrographische Mitteilungen*, **80**, 93–105.
- SCHMIDT MUMM, A., OBERTHÜR, T., VETTER, U., & BLENKINSOP, T.G. 1997. High CO₂ content of fluid inclusions in gold mineralisations in the Ashanti Belt, Ghana: A new category of ore forming fluids? *Mineralium Deposita*, **32**, 107–118.
- SELBY, D., & CREASER, R.A. 2001. Re-Os geochronology and systematics in molybdenite from the Endako porphyry molybdenum deposit, British Columbia, Canada. *Economic Geology*, **96**, 197–204.
- SELBY, D., & CREASER, R.A. 2003. Re-Os geochronology of organic rich sediments: an evaluation of organic matter analysis methods. *Chemical Geology*, **200**, 225–240.
- SELBY, D., CREASER, R.A., STEIN, H.J., MARKEY, R.J., & HANNAH, J.L. 2007. Assessment of the ¹⁸⁷Re decay constant by cross calibration of Re-Os molybdenite and U-Pb zircon chronometers in magmatic ore systems. *Geochimica et Cosmochimica Acta*, **71**, 1999–2013.
- SELBY, D., KELLEY, K.D., HITZMAN, M.W., & ZIEG, J. 2009. Re-Os Sulfide (Bornite, Chalcopyrite and Pyrite) Systematics of the Carbonate-Hosted Copper Deposits at Ruby Creek, Southern Brooks Range, Alaska. *Economic Geology*, **104**(3), 437–444.
- SESTINI, G. 1973. Sedimentology of a palaeoplacer: The gold-bearing Tarkwaian of Ghana. *Pages 275–305 of: AMSTUTZ, G.C., & BERNARD, A.J. (eds), Ore in Sediments*. Berlin, Germany: Springer-Verlag.
- SHAW, D.M. 1956. Geochemistry of pelitic rocks. Part III: Major elements and general geochemistry. *Geological Society of America Bulletin*, **67**(7), 919–934.
- SHIREY, S.B., & WALKER, R.J. 1995. Carius tube digestion for low-blank rhenium-osmium analysis. *Analytical Chemistry*, **67**(13), 2136–2141.
- SIAME, L.L., ANGELIER, J., CHEN, R.-F., GODARD, V., DERRIEUX, F., BOURLÈS, D.L., BRAUCHER, R., CHANG, K.-J., CHU, H.-T., & LEE, J.-C. 2011. Erosion rates in an active orogen (NE-Taiwan): A confrontation of cosmogenic measurements with river suspended loads. *Quaternary Geochronology*, **6**, 246–260.

- SLACK, J.F., HERRIMAN, N., BARNES, R.G., & PLIMER, I.R. 1984. Stratiform tourmalinites in metamorphic terranes and their geologic significance. *Geology*, **12**, 713–716.
- SMITH, H.A., & BARREIRO, B. 1990. Monazite U-Pb dating of staurolite grade metamorphism in pelitic schists. *Contributions to Mineralogy and Petrology*, **105**, 602–615.
- SMYE, A.J., BICKLE, M.J., HOLLAND, T.J.B., PARRISH, R.R., & D.J.CONDON. 2011. Rapid formation and exhumation of the youngest Alpine eclogites: A thermal conundrum to Barrovian metamorphism. *Earth and Planetary Science Letters*, **306**, 4193–204.
- SPEAR, F.S. 1995. *Metamorphic phase equilibria and pressure-temperature-time paths*. 2nd edn. New York, USA: Mineralogical Society of America.
- SPEAR, F.S. 2010. Monazite–allanite phase relations in metapelites. *Chemical Geology*, **279**(1-2), 55–62.
- SPEAR, F.S., & PYLE, J.M. 2002. Apatite, monazite, and xenotime in metamorphic rocks. *Pages 293–335 of: KOHN, M.J., RAKOVAN, J., & HUGHES, J.M. (eds), Phosphates: Geochemical, Geobiological, and Materials Importance*, vol. 48. Reviews in Mineralogy and Geochemistry, Mineralogical Society of America, Washington D.C.
- STÖCKLI, R., VERMOTE, E., SALEOUS, N., SIMMON, R., & HERRING, D. 2005. *The Blue Marble Next Generation - A true color earth dataset including seasonal dynamics from MODIS*. NASA Earth Observatory.
- STROGEN, P. 1988. The structure and sedimentology of the Tarkwaian, and its relevance to gold mining exploration and development. *In: Internatinal conference and workshop on the geology of Ghana with special emphasis on gold. Programme and abstracts*.
- STÜWE, K. 1998. Tectonic constraints on the timing relationships of metamorphism, fluid production and gold-bearing quartz vein emplacement. *Ore Geology Reviews*, **13**, 219–228.
- STÜWE, K., WILL, T.M., & ZHOU, S. 1993. On the timing relationship between fluid production and metamorphism in metamorphic piles: some implications for the origin of post-metamorphic gold mineralisation. *Earth and Planetary Science Letters*, **114**(4), 417–430.
- SUZUKI, K., & ADACHI, M. 1991. Precambrian provenance and Silurian metamorphism of the Tsubonosawa paragneiss in the South Kitakami terrane, Northeast Japan, revealed by the chemical Th-U-total Pb isochron ages of monazite, zircon and xenotime. *Chemical Geology*, **25**, 357–376.
- SUZUKI, K., ADACHI, M., & TANAKA, T. 1991. Middle Precambrian provenance of Jurassic sandstone in the Mino Terrane, central Japan: Th-U-total Pb evidence from an electron microprobe monazite study. *Sedimentary Geology*, **75**(1-2), 141–147.

- SUZUKI, K., ADACHI, M., & KAJIZUKA, I. 1994. Electron microprobe observations of Pb diffusion in metamorphosed detrital monazites. *Earth and Planetary Science Letters*, **128**(3-4), 391–405.
- SWALF, P., CRÓSTA, A.P., & FILHO, C.R. DE SOUZA. 2003. Remote sensing signature of the Morro Do Ouro gold deposit, Minas Gerais, Brazil, using reflectance spectrometry: application to mineral exploration using spaceborne multispectral sensors. *Revista Brasileira de Geociências*, **33**(2), 221–227.
- SYLVESTER, P.J., & ATTOH, K. 1992. Lithostratigraphy and composition of 2.1 Ga greenstone belts of the West African craton and their bearing on crustal evolution and the Archean-Proterozoic boundary. *Journal of Geology*, **100**, 377–393.
- TAYLOR, B.E. 2007. Epithermal gold deposits. *Pages 113–139 of: Mineral Deposits of Canada: A synthesis of major deposit-types, district metallogeny, the evolution of geological provinces and exploration methods*. Geological Association of Canada, Mineral Deposits Division, Special Publication No. 5.
- TAYLOR, P., MOORBATH, S., LEUBE, A., & HIRDES, W. 1992. Early Proterozoic crustal evolution in the Birimian of Ghana: constraints from geochronology and isotope geochemistry. *Precambrian Research*, **56**, 97–111.
- THOMPSON, A.B., & ENGLAND, P.C. 1984. Pressure–Temperature–Time Paths of Regional Metamorphism II. Their Inference and Interpretation using Mineral Assemblages in Metamorphic Rocks. *Journal of Petrology*, **25**(4), 929–955.
- THOMPSON, A.J.B., HAUFF, P.L., & ROBITAILLE, A.J. 1999. Alteration mapping in exploration: application of short-wave infrared (SWIR) spectroscopy. *Society of Economic Geologists Newsletter*, **39**.
- THOMPSON, J.B. 1955. The thermodynamic basis for the mineral facies concept. *American Journal of Science*, **253**, 65–103.
- TOMKINS, A.G., & GRUNDY, C. 2009. Upper Temperature Limits of Orogenic Gold Deposit Formation: Constraints from the Granulite-Hosted Griffin’s Find Deposit, Yilgarn Craton. *Economic Geology*, **104**(5), 669.
- TOMKINS, H.S., & PATTISON, D.R.M. 2007. Accessory phase petrogenesis in relation to major phase assemblages in pelites from the Nelson contact aureole, southern British Columbia. *Journal of Metamorphic Geology*, **25**, 401–421.
- TOWNSEND, K.J., MILLER, C.F., D’ANDREA, J.L., AYERS, J.C., HARRISON, T.M., & COATH, C.D. 2000. Low temperature replacement of monazite in the Ireteba granite, Southern Nevada: geochronological implications. *Chemical geology*, **172**, 95–112.

- TRELOAR, P.J., RANKIN, A.H., LAWRENCE, D.M., FRENCH, C.M., BOYCE, A., SENGHOR, D., & HARBRIDGE, P. 2011. Orogenic gold mineralization in the Birimian of West Africa: Is it as simple as it seems? (Abstract). *Page 30 of: Fermor 2011: Ore deposits in an evolving Earth*.
- TUNKS, A., SELLEY, D., ROGERS, J., & BRABHAM, G. 2004. Vein mineralization at the Damang Gold Mine, Ghana: controls on mineralization. *Journal of Structural Geology*, **26**(6-7), 1257–1273.
- TURNER, F.J. 1968. *Metamorphic petrology: Mineralogical and field aspects*. New York, USA: McGraw-Hill Book Company, Inc.
- TURNER, F.J., & VERHOOGEN, J. 1960. *Igneous and metamorphic petrology*. New York, USA: McGraw-Hill Book Company, Inc.
- VACHETTE, M., & YACE, I. 1969. Sur l'âge au strontium des laves acides du Precambrien de la région de Roumoudi en Côte d'Ivoire. *Comptes Rendus de l'Académie des Sciences de Paris D*, **268**, 2235–2236.
- WATERS, D.J. 1990. Thermal history and tectonic setting of the Namaqualand granulites, Southern Africa: Clues to Proterozoic crustal development. *Pages 243–256 of: Granulites and Crustal Evolution*. Dordrecht, The Netherlands: NATO ASI Series C-311, Kluwer.
- WATERS, D.J. 2004 (October). *Practical aspects of mineral thermobarometry*. [Online]. Available at: [http://www.earth.ox.ac.uk/\\$\sim\\$davewa/pt/pt05.html](http://www.earth.ox.ac.uk/$\sim$davewa/pt/pt05.html) [accessed 7 December 2011].
- WATERS, D.J., & LOVEGROVE, D.P. 2002. Assessing the extent of disequilibrium and overstepping of prograde metamorphic reactions in metapelites from the Bushveld Complex aureole, South Africa. *Journal of Metamorphic Geology*, **20**, 135–149.
- WETHERILL, G. W. 1956. Discordant uranium-lead ages. *Transactions of the American Geophysical Union*, **37**, 320–326.
- WHEELER, J. 1996. DIFFARG: A program for simulating argon diffusion profiles in minerals. *Computers & Geosciences*, **22**(8), 919–929.
- WHITE, A.J.R., ROBB, V.M., ROBB, L.J., & WATERS, D.J. 2010. Portable infrared spectroscopy as a tool for the exploration of gold deposits in tropical terrains: A case study at the Damang deposit, Ghana. *Pages 67–84 of: GOLDFARB, R.J., MARSH, E.E., & MONECKE, T. (eds), The challenge of finding new mineral resources: Global metallogeny, innovative exploration, and new discoveries*. Society of Economic Geologists Special Publication 15.

- WHITE, R., POWELL, R., & CLARKE, G. 2002. The interpretation of reaction textures in Fe-rich metapelitic granulites of the Musgrave Block, central Australia: constraints from mineral equilibria calculations in the system K_2O – FeO – MgO – Al_2O_3 – SiO_2 – H_2O – TiO_2 – Fe_2O_3 . *Journal of Metamorphic Geology*, **20**, 41–55.
- WHITE, R., POWELL, R., & PHILLIPS, G.N. 2003. A mineral equilibria study of the hydrothermal alteration in mafic greenschist facies rocks at Kalgoorlie, Western Australia. *Journal of Metamorphic Geology*, **21**, 455–468.
- WHITE, R., POMROY, N., & POWELL, R. 2005. An in situ metatexite–diatexite transition in upper amphibolite facies rocks from Broken Hill, Australia. *Journal of Metamorphic Geology*, **23**, 579–602.
- WHITE, R.W., POWELL, R., HOLLAND, T.J.B., & WORLEY, B.A. 2000. The effect of TiO_2 and Fe_2O_3 on metapelitic assemblages at greenschist and amphibolite facies conditions: mineral equilibria calculations in the system K_2O – FeO – MgO – Al_2O_3 – SiO_2 – H_2O – TiO_2 – Fe_2O_3 . *Journal of Metamorphic Geology*, **18**(5), 497–511.
- WHITE, R.W., POWELL, R., & HOLLAND, T. 2007. Progress relating to calculation of partial melting equilibria for metapelites. *Journal of Metamorphic Geology*, **25**(5), 511.
- WHITEHOUSE, M.J., & KAMBER, B.S. 2005. Assigning Dates to Thin Gneissic Veins in High-Grade Metamorphic Terranes: A Cautionary Tale from Akilia, Southwest Greenland. *Journal of Petrology*, **46**(2), 291–318.
- WHITEHOUSE, M.J., CLAESSON, S., SUNDE, T., & VESTIN, J. 1997. Ion microprobe U–Pb zircon geochronology and correlation of Archaean gneisses from the Lewisian Complex of Gruinard Bay, northwestern Scotland. *Geochimica et Cosmochimica Acta*, **61**(20), 4429–4438.
- WHITEHOUSE, M.J., KAMBER, B.S., & MOORBATH, S. 1999. Age significance of U–Th–Pb zircon data from early Archaean rocks of west Greenland - a reassessment based on combined ion-microprobe and imaging studies. *Chemical Geology*, **160**, 201–224.
- WHITELAW, O.A.L. 1929. Geological mining and features of the Tarkwa-Abosso goldfield. *Gold Coast Geological Survey Memoir*, **1**, 46.
- WHITNEY, D. 2011 (October). “Classical” Thermobarometry. [Online]. Available at: http://www.serc.carleton.edu/research_education/equilibria/classicalthermobarometry.html [accessed 7 December 2011].
- WHITNEY, G., ABRAMS, M.J., & GOETZ, A.F.H. 1983. Mineral discrimination using a portable radio-determining radiometer. *ECONOMIC GEOLOGY*, **78**, 688–698.

- WIEDENBECK, M., ALLÉ, P., CORFU, F., GRIFFIN, W.L., MEIER, M., OBERLI, F., VON QUADT, A., RODDICK, J.C., & SPIEGEL, W. 1995. Three natural zircon standards for U-Th-Pb, Lu-Hf, trace element and REE analysis. *Geostandards Newsletter*, **19**, 1–23.
- WIEWORA, A., & WEISS, Z. 1990. Crystallographic classifications of phyllosilicates based on the unified system of projection of chemical compositions II: The chlorite group. *Clay Minerals*, **25**, 83–92.
- WILKINSON, B.H., & KESLER, S.E. 2010. Tectonic-diffusion estimate of orogenic gold resources. *Economic Geology*, **105**, 1321–1334.
- WILLE, S., & KLEMD, R. 2004. Fluid inclusion studies of the Abawso gold prospect, near the Ashanti Belt, Ghana. *Mineralium Deposita*, **39**(1), 31–45.
- WILLIAMS, M.L., & JERCINOVIC, M.J. 2002. Microprobe monazite geochronology: putting absolute time into microstructural analysis. *Journal of Structural Geology*, **24**(6-7), 1013–1028.
- WILLNER, A.P. 1992. Tourmalinites from the stratiform peraluminous metamorphic suite of the Central Namaqua Mobile Belt (South Africa). *Mineralium Deposita*, **27**, 304–313.
- WILLNER, A.P., W, SCHREYER, & MOORE, J.M. 1990. Peraluminous metamorphic rocks from the Namaqualand Metamorphic Complex (South Africa): Geochemical evidence for an exhalation-related, sedimentary origin in a Mid-Proterozoic rift system. *Chemical Geology*, **81**, 221–240.
- WILSON, M. 1989. *Igneous Petrogenesis*. London, UK: Chapman & Hall.
- WING, B.A., FERRY, J.M., & HARRISON, T.M. 2003. Prograde destruction and formation of monazite and allanite during contact and regional metamorphism of pelites: petrology and geochronology. *Contributions to Mineralogy and Petrology*, **145**, 228–250.
- WOOD, B.J., & FRASER, D.G. 1977. *Elementary thermodynamics for geologists*. Oxford, UK: Oxford University Press.
- WOODFIELD, P.D. 1966. Geology of field sheet 91. *Geological Survey of Ghana Bulletin*, **30**, 66.
- WORLEY, B., & POWELL, R. 2000. High-precision relative thermobarometry: theory and a worked example. *Journal of Metamorphic Geology*, **18**, 91–101.
- WRIGHT, N., LAYER, P.W., & YORK, D. 1991. New insights into thermal history from single grain $^{40}\text{Ar}/^{39}\text{Ar}$ analysis of biotite. *Earth and Planetary Science Letters*, **104**(1), 70–79.

- YAO, Y., & ROBB, L.J. 1999. Gold mineralization associated with the Birimian granitoids of Ghana. *Johannesburg, University of the Witwatersrand, final report to Gold Fields of South Africa Ltd*, 46.
- YAO, Y., MURPHY, P.J., & ROBB, L.J. 2001. Fluid characteristics of granitoid-hosted gold deposits in the Birimian Terrane of Ghana: a fluid inclusion microthermometric and Raman spectroscopic study. *ECONOMIC GEOLOGY*, **96**(7), 1611–1643.
- YARDLEY, B., & CLEVERLEY, J. 2011. Metamorphic fluids: Do they contribute to ore systems? (Abstract). *Page 29 of: Fermor 2011: Ore deposits in an evolving Earth*.
- YARDLEY, B.W.D. 1989. *An Introduction to Metamorphic Petrology*. London, UK: Prentice Hall, an imprint of Pearson Education.

Appendices



On the road to Damang. The Gold Coast looking towards Cape Coast

Appendix A

Sample Petrography and Location Information

Notes for tables [A.1](#) and [A.2](#):

Table [A.1](#) contains location details of samples prepared as thin sections in this study. Localities are given in two coordinate systems, the Abosso Goldfields Ltd. grid (AGL) and the Universal Transverse Mercator (UTM) grid. The UTM zone for the Damang region is 30 N, using WGS 84. The AGL grid is primarily used in and immediately around the DPCB (Damang Pit Cut-Back) and is based around a N-S line that is parallel to the long-axis of the pit. ‘RL’ refers to the bench level in the DPCB and is measured in metres above sea-level. Those cores without collar coordinates are old drill cores for which the coordinates are not in the Abosso Goldfields Ltd. database. All of these cores, however, are from the main Damang pit area.

Table [A.2](#) presents the mineralogy of samples as determined through optical and scanning electron microscopy. The following symbols are used:

- x = major phase (primary and/or metamorphic in origin).
- h = phase is unequivocally part of a post-metamorphic hydrothermal alteration assemblage, i.e. is associated with veins or similar.
- s = phase is secondary, formed during post-peak metamorphic retrogression and not associated with a major hydrothermal episode.
- (...) = minor phase of varying origins.

Table A.1: Sample locality details.

Sample	Location/ Project	Strat. ID	Core/ Pit	AGL Grid E (m) N (m)	UTM Grid E (m) N (m)	Depth (m)
AWDH1	Damang	Huni Sandstone	Pit	10236 25440	628420 609620	840RL
AWDH2	Damang	Huni Sandstone	Pit	9975 25202	628078 609529	831RL
AWDH3	Damang	Huni Sandstone	Pit	9975 25201	628078 609528	831RL
AWDH4	Damang	Huni Sandstone	Pit	10243 25440	628426 609616	840RL
AWDP1	Damang	Tarkwa Phyllite	Pit	10216 25440	628402 609629	840RL
AWDP2	Damang	Tarkwa Phyllite	Pit	10205.5 25440	628393 609633	840RL
AWDP3	Damang	Tarkwa Phyllite	Pit	10224.5 25440	628410 609625	840RL
AWDP4	Damang	Tarkwa Phyllite	Pit	10228.5 25440	628413 609623	840RL
AWDP5	Damang	Tarkwa Phyllite	DRC220D			69.9
AWDP6	Damang	Tarkwa Phyllite	DRC1394D	10314 25750	628632 609859	206.85
AWDP7	Damang	Tarkwa Phyllite	DRC1394D	10314 25750	628632 609859	234.2
AWDDo1	Damang	Dolerite	Pit	10165 25515	628391 609719	840RL
AWDDo2	Damang	Dolerite	Pit	9975 25200	628077 609527	831RL
AWDDo3	Damang	Dolerite	Pit	10220 25360	628369 609556	840RL
AWDDo4	Damang	Dolerite	Pit	10170 25455	628368 609663	840RL
AWDDo5	Damang	Dolerite	Pit	10190 25355	628340 609565	840RL
AWDDo6	Damang	Dolerite	Pit	10180 25550	628421 609743	840RL
AWDDo7	Damang	Dolerite	DRC217D			79.6
AWDDo8	Damang	Dolerite	DRC217D			83
AWDDo9	Damang	Dolerite	DRC217D			95.2
AWDDo10	Damang	Dolerite	DRC217D			95.8
AWDDo11	Damang	Dolerite	DRC217D			99.4
AWDDo12	Damang	Dolerite	DRC217D			99
AWDDo13	Damang	Dolerite	DRC220D			78
AWDDo14	Damang	Dolerite	DRC220D			77.8
AWDDo15	Damang	Dolerite	DRC220D			86.6
AWDDo16	Damang	Dolerite	DRC220D			88.6
PS02	Damang	Dolerite	DRC1913D	10032 25075	628070 609390	104.9
PS04	Damang	Dolerite	DRC1888D	9985 24773	627889 609144	131.3
PS05	Damang	Dolerite	DRC1888D	9985 24773	627889 609144	140.8
PS08	Damang	Dolerite	DRC1897D	10074 25089	628114 609383	313.2
PS09	Damang	Dolerite	DDD044	10340 24450	628055 608693	211.5
PS11	Damang	Dolerite	DDD046	9892 24291	627584 608759	179.5
AWDBm1	Damang	Tarkwa Phyllite	Pit	10110 25850	628497 610041	831RL
AWDBq1	Damang	Banket Quartzite	Pit	10085 25825	628464 610031	831RL
AWDBh1	Damang	Banket Hangingwall	DRC428D			33.5
AWDCm1	Damang	Malta Conglomerate	DRC1517D	10368 25449	628541 609567	152.6
AWDCs1	Damang	Star Conglomerate	Pit	10075 25835	628459 610044	831RL
AWDCs2	Damang	Star Conglomerate	DRC1517D	10368 25449	628541 609567	346.2
AWDCs3	Damang	Star Conglomerate	DRC1517D	10368 25449	628541 609567	346
AWABf1	Amoanda	Banket Footwall	Pit	8252 18605	623503 604474	
AWABf2	Damang	Banket Footwall	Pit	10050 25195	628142 609488	831RL
AWABf3	Amoanda	Banket Footwall	AMDD113	8280 18781	623609 604617	88.2
AWABf4	Amoanda	Banket Footwall	AMDD113	8280 18781	623609 604617	88.4
AWDBf1	Damang	Banket Footwall	DRC423D			93.75
AWDBf2	Damang	Banket Footwall	DRC428D			72.8
AWDBf3	Damang	Banket Footwall	DRC423D			96.7
AWDBf4	Damang	Banket Footwall	DRC423D			88.75
AWDBf5	Damang	Banket Footwall	DRC423D			85.75
AWDBf6	Damang	Banket Footwall	DRC428D			68.5
AWDBf7	Damang	Banket Footwall	DRC428D			69.55
AWDBf8	Damang	Banket Footwall	DRC428D			69.8
AWAK1	Amoanda	Kawere Sandstone	Pit	8211 18660	623492 604542	
AWADi1	Amoanda	Diorite Porphyry	AMDD057	8478 18793	623790 604536	58.3
AWADi2	Amoanda	Diorite Porphyry	AMDD057	8478 18793	623790 604536	48.35
AWABv1	Amoanda	Birimian Volcaniclastic	AMDD049	8517 18837	623845 604557	56
AWABbs	Amoanda	Birimian Black Shale	AMDD049	8517 18837	623845 604557	37.5
AWABs	Amoanda	Birimian Siltstone	AMDD049	8517 18837	623845 604557	51.1
AWDF1	Damang	Damang Fault	DRC1834D	10092 25711	628417 609926	255
AWDF2	Damang	Damang Fault	DRC1834D	10092 25711	628417 609926	255.2
AWDF3	Damang	Damang Fault	DRC1777D	10253 27101	629202 611085	228.1
AWT1	Tarkwa - Pepe	Banket Conglomerate	Pit			
AWT2	Tarkwa - Pepe	Banket Conglomerate	Pit			
AWTC1	Tarkwa - Akontansi Central	Banket Conglomerate	Pit			
AWTC2	Tarkwa - Akontansi Central	Banket Conglomerate	Pit			
AWTC3	Tarkwa - Akontansi Ridge	Banket Conglomerate	Pit			

Table A.2: Sample petrography.

Sample	Strat ID	Hydrothermal Alteration Style	Mineralogy																										
			Qtz	Pl	Kfs	Ms	Pg	Chl	Bt	Grt	St	Ky	Hbl	Act	Cum	Ep	Ank	Cal	Tur	Ilm	Rt	Mag	Hem	Sph	Py	Po	Ccp	Accessory	
AWDH1	Metapelite	PH	x	x		x		(h)	x	(x)							(h)		(x)										
AWDH2	Metapelite		x	x	x	x	(x)		(h)								(h)	h		(x)									
AWDH3	Metapelite		x	x	x				(h)								(h)	h						x					
AWDH4	Metapelite		x	x	x				(h)								(h)		h										
AWDP1	Metapelite	Mineralized	x	x	x	x	s	x	x	(x)	x						h	(h)	x	x		x					h		Mnz
AWDP2	Metapelite		x	x	x	x		xh		(x)	x		h	h			h			x	x		x						Ap, Ahn
AWDP3	Metapelite		x	x	x	x	s	x				x							x	x	x		x			h			Mnz
AWDP4	Metapelite		x	x	x	x	x	x	(x)			x							(x)	x	x					h			Ap, Ahn
AWDP5	Metapelite	Hornfelsed	x	x	x	x	x	x	x	(x)									x	x					h				
AWDP6	Metapelite		x	x	x	x	x												x	x									
AWDP7	Metapelite		x	x	x	x	x	x	x	x									x	x					(x)	h			
AWDDo1	Metadolomite		x	x	x		(x)	h					x		h		h				x								
AWDDo2	Metadolomite	Mineralized	x	x			(x)	h				x			h		h				x								
AWDDo3	Metadolomite		(x)	x			(x)						x			(x)		h											
AWDDo4	Metadolomite		x	x			x	h					x				(x)	h	(h)	h	x		x			h	h		
AWDDo5	Metadolomite		x	x			(x)	h					x				h	h			(x)		x			h	h		
AWDDo6	Metadolomite	Mineralized	(x)	x			h		h			x				h	h			(x)									
AWDDo7	Metadolomite		x	x				(h)					x				(h)												
AWDDo8	Metadolomite		x	x		(h)		h		h			(x)				h	h								h	h		
AWDDo9	Metadolomite		x	x		(h)				h	x						h	h			(x)		x			h	h		
AWDDo10	Metadolomite	PH	x	x			(x)	h		x						h	h			x		x			h	h			
AWDDo11	Metadolomite		x	x													(h)	h			(x)		x		h	h			
AWDDo12	Metadolomite		x	x				x									(h)	h	h		(x)		x			h	h		
AWDDo13	Metadolomite		x	x						h							h				x		(x)		x	h			
AWDDo14	Metadolomite	Mineralized	(x)	x		(x)			(h)				x				(h)												
AWDDo15	Metadolomite		x	x													(x)	h			(x)								
AWDDo16	Metadolomite		(x)	x			(x)	h		(h)			x				h	h											
PS02	Metadolomite		Altered	x	x			(h)										(h)											
PS04	Metadolomite	x		x				h					x				h	h											
PS05	Metadolomite	x		x		x		h									h	h											
PS08	Metadolomite	x		x				(h)		h							h	h											
PS09	Metadolomite	Mineralized	x	x					h							h	h									h			
PS11	Metadolomite		x	x				h		h			x				h	h								(h)			

Sample	Strat ID	Hydrothermal Alteration Style	Qtz	Pl	Kfs	Ms	Pg	Chl	Bt	Grt	St	Ky	Hbl	Act	Cum	Ep	Ank	Cal	Tur	Ilm	Rt	Mgt	Hem	Sph	Py	Po	Ccp	Accessory
AWDBm1	Metapelite		x	x	x	x	x	s	x	x	x								x	x		x						Ctd, Ap, Mnz
AWDBq1	Metaquartzite	Mineralized	x	x		x		(h)	(x)	(x)	(x)	(x)					(h)			(x)					h			
AWDBh1	Metaquartzite	Mineralized	x	x				sh	x	x							h								h			
AWDCm1	Metaquartzite		x	x	(x)	x	x	x	x	x							(h)			x		x						Zrn
AWDCs1	Metaquartzite	Mineralized	x	x	x	x	x	x	x	x									(x)	x		x			h			Zrn
AWDCs2	Metaquartzite	Mineralized	x	(x)	(x)	x	(x)	(x)	x	(x)							h								h			
AWDCs3	Metaquartzite	Mineralized	x	(x)	(x)	x	(x)	(x)	x	(x)	(x)						(h)								h			
AWABf1	Metaquartzite		x	x	x	x	x										(h)											
AWABf2	Metaquartzite	PH	x	(x)		x	(x)	(x)									(h)	h				(x)			h	(h)		Ap
AWABf3	Metaquartzite	Mineralized	x	x		xh	x	x	(x)								h		(x)	x		x			h	h		
AWABf4	Metaquartzite	Mineralized	x	x	x	xh	x	x	(x)								(h)		(x)	x		x			h	h		
AWDBf1	Metaquartzite		x	x	x	x			(x)										(x)	(x)								
AWDBf2	Metaquartzite	PH	x	x		x	(x)	(x)	(x)	(x)	(x)						(h)	h				x		(x)				
AWDBf3	Metaquartzite		x	x		x	x	x									h	h										
AWDBf4	Metaquartzite	PH	x	x	(x)	x											(h)	h				x		(x)				
AWDBf5	Metaquartzite	Mineralized	x	(x)		xh	x	(x)				x							(x)						h			
AWDBf6	Metaquartzite	Mineralized	x	x		xh	x	(x)	x	(x)	(x)														h			
AWDBf7	Metaquartzite	Mineralized	x	x		xh	x	(x)	(x)	(x)															h	(x)		
AWDBf8	Metaquartzite		x	x	(x)	x		(x)	x	x							(h)					x						
AWAK1	Sandstone	Weathered	x	x		x	x	x	x													(x)	s					
AWADh1	Diorite		x	x				h	x								h								h			
AWADi2	Diorite		x	x		x		(h)	x			(x)					h								h			
AWABv1	Volcaniclastic		x	x		x	x	x	x								x			x								
AWABbs	Black Shale		x			x	x	(h)	x																			
AWABs	Metapelite		x			x	x	(h)	x								(h)								h			
AWDF1	Metapelite	Altered	x			(x)	x	(x)	x	(x)							h		(x)	x					h			Apy, graphite
AWDF2	Metapelite	Altered	x	(x)		x	x	h	x								h		xh						x			Apy
AWDF3	Metapelite	Altered	x	x		x		h									h			(x)					h			
AWT1	Metaquartzite	Altered	x			x		h		(x)									h			x			h			
AWT2	Metaquartzite	Altered	x																h						h			
AWTC1	Metaquartzite		x	(x)	(x)	x		(x)	(x)													x			h			
AWTC2	Metaquartzite		x	(x)	(x)	x		(x)	(x)													x			h			
AWTC3	Metaquartzite		x	(x)	(x)	x		(x)	(x)													x			h			

Appendix B

X-ray Fluorescence Data

Whole-rock major and trace element X-ray fluorescence (XRF) data collected as part of this study were obtained with the assistance of John Watson at The Open University (OU) and Alastair Bewsher and Paul Lythgoe at the University of Manchester (MU). Additional XRF data from [Brabham \(1998\)](#) and [Pigois \(2003\)](#) were also used. A blank entry denotes an element that was not analysed for, while a '-' denotes concentration was below detection levels.

Analyses from the OU present titanium as an oxide, as per the other major elements, while analyses from MU present titanium as elemental, as per the trace elements. Altered versus unaltered lithologies were defined on the basis of petrology and mineralogy, not geochemistry. Some 'unaltered' dolerites do show elevated levels of certain Au-associated elements, such as Hg and Te (e.g. PS09, AWDDoU1), which suggests that these may have experienced minor geochemical alteration that is not detectable in the sample mineralogy.

Table B.1: X-ray fluorescence data for a range of rocks from Damang.

Sample Lab ¹	AWDDo2 OU	AWDDo3 OU	AWDDo7 OU	PS02 MU	PS09 MU	AWDDoU1 MU	AWDDoU2 MU
Lithology	Unaltered Dolerite	Unaltered Dolerite	Unaltered Dolerite	Unaltered Dolerite	Unaltered Dolerite	Unaltered Dolerite	Unaltered Dolerite
Wt %							
SiO ₂	52.77	49.40	53.56	53.81	49.60	46.10	54.35
TiO ₂	0.89	0.46	0.56				
Ti				0.63	0.38	0.43	0.42
Al ₂ O ₃	16.23	17.42	15.79	14.38	15.56	16.90	14.93
Fe ₂ O ₃	10.74	9.37	8.74	12.85	11.10	12.92	11.05
MnO	0.16	0.15	0.15				
Mn				0.18	0.15	0.16	0.12
MgO	5.73	8.92	4.95	3.52	6.59	6.66	5.95
CaO	6.78	10.07	8.39	9.16	10.01	11.46	7.83
Na ₂ O	3.66	2.04	4.08	2.08	2.14	2.16	1.75
K ₂ O	0.41	0.03	0.18	0.02	0.84	0.44	1.63
P ₂ O ₅	0.15	0.06	0.13	0.27	0.12	0.13	0.8
SO ₃				0.54	0.02	0.15	0.05
LOI	2.24	2.18	3.96	2.38	3.26	2.37	1.48
TOTAL	99.75	100.09	100.48	99.83	99.76	99.87	99.75
ppm							
Rb	8.0	0.6	2.5	1.3	24.2	8.5	31.3
Sr	576.8	233.7	339.2	310.5	363.8	283.7	268.3
Y	26.7	14.4	18.5	21.9	14.9	15.9	13.9
Zr	102.4	45.1	76.8	81.4	58.9	51.6	61.5
Nb	4.7	2.7	3.8	2.4	1.5	1.2	1.7
Ba	124.0	13.0	175.7	25.7	235.8	72.1	412.3
Pb	16.3	2.7	6.7	3.3	6.7	3.8	8.7
Th	3.3	1.3	0.9	2.9	2.0	1.5	2.0
U	-	-	-	2.2	2.6	2.1	2.3
Sc	35.9	28.4	31.9	28.4	19.8	20.2	18.3
V	180.7	140.0	182.0	222.6	131.3	143.7	107.0
Cr	147.7	66.0	98.4	30.7	321.3	44.3	295.5
Co	40.4	54.5	31.7	44.3	45.7	55.4	40.8
Ni	65.9	185.7	49.7	26.9	132.1	89.9	139.4
Cu	114.9	32.1	9.7	65.1	23.4	41.7	8.3
Zn	79.8	65.2	61.3	87.8	69.8	80.4	57.8
Ga	17.1	13.9	15.3	16.7	15.1	15.1	13.1
Mo	-	-	0.1	0.3	1.3	-	-
As	-	7.3	-	5.1	-	-	-
Ge	-	-	-	2.3	2.2	2.3	1.9
Hg	-	-	-	7.9	7.1	6.9	8.3
Tl	-	-	-	-	-	-	-
Te	-	-	-	-	6.1	3.9	-
Cs	-	-	-	-	-	-	9.1
La	-	-	-	16.1	-	-	6.2
Hf	-	-	-	5.7	4.2	4.1	3.3
Ce	-	-	-	32.9	25.8	17.6	16.2
Nd	-	-	-	10.3	8.3	-	6.9
Sm	-	-	-	-	-	-	-
S	54.7	6.2	18.6				

¹OU = The Open University, MU = The University of Manchester

Table B.1 cont.: X-ray fluorescence data for a range of rocks from Damang.

Sample Lab ¹	AWDDoU3 MU	AWDDoU4 MU	AWDDoU5 MU	AWDDoU6 MU	AWDDo1 OU	AWDDogb1 OU	AWDDo16 MU
Lithology	Unaltered Dolerite	Unaltered Dolerite	Unaltered Dolerite	Unaltered Dolerite	Mineralized Dolerite	Mineralized Dolerite	Mineralized Dolerite
Wt %							
SiO ₂	48.13	41.60	49.69	46.85	50.42	47.82	44.33
TiO ₂					0.62	0.62	
Ti	0.27	0.38	0.38	0.36			0.32
Al ₂ O ₃	13.82	13.93	15.30	15.77	15.81	12.81	13.66
Fe ₂ O ₃	11.88	10.46	11.07	10.95	9.57	9.38	10.32
MnO					0.15	0.17	
Mn	0.16	0.18	0.15	0.14			0.17
MgO	9.71	6.18	6.48	7.12	4.05	3.80	6.71
CaO	10.07	16.54	9.32	11.44	8.41	9.65	14.64
Na ₂ O	1.18	2.32	2.45	2.67	3.08	2.92	2.26
K ₂ O	0.43	0.16	0.75	0.28	1.52	1.07	0.42
P ₂ O ₅	0.07	0.16	0.13	0.08	0.11	0.10	0.09
SO ₃	0.03	0.43	0.01	0.16			0.12
LOI	3.96	7.47	4.06	3.99	4.91	9.42	6.82
TOTAL	99.71	99.81	99.79	99.80	98.65	97.76	99.87
ppm							
Rb	10.7	3.2	18.9	4.5	49.2	39.4	8.9
Sr	421.9	440.6	354.4	387.9	408.4	323.8	317.6
Y	12.4	13.9	14.7	14.6	17.6	20.3	11.0
Zr	37.8	47.3	49.2	52.6	82.9	83.1	34.7
Nb	1.0	0.6	1.3	1.7	3.4	3.7	0.6
Ba	103.6	75.9	284.7	124.7	669.1	625.8	52.7
Pb	7.9	5.0	4.3	4.5	9.6	8.7	3.6
Th	1.8	1.5	1.6	2.1	3.8	1.1	1.6
U	2.1	2.6	2.1	2.5	0.9	-	2.9
Sc	24.4	22.6	29.1	28.0	45.5	40.6	19.4
V	133.2	122.3	146.7	143.1	256.9	208.1	109.9
Cr	743.7	322.2	219.8	313.6	18.1	6.7	209.3
Co	57.7	32.9	40.8	36.4	19.4	24.7	38.3
Ni	294.8	53.9	67.1	75.9	29.5	20.6	65.7
Cu	32.2	75.1	2.9	19.9	94.5	139.1	41.0
Zn	65.1	48.5	60.6	64.9	70.3	53.0	55.5
Ga	12.5	12.5	14.3	14.2	17.5	13.8	11.6
Mo	-	-	-	-	-	-	-
As	-	-	-	-	1.6	0.8	-
Ge	2.2	2.0	2.1	2.5	-	-	1.8
Hg	7.0	8.5	6.8	8.7	-	-	8.7
Tl	-	-	-	-	-	-	-
Te	-	-	-	-	-	-	-
Cs	-	-	-	-	-	-	3.1
La	-	11.5	7.5	9.3	-	-	-
Hf	3.3	2.8	3.8	5.6	-	-	3.6
Ce	15.9	18.7	24.8	17.1	-	-	17.0
Nd	-	6.1	6.6	6.8	-	-	6.1
Sm	-	6.8	-	-	-	-	-
S					4850.7	2530.2	-

¹OU = The Open University, MU = The University of Manchester

Table B.1 cont.: X-ray fluorescence data for a range of rocks from Damang.

Sample Lab ¹	AWDDo8 MU	AWDDo10 OU	AWDDo12 OU	AWABv1 OU	AWADi1 OU	AWDBf1 OU	AWDBf2 OU
Lithology	Mineralized Dolerite	Pink Haem. Doelrite	Pink Haem. Dolerite	Birimian Volcan.	Diorite Porphyry	Unaltered Banket	Unaltered Banket
Wt %							
SiO ₂	51.55	64.70	64.89	62.53	56.77	81.86	70.90
TiO ₂		0.52	0.56	0.49	0.88	0.26	0.53
Ti	0.53						
Al ₂ O ₃	15.16	13.95	14.01	18.59	13.52	9.51	14.68
Fe ₂ O ₃	11.47	9.24	7.88	5.77	7.57	3.32	5.84
MnO		0.17	0.10	0.08	0.11	0.05	0.09
Mn	0.11						
MgO	3.25	0.91	0.78	2.62	3.67	0.20	0.60
CaO	7.74	2.41	3.06	2.37	4.68	0.80	1.20
Na ₂ O	1.30	5.38	5.50	2.69	4.37	1.21	1.57
K ₂ O	2.39	0.26	0.27	3.50	2.11	1.83	2.72
P ₂ O ₅	0.17	0.23	0.21	0.15	0.28	0.04	0.07
SO ₃	1.97						
LOI	4.13	3.05	2.55	1.92	5.46	1.12	1.57
TOTAL	99.75	100.80	99.80	100.70	99.41	100.18	99.77
ppm							
Rb	63.0	8.4	9.5	75.6	50.5	43.2	79.6
Sr	188.1	110.6	154.1	239.3	568.2	280.2	535.7
Y	20.3	37.4	40.9	11.7	16.4	10.1	21.9
Zr	79.2	160.9	169.7	109.1	101.9	97.6	167.0
Nb	2.3	7.4	6.7	6.0	7.4	3.8	6.8
Ba	670.0	204.5	532.3	777.8	731.5	848.3	1421.0
Pb	7.7	8.4	3.0	6.8	7.9	9.5	16.2
Th	2.9	3.3	4.4	2.5	2.6	1.0	4.6
U	2.4	1.9	1.0	1.9	0.8	2.5	0.9
Sc	34.2	18.6	21.4	15.0	18.3	5.1	12.8
V	213.3	16.9	23.6	102.7	116.3	43.9	71.6
Cr	58.0	3.1	3.8	18.3	138.3	30.7	88.8
Co	28.2	12.9	20.0	12.5	13.3	16.8	21.7
Ni	35.4	3.5	3.4	13.5	27.9	9.8	34.1
Cu	69.0	31.4	16.5	6.1	104.7	2.8	5.1
Zn	88.4	314.8	55.6	74.5	66.7	20.5	41.8
Ga	17.4	17.1	16.2	20.6	13.5	10.8	17.2
Mo	-	4.2	1.2	-	2.5	0.8	0.2
As	-	2.6	0.4	3.5	3.6	2.0	1.6
Ge	1.8	-	-	-	-	-	-
Hg	6.8	-	-	-	-	-	-
Tl	1.1	-	-	-	-	-	-
Te	-	-	-	-	-	-	-
Cs	5.6	-	-	-	-	-	-
La	7.0	-	-	-	-	-	-
Hf	4.3	-	-	-	-	-	-
Ce	19.0	-	-	-	-	-	-
Nd	5.0	-	-	-	-	-	-
Sm	-	-	-	-	-	-	-
S	-	6305.3	151.3	9.6	1108.8	41.5	65.0

¹OU = The Open University, MU = The University of Manchester

Table B.1 cont.: X-ray fluorescence data for a range of rocks from Damang.

Sample Lab ¹	AWDBf6 OU	AWDBfpq1 OU	AWDBf4 OU	AWDHu1 OU	AWDH3 OU
Lithology	Mineralized Banket	Mineralized Banket	Pink Haem. Banket	Unalterd Huni	Pink Haem. Huni
Wt %					
SiO ₂	82.30	82.18	83.71	68.49	74.57
TiO ₂	0.19	0.18	0.18	0.63	0.32
Ti					
Al ₂ O ₃	8.22	10.55	9.05	14.79	9.76
Fe ₂ O ₃	2.42	1.94	2.06	4.48	2.47
MnO	0.15	0.03	0.01	0.11	0.12
Mn					
MgO	0.75	0.16	0.07	1.36	1.39
CaO	1.61	0.43	0.40	2.84	3.35
Na ₂ O	0.96	0.74	2.08	2.90	3.20
K ₂ O	1.16	2.37	1.47	2.38	0.55
P ₂ O ₅	0.04	0.03	0.02	0.09	0.03
SO ₃					
LOI	2.25	1.91	1.08	2.04	4.52
TOTAL	100.05	100.51	100.12	100.11	100.27
ppm					
Rb	40.3	54.9	34.4	72.5	19.1
Sr	120.9	311.5	128.5	301.3	136.8
Y	9.0	11.7	9.7	17.7	15.4
Zr	64.4	59.9	55.4	198.3	64.5
Nb	3.3	2.9	2.7	9.0	4.7
Ba	774.5	1018.8	959.3	1054.6	199.5
Pb	5.6	11.9	6.0	16.1	11.2
Th	1.9	1.6	1.0	5.7	1.1
U	0.8	0.3	1.2	1.6	0.8
Sc	5.5	5.2	3.4	13.7	6.8
V	30.8	45.3	31.9	105.2	41.6
Cr	28.8	12.1	16.7	80.7	41.0
Co	23.4	15.0	15.2	17.1	19.0
Ni	23.3	10.1	8.7	33.2	19.0
Cu	4.7	13.3	3.4	4.0	39.2
Zn	33.5	11.6	10.6	47.7	28.2
Ga	9.1	12.9	10.7	16.8	10.4
Mo	1.1	0.8	1.1	0.4	1.0
As	3.6	3.1	1.4	-	4.1
Ge	-	-	-	-	-
Hg	-	-	-	-	-
Tl	-	-	-	-	-
Te	-	-	-	-	-
Cs	-	-	-	-	-
La	-	-	-	-	-
Hf	-	-	-	-	-
Ce	-	-	-	-	-
Nd	-	-	-	-	-
Sm	-	-	-	-	-
S	866.2	1501.8	153.2	8.2	21.0

¹OU = The Open University, MU = The University of Manchester

Appendix C

Averaged Mineral Compositions

Mineral chemical data were collected according to the details given in section 3.7, and are presented here as averaged analyses per mineral, per rock, and given as weight percent oxide and recalculated cations. All Fe is measured as FeO with the amount of ferric Fe estimated, where appropriate, using the equation of [Droop \(1987\)](#):

$$F = 2X(1 - T/S) \quad (\text{C.1})$$

where F is the number of Fe^{3+} ions per X oxygens, T is the ideal number of cations per formula unit and S is the cation total obtained when all Fe is assumed to be Fe^{2+} . This technique is therefore only suitable for minerals in which Fe is the only element present with a variable oxidation state, where oxygen is the only anion and where there are no vacancies in the crystal lattice. All Fe in epidote is inferred to be Fe^{3+} . Cation recalculation, including Fe^{3+} determination, for amphiboles follows the procedure of [Holland & Blundy \(1994\)](#).

Garnets in the Tarkwa Phyllite and Banket Series are generally small and consequently show negligible zoning, with only a decrease in spessartine content of less than 5% from core to rim, balanced by an increasing almandine proportion. As such, a single garnet analysis is given per relevant sample, with no distinction between core and rim compositions.

Table C.1: Averaged mineral analyses for sample AWDH3

AWDH3	Ank	Cal	Chl	Hem	Ilm	Mag	Ms	Pl
Wt %								
SiO ₂	0.18	0.36	25.97	0.13	0.29	0.09	46.11	66.88
TiO ₂	0.04	0.03	0.23	0.13	67.41	0.12	0.69	0.02
Al ₂ O ₃	0.15	0.23	23.11	0.31	0.47	0.30	34.73	21.91
FeO	11.09	0.78	29.64	89.91	26.73	93.91	2.56	0.22
MnO	1.31	0.54	0.04	0.02	0.09	0.01	0.01	0.00
MgO	14.02	0.17	8.73	0.02	0.18	0.09	0.71	0.00
CaO	29.21	54.27	0.14	0.08	0.08	0.05	0.04	2.30
Na ₂ O	0.00	0.00	0.06	0.00	0.00	0.00	0.48	9.19
K ₂ O	0.02	0.03	0.11	0.00	0.04	0.03	9.66	0.04
TOTAL	56.04	56.41	88.03	90.61	95.30	94.60	94.98	100.56
Recalculated Ions								
Oxygens	2	1	28	3	3	4	22	8
Si	0.01	0.01	5.54	0.00	0.01	0.00	6.17	2.91
Ti	0.00	0.00	0.04	0.00	1.21	0.00	0.07	0.00
Al	0.01	0.00	5.81	0.01	0.01	0.01	5.47	1.12
Fe ³⁺	0.00	0.00	0.00	1.97	0.00	1.97	0.00	0.00
Fe ²⁺	0.29	0.01	5.29	0.00	0.53	0.98	0.29	0.01
Mn	0.04	0.01	0.01	0.00	0.00	0.00	0.00	0.00
Mg	0.66	0.00	2.78	0.00	0.01	0.00	0.14	0.00
Ca	0.99	0.96	0.03	0.00	0.00	0.00	0.01	0.11
Na	0.00	0.00	0.03	0.00	0.00	0.00	0.12	0.77
K	0.00	0.00	0.03	0.00	0.00	0.00	1.65	0.00
TOTAL	1.99	0.99	19.55	2.00	1.78	2.99	13.91	4.92

Table C.1 cont.: Averaged mineral analyses for sample AWDP1

AWDP1	Bt	Chl	Grt	St	Pg	Ms	Ab	Mag	Ilm	Tur
Wt %										
SiO ₂	35.33	24.95	37.07	27.48	47.29	45.67	67.25	0.24	2.17	37.81
TiO ₂	1.79	0.08	0.04	0.48	0.17	0.33	0.01	0.31	54.08	0.26
Al ₂ O ₃	19.07	23.72	21.00	54.13	39.46	37.26	20.99	0.39	1.29	32.87
FeO	18.06	23.05	32.87	14.24	0.86	1.90	0.02	95.05	45.00	8.48
MnO	0.10	0.15	7.52	0.37	0.01	0.02	0.01	0.09	0.25	0.03
MgO	11.39	16.88	2.35	1.54	0.00	0.76	0.00	0.03	1.33	5.12
CaO	0.09	0.03	1.01	0.01	0.19	0.05	1.46	0.02	0.04	0.14
Na ₂ O	0.29	0.14	0.00	0.00	4.55	1.71	10.08	0.00	0.01	1.42
K ₂ O	10.21	0.01	0.00	0.01	1.66	8.67	0.00	0.01	0.18	0.04
TOTAL	96.32	89.01	101.86	98.26	94.19	96.36	99.82	96.12	104.34	86.16
Recalculated Ions										
Oxygens	22	28	12	46	22	22	8	4	3	49
Si	5.32	5.11	2.97	7.62	6.08	5.99	2.94	0.01	0.05	12.31
Ti	0.20	0.01	0.00	0.10	0.02	0.03	0.00	0.01	0.95	0.06
Al	3.39	5.73	1.98	17.69	5.98	5.76	1.08	0.02	0.04	12.62
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.95	0.00	0.00
Fe ²⁺	2.27	3.95	2.20	3.30	0.09	0.21	0.00	0.99	0.88	2.31
Mn	0.01	0.03	0.51	0.09	0.00	0.00	0.00	0.00	0.00	0.01
Mg	2.56	5.15	0.28	0.63	0.00	0.15	0.00	0.00	0.05	2.49
Ca	0.01	0.01	0.09	0.00	0.03	0.01	0.07	0.00	0.00	0.05
Na	0.08	0.05	0.00	0.00	1.13	0.43	0.85	0.00	0.00	0.89
K	1.96	0.00	0.00	0.00	0.27	1.45	0.00	0.00	0.01	0.02
TOTAL	15.81	20.04	8.04	29.44	13.61	14.04	4.95	2.97	1.98	30.77

Table C.1 cont.: Averaged mineral analyses for sample AWDP2

AWDP2	Act	Ank	Bt	Cal	Hbl	Ilm	Ms	Pl
Wt %								
SiO ₂	52.83	0.10	36.33	0.08	43.92	0.19	45.07	55.21
TiO ₂	0.17	0.05	1.85	0.01	0.54	52.17	0.44	0.03
Al ₂ O ₃	4.30	0.09	17.12	0.06	13.08	0.25	35.21	28.51
FeO	10.93	9.88	15.45	1.82	14.83	44.53	2.10	0.14
MnO	0.22	0.67	0.15	0.54	0.19	3.22	0.02	0.02
MgO	16.63	14.06	13.49	0.79	11.12	0.30	0.58	0.00
CaO	12.45	29.31	0.06	46.13	12.07	0.17	0.03	10.44
Na ₂ O	0.14	0.00	0.00	0.00	0.84	0.00	0.55	5.29
K ₂ O	0.07	0.02	9.00	0.01	0.37	0.03	10.23	0.02
TOTAL	97.73	54.18	93.46	49.44	96.97	100.85	94.24	99.66
Recalculated Ions								
Oxygens	23	2	22	3	23	3	22	8
Si	7.54	0.00	5.52	0.00	6.52	0.00	6.09	2.49
Ti	0.02	0.00	0.21	0.00	0.06	0.98	0.05	0.00
Al	0.72	0.00	3.07	0.00	2.29	0.01	5.61	1.52
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	1.30	0.27	1.96	0.03	1.84	0.93	0.24	0.01
Mn	0.03	0.02	0.02	0.01	0.02	0.07	0.00	0.00
Mg	3.54	0.68	3.06	0.02	2.46	0.01	0.12	0.00
Ca	1.90	1.02	0.01	0.93	1.92	0.00	0.00	0.50
Na	0.04	0.00	0.00	0.00	0.24	0.00	0.14	0.46
K	0.01	0.00	1.75	0.00	0.07	0.00	1.76	0.00
TOTAL	15.11	1.99	15.60	1.00	15.43	2.01	14.01	4.98

Table C.1 cont.: Averaged mineral analyses for sample AWDP4

AWDP4	Bt	Chl	St	Tur	Ilm	Ms	Pl
Wt %							
SiO ₂	35.76	25.37	26.40	36.27	0.08	45.35	59.98
TiO ₂	1.73	0.10	0.50	0.39	52.61	0.49	0.01
Al ₂ O ₃	18.80	20.77	53.63	32.14	0.22	35.85	25.66
FeO	16.91	27.86	14.10	8.39	46.96	1.63	0.19
MnO	0.11	0.20	0.48	0.02	1.68	0.02	0.01
MgO	11.90	11.87	1.72	5.72	0.17	0.33	0.00
CaO	0.03	0.10	0.03	0.45	0.04	0.06	6.86
Na ₂ O	0.00	0.00	0.00	1.46	0.00	0.97	7.11
K ₂ O	8.97	0.08	0.00	0.03	0.06	8.26	0.02
TOTAL	94.21	86.35	96.85	84.88	101.81	92.96	99.85
Recalculated Ions							
Oxygens	22	28	46	49	3	22	8
Si	5.42	5.51	7.44	12.05	0.00	6.12	2.67
Ti	0.20	0.02	0.11	0.10	0.98	0.05	0.00
Al	3.36	5.32	17.82	12.59	0.01	5.70	1.35
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	2.14	5.06	3.32	2.33	0.98	0.18	0.01
Mn	0.01	0.04	0.12	0.01	0.04	0.00	0.00
Mg	2.69	3.84	0.72	2.83	0.01	0.07	0.00
Ca	0.01	0.02	0.01	0.16	0.00	0.01	0.33
Na	0.00	0.00	0.00	0.94	0.00	0.25	0.61
K	1.74	0.02	0.00	0.01	0.00	1.42	0.00
TOTAL	15.57	19.83	29.54	31.03	2.01	13.81	4.97

Table C.1 cont.: Averaged mineral analyses for sample AWDP5

AWDP5	Bt	Chl	St	Ilm	Mag	Ms	Pl	Tur
Wt %								
SiO ₂	36.16	25.31	27.16	0.17	0.19	46.17	64.85	36.74
TiO ₂	1.74	0.09	0.37	51.21	0.09	0.44	0.02	0.40
Al ₂ O ₃	19.18	22.96	53.44	0.21	0.38	36.20	22.96	32.48
FeO	18.92	23.10	15.45	51.53	93.40	2.09	0.11	8.60
MnO	0.10	0.21	0.38	1.21	0.05	0.01	0.01	0.00
MgO	10.54	16.11	1.70	0.01	0.01	0.23	0.00	5.82
CaO	0.02	0.02	0.01	0.02	0.02	0.02	3.72	0.48
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.95	8.93	1.47
K ₂ O	9.62	0.02	0.00	0.05	0.01	8.56	0.01	0.02
TOTAL	96.27	87.82	98.52	104.41	94.16	94.67	100.61	86.01
Recalculated Ions								
Oxygens	22	28	46	3	4	22	8	49
Si	5.43	5.25	7.56	0.00	0.01	6.14	2.83	12.05
Ti	0.20	0.01	0.08	0.95	0.00	0.04	0.00	0.10
Al	3.39	5.62	17.55	0.01	0.02	5.67	1.18	12.56
Fe ³⁺	0.00	0.00	0.00	0.00	1.96	0.00	0.00	0.00
Fe ²⁺	2.37	4.01	3.60	1.06	0.99	0.23	0.00	2.36
Mn	0.01	0.04	0.09	0.03	0.00	0.00	0.00	0.00
Mg	2.36	4.98	0.71	0.00	0.00	0.05	0.00	2.85
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.17	0.17
Na	0.00	0.00	0.00	0.00	0.00	0.25	0.76	0.93
K	1.84	0.00	0.00	0.00	0.00	1.45	0.00	0.01
TOTAL	15.60	19.93	29.59	2.05	2.98	13.83	4.95	31.04

Table C.1 cont.: Averaged mineral analyses for sample AWDP6

AWDP6	Cal	Chl	Grt	Ilm	Ms	Pl	Tur
Wt %							
SiO ₂	0.32	25.42	37.17	0.11	46.39	66.64	36.98
TiO ₂	1.20	0.09	0.07	54.00	0.40	0.02	0.48
Al ₂ O ₃	0.40	22.18	21.28	0.19	36.48	22.27	32.38
FeO	0.74	31.68	28.39	48.22	1.72	0.10	7.83
MnO	1.40	0.17	11.06	1.19	0.01	0.01	0.05
MgO	0.14	8.89	2.77	0.03	0.31	0.00	6.52
CaO	54.01	0.05	1.23	0.04	0.02	2.63	0.50
Na ₂ O	0.00	0.00	0.00	0.00	1.28	9.59	1.61
K ₂ O	0.01	0.02	0.01	0.03	8.07	0.01	0.02
TOTAL	58.21	88.51	101.98	103.82	94.69	101.28	86.36
Recalculated Ions							
Oxygens	1	28	12	3	22	8	49
Si	0.01	5.47	2.96	0.00	6.14	2.88	12.05
Ti	0.01	0.01	0.00	0.99	0.04	0.00	0.12
Al	0.01	5.63	2.00	0.01	5.69	1.14	12.44
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.01	5.70	1.89	0.98	0.19	0.00	2.13
Mn	0.02	0.03	0.75	0.02	0.00	0.00	0.01
Mg	0.00	2.85	0.33	0.00	0.06	0.00	3.17
Ca	0.92	0.01	0.11	0.00	0.00	0.12	0.18
Na	0.00	0.00	0.00	0.00	0.33	0.80	1.02
K	0.00	0.01	0.00	0.00	1.36	0.00	0.01
TOTAL	0.98	19.71	8.04	2.01	13.82	4.95	31.12

Table C.1 cont.: Averaged mineral analyses for sample AWDDo1

AWDDo1	Ank	Bt	Chl	Hbl	Ilm	Pl
Wt %						
SiO ₂	0.36	36.27	26.07	43.17	0.35	59.85
TiO ₂	0.03	2.41	0.15	0.36	54.76	0.01
Al ₂ O ₃	0.17	18.03	22.80	15.18	0.33	26.45
FeO	12.47	18.21	22.96	16.77	49.39	0.07
MnO	0.83	0.06	0.10	0.37	0.79	0.01
MgO	13.64	10.69	17.59	9.76	0.23	0.29
CaO	29.78	0.06	0.05	11.32	0.05	7.17
Na ₂ O	0.19	0.00	0.26	1.27	0.00	7.15
K ₂ O	0.04	9.65	0.11	0.30	0.10	0.03
TOTAL	57.51	95.38	90.10	98.51	105.99	101.03
Recalculated Ions						
Oxygens	2	22	28	23	3	8
Si	0.01	5.48	5.27	6.28	0.01	2.64
Ti	0.00	0.27	0.02	0.04	0.98	0.00
Al	0.01	3.21	5.43	2.60	0.01	1.37
Fe ³⁺	0.00	0.00	0.00	0.60	0.00	0.00
Fe ²⁺	0.32	2.30	3.88	1.44	0.98	0.00
Mn	0.02	0.01	0.02	0.05	0.02	0.00
Mg	0.63	2.41	5.30	2.11	0.01	0.02
Ca	0.99	0.01	0.01	1.76	0.00	0.34
Na	0.01	0.00	0.10	0.36	0.00	0.61
K	0.00	1.86	0.03	0.06	0.00	0.00
TOTAL	1.99	15.56	20.06	15.29	2.01	4.98

Table C.1 cont.: Averaged mineral analyses for sample AWDDo2

AWDDo2	Ank	Bt	Chl	Cum	Hbl	Ilm	Mag	Pl
Wt %								
SiO ₂	0.37	38.10	27.45	54.68	43.71	0.57	0.31	59.48
TiO ₂	0.03	1.65	0.08	0.05	0.43	50.98	0.12	0.01
Al ₂ O ₃	0.14	18.03	21.70	1.79	14.50	0.33	0.48	26.24
FeO	9.13	14.17	16.58	20.40	15.73	49.70	93.12	0.18
MnO	1.03	0.04	0.09	0.88	0.31	1.29	0.01	0.01
MgO	15.64	15.85	21.73	18.70	10.25	0.32	0.14	0.00
CaO	29.33	0.16	0.14	1.50	11.36	0.49	0.08	7.79
Na ₂ O	0.19	0.27	0.14	0.00	0.86	0.00	0.00	6.71
K ₂ O	0.02	9.30	0.03	0.02	0.24	0.02	0.01	0.02
TOTAL	55.90	97.59	87.94	98.02	97.39	103.70	94.27	100.44
Recalculated Ions								
Oxygens	2	22	28	23	23	3	4	8
Si	0.01	5.50	5.49	7.87	6.38	0.01	0.01	2.64
Ti	0.00	0.18	0.01	0.01	0.05	0.94	0.00	0.00
Al	0.01	3.07	5.12	0.30	2.49	0.01	0.02	1.37
Fe ³⁺	0.00	0.00	0.00	0.00	0.60	0.00	1.95	0.00
Fe ²⁺	0.24	1.71	2.77	2.45	1.32	1.02	0.98	0.01
Mn	0.03	0.00	0.01	0.11	0.04	0.03	0.00	0.00
Mg	0.72	3.41	6.47	4.01	2.23	0.01	0.01	0.00
Ca	0.97	0.03	0.03	0.23	1.78	0.01	0.00	0.37
Na	0.01	0.08	0.05	0.00	0.24	0.00	0.00	0.58
K	0.00	1.71	0.01	0.00	0.04	0.00	0.00	0.00
TOTAL	1.99	15.68	19.97	14.98	15.17	2.04	2.97	4.96

Table C.1 cont.: Averaged mineral analyses for sample AWDDo3

AWDDo3	Chl	Ep	Hbl	Pl
Wt %				
SiO ₂	27.54	38.83	45.64	58.16
TiO ₂	0.08	0.08	0.32	0.02
Al ₂ O ₃	22.33	26.30	13.46	27.02
FeO	13.76	9.94	12.45	0.14
MnO	0.11	0.13	0.22	0.00
MgO	23.86	0.02	12.19	0.00
CaO	0.13	24.68	12.19	8.82
Na ₂ O	0.15	0.00	0.61	6.14
K ₂ O	0.01	0.01	0.02	0.00
TOTAL	87.98	99.98	97.11	100.30
Recalculated Ions				
Oxygens	28	25	23	8
Si	5.43	6.03	6.58	2.59
Ti	0.01	0.01	0.04	0.00
Al	5.19	4.82	2.29	1.42
Fe ³⁺	0.00	1.04	0.43	0.00
Fe ²⁺	2.27	0.00	1.07	0.01
Mn	0.02	0.02	0.03	0.00
Mg	7.00	0.00	2.62	0.00
Ca	0.03	4.11	1.88	0.42
Na	0.06	0.00	0.17	0.53
K	0.00	0.00	0.00	0.00
TOTAL	20.00	16.03	15.11	4.97

Table C.1 cont.: Averaged mineral analyses for sample AWDDo4

AWDDo4	Ank	Bt	Chl	Hbl	Ilm	Mag	Pl	Tur
Wt %								
SiO ₂	0.07	36.24	27.31	40.98	0.11	0.07	55.84	36.67
TiO ₂	0.02	1.44	0.10	0.50	52.88	0.24	0.01	0.45
Al ₂ O ₃	0.13	19.52	20.63	18.34	0.11	0.29	29.48	32.80
FeO	12.23	16.85	25.10	15.45	48.78	92.57	0.19	7.11
MnO	0.62	0.05	0.10	0.37	0.99	0.03	0.01	0.01
MgO	13.42	11.91	14.09	9.26	0.18	0.05	0.00	6.62
CaO	29.24	0.07	0.17	11.64	0.00	0.06	11.14	0.53
Na ₂ O	0.00	0.00	0.00	1.29	0.00	0.00	5.13	1.59
K ₂ O	0.02	9.57	0.04	0.32	0.02	0.01	0.03	0.01
TOTAL	55.75	95.66	87.54	98.15	103.06	93.33	101.83	85.80
Recalculated Ions								
Oxygens	2	22	28	23	3	4	8	49
Si	0.00	5.41	5.73	5.97	0.00	0.00	2.47	11.98
Ti	0.00	0.16	0.02	0.05	0.98	0.01	0.00	0.11
Al	0.00	3.44	5.10	3.15	0.00	0.01	1.54	12.64
Fe ³⁺	0.00	0.00	0.00	0.51	0.00	2.00	0.00	0.00
Fe ²⁺	0.33	2.11	4.40	1.37	1.00	0.95	0.01	1.94
Mn	0.02	0.01	0.02	0.05	0.02	0.00	0.00	0.00
Mg	0.64	2.65	4.40	2.01	0.01	0.00	0.00	3.22
Ca	1.00	0.01	0.04	1.82	0.00	0.00	0.53	0.19
Na	0.00	0.00	0.00	0.36	0.00	0.00	0.44	1.01
K	0.00	1.82	0.01	0.06	0.00	0.00	0.00	0.00
TOTAL	2.00	15.62	19.71	15.35	2.02	2.98	4.98	31.09

Table C.1 cont.: Averaged mineral analyses for sample AWDDo12

AWDDo12	Ank	Cal	Chl	Ilm	Mag	Ms	Pl
Wt %							
SiO ₂	0.11	0.36	24.32	0.87	0.20	45.22	67.20
TiO ₂	0.01	0.01	0.15	74.92	0.37	1.25	0.02
Al ₂ O ₃	0.16	0.25	22.84	0.73	0.31	34.58	21.08
FeO	15.15	0.97	35.42	20.03	92.49	2.64	0.12
MnO	1.78	2.85	0.27	0.11	0.02	0.01	0.01
MgO	10.82	0.11	5.97	0.10	0.01	0.66	0.00
CaO	28.63	53.05	0.09	0.33	0.02	0.04	1.67
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.42	9.37
K ₂ O	0.01	0.03	0.06	0.04	0.01	9.34	0.00
TOTAL	56.67	57.63	89.12	97.11	93.44	94.17	99.49
Recalculated Ions							
Oxygens	2	1	28	3	4	22	8
Si	0.00	0.01	5.31	0.02	0.01	6.10	2.94
Ti	0.00	0.00	0.02	1.27	0.01	0.13	0.00
Al	0.01	0.00	5.88	0.02	0.01	5.50	1.09
Fe ³⁺	0.00	0.00	0.00	0.00	1.95	0.00	0.00
Fe ²⁺	0.41	0.01	6.47	0.38	0.99	0.30	0.00
Mn	0.05	0.04	0.05	0.00	0.00	0.00	0.00
Mg	0.52	0.00	1.95	0.00	0.00	0.13	0.00
Ca	1.00	0.93	0.02	0.01	0.00	0.01	0.08
Na	0.00	0.00	0.00	0.00	0.00	0.11	0.80
K	0.00	0.00	0.02	0.00	0.00	1.61	0.00
TOTAL	1.99	0.99	19.73	1.70	2.97	13.88	4.91

Table C.1 cont.: Averaged mineral analyses for sample AWDBm1

AWDBm1	Bt	Chl	Grt	St	Pg	Ms	Ab	Mag	Ilm	Tur
Wt %										
SiO ₂	35.82	24.67	35.63	27.46	47.06	45.86	68.58	0.28	0.76	36.07
TiO ₂	1.62	0.08	1.86	0.49	0.12	0.32	0.01	0.06	51.02	0.31
Al ₂ O ₃	18.86	23.61	20.32	54.14	40.23	37.16	20.68	0.46	0.64	32.15
FeO	19.57	25.64	35.93	15.49	0.92	1.91	0.05	95.70	49.34	8.61
MnO	0.05	0.08	4.84	0.17	0.02	0.01	0.01	0.00	0.80	0.01
MgO	10.11	14.87	2.08	1.39	0.45	0.76	0.00	0.08	0.21	4.97
CaO	0.02	0.04	1.38	0.01	0.23	0.05	1.04	0.03	0.03	0.20
Na ₂ O	0.00	0.08	0.00	0.00	5.29	1.72	10.32	0.03	0.03	1.31
K ₂ O	9.59	0.06	0.00	0.00	1.95	8.87	0.00	0.01	0.07	0.02
TOTAL	95.64	89.14	102.05	99.14	96.27	96.67	100.68	96.66	102.89	83.65
Recalculated Ions										
Oxygens	22	28	12	46	22	22	8	4	3	49
Si	5.43	5.12	2.87	7.58	5.97	6.01	2.97	0.01	0.02	12.14
Ti	0.18	0.01	0.11	0.10	0.01	0.03	0.00	0.00	0.95	0.08
Al	3.37	5.77	1.93	17.63	6.01	5.73	1.05	0.02	0.02	12.76
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.96	0.00	0.00
Fe ²⁺	2.48	4.45	2.42	3.58	0.10	0.21	0.00	0.98	1.02	2.42
Mn	0.01	0.01	0.33	0.04	0.00	0.00	0.00	0.00	0.02	0.00
Mg	2.29	4.59	0.25	0.57	0.09	0.15	0.00	0.00	0.01	2.49
Ca	0.00	0.01	0.12	0.00	0.03	0.01	0.05	0.00	0.00	0.07
Na	0.00	0.03	0.00	0.00	1.30	0.43	0.87	0.00	0.00	0.86
K	1.86	0.02	0.00	0.00	0.31	1.47	0.00	0.00	0.00	0.01
TOTAL	15.62	20.01	8.05	29.50	13.82	14.04	4.94	2.98	2.03	30.83

Table C.1 cont.: Averaged mineral analyses for sample AWDCs1

AWDCs1	Bt	Chl	Grt	Ms	Pl
Wt %					
SiO ₂	34.74	25.47	37.05	46.40	58.68
TiO ₂	1.68	0.05	0.06	0.63	0.01
Al ₂ O ₃	19.55	20.70	20.97	35.30	26.27
FeO	21.91	29.76	27.21	2.76	0.12
MnO	0.11	0.16	10.02	0.02	0.02
MgO	10.38	12.58	2.38	1.09	0.00
CaO	0.25	0.14	3.66	0.09	8.03
Na ₂ O	0.19	0.15	0.00	0.58	6.46
K ₂ O	8.53	0.15	0.01	10.57	0.02
TOTAL	97.34	89.17	101.36	97.44	99.61
Recalculated Ions					
Oxygens	22	28	12	22	8
Si	5.22	5.41	2.96	6.09	2.63
Ti	0.19	0.01	0.00	0.06	0.00
Al	3.46	5.18	1.98	5.47	1.39
Fe ³⁺	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	2.75	5.29	1.82	0.30	0.00
Mn	0.01	0.03	0.68	0.00	0.00
Mg	2.33	3.98	0.28	0.21	0.00
Ca	0.04	0.03	0.31	0.01	0.38
Na	0.05	0.06	0.00	0.15	0.56
K	1.64	0.04	0.00	1.77	0.00
TOTAL	15.70	20.04	8.04	14.07	4.96

Table C.1 cont.: Averaged mineral analyses for sample AWABf1

AWABf1	Ab	Ank	Ilm	Mag	Ms	Pl
Wt %						
SiO ₂	68.79	0.18	0.42	0.24	45.19	64.60
TiO ₂	0.01	0.01	53.88	0.05	0.39	0.02
Al ₂ O ₃	20.94	0.05	0.20	0.20	35.83	23.44
FeO	0.05	11.83	46.66	94.29	2.71	0.11
MnO	0.02	1.88	2.56	0.01	0.09	0.02
MgO	0.27	13.11	0.11	0.06	0.78	0.24
CaO	0.46	28.56	0.03	0.02	0.09	3.52
Na ₂ O	10.33	0.25	0.00	0.00	0.85	9.08
K ₂ O	0.02	0.02	0.04	0.02	10.57	0.04
TOTAL	100.88	55.90	103.91	94.88	96.51	101.04
Recalculated Ions						
Oxygens	8	2	3	4	22	8
Si	2.97	0.01	0.01	0.01	6.01	2.81
Ti	0.00	0.00	0.98	0.00	0.04	0.00
Al	1.06	0.00	0.01	0.01	5.62	1.20
Fe ³⁺	0.00	0.00	0.00	1.97	0.00	0.00
Fe ²⁺	0.00	0.32	0.95	0.99	0.30	0.00
Mn	0.00	0.05	0.05	0.00	0.01	0.00
Mg	0.02	0.63	0.00	0.00	0.15	0.02
Ca	0.02	0.98	0.00	0.00	0.01	0.16
Na	0.86	0.02	0.00	0.00	0.22	0.77
K	0.00	0.00	0.00	0.00	1.79	0.00
TOTAL	4.93	2.00	2.00	2.99	14.15	4.97

Table C.1 cont.: Averaged mineral analyses for sample AWABf3

AWABf3	Ab	Ank	Chl	Ilm	Ms	Pl
Wt %						
SiO ₂	68.70	0.21	24.60	0.25	45.19	62.79
TiO ₂	0.01	0.02	0.04	54.73	0.48	0.02
Al ₂ O ₃	20.02	0.10	21.97	0.23	35.20	23.82
FeO	0.05	11.89	36.60	48.95	2.98	0.05
MnO	0.00	2.41	0.14	0.76	0.33	0.03
MgO	0.00	12.79	5.29	0.16	0.73	0.00
CaO	0.33	29.16	0.13	0.07	0.08	4.83
Na ₂ O	10.76	0.20	0.15	0.00	0.55	8.24
K ₂ O	0.00	0.03	0.69	0.03	10.60	0.02
TOTAL	99.87	56.80	89.61	105.19	96.13	99.80
Recalculated Ions						
Oxygens	8	2	28	3	22	8
Si	2.99	0.01	5.40	0.01	6.04	2.78
Ti	0.00	0.00	0.01	0.99	0.05	0.00
Al	1.03	0.00	5.69	0.01	5.55	1.24
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.00	0.31	6.73	0.98	0.33	0.00
Mn	0.00	0.06	0.03	0.02	0.04	0.00
Mg	0.00	0.60	1.73	0.01	0.14	0.00
Ca	0.02	0.99	0.03	0.00	0.01	0.23
Na	0.91	0.01	0.06	0.00	0.14	0.71
K	0.00	0.00	0.19	0.00	1.81	0.00
TOTAL	4.95	2.00	19.87	2.00	14.11	4.96

Table C.1 cont.: Averaged mineral analyses for sample AWABf4

AWABf4	Ab	Ank	Chl	Ilm	Mag	Ms	Pl	Tur
Wt %								
SiO ₂	69.25	0.25	28.83	0.26	0.23	45.58	62.25	37.36
TiO ₂	0.03	0.02	0.39	54.88	0.09	0.51	0.01	0.32
Al ₂ O ₃	20.10	0.14	21.88	0.19	0.24	35.46	23.87	33.73
FeO	0.04	13.20	31.54	48.65	94.63	2.88	0.08	7.96
MnO	0.01	2.51	0.14	1.05	0.00	0.64	0.01	0.09
MgO	0.00	11.71	9.82	0.16	0.07	0.37	0.00	7.35
CaO	0.36	28.78	0.06	0.05	0.02	0.07	5.08	0.56
Na ₂ O	10.71	0.13	0.21	0.00	0.03	0.51	7.86	1.91
K ₂ O	0.00	0.04	0.36	0.03	0.01	10.62	0.03	0.01
TOTAL	100.49	56.78	93.25	105.29	95.33	96.63	99.17	89.29
Recalculated Ions								
Oxygens	8	2	28	3	4	22	8	49
Si	3.00	0.01	5.83	0.01	0.01	6.06	2.77	11.81
Ti	0.00	0.00	0.06	0.99	0.00	0.05	0.00	0.08
Al	1.03	0.01	5.21	0.01	0.01	5.56	1.25	12.57
Fe ³⁺	0.00	0.00	0.00	0.00	1.97	0.00	0.00	0.00
Fe ²⁺	0.00	0.35	5.33	0.97	0.99	0.32	0.00	2.10
Mn	0.00	0.07	0.02	0.02	0.00	0.07	0.00	0.02
Mg	0.00	0.56	2.96	0.01	0.00	0.07	0.00	3.46
Ca	0.02	0.99	0.01	0.00	0.00	0.01	0.24	0.19
Na	0.90	0.01	0.08	0.00	0.00	0.13	0.68	1.17
K	0.00	0.00	0.09	0.00	0.00	1.80	0.00	0.00
TOTAL	4.94	1.99	19.60	2.00	2.98	14.08	100.49	99.17

Table C.1 cont.: Averaged mineral analyses for sample AWDBf1

AWDBf1	Ms	Pl	Mag
Wt %			
SiO ₂	45.93	64.32	0.17
TiO ₂	0.51	0.02	0.15
Al ₂ O ₃	35.88	23.76	0.24
FeO	2.57	0.10	94.83
MnO	0.02	0.01	0.02
MgO	0.20	0.00	0.05
CaO	0.02	4.46	0.02
Na ₂ O	0.62	8.51	0.00
K ₂ O	9.72	0.02	0.03
TOTAL	95.47	101.19	95.50
Recalculated Ions			
Oxygens	22	8	4
Si	6.11	2.80	0.01
Ti	0.05	0.00	0.00
Al	5.63	1.22	0.01
Fe ³⁺	0.00	0.00	1.97
Fe ²⁺	0.29	0.00	0.99
Mn	0.00	0.00	0.00
Mg	0.04	0.00	0.00
Ca	0.00	0.21	0.00
Na	0.16	0.72	0.00
K	1.65	0.00	0.00
TOTAL	13.93	4.95	2.99

Table C.1 cont.: Averaged mineral analyses for sample AWDBf3

AWDBf3	Ank	Chl	Fe-Ti ox	Ms	Pl
Wt %					
SiO ₂	0.39	22.52	0.96	45.81	66.29
TiO ₂	0.15	0.04	7.75	0.46	0.02
Al ₂ O ₃	0.19	23.71	0.57	35.58	22.32
FeO	14.42	41.85	81.68	3.08	0.11
MnO	2.81	0.15	0.01	0.02	0.01
MgO	10.89	1.99	0.08	0.78	0.18
CaO	28.21	0.08	0.04	0.10	2.25
Na ₂ O	0.13	0.01	0.00	0.72	9.55
K ₂ O	0.01	0.03	0.12	10.19	0.09
TOTAL	57.20	90.39	91.21	96.73	100.82
Recalculated Ions					
Oxygens	2	28	3	22	8
Si	0.01	5.04	0.02	6.06	2.88
Ti	0.00	0.01	0.14	0.05	0.00
Al	0.01	6.25	0.02	5.55	1.14
Fe ³⁺	0.00	0.00	1.63	0.00	0.00
Fe ²⁺	0.39	7.82	0.01	0.34	0.00
Mn	0.08	0.03	0.00	0.00	0.00
Mg	0.52	0.66	0.00	0.15	0.01
Ca	0.97	0.02	0.00	0.01	0.10
Na	0.01	0.01	0.00	0.19	0.80
K	0.00	0.01	0.00	1.72	0.01
TOTAL	1.98	19.84	1.83	14.07	4.95

Table C.1 cont.: Averaged mineral analyses for sample AWDBf4

AWDBf4	Ank	Cal	Chl	Ilm	Mag	Ms	Pl
Wt %							
SiO ₂	0.17	0.29	28.63	0.58	0.16	46.03	67.86
TiO ₂	0.03	0.02	0.02	34.48	0.06	0.55	0.01
Al ₂ O ₃	0.09	0.23	22.52	0.28	0.22	35.32	20.66
FeO	15.63	1.53	36.07	61.90	93.53	3.04	0.09
MnO	4.34	2.23	0.13	0.02	0.02	0.01	0.02
MgO	8.86	0.14	2.20	0.01	0.01	0.15	0.00
CaO	28.32	51.09	0.20	0.23	0.01	0.01	1.13
Na ₂ O	0.00	0.00	0.91	0.00	0.00	0.53	9.71
K ₂ O	0.09	0.06	0.12	0.02	0.01	9.31	0.01
TOTAL	57.52	55.59	90.81	97.53	94.01	94.95	99.49
Recalculated Ions							
Oxygens	2	1	28	3	4	22	8
Si	0.01	0.00	6.08	0.02	0.01	6.15	2.97
Ti	0.00	0.00	0.00	0.74	0.00	0.06	0.00
Al	0.00	0.00	5.64	0.01	0.01	5.57	1.06
Fe ³⁺	0.00	0.00	0.00	0.00	1.98	0.00	0.00
Fe ²⁺	0.43	0.02	6.41	1.47	0.99	0.34	0.00
Mn	0.12	0.03	0.02	0.00	0.00	0.00	0.00
Mg	0.43	0.00	0.70	0.00	0.00	0.03	0.00
Ca	1.00	0.93	0.05	0.01	0.00	0.00	0.05
Na	0.00	0.00	0.38	0.00	0.00	0.14	0.82
K	0.00	0.00	0.03	0.00	0.00	1.59	0.00
TOTAL	1.99	0.99	19.30	2.24	2.99	13.87	4.91

Table C.1 cont.: Averaged mineral analyses for sample AWDBf6

AWDBf6	Ank	Bt	Chl	Grt	St	Ilm	Ms	Pl
Wt %								
SiO ₂	0.13	36.41	25.94	37.61	27.43	0.33	45.78	59.51
TiO ₂	0.03	2.27	0.12	0.15	0.46	52.37	0.77	0.01
Al ₂ O ₃	0.16	19.11	22.48	21.13	54.53	0.27	34.93	26.38
FeO	9.65	16.91	21.53	25.09	13.44	47.96	2.53	0.08
MnO	1.91	0.09	0.14	11.91	0.65	1.71	0.02	0.01
MgO	14.34	11.47	17.09	2.42	1.52	0.20	0.62	0.00
CaO	29.08	0.09	0.08	4.28	0.02	0.19	0.05	7.80
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.46	6.76
K ₂ O	0.02	9.49	0.01	0.01	0.00	0.03	10.29	0.03
TOTAL	55.32	95.85	87.40	102.59	98.04	103.06	95.45	100.58
Recalculated Ions								
Oxygens	2	22	28	12	46	3	22	8
Si	0.00	5.43	5.36	2.97	7.60	0.01	6.12	2.63
Ti	0.00	0.26	0.02	0.01	0.10	0.97	0.08	0.00
Al	0.01	3.36	5.48	1.97	17.81	0.01	5.50	1.38
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.26	2.11	3.72	1.66	3.11	0.99	0.28	0.00
Mn	0.05	0.01	0.02	0.80	0.15	0.04	0.00	0.00
Mg	0.68	2.55	5.26	0.28	0.63	0.01	0.12	0.00
Ca	0.99	0.01	0.02	0.36	0.00	0.01	0.01	0.37
Na	0.00	0.00	0.00	0.00	0.00	0.00	0.12	0.58
K	0.00	1.81	0.00	0.00	0.00	0.00	1.75	0.00
TOTAL	1.99	15.54	19.89	8.04	29.40	2.02	13.99	4.97

Table C.1 cont.: Averaged mineral analyses for sample AWABv1

AWABv1	Ank	Bt	Chl	Ilm	Ms	Pl
Wt %						
SiO ₂	0.55	35.03	26.00	0.17	47.50	62.19
TiO ₂	0.01	1.80	0.09	55.40	0.44	0.02
Al ₂ O ₃	0.05	19.36	23.84	0.21	35.37	24.99
FeO	12.80	20.99	25.60	44.65	1.36	0.11
MnO	2.37	0.23	0.38	3.84	0.03	0.01
MgO	11.44	9.39	14.41	0.12	1.25	0.00
CaO	28.89	0.06	0.06	0.14	0.44	6.07
Na ₂ O	0.19	0.00	0.23	0.00	0.50	7.65
K ₂ O	0.02	9.43	0.27	0.03	9.79	0.04
TOTAL	56.32	96.29	90.89	104.56	96.69	101.10
Recalculated Ions						
Oxygens	2	22	28	3	22	8
Si	0.02	5.32	5.28	0.00	6.20	2.72
Ti	0.00	0.21	0.01	1.00	0.04	0.00
Al	0.00	3.47	5.70	0.01	5.45	1.29
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.34	2.67	4.35	0.90	0.15	0.00
Mn	0.06	0.03	0.07	0.08	0.00	0.00
Mg	0.55	2.13	4.36	0.00	0.24	0.00
Ca	1.00	0.01	0.01	0.00	0.06	0.28
Na	0.01	0.00	0.09	0.00	0.13	0.65
K	0.00	1.83	0.07	0.00	1.63	0.00
TOTAL	1.99	15.65	19.94	1.99	13.91	4.96

Table C.1 cont.: Averaged mineral analyses for sample AWT1

AWT1	Tur
Wt %	
SiO ₂	34.91
TiO ₂	0.51
Al ₂ O ₃	31.66
FeO	13.69
MnO	0.03
MgO	1.92
CaO	0.14
Na ₂ O	1.12
K ₂ O	0.01
TOTAL	83.99
Recalculated Ions	
Oxygens	49
Si	12.03
Ti	0.13
Al	12.86
Fe ³⁺	0.00
Fe ²⁺	3.95
Mn	0.01
Mg	0.99
Ca	0.05
Na	0.75
K	0.00
TOTAL	30.78

Table C.1 cont.: Averaged mineral analyses for sample AWDF1

AWDF1	Ank	Bt	Grt	Ms	Pl	Tur
Wt %						
SiO ₂	0.07	35.58	37.48	45.80	60.48	36.69
TiO ₂	0.02	2.19	0.07	0.61	0.03	0.58
Al ₂ O ₃	0.13	18.93	21.28	35.28	25.95	32.28
FeO	12.82	19.21	27.52	2.59	0.09	9.81
MnO	1.62	0.10	10.37	0.02	0.01	0.02
MgO	12.33	10.11	2.33	0.40	0.00	5.01
CaO	29.61	0.05	3.50	0.04	7.33	0.33
Na ₂ O	0.00	0.00	0.00	0.15	6.95	1.53
K ₂ O	0.01	9.83	0.00	10.00	0.02	0.01
TOTAL	56.61	96.00	102.57	94.90	100.86	86.27
Recalculated Ions						
Oxygens	2	22	12	22	8	49
Si	0.00	5.38	2.96	6.13	2.66	2.66
Ti	0.00	0.25	0.00	0.06	0.00	0.00
Al	0.00	3.38	1.98	5.57	1.35	1.35
Fe ³⁺	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.34	2.43	1.82	0.29	0.00	0.00
Mn	0.04	0.01	0.69	0.00	0.00	0.00
Mg	0.59	2.28	0.28	0.08	0.00	0.00
Ca	1.01	0.01	0.30	0.01	0.35	0.35
Na	0.00	0.00	0.00	0.04	0.59	0.59
K	0.00	1.90	0.00	1.71	0.00	0.00
TOTAL	2.00	15.63	8.04	13.89	4.96	4.96

Table C.1 cont.: Averaged mineral analyses for sample AWDF2

AWDF2	Chl	Grt	Ms	Pl	Tur
Wt %					
SiO ₂	24.84	37.64	45.50	68.89	36.42
TiO ₂	0.11	0.07	0.56	0.00	0.58
Al ₂ O ₃	22.15	21.32	35.16	20.15	32.51
FeO	33.14	25.99	2.31	0.12	9.02
MnO	0.15	11.64	0.02	0.03	0.03
MgO	7.98	2.40	0.38	0.00	5.33
CaO	0.09	3.63	0.03	0.39	0.61
Na ₂ O	0.00	0.00	0.25	9.50	1.35
K ₂ O	0.06	0.00	10.49	0.16	0.02
TOTAL	88.52	102.70	94.69	99.24	85.87
Recalculated Ions					
Oxygens	28	12	22	8	49
Si	5.40	2.97	6.12	3.01	12.00
Ti	0.02	0.00	0.06	0.00	0.14
Al	5.67	1.98	5.58	1.04	12.63
Fe ³⁺	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	6.02	1.71	0.26	0.00	2.49
Mn	0.03	0.78	0.00	0.00	0.01
Mg	2.58	0.28	0.08	0.00	2.62
Ca	0.02	0.31	0.00	0.02	0.21
Na	0.00	0.00	0.07	0.80	0.86
K	0.02	0.00	1.80	0.01	0.01
TOTAL	19.76	8.04	13.97	4.88	30.97

Appendix D

THERMOCALC Pseudosection Bulk Compositions

Bulk compositions for samples used in THERMOCALC modelling (Table 4.2) were obtained by combining average mineral compositions (Appendix C), which have been converted to ‘ideal’ analyses (Powell & Holland, 2008), with optically determined modal proportions. These ideal mineral compositions and modal proportions are presented here.

The Ideal Analysis Approach

The principles behind the ideal analysis approach are discussed in section 4.2.3 and by Powell & Holland (2008). An ideal analysis for a mineral is obtained by running a least squares minimization between the measured mineral analysis and the ideal analysis that is constrained by the appropriate a - x model (Appendix F). An example is given here for the average staurolite analysis in sample AWDP1.

The model system for sample AWDP1 is MnNCKFMASHTO, with the staurolite a - x model in the system MnFMASH. The composition of staurolite in this model system, is given by:



Compositional parameters are defined as:

$$\begin{aligned} x(\text{st}) &= \text{Fe}/(\text{Mg}+\text{Fe}) \\ m(\text{st}) &= \text{Mn}/(\text{Mg}+\text{Mn}+\text{Fe}) \end{aligned}$$

Therefore, in the ideal analysis, the numbers of Si and Al cations are fixed, cation exchange is limited to between Fe-Mg-Mn and no other elements are permitted. The numbers of Fe, Mg and Mn cations are calculated via the compositional parameters above according to the equations in column ‘Ideal calculation’ in table D.1. The ideal analysis is achieved by

Table D.1: Example of an ideal analysis calculation. See text for descriptions of columns.

	Obs. cats. ¹	Ideal calculation	Calc. cats. ²	diff. ^{2*}	weight [†]	lsq. [‡]
Si	7.62	7.5	7.50	1.44	0.36	0.52
Ti	0.10	-	0.00	1.00	3.16	3.16
Al	17.69	18	18.00	9.61	0.24	2.28
Fe ³⁺	0.00	-	0.00	0.00	5.00	0.00
Fe ²⁺	3.30	$4 \times x(\text{st}) \times (1 - m(\text{st}))$	3.29	0.02	0.55	0.01
Mn	0.09	$4 \times m(\text{st})$	0.09	0.00	3.33	0.00
Mg	0.63	$4 \times (1 - x(\text{st})) \times (1 - m(\text{st}))$	0.63	0.00	1.26	0.00
Ca	0.00	-	0.00	0.00	5.00	0.00
Na	0.00	-	0.00	0.00	5.00	0.00
K	0.00	-	0.00	0.00	5.00	0.00
Sum	29.43	-	29.50	12.07		5.98
x(st)	0.84	-	0.84			
m(st)	0.02	-	0.02			

¹Measured cations, as per appendix C.²Recalculated, ‘ideal’ cations.*(Obs. – Calc.)² × 100. Factor of 100 is a scalar to produce sensible numbers.[†]1/SQRT(Obs. cats.). If the observed cations are less than a cutoff of 0.04, then the weighting is given by 1/SQRT(0.04).[‡]diff.² × weight.

minimizing the least squares (lsq.) sum, by iteratively changing the calculated compositional parameters $x(\text{st})$ and $m(\text{st})$.

Adjustment of the Bulk Composition for AWD P1

The bulk composition for sample AWD P1 given in table 4.2 has been adjusted from that calculated through ideal mineral analyses. The bulk composition obtained by recalculating ideal analyses, as above, are given in table D.2. Early attempts at modelling sample AWD P1 using this bulk composition, however, showed that the characteristic isopleths did not intersect within the calculated peak assemblage field. This is because, although they converged towards low pressure, they were truncated by the stability limit of garnet, which marks the low pressure boundary to the field.

The position of the garnet-in line was ultimately found to be highly sensitive to total MnO content, with higher amounts dramatically increasing the stability range of garnet (Fig. D.1). This phenomenon has been well documented by White *et al.* (2007); Powell & Holland (2008) and Pattison & Tinkham (2009). Figure D.1 shows the position of the peak assemblage field (Bt-Grt-St-Ms-Pg-Pl-Ilm-Mag-Qtz) as a function of varying bulk MnO content. It can be seen that with higher MnO contents, the garnet-in line occurs at progressively lower pressures as Mn stabilizes garnet. In contrast, the positions of other field boundaries and isopleths do not change (all are within error) since these equilibria are not sensitive to variations in

minor element concentrations.

Furthermore, this manual adjustment is justified due to the significant uncertainty in determining effective bulk MnO content. The uncertainty arises as garnet is the major host to Mn in the sample, but accurately determining the modal proportion of garnet is difficult due to its very low abundance. The apparent abundance of garnet is, of course, sensitive to exactly which parts of which garnets appear in the thin section.

Ultimately, Mn-insensitive isopleths are used to locate the equilibrium position within the field, while the modelled garnet Mn content matches well with the observed composition.

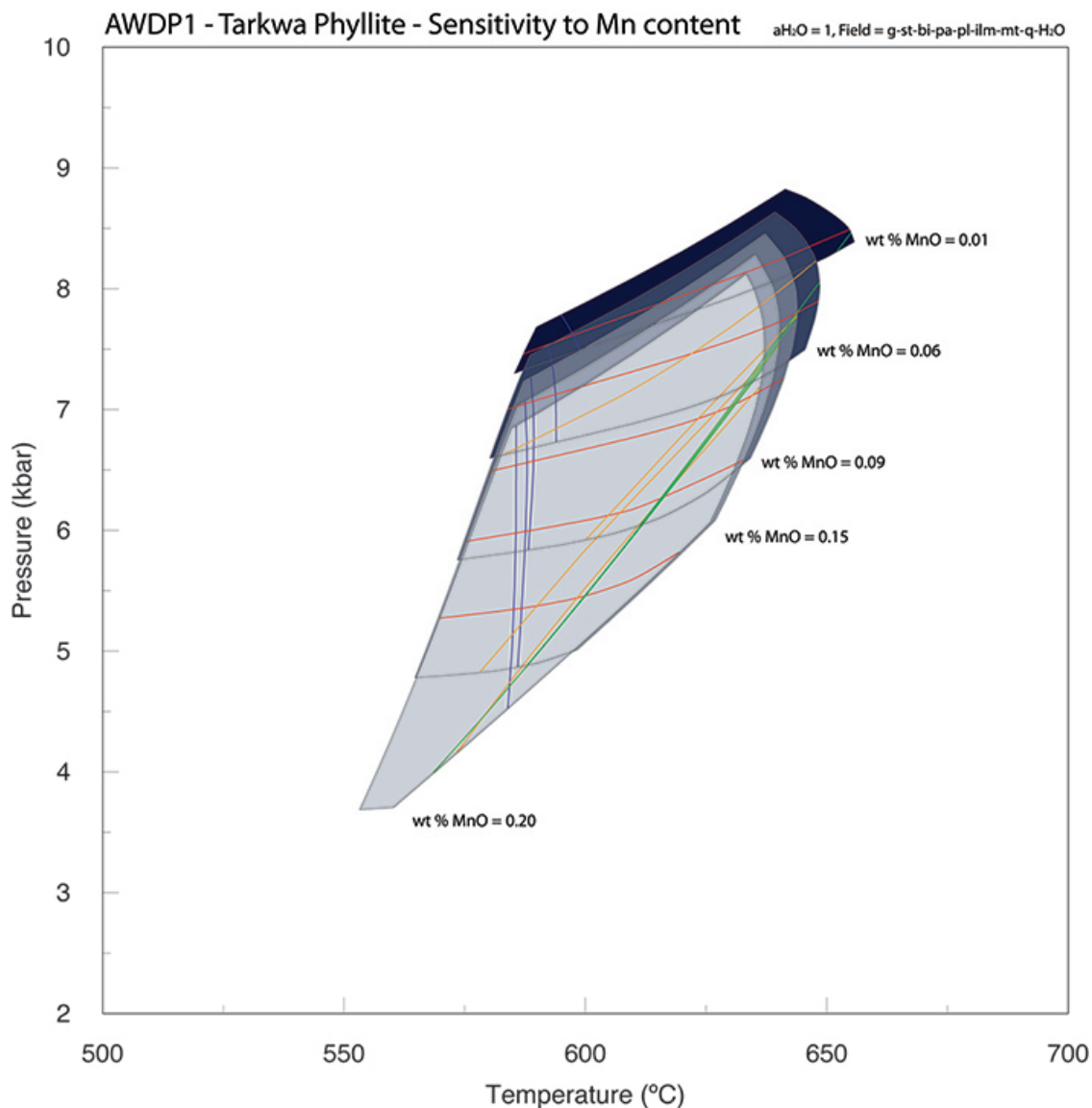


Figure D.1: Sensitivity of the position of the field Bt-Grt-St-Ms-Pg-Pl-Ilm-Mag-Qtz in sample AWDP1 as a function of bulk rock MnO content. Increasing bulk MnO results in garnet stabilizing at lower pressures, while other field boundaries are effectively insensitive to changing MnO content. Calculated compositional and modal isopleths (blue, $\text{Grt}_{\text{Prp}} = 0.116$; red, $\text{Grt}_{\text{Sps}} = 0.152$; green, $\text{Grt}_{\text{Grs}} = 0.025$; yellow, St mode = 5.5%) are shown and do not vary outside of error between different compositions.

Table D.2: Ideal mineral analyses, mineral modal proportions and bulk composition for sample AWDP1

	Grt	St	Chl	Bt	Ms	Pl	Pg	Mag	Ilm	Qtz
Si	3.00	7.50	2.57	2.65	3.07	2.92	3.00			1.00
Ti				0.11				0.01	0.99	
Al	2.00	18.00	2.87	1.70	2.81	1.08	3.00	0.02		
Fe ³⁺	0.00			0.00				1.97	0.01	
Fe ²⁺	2.16	3.29	1.98	1.14	0.08			1.00	0.97	
Mn	0.49	0.09	0.02	0.01					0.02	
Mg	0.27	0.63	2.58	1.29	0.05			0.00		
Ca	0.08					0.08				
Na					0.24	0.92	0.77			
K				1.00	0.76	0.00	0.23			
H		4.00	8.00	2.00	2.00		2.00			
Vol. %	0.4	4.7	6.3	7.0	34.8	7.5	9.2	3.5	0.5	26.2

	SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O	TiO ₂	MnO	Fe ₂ O ₃
Orig.	62.45	16.02	0.15	3.76	10.05	2.99	2.15	0.48	0.09	1.85
Adjust.	62.39	16.00	0.15	3.76	10.04	2.99	2.15	0.48	0.20	1.85

Table D.2 cont.: Ideal mineral analyses, mineral modal proportions and bulk composition for sample AWDBm1

	Grt	St	Chl	Bt	Ms	Pl	Pg	Mag	Ilm	Qtz
Si	3.00	7.50	2.56	2.64	3.07	2.90	3.00			1.00
Ti				0.10				0.00	1.00	
Al	2.00	18.00	2.89	1.72	2.80	1.10	3.00	0.02		
Fe ³⁺	0.00			0.00				1.98	0.00	
Fe ²⁺	2.27	3.24	2.24	1.22	0.08			0.99	0.99	
Mn	0.33	0.01	0.01	0.01					0.01	
Mg	0.30	0.75	2.31	1.21	0.05			0.01		
Ca	0.10					0.10				
Na					0.23	0.89	0.79			
K				1.00	0.77	0.01	0.21			
H		4.00	8.00	2.00	2.00		2.00			
Vol. %	1.0	3.0	6.0	3.5	38.5	15.5	3.0	1.0	0.5	28.0

	SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O	TiO ₂	MnO	Fe ₂ O ₃
	68.55	15.59	0.41	2.88	5.86	2.95	2.69	0.44	0.09	0.54

Table D.2 cont.: Ideal mineral analyses, mineral modal proportions and bulk composition for sample AWDDo3

	Hbl	Ep	Chl	Pl	Sph	Qtz	
Si	6.46	3.00	2.70	2.48	1.00	1.00	
Ti							
Al	2.36	2.94	2.60	1.52	1.00		
Fe ³⁺	0.01	0.06					
Fe ²⁺	1.51		1.16				
Mg	2.68		3.54				
Ca	1.80	2.00		0.52	1.00		
Na	0.34			0.48			
H	2.00	2.00	8.00				
Vol. %	47.0	5.0	2.5	44.0	38.5	0.5	1.0
SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	Na ₂ O	TiO ₂	Fe ₂ O ₃
68.55	15.59	0.41	2.88	5.86	2.69	0.44	0.54

Table D.2 cont.: Ideal mineral analyses, mineral modal proportions and bulk composition for sample AWDDo1

	Hbl	Chl	Bt	Pl	Ank	Ilm	Qtz	
Si	6.31	2.64	2.66	2.61			1.00	
Ti			0.13			1.00		
Al	2.57	2.73	1.68	1.39				
Fe ³⁺	0.01		0.00			0.00		
Fe ²⁺	2.06	1.96	1.20		0.34	1.00		
Mg	2.11	2.67	1.33		0.66			
Ca	1.77			0.39	1.00			
Na	0.35			0.61				
K			1.00	0.01				
C					2.00			
H	2.00	8.00	1.74					
Vol. %	1.0	1.0	17.0	41.0	17.0	1.0	22.0	
SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O	TiO ₂	Fe ₂ O ₃
58.62	9.66	10.63	8.56	6.84	1.44	3.10	1.15	0.0033

Table D.2 cont.: Ideal mineral analyses, mineral modal proportions and bulk composition for sample AWABf3

	Chl	Ms	Ab	Pl	Ank	Ilm	Qtz	
Si	2.56	3.10	3.00	2.75			1.00	
Ti						1.00		
Al	2.89	2.70	1.00	1.25				
Fe ³⁺						0.00		
Fe ²⁺	3.44	0.13			0.22	1.00		
Mg	1.12	0.07			0.78			
Ca			0.00	0.25	1.00			
Na		0.08	1.00	0.74				
K		0.92	0.00	0.01				
C					2.00			
H	8.00	2.00						
Vol. %	1.5	23.0	5.0	6.0	4.0	0.5	60.0	

SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O	TiO ₂	Fe ₂ O ₃
83.94	7.16	1.87	1.67	1.82	1.84	1.30	0.39	0.00056

Table D.2 cont.: Ideal mineral analyses, mineral modal proportions and bulk composition for sample AWDDo2

	Hbl	Cum	Chl	Bt	Pl	Ank	Mag	Ilm
Si	6.32	7.90	2.72	2.73	2.64			
Ti				0.10			0.01	1.00
Al	2.57	0.23	2.57	1.54	1.36		0.00	
Fe ³⁺	0.01	0.00		0.01			1.99	0.00
Fe ²⁺	1.89	2.55	1.42	0.88		0.25	1.01	1.00
Mg	2.35	4.07	3.29	1.74		0.75		
Ca	1.70	0.20			0.36	1.00		
Na	0.35	0.03			0.63			
K				1.00	0.01			
C						2.00		
H	2.00	2.00	8.00	1.90				
Vol. %	36.0	12.0	4.5	1.6	39.2	5.5	0.7	0.5

SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O	TiO ₂	Fe ₂ O ₃
50.20	10.84	10.58	13.66	10.56	0.14	3.28	0.37	0.36

Table D.2 cont.: Ideal mineral analyses, mineral modal proportions and bulk composition for sample AWDP4

	Chl	Bt	Ms	Pl	Qtz	
Si	2.58	2.66	3.06	2.64	1.00	
Al	2.84	1.67	2.81	1.36		
Fe ²⁺	1.81	1.09	0.09			
Mg	2.77	1.36	0.03			
Ca				0.36		
Na			0.20	0.63		
K		1.00	0.80	0.01		
C						
H	8.00	2.00	2.00			
Vol. %	1.3	4.0	60.9	32.5	1.3	
SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O
61.44	22.78	3.08	1.79	2.10	4.95	3.86

Appendix E

THERMOCALC Average P-T results

Results of THERMOCALC average P-T calculations, as per table 4.3, with calculated end-member activities. ‘correl.’ gives the correlation between uncertainties on P and T due to shared end-members between multiple equilibria. ‘fit’ is $\sqrt{\text{MSWD}}$ and for 95% confidence this value should, in general, be less than ‘max. fit’, which is determined from a χ^2 test. N_R is the number of independent reactions (equilibria) used to calculate the values of P and T. In certain cases particular end-members are excluded from calculations. This is because their activity residual (the difference between the measured activity of an end-member, based on the mineral analysis, and the calculated activity required for all equilibria to intersect the average P-T) is large. A more comprehensive discussion of these ‘diagnostic’ values is given in [Powell & Holland \(1994\)](#).

Table E.1: Detailed results of THERMOCALC average P-T calculations

Sample	AWDP1	AWDBm1	AWDF1	AWDCs1	AWDBf6	AWABv1	AWDDo3
Activites							
Chlorite							
clin	0.041	0.026	-	0.0119	0.053	-	0.19
daph	0.0103	0.022	-	0.043	0.0103	0.09	8.1E ⁻⁴
ames	0.054	0.035	-	0.0161	0.049	2.0E ⁻⁵	0.132
Biotite							
phl	0.07	0.055	0.049	0.031	0.064	-	-
ann	0.04	0.05	0.053	0.033	0.033	-	-
east	0.066	0.055	0.048	0.055	0.061	3.0E ⁻⁵	-
Garnet							
py	0.0028	0.0019	0.00137	0.0025	0.00161	-	-
gr	3.4E ⁻⁵	7.0E ⁻⁵	0.00122	0.0018	0.0021	-	-
alm	0.29	0.36	0.17	0.2	0.12	-	-
Staurolite							
mst	0.0035	0.0019	-	-	0.0016	-	-
fst	0.33	0.39	-	-	0.41	-	-
Amphibole							
tr	-	-	-	-	-	-	0.149
fact	-	-	-	-	-	-	3.23E ⁻⁴
ts	-	-	-	-	-	-	0.028
parg	-	-	-	-	-	-	0.002
gl	-	-	-	-	-	-	0.00324
Epidote							
cz	-	-	-	-	-	-	0.46
Muscovite							
mu	0.73	0.71	0.74	0.69	0.71	0.95	-
cel	-	-	-	0.107	0.0087	-	-
fcel	-	-	-	-	0.017	-	-
pa	0.39	0.1	0.048	0.193	0.208	0.286	-
Plagioclase							
an	0.083	0.119	0.56	0.086	0.61	0.53	0.68
ab	0.95	0.92	0.64	0.95	0.62	0.7	0.57
Ilm-Hem							
ilm	0.55	0.86	-	-	-	-	-
hem	-	0.0015	-	-	-	-	-
geik	0.0044	0.0071	-	-	-	-	-
Spinel							
mag	-	-	-	-	-	0.366	-
Carbonate							
dol	-	-	-	-	0.7	-	-
ank	-	-	-	-	0.33	-	-
sid	-	-	-	-	-	0.63	-
cc	-	-	-	-	-	0.79	-
Others	q, H ₂ O	q, H ₂ O	q, H ₂ O	q, H ₂ O	q, H ₂ O	q, H ₂ O	q, H ₂ O
Results							
T°C	584	578	615	550	580	501	594
σ (T)	18	17	129	37	20	20	64
P (kbar)	6.9	6.7	6.0	3.6	6.8	-	7.5
σ (P)	1.5	1.3	1.6	3.4	1.2	-	2.2
correl.	0.383	0.376	0.866	0.878	0.359	-	0.963
fit	1.36	1.31	0.49	1.74	1.64	1.41	1.31
max. fit	1.42	1.42	1.73	1.49	1.35	1.61	1.73
N _R	9	9	4	7	12	4	4
eliminated end-members	mst	mst, hem	-	an	-	-	gl, ames

Appendix F

Activity-Composition Models

Presented here are the activity-composition (a - x) relationships for solid solution minerals used in pseudosection calculations with THERMOCALC.

```
% =====
% Staurolite: MnFMASH: Holland & Powell (1998)

% x(st) = Fe/(Fe + Mg)
% m(st) = Mn/(Fe + Mg + Mn)

st 3 % mineral abbreviation and number of end-members
x(st) 0.8153 % Starting guesses
m(st) 0.003140

%-----
% definitions of end-member proportions
p(mst) 1 2 1 1 -1 x 1 1 -1 m
p(fst) 1 2 0 1 1 x 1 1 -1 m
p(mnst) 1 1 0 1 1 m
%-----
% type of mixing model
sf
% interaction energies
w(mst,fst) -8 0 0
w(mst,mnst) 0 0 0
w(fst,mnst) 0 0 0
%-----
3 % number of site terms
% definitions of site occupancies
x(Mg) 1 2 1 1 -1 x 1 1 -1 m
x(Fe) 1 2 0 1 1 x 1 1 -1 m
x(Mn) 1 1 0 1 1 m
%-----
% mixing activities
mst 1 1 x(Mg) 4
fst 1 1 x(Fe) 4
mnst 1 1 x(Mn) 4
```

```
% =====
% Cordierite: MnFMASH: Holland & Powell (1998)

% x(cd) = Fe/(Fe + Mg)
% m(cd) = Mn/(Fe + Mg + Mn)
% h(cd) = H2O

cd 4
x(cd) 0.3
m(cd) 0.02
h(cd) 0.5

%-----
p(crd) 2 1 1 2 -1 m -1 h
2 0 1 -1 x 1 1 -1 m
p(fcrd) 1 2 0 1 1 x 1 1 -1 m
p(mncrd) 1 1 0 1 1 m
p(hcrd) 1 1 0 1 1 h
%-----
ideal
%-----
5
x(Mg) 1 2 1 1 -1 x 1 1 -1 m
x(Fe) 1 2 0 1 1 x 1 1 -1 m
x(Mn) 1 1 0 1 1 m
h 1 1 0 1 1 h
noth 1 1 1 1 -1 h
%-----
% mixing activities
crd 1 2 x(Mg) 2 noth 1
fcrd 1 2 x(Fe) 2 noth 1
mncrd 1 2 x(Mn) 2 noth 1
hcrd 1 2 x(Mg) 2 h 1
```

```

% =====
% Epidote: CFMASH0: Holland & Powell (1998)

% f(ep) = Fe3+(m13)/(Fe3+(m13) + Al(m13))
% Q(ep) = 1/2(Fe3+(m3) - Fe3+(m1))

ep 3
  f(ep) 0.117
  Q(ep) 0.116 range 0 0.5
% -----
p(cz) 1 1 1 2 -1 f -1 Q
p(ep) 1 1 0 1 2 Q
p(fep) 1 1 0 2 1 f -1 Q
% -----
sf
W(cz,ep) 0 0 0
W(cz,fep) 15.4 0 0
W(ep,fep) 3 0 0
% -----
4
xFeM1 1 1 0 2 1 f -1 Q
xAlM1 1 1 1 2 -1 f 1 Q
xFeM3 1 1 0 2 1 f 1 Q
xAlM3 1 1 1 2 -1 f -1 Q
% -----
cz 1 2 xAlM1 1 xAlM3 1
ep 1 2 xAlM1 1 xFeM3 1
fep 1 2 xFeM1 1 xFeM3 1
% =====
% Chlorite: MnFMASH: Holland et al. (1998)

% x(chl) = Fe/(Fe + Mg)
% y(chl) 2y = x(Al,M4) + x(Al,M1)
% m(chl) = x(Mn,M23) = x(Mn,M1)

chl 4
  x(chl) 0.3356
  y(chl) 0.5371
  m(chl) 0.0009870
% -----
p(clin) 2 1 2 2 -2 y -1 m
        2 0 1 -2/5 x 3 2 -1 y -1 m
p(daph) 1 2 0 1 2/5 x 3 2 -1 y -1 m
p(ames) 1 1 -1 1 2 y
p(mnchl) 1 1 0 1 1 m
% -----
sf
W(clin,daph) 2.5 0 0
W(clin,ames) 18 0 0
W(clin,mnchl) 0 0 0
W(daph,ames) 13.5 0 0
W(daph,mnchl) 0 0 0
W(ames,mnchl) 0 0 0
% -----
9
x(Fe,M23) 1 2 0 1 1 x 1 1 -1 m
x(Mg,M23) 1 2 1 1 -1 x 1 1 -1 m
x(Mn,M23) 1 1 0 1 1 m

x(Al,M1) 1 1 -1 1 2 y
x(Fe,M1) 1 2 0 1 1 x 2 2 -2 y -1 m
x(Mg,M1) 1 2 1 1 -1 x 2 2 -2 y -1 m
x(Mn,M1) 1 1 0 1 1 m
x(Al,T2) 1 1 0 1 1 y
x(Si,T2) 1 1 1 1 -1 y
% -----
clin 4 4 x(Mg,M23) 4 x(Mg,M1) 1 x(Al,T2) 1 x(Si,T2) 1
daph 4 4 x(Fe,M23) 4 x(Fe,M1) 1 x(Al,T2) 1 x(Si,T2) 1
ames 1 3 x(Mg,M23) 4 x(Al,M1) 1 x(Al,T2) 2
mnchl 4 4 x(Mn,M23) 4 x(Mn,M1) 1 x(Al,T2) 1 x(Si,T2) 1
% =====
% Chlorite: FMASH: Holland et al. (1998)

% x(chl) = Fe/(Fe + Mg)
% y(chl) 2y = x(Al,M4) + x(Al,M1)

chl 4
  x(chl) 0.414
  y(chl) 0.508
  Q(chl) 0.4905
% -----
p(afchl) 1 1 1 2 -1 y -1 Q
p(clin) 2 1 0 1 2 Q
        2 0 1 -2/5 x 3 1 -1 y
p(daph) 1 2 0 1 2/5 x 3 1 -1 y
p(ames) 1 1 0 2 1 y -1 Q
% -----
sf
W(afchl,clin) 18 0 0
W(afchl,daph) 14.5 0 0
W(afchl,ames) 20 0 0
W(clin,daph) 2.5 0 0
W(clin,ames) 18 0 0
W(daph,ames) 13.5 0 0
% =====
10
x(Fe,M23) 1 1 0 1 1 x
x(Mg,M23) 1 1 1 1 -1 x

x(Al,M1) 1 1 0 2 1 y -1 Q
x(Fe,M1) 1 2 0 1 1 x 1 2 -1 y 1 Q
x(Mg,M1) 1 2 1 1 -1 x 1 2 -1 y 1 Q

x(Al,M4) 1 1 0 2 1 y 1 Q
x(Fe,M4) 1 2 0 1 1 x 1 2 -1 y -1 Q
x(Mg,M4) 1 2 1 1 -1 x 1 2 -1 y -1 Q

x(Al,T2) 1 1 0 1 1 y
x(Si,T2) 1 1 1 1 -1 y
% -----
afchl 1 4 x(Mg,M23) 4 x(Mg,M1) 1 x(Mg,M4) 1 x(Si,T2) 2
check 0 0 0
clin 4 5 x(Mg,M23) 4 x(Mg,M1) 1 x(Al,M4) 1 x(Al,T2) 1 x(Si,T2) 1
check 0 1/2 1/2
daph 4 5 x(Fe,M23) 4 x(Fe,M1) 1 x(Al,M4) 1 x(Al,T2) 1 x(Si,T2) 1
check 1 1/2 1/2
ames 1 4 x(Mg,M23) 4 x(Al,M1) 1 x(Al,M4) 1 x(Al,T2) 2
check 0 1 0
% =====
% Magnetite: FT0: White et al. (2000)

% x(mt) = prop "mt"
% Q (mt) = x(Fe3,tet)

mt 3
  x(mt) 0.9704
  Q(mt) 0.7472
% -----
p(int) 1 1 0 2 -2 x 3 Q
p(dmt) 1 1 0 2 3 x -3 Q
p(usp) 1 1 1 1 -1 x
% -----
sf
W(int,dmt) 2.4 0 0
W(int,usp) 1 0 0
W(dmt,usp) -5 0 0
% -----
5
x(Ti,oct) 1 1 1/2 1 -1/2 x
x(Fe3,oct) 1 1 0 2 1 x -1/2 Q
x(Fe2,oct) 1 1 1/2 2 -1/2 x 1/2 Q
x(Fe3,tet) 1 1 0 1 1 Q
x(Fe2,tet) 1 1 1 1 -1 Q
% -----
int 4 3 x(Fe3,oct) 1 x(Fe2,oct) 1 x(Fe3,tet) 1
make 1 mt 1
DQF -1.8595 0.003166 0 % from dmt, - dHid - RT Log[16/27]
check 1 1 % dHid = 3.19
dmt 27/4 4 x(Fe3,oct) 4/3 x(Fe2,oct) 2/3 x(Fe3,tet) 2/3 x(Fe2,tet) 1/3
make 1 mt 1
DQF 1.3305 -0.0011845 0 % taking off config dqf
check 1 2/3
usp 4 3 x(Ti,oct) 1 x(Fe2,oct) 1 x(Fe2,tet) 1
check 0 0
% =====
% Spinel: FMAT0: White et al (2002)

% x(he) = Fe2/(Mg + Fe2)
% y(he) = Al/(Al + Fe3 + 2Ti)
% z(he) = 2Ti/(Al + Fe3 + 2Ti)

sp 4
  x(sp) 0.93 % Fe2/(Mg + Fe2)
  y(sp) 0.80 % Al/(Al + Fe3 + 2Ti)
  z(sp) 0.01 % 2Ti/(Al + Fe3 + 2Ti)
% -----
p(herc) 2 1 0 1 1 y
        2 -1 1 1 x 1 1 1 z
p(sp) 1 2 1 1 -1 x 1 1 1 z
p(mt) 1 1 1 2 -1 y -1 z
p(usp) 1 1 0 1 1 z
% -----
sf
W(herc,sp) 0 0 0
W(herc,mt) 18.5 0 0
W(herc,usp) 27 0 0
W(sp,mt) 40 0 0
W(sp,usp) 30 0 0 % * free
W(mt,usp) 0 0 0
% -----
5
x(Al) 1 1 0 1 1 y
x(Fe3) 1 1 1 2 -1 y -1 z
x(Ti) 1 1 0 1 1 z
x(Mg) 1 1 1 1 -1 x
x(Fe2) 1 1 0 1 1 x
% -----
herc 1 2 x(Al) 1 x(Fe2) 1
sp 1 2 x(Al) 1 x(Mg) 1
mt 1 2 x(Fe3) 1 x(Fe2) 1
usp 1 2 x(Ti) 1 x(Fe2) 1

```

```

% =====
% K-Feldspar: NCKAS: Holland & Powell (2003)

% na(ksp) = Na/(Na + Ca + K)
% ca(ksp) = Ca/(Na + Ca + K)

ksp 3
na(ksp) 0.15
ca(ksp) 0.004
% -----
p(san) 1 1 1 2 -1 na -1 ca
p(ab) 1 1 0 1 1 na
p(an) 1 1 0 1 1 ca
% -----
%
asf
w(sanab) 25.1 -0.0108 0.338
w(sanan) 40 0 0
w(aban) 3.1 0 0

san 1.0 0 0
ab 0.643 0 0
an 1.0 0 0
% -----
3
x(K) 1 1 1 2 -1 na -1 ca
x(Na) 1 1 0 1 1 na
x(Ca) 1 1 0 1 1 ca
% -----
%
san 1 1 x(K) 1
abn 1 1 x(Na) 1
an 1 1 x(Ca) 1
DQF 7.03 -0.00466 0
% -----
% =====
% Plagioclase Feldspar: NCKAS: Holland & Powell (2003)

% ca(pl) = Ca/(Na + Ca + K)
% k(pl) = k/(Na + Ca + K)

pl 3
ca(pl) 0.5
k(pl) 0.02
% -----
p(ab) 1 1 1 2 -1 k -1 ca
p(an) 1 1 0 1 1 ca
p(san) 1 1 0 1 1 k
% -----
%
asf
w(aban) 3.1 0 0
w(sanab) 25.1 -0.0108 0.338
w(sanan) 40 0 0

ab 0.643 0 0
an 1.0 0 0
san 1.0 0 0
% -----
3
x(K) 1 1 0 1 1 k
x(Na) 1 1 1 2 -1 k -1 ca
x(Ca) 1 1 0 1 1 ca
% -----
%
abn 1 1 x(Na) 1
an 1 1 x(Ca) 1
DQF 7.03 -0.00466 0
san 1 1 x(K) 1
% -----
% =====
% Plagioclase: NCAS: Holland & Powell (2003)

% ca(pl) = Ca/(Na + Ca)

pl 2
ca(pl) 0.4
% -----
p(abh) 1 1 1 1 -1 ca
p(an) 1 1 0 1 1 ca
% -----
%
asf
w(abhan) 3.1 0 0

abh 0.643 0 0
an 1.0 0 0
% -----
2
x(Na) 1 1 1 1 -1 ca
x(Ca) 1 1 0 1 1 ca
% -----
%
abh 1 1 x(Na) 1
an 1 1 x(Ca) 1
DQF 7.03 -0.00466 0
% -----
% =====
% Dolomite: CoFM: White et al (2003)

% x = Fe/(Fe+Mg)

dol 2
x(dol) 0.1
% -----
p(dol) 1 1 1 1 -1 x
p(ank) 1 1 0 1 1 x
% -----
%
sf
w 3 0 0
% -----
2
x(Mg) 1 1 1 1 -1 x
x(Fe) 1 1 0 1 1 x
% -----
%
dol 1 1 x(Mg) 1
ank 1 1 x(Fe) 1
DQF 5 0 0
% -----
% =====
% siderite: FMC: White et al (2003)

% x = Fe/(Fe+Mg)

sid 2
x(sid) 0.9
% -----
p(mag) 1 1 1 1 -1 x
p(sid) 1 1 0 1 1 x
% -----
%
sf
w 4 0 0
% -----
2
x(Mg) 1 1 1 1 -1 x
x(Fe) 1 1 0 1 1 x
% -----
%
mag 1 1 x(Mg) 1
sid 1 1 x(Fe) 1
DQF 1.831 0 0
% -----
% =====
% Clin amphibole: NCFMASHO: Diener et al (2007)

% A M13 M2 M4 T1
% v Na Mg Fe Mg Fe Al Fe3 Ca Mg Fe Na Si Al
% tr 1 0 3 0 2 0 0 0 2 0 0 0 4 0
% ts 1 0 3 0 0 0 2 0 2 0 0 0 2 2
% parg 0 1 3 0 1 0 1 0 2 0 0 0 2 2
% gl 1 0 3 0 0 0 2 0 0 0 0 2 4 0
% cumm 1 0 3 0 2 0 0 0 0 2 0 0 4 0
% grun 1 0 0 3 0 2 0 0 0 0 2 0 4 0
% a 1 0 3 0 0 2 0 0 0 0 2 0 4 0
% b 1 0 0 3 2 0 0 0 0 0 2 0 4 0
% mrb 1 0 3 0 0 0 0 2 0 0 0 2 4 0
%
% 3 xFeM13 + 2 xFeM2 + 2 xFeM4
% x ->
% 3 xFeM13 + 2 xFeM2 + 2 xFeM4 + 3 xMgM13 + 2 xMgM2 + 2 xMgM4
%
% y -> xAlM2
%
% z -> xNaM4
%
% a -> xNaA
%
% c -> xCaM4
%
% f -> xFe3M2
%
% xFeM13
% Q1 -> x -
% xFeM13 + xMgM13
%
% xFeM2
% Q2 -> x -
% xFeM2 + xMgM2
% -----
hb 9 % the same coding is used for gl, oct, cu

x(hb) 0.4146
y(hb) 0.6312
z(hb) 0.4302
a(hb) 0.4060
c(hb) 0.5607
f(hb) 0.07588
Q1(hb) -0.02510 range -1 1
Q2(hb) 0.1443 range -1 1

```

```

% -----
p(tr) 1 1 0 5 -1/2 a 1 c -1 f -1 y 1 z
p(ts) 1 1 0 4 -1/2 a 1 f 1 y -1 z
p(parg) 1 1 0 1 1 a
p(gl) 1 1 0 2 -1 f 1 z
p(cum) 5 1 1 5 -1 c -1 Q2 -1 x -1 z -3/2 Q1
      2 0 1 1 f 0 1 1 Q2
      2 0 1 1 c 0 1 1 x
      2 0 1 1 Q2 0 1 1 y
      2 0 1 1 x 0 1 1 z
p(grun) 7 1 0 3 1 x -2 Q2 -5/2 Q1
      2 0 1 2 f 0 1 1 Q2
      2 0 1 1 c 0 1 1 x
      2 0 1 -1 f 0 1 1 x
      2 0 1 2 Q2 0 1 1 y
      2 0 1 -1 x 0 1 1 y
      2 0 1 1 x 0 1 1 z
p(a) 5 1 0 2 1 Q2 5/2 Q1
      2 0 1 -1 f 0 1 1 Q2
      2 0 1 -1 c 0 1 1 x
      2 0 1 -1 Q2 0 1 1 y
      2 0 1 -1 x 0 1 1 z
p(b) 7 1 0 2 2 Q2 3/2 Q1
      2 0 1 -2 f 0 1 1 Q2
      2 0 1 -1 c 0 1 1 x
      2 0 1 1 f 0 1 1 x
      2 0 1 -2 Q2 0 1 1 y
      2 0 1 1 x 0 1 1 y
      2 0 1 -1 x 0 1 1 z
p(mrb) 1 1 0 1 1 f
% -----
osf
W(tr,ts) 20 0 0
W(tr,parg) 25 0 0
W(tr,gl) 65 0 0
W(tr,cum) 45 0 0
W(tr,grun) 75 0 0
W(tr,a) 57 0 0
W(tr,b) 63 0 0
W(tr,mrb) 65 0 0 %
W(ts,parg) -40 0 0
W(ts,gl) 25 0 0
W(ts,cum) 70 0 0
W(ts,grun) 80 0 0
W(ts,a) 70 0 0
W(ts,b) 72.5 0 0
W(ts,mrb) 25 0 0 %
W(parg,gl) 50 0 0
W(parg,cum) 50 0 0
W(parg,grun) 106.7 0 0
W(parg,a) 94.8 0 0
W(parg,b) 94.8 0 0
W(parg,mrb) 50 0 0 %
W(gl,cum) 100 0 0
W(gl,grun) 113.5 0 0
W(gl,a) 100 0 0
W(gl,b) 111.2 0 0
W(gl,mrb) 0 0 0 %
W(cum,grun) 33 0 0
W(cum,a) 18 0 0
W(cum,b) 23 0 0
W(cum,mrb) 100 0 0 %
W(grun,a) 12 0 0
W(grun,b) 8 0 0
W(grun,mrb) 113.5 0 0 %
W(a,b) 20 0 0
W(a,mrb) 100 0 0
W(b,mrb) 111.2 0 0 %

tr 1 0 0
ts 1.5 0 0
parg 1.7 0 0
gl 0.8 0 0
cum 1 0 0
grun 1 0 0
a 1 0 0
b 1 0 0
mrb 0.8 0 0

```

```

% -----
14
xvA 1 1 1 1 -1 a
xNaA 1 1 0 1 1 a
xMgM13 1 1 1 2 1 Q1 -1 x
xFeM13 1 1 0 2 -1 Q1 1 x
xMgM2 5 1 1 4 -1 f 1 Q2 -1 x -1 y
      2 0 1 -1 f 0 1 1 Q2
      2 0 1 1 f 0 1 1 x
      2 0 1 -1 Q2 0 1 1 y
      2 0 1 1 x 0 1 1 y
xFeM2 5 1 0 2 -1 Q2 1 x
      2 0 1 1 f 0 1 1 Q2
      2 0 1 -1 f 0 1 1 x
      2 0 1 1 Q2 0 1 1 y
      2 0 1 -1 x 0 1 1 y
xAlM2 1 1 0 1 1 y
xFe3M2 1 1 0 1 1 f
xCoM4 1 1 0 1 1 c
xMgM4 5 1 1 5 -1 c -1 Q2 -1 x -1 z -3/2 Q1
      2 0 1 1 f 0 1 1 Q2
      2 0 1 1 c 0 1 1 x
      2 0 1 1 Q2 0 1 1 y
      2 0 1 1 x 0 1 1 z
xFeM4 5 1 0 3 1 Q2 1 x 3/2 Q1
      2 0 1 -1 f 0 1 1 Q2
      2 0 1 -1 c 0 1 1 x
      2 0 1 -1 Q2 0 1 1 y
      2 0 1 -1 x 0 1 1 z
xNaM4 1 1 0 1 1 z
xSiT1 1 1 1 4 -1/2 f -1/2 y 1/2 z -1/4 a
xAlT1 1 1 0 4 1/2 f 1/2 y -1/2 z 1/4 a
% -----
tr 1 5 xvA 1 xMgM13 3 xMgM2 2 xCoM4 2 xSiT1 1
ts 2 6 xvA 1 xMgM13 3 xAlM2 2 xCoM4 2 xSiT1 1/2 xAlT1 1/2
DQF 10 0 0
parg 8 7 xNaA 1 xMgM13 3 xMgM2 1 xAlM2 1 xCoM4 2 xSiT1 1/2 xAlT1 1/2
DQF 15 0 0
gl 1 5 xvA 1 xMgM13 3 xAlM2 2 xNaM4 2 xSiT1 1
DQF 3 0 0
cum 1 5 xvA 1 xMgM13 3 xMgM2 2 xMgM4 2 xSiT1 1
DQF -6.4 0 0
grun 1 5 xvA 1 xFeM13 3 xFeM2 2 xFeM4 2 xSiT1 1
DQF -5 0 0
a 1 5 xvA 1 xMgM13 3 xFeM2 2 xFeM4 2 xSiT1 1
make 2 cum 3/7 grun 4/7
DQF -15.1 0 0
b 1 5 xvA 1 xFeM13 3 xMgM2 2 xFeM4 2 xSiT1 1
make 2 cum 2/7 grun 5/7
DQF -17.1 0 0
mrb 1 5 xvA 1 xMgM13 3 xFe3M2 2 xNaM4 2 xSiT1 1
make 3 gl 1 jd -2 acm 2
DQF 8 0 0
% =====
% Orthoamphibole: NCFMASHO: Diener et al (2007)
%
% A M4 M13 M2 T1
% v Na Ca Na Mg Fe Mg Fe Al Fe3 Mg Fe Al Si
% anth 1 0 0 0 2 0 3 0 0 0 2 0 0 4
% ged 1 0 0 0 2 0 3 0 2 0 0 0 2 2
% ompa 0 1 0 0 2 0 3 0 1 0 1 0 2 2
% omgl 1 0 0 2 0 0 3 0 2 0 0 0 0 4
% otr 1 0 2 0 0 0 3 0 0 0 2 0 0 4
% fanth 1 0 0 0 0 2 0 3 0 0 0 2 0 4
% omrb 1 0 0 2 0 0 3 0 0 2 0 0 0 4
% a 1 0 0 0 0 2 3 0 0 0 0 2 0 4
% b 1 0 0 0 0 2 0 3 0 0 2 0 0 4
%
% 3 xFeM13 + 2 xFeM2 + 2 xFeM4
% x -> -----
% 3 xFeM13 + 2 xFeM2 + 2 xFeM4 + 3 xMgM13 + 2 xMgM2 + 2 xMgM4
%
% y -> xAlM2
%
% z -> xNaM4
%
% a -> xNaA
%
% c -> xCoM4
%
% f -> xFe3M2
%
% xFeM13
% Q1 -> x -> -----
% xFeM13 + xMgM13
%
% xFeM2
% Q2 -> x -> -----
% xFeM2 + xMgM2
% -----

```



```
%
% 3 xFeM13 + 2 xFeM2 + 2 xFeM4
% x -> -----
% 3 xFeM13 + 2 xFeM2 + 2 xFeM4 + 3 xMgM13 + 2 xMgM2 + 2 xMgM4
%
% y -> xAlM2
%
% z -> xNaM4
%
% a -> xNaA
%
% c -> xCaM4
%
% f -> xFe3M2
%
%
% xFeM13
% Q1 -> x -> -----
% xFeM13 + xMgM13
%
% xFeM2
% Q2 -> x -> -----
% xFeM2 + xMgM2
%
% -----
```

anth 9 % the same coding is used for ged

```
x(anth) 0.28
y(anth) 0.11
z(anth) 0.05
a(anth) 0.09
c(anth) 0.05
f(anth) 0.05
Q1(anth) 0 range -1 1
Q2(anth) 0 range -1 1
%
% -----
p(anth) 5 1 1 7 -1/2 a -1 c -1 f -1 Q2 -1 x -1 y -3/2 Q1
2 0 1 1 f 0 1 1 Q2
2 0 1 1 c 0 1 1 x
2 0 1 1 Q2 0 1 1 y
2 0 1 1 x 0 1 1 z
p(ged) 1 1 0 4 -1/2 a 1 f 1 y -1 z
p(ompa) 1 1 0 1 1 a
p(ongl) 1 1 0 2 -1 f 1 z
p(otr) 1 1 0 1 1 c
p(fanth) 7 1 0 3 1 x -2 Q2 -5/2 Q1
2 0 1 2 f 0 1 1 Q2
2 0 1 1 c 0 1 1 x
2 0 1 -1 f 0 1 1 x
2 0 1 2 Q2 0 1 1 y
2 0 1 -1 x 0 1 1 y
2 0 1 1 x 0 1 1 z
p(omrb) 1 1 0 1 1 f
p(a) 5 1 0 2 1 Q2 5/2 Q1
2 0 1 -1 f 0 1 1 Q2
2 0 1 -1 c 0 1 1 x
2 0 1 -1 Q2 0 1 1 y
2 0 1 -1 x 0 1 1 z
p(b) 7 1 0 2 2 Q2 3/2 Q1
2 0 1 -2 f 0 1 1 Q2
2 0 1 -1 c 0 1 1 x
2 0 1 1 f 0 1 1 x
2 0 1 -2 Q2 0 1 1 y
2 0 1 1 x 0 1 1 y
2 0 1 -1 x 0 1 1 z
%
% -----
```

```
asf
W(anth,ged) 25 0 0
W(anth,ompa) 25 0 0
W(anth,ongl) 65 0 0
W(anth,otr) 45 0 0
W(anth,fanth) 33 0 0
W(anth,omrb) 65 0 0
W(anth,a) 18 0 0
W(anth,b) 23 0 0
W(ged,ompa) -40 0 0
W(ged,ongl) 25 0 0
W(ged,otr) 70 0 0
W(ged,fanth) 39.5 0 0
W(ged,omrb) 25 0 0
W(ged,a) 29 0 0
W(ged,b) 34.6 0 0
W(ompa,ongl) 50 0 0
W(ompa,otr) 90 0 0
W(ompa,fanth) 45 0 0
W(ompa,omrb) 50 0 0
W(ompa,a) 33.2 0 0
W(ompa,b) 36 0 0
W(ongl,otr) 65 0 0
W(ongl,fanth) 81.2 0 0
W(ongl,omrb) 0 0 0
W(ongl,a) 65.5 0 0
W(ongl,b) 78.4 0 0
W(otr,fanth) 75 0 0
W(otr,omrb) 65 0 0 %
W(otr,a) 57 0 0
W(otr,b) 63 0 0
W(fanth,omrb) 81.2 0 0 %
W(fanth,a) 12 0 0
W(fanth,b) 8 0 0
W(omrb,a) 65.5 0 0 %
W(omrb,b) 78.4 0 0 %
```

W(a,b) 20 0 0

```
anth 1 0 0
ged 1.5 0 0
ompa 1.7 0 0
ongl 0.8 0 0
otr 1 0 0
fanth 1 0 0
omrb 0.8 0 0
a 1 0 0
b 1 0 0
```

```
%
% -----
14
xvA 1 1 1 1 -1 a
xNaA 1 1 0 1 1 a
xCaM4 1 1 0 1 1 c
xNaM4 1 1 0 1 1 z
xMgM4 5 1 1 5 -1 c -1 Q2 -1 x -1 z -3/2 Q1
2 0 1 1 f 0 1 1 Q2
2 0 1 1 c 0 1 1 x
2 0 1 1 Q2 0 1 1 y
2 0 1 1 x 0 1 1 z
xFeM4 5 1 0 3 1 Q2 1 x 3/2 Q1
2 0 1 -1 f 0 1 1 Q2
2 0 1 -1 c 0 1 1 x
2 0 1 -1 Q2 0 1 1 y
2 0 1 -1 x 0 1 1 z
xMgM13 1 1 1 2 1 Q1 -1 x
xFeM13 1 1 0 2 -1 Q1 1 x
xAlM2 1 1 0 1 1 y
xFe3M2 1 1 0 1 1 f
xMgM2 5 1 1 4 -1 f 1 Q2 -1 x -1 y
2 0 1 -1 f 0 1 1 Q2
2 0 1 1 f 0 1 1 x
2 0 1 -1 Q2 0 1 1 y
2 0 1 1 x 0 1 1 y
xFeM2 5 1 0 2 -1 Q2 1 x
2 0 1 1 f 0 1 1 Q2
2 0 1 -1 f 0 1 1 x
2 0 1 1 Q2 0 1 1 y
2 0 1 -1 x 0 1 1 y
xAlT1 1 1 0 4 1/2 f 1/2 y -1/2 z 1/4 a
xSiT1 1 1 1 4 -1/2 f -1/2 y 1/2 z -1/4 a
%
% -----
```

```
anth 1 5 xvA 1 xMgM4 2 xMgM13 3 xMgM2 2 xSiT1 1
check 0 0 0 0 0 0 0
ged 2 6 xvA 1 xMgM4 2 xMgM13 3 xAlM2 2 xAlT1 1/2 xSiT1 1/2
check 0 1 0 0 0 0 0
DQF 22 0 0
ompa 8 7 xNaA 1 xMgM4 2 xMgM13 3 xAlM2 1 xMgM2 1 xAlT1 1/2 xSiT1 1/2
check 0 1/2 0 1 0 0 0
make 3 pang 1 tr -1 anth 1
DQF 27 0 0
ongl 1 5 xvA 1 xNaM4 2 xMgM13 3 xAlM2 2 xSiT1 1
check 0 1 1 0 0 0 0
make 1 gl 1
DQF 15 0 0
otr 1 5 xvA 1 xCaM4 2 xMgM13 3 xMgM2 2 xSiT1 1
check 0 0 0 0 1 0 0
make 1 tr 1
DQF 0 0 0
fanth 1 5 xvA 1 xFeM4 2 xFeM13 3 xFeM2 2 xSiT1 1
check 1 0 0 0 0 0 0
DQF 7 0 0
omrb 1 5 xvA 1 xNaM4 2 xMgM13 3 xFe3M2 2 xSiT1 1
check 0 0 1 0 0 1 0
make 3 gl 1 jd -2 acm 2
DQF 33 0 0
a 1 5 xvA 1 xFeM4 2 xMgM13 3 xFeM2 2 xSiT1 1
check 4/7 0 0 0 0 0 4/7 -3/7
make 2 anth 3/7 fanth 4/7
DQF -5.5 0 0
b 1 5 xvA 1 xFeM4 2 xFeM13 3 xMgM2 2 xSiT1 1
check 5/7 0 0 0 0 0 -2/7 5/7
make 2 anth 2/7 fanth 5/7
DQF -6.7 0 0
```

```

% =====
% Clinopyroxene: NCFMAS0: Green et al (2007)
%
%      M1m      M1a      M2c      M2n
%      Mg Fe Fe3 Al  Mg Fe Fe3 Al  Na Ca  Na Ca
% jd  0 0 0 1/2  0 0 0 1/2  1/2 0  1/2 0
% di  1/2 0 0 0  1/2 0 0 0  0 1/2  0 1/2
% hed 0 1/2 0 0  0 1/2 0 0  0 1/2  0 1/2
% ocm 0 0 1/2 0  0 0 1/2 0  1/2 0  1/2 0
% om  1/2 0 0 0  0 0 0 1/2  0 1/2  1/2 0
% cfm 0 1/2 0 0  1/2 0 0 0  0 1/2  0 1/2
% jac 0 0 0 1/2  0 0 1/2 0  1/2 0  1/2 0
%
%      xFe3M1a + xFe3M1m
% f -> -----
%      xAlM1a + xAlM1m + xFe3M1a + xFe3M1m
%
%      xFeM1a + xFeM1m
% x -> -----
%      xFeM1a + xFeM1m + xMgM1a + xMgM1m
%
%      xNaM2c + xNaM2n
% j -> -----
%      2
%
%      -xNaM2c + xNaM2n
% Q -> -----
%      2
%
%      xFe3M1a - xFe3M1m
% Qaf -> -----
%      2
%
%      xFeM1a
% Qfm -> -x + -----
%      xFeM1a + xMgM1a
%
% -----
di 7
x(di)      0.3481
j(di)      0.1747
f(di)      0.3345
Q(di)      0.02126
Qaf(di)    0.009286 range -0.5 0.5
Qfm(di)    -0.1135 range -0.5 0.5
%
p(jd)      2 1 0 3 1 j -1 Q -1 Qaf
           2 0 1 -1 f 0 1 1 j
p(di)      5 1 1 4 -1 j -1 Q 1 Qfm -1 x
           2 0 1 -1 j 0 1 1 Qfm
           2 0 1 -1 Q 0 1 1 Qfm
           2 0 1 1 j 0 1 1 x
           2 0 1 -1 Q 0 1 1 x
p(hed)     5 1 0 2 1 Qfm 1 x
           2 0 1 -1 j 0 1 1 Qfm
           2 0 1 -1 Q 0 1 1 Qfm
           2 0 1 -1 j 0 1 1 x
           2 0 1 -1 Q 0 1 1 x
p(ocm)     2 1 0 1 -1 Qaf
           2 0 1 1 f 0 1 1 j
p(om)      1 1 0 1 2 Q
p(cfm)     4 1 0 1 -2 Qfm
           2 0 1 2 j 0 1 1 Qfm
           2 0 1 2 Q 0 1 1 Qfm
           2 0 1 2 Q 0 1 1 x
p(jac)     1 1 0 1 2 Qaf
%
sf
W(jd,di)   26 0 0
W(jd,hed)  24 0 0
W(jd,ocm)   5 0 0
W(jd,om)    15.5 0 0
W(jd,cfm)   25.2 0 0 % 14.8 + Wjdfm - y
W(jd,jac)   3 0 0
W(di,hed)   4 0 0
W(di,ocm)   15 0 0
W(di,om)    15.75 0 0
W(di,cfm)   2 0 0
W(di,jac)   21.05 0 0
W(hed,ocm)  14 0 0
W(hed,om)   17.2 0 0 % 6.3 + Whedfom - y
W(hed,cfm)  2 0 0
W(hed,jac)  20.1 0 0
W(ocm,om)   12.8 0 0
W(ocm,cfm)  15.5 0 0
W(ocm,jac)  3 0 0
W(om,cfm)   18.45 0 0
W(om,jac)   19.3 0 0
W(cfm,jac)  21.05 0 0

```

```

% -----
12
xMgM1m     5 1 1 4 -1 j 1 Q 1 Qfm -1 x
           2 0 1 -1 j 0 1 1 Qfm
           2 0 1 -1 Q 0 1 1 x
           2 0 1 -1 Q 0 1 1 x
xFeM1m     5 1 0 2 -1 Qfm 1 x
           2 0 1 1 j 0 1 1 Qfm
           2 0 1 1 Q 0 1 1 Qfm
           2 0 1 -1 j 0 1 1 x
           2 0 1 1 Q 0 1 1 x
xFe3M1m    2 1 0 1 -1 Qaf
           2 0 1 1 f 0 1 1 j
xAlM1m     2 1 0 3 1 j -1 Q 1 Qaf
           2 0 1 -1 f 0 1 1 j
xMgM1a     5 1 1 4 -1 j -1 Q -1 Qfm -1 x
           2 0 1 1 j 0 1 1 Qfm
           2 0 1 1 Q 0 1 1 Qfm
           2 0 1 1 j 0 1 1 x
           2 0 1 1 Q 0 1 1 x
xFeM1a     5 1 0 2 1 Qfm 1 x
           2 0 1 -1 j 0 1 1 Qfm
           2 0 1 -1 Q 0 1 1 Qfm
           2 0 1 -1 j 0 1 1 x
           2 0 1 -1 Q 0 1 1 x
xFe3M1a    2 1 0 1 1 Qaf
           2 0 1 1 f 0 1 1 j
xAlM1a     2 1 0 3 1 j 1 Q -1 Qaf
           2 0 1 -1 f 0 1 1 j
xNaM2c     1 1 0 2 1 j -1 Q
xCaM2c     1 1 1 2 -1 j 1 Q
xNaM2n     1 1 0 2 1 j 1 Q
xCaM2n     1 1 1 2 -1 j -1 Q
%
jd 1 4 xAlM1m 1/2 xAlM1a 1/2 xNaM2c 1/2 xNaM2n 1/2
di 1 4 xMgM1m 1/2 xMgM1a 1/2 xCaM2c 1/2 xCaM2n 1/2
hed 1 4 xFeM1m 1/2 xFeM1a 1/2 xCaM2c 1/2 xCaM2n 1/2
ocm 1 4 xFe3M1m 1/2 xFe3M1a 1/2 xNaM2c 1/2 xNaM2n 1/2
om 1 4 xMgM1m 1/2 xAlM1a 1/2 xCaM2c 1/2 xNaM2n 1/2
make 2 jd 1/2 di 1/2
DQF -2.9 0 0
cfm 1 4 xFeM1m 1/2 xMgM1a 1/2 xCaM2c 1/2 xCaM2n 1/2
make 2 di 1/2 hed 1/2
DQF -1.5 0 0
jac 1 4 xAlM1m 1/2 xFe3M1a 1/2 xNaM2c 1/2 xNaM2n 1/2
make 2 jd 1/2 ocm 1/2
DQF -1.0 0 0
%
% -----
% Melt: White et al (2007)
%
% q(L) = Si408
% fsp(L) = KAlSi308 + NaAlSi308
% na(L) = Na/(Na + K)
% an(L) = CaAl2Si208
% ol(L) = Fe4Si208
% x(L) = Fe/(Fe + Mg)
% h2o(L) = H2O
%
liq 8
q(L) 0.1968
fsp(L) 0.3249
na(L) 0.3919
an(L) 0.01342
ol(L) 0.005044
x(L) 0.8241
h2o(L) 0.4357
%
p(q) 1 1 0 1 1 q
p(ob) 1 2 0 1 1 fsp 0 1 1 na
p(ksp) 1 2 0 1 1 fsp 1 1 -1 na
p(an) 1 1 0 1 1 an
p(sil) 1 1 1 5 -1 q -1 fsp -1 an -1 ol -1 h2o
p(fo) 1 2 0 1 1 ol 1 1 -1 x
p(fa) 1 2 0 1 1 ol 0 1 1 x
p(h2o) 1 1 0 1 1 h2o

```

```

% -----
sf
W(qab)      12 0 -0.4
W(qksp)     -2 0 -0.5
W(qan)     -10 0 0
W(qsil)     12 0 0
W(qfo)      12 0 -0.4
W(qfo)      14 0 0
W(qh2o)     15 0 0

W(abksp)    -6 0 3.0
W(aban)      0 0 0
W(absil)     12 0 0
W(abfo)     10 0 0
W(abfo)      2 0 0
W(abh2o)     1 0 -0.2

W(kspan)     0 0 -1.0
W(kspsil)    12 0 0
W(kspfo)     12 0 0
W(kspfo)     12 0 0
W(ksph2o)    11 0 -0.45

W(ansil)     0 0 0
W(anfo)      0 0 0
W(anfo)      0 0 0
W(anh2o)     9 0 -0.85

W(silfo)     12 0 0
W(silfo)     12 0 0
W(silh2o)    16 0 0

W(fofo)      18 0 0
W(foh2o)     11 0 -0.5

W(fah2o)     12 0 0
% -----
10
fac 1 1 1 1 -1 h2o
pq 1 1 0 1 1 q
xob 1 2 0 1 1 fsp 0 1 1 na
xksp 1 2 0 1 1 fsp 1 1 -1 na
pan 1 1 0 1 1 an
psil 1 1 1 5 -1 q -1 fsp -1 an -1 ol -1 h2o
pol 1 1 0 1 1 ol
xFe 1 1 0 1 1 x
xMg 1 1 1 1 -1 x
ph2o 1 1 0 1 1 h2o
% -----
ql 1 2 fac 1 pq 1
make 1 ql 4
obl 1 2 fac 1 xob 1
kspl 1 2 fac 1 xksp 1
anl 1 2 fac 1 pan 1
sill 1 2 fac 1 psil 1
make 1 sill 8/5
DQF -10 0 0
fol 1 3 fac 1 pol 1 xMg 5
make 1 fol 2
DQF -10 0 0
fal 1 3 fac 1 pol 1 xFe 5
make 1 fal 2
DQF -9 0 -1.3
h2ol 1 1 ph2o 2
% -----
% =====
% Ilmenite: MnFTO: White et al (2005)

% x(iln) = prop "ilm"
% Q(iln) = x(Fe2,A) - x(Fe2,B)
% m(iln) = x(Mn,A)

ilm 4
x(iln) 0.96
Q(iln) 0.90
m(iln) 0.02
% -----
p(o) 1 1 0 1 1 Q
p(d) 1 1 0 3 1 x -1 Q -1 m
p(h) 1 1 1 1 -1 x
p(m) 1 1 0 1 1 m
% -----
sf
W(o,d) 15.6 0 0
W(o,h) 26.6 0 0
W(o,m) 1.76 0 0
W(d,h) 11 0 0
W(d,m) 0 0 0
W(h,m) 0 0 0
% -----

```

```

% -----
7
x(Fe2,A) 1 1 0 3 1/2 x 1/2 Q -1/2 m
x(Ti,A) 1 1 0 3 1/2 x -1/2 Q -1/2 m
x(Fe3,A) 1 1 1 1 -1 x
x(Mn,A) 1 1 0 1 1 m
x(Fe2,B) 1 1 0 3 1/2 x -1/2 Q -1/2 m
x(Ti,B) 1 1 0 3 1/2 x 1/2 Q 1/2 m
x(Fe3,B) 1 1 1 1 -1 x
% -----
oiln 1 2 x(Fe2,A) 1 x(Ti,B) 1
make 1 ilm 1
DQF -13.6075 0.009426 0 % DQF = dH + R Log[4]; dH = 15.6
check 1 1 0
dilm 4 4 x(Fe2,A) 1/2 x(Ti,A) 1/2 x(Fe2,B) 1/2 x(Ti,B) 1/2
make 1 ilm 1
DQF 1.9928 -0.0021 0 % DQF = G(equil,Landau) - G(equil,SF)
check 1 0 0
hem 1 2 x(Fe3,A) 1 x(Fe3,B) 1
check 0 0 0
make 1 hem 1
pnt 1 2 x(Mn,A) 1 x(Ti,B) 1
check 0 1 1
%make 1 pnt 1
% =====
% Ilmenite: FT0: White et al (2000)

% x(ilm) = prop "ilm"
% Q(ilm) = x(Fe2,A) - x(Fe2,B)

ilm 3
x(ilm) 0.93
Q(ilm) 0.89 range -0.99 0.99
% -----
p(oilm) 1 1 0 1 1 Q
p(dilm) 1 1 0 2 1 x -1 Q
p(hem) 1 1 1 1 -1 x
% -----
sf
W(oilm,dilm) 15.6 0 0
W(oilm,hem) 26.6 0 0
W(dilm,hem) 11 0 0
% -----
6
x(Fe2,A) 1 1 0 2 1/2 x 1/2 Q
x(Ti,A) 1 1 0 2 1/2 x -1/2 Q
x(Fe3,A) 1 1 1 1 -1 x
x(Fe2,B) 1 1 0 2 1/2 x -1/2 Q
x(Ti,B) 1 1 0 2 1/2 x 1/2 Q
x(Fe3,B) 1 1 1 1 -1 x
% -----
oiln 1 2 x(Fe2,A) 1 x(Ti,B) 1
make 1 ilm 1
DQF -13.6075 0.009426 0 % DQF = dH + R Log[4]; dH = 15.6
check 1 1
dilm 4 4 x(Fe2,A) 1/2 x(Ti,A) 1/2 x(Fe2,B) 1/2 x(Ti,B) 1/2
make 1 ilm 1
DQF 1.9928 -0.0021 0 % DQF = G(equil,Landau) - G(equil,SF)
check 1 0
hem 1 2 x(Fe3,A) 1 x(Fe3,B) 1
check 0 0
make 1 hem 1
% =====
% Garnet: CFMASO: DJW asf formulation with Fe3+ and dependent alm

% prp = % Mg/(Fe+Mg+Ca+Mn)
% grs = % Ca/(Fe+Mg+Ca+Mn)
% sps = % Mn/(Fe+Mg+Ca+Mn)
% fe3 = % Fe3+/(Al + Fe3+) = Fe3+/2

g 5
prp 0.10
grs 0.10
sps 0.15
fe3 0.005
% -----
p(gr) 1 1 0 2 1 grs -1 fe3
p(alm) 1 1 1 3 -1 prp -1 grs -1 sps
p(py) 1 1 0 1 1 prp
p(spss) 1 1 0 1 1 sps
p(andr) 1 1 0 1 1 fe3
% -----
asf
W(gr,alm) 10 0 0
W(gr,py) 45 0 0
W(gr,spss) 0 0 0
W(gr,andr) 0 0 0

W(alm,py) 2.5 0 0
W(alm,spss) 0 0 0
W(alm,andr) 75 0 0

W(py,spss) 0 0 0
W(py,andr) 90 0 0

```

```

W(spss,andr) 75 0 0

gr 3 0 0
alm 1 0 0
py 1 0 0
spss 1 0 0
andr 3 0 0

% -----
6
xFeX 1 1 1 3 -1 prp -1 grs -1 sps
xMgX 1 1 0 1 1 prp
xCaX 1 1 0 1 1 grs
xMnX 1 1 0 1 1 sps
xAlY 1 1 1 1 -1 fe3
xFeY 1 1 0 1 1 fe3

% -----
gr 1 2 xCaX 3 xAlY 2
alm 1 2 xFeX 3 xAlY 2
py 1 2 xMgX 3 xAlY 2
spss 1 2 xMnX 3 xAlY 2
andr 1 2 xCaX 3 xFeY 2

% =====
% Biotite: MnKFMASHTO: White et al (2005)

% x(bi) = Fe/(Fe + Mg)
% y(bi) = Al/M1
% Q(bi) = 3(x - x(Fe,M2))
% f(bi) = x(Fe3,M1)
% t(bi) = Ti/M1
% Q(bi) = 3(x(bi) - x(Fe,M2)) = 3(x(bi) - FeM2/(FeM2 + MgM2))

bi 7
x(bi) 0.6737
y(bi) 0.5158
m(bi) 0.0500
f(bi) 0.03729
t(bi) 0.04232
Q(bi) 0.1664

% -----
p(phl) 3 1 1 4 -1 y -1 f -1 t -1 m
2 0 1 -1 x 1 4 -1 y -1 f -1 t -1 m
2 0 1 -2/3 Q 1 2 -1 m -1/2 t
p(ann) 1 2 1 2 -1 m -1/2 t 0 2 1 x -1/3 Q
p(east) 1 1 0 1 1 y
p(fbi) 1 1 0 1 1 f
p(tbi) 1 1 0 1 1 t
p(mnbi) 1 1 0 1 1 m
p(obi) 2 2 0 1 -1 x 0 3 1 y 1 f 1/2 t
2 0 1 1 Q 1 2 -1 m -1/2 t

% -----
sf
W(phl,ann) 9 0 0
W(phl,east) 10 0 0
W(phl,fbi) 0 0 0
W(phl,tbi) -10 0 0
W(phl,mnbi) 0 0 0
W(phl,obi) 3 0 0

W(ann,east) -1 0 0
W(ann,fbi) 0 0 0
W(ann,tbi) 12 0 0
W(ann,mnbi) 0 0 0
W(ann,obi) 6 0 0

W(east,fbi) 0 0 0
W(east,tbi) 0 0 0
W(east,mnbi) 0 0 0
W(east,obi) 10 0 0

W(fbi,tbi) 0 0 0
W(fbi,mnbi) 0 0 0
W(fbi,obi) 0 0 0

W(tbi,mnbi) 0 0 0
W(tbi,obi) 0 0 0

W(mnbi,obi) 0 0 0

% -----
12
x(Al,M1) 1 1 0 1 1 y
x(Fe3,M1) 1 1 0 1 1 f
x(Ti,M1) 1 1 0 1 1 t
x(Mn,M1) 1 1 0 1 1 m
x(Fe,M1) 2 2 0 1 2/3 Q 1 2 -1 m -1/2 t
2 0 1 1 x 1 4 -1 y -1 f -1 t -1 m
x(Mg,M1) 3 1 1 4 -1 y -1 f -1 t -1 m
2 0 1 -1 x 1 4 -1 y -1 f -1 t -1 m
2 0 1 -2/3 Q 1 2 -1 m -1/2 t
x(Fe,M2) 1 2 1 2 -1 m -1/2 t 0 2 1 x -1/3 Q
x(Mg,M2) 2 1 1 2 -1 m -1/2 t
2 1 2 -1 m -1/2 t 0 2 -1 x 1/3 Q
x(v,M2) 1 1 0 1 1/2 t
x(Mn,M2) 1 1 0 1 1 m
x(Al,T1) 1 1 1/2 2 1/2 y 1/2 f
x(Si,T1) 1 1 1/2 2 -1/2 y -1/2 f

phl 4 4 x(Mg,M1) 1 x(Mg,M2) 2 x(Al,T1) 1 x(Si,T1) 1
check 0 0 0 0 0 0
ann 4 4 x(Fe,M1) 1 x(Fe,M2) 2 x(Al,T1) 1 x(Si,T1) 1
check 1 0 0 0 0 0
east 1 3 x(Al,M1) 1 x(Mg,M2) 2 x(Al,T1) 2
check 0 1 0 0 0 0
fbi 1 3 x(Fe3,M1) 1 x(Mg,M2) 2 x(Al,T1) 2
make 3 east 1 cor -1/2 hem 1/2
DQF 13.45 0 0
check 0 0 0 1 0 0
tbi 16 5 x(Ti,M1) 1 x(v,M2) 1 x(Mg,M2) 1 x(Al,T1) 1 x(Si,T1) 1
make 4 phl 1 fo -1 ru 1 q 1
DQF 45 0 0
check 0 0 0 1 0 0
mnbi 4 4 x(Mn,M1) 1 x(Mn,M2) 2 x(Al,T1) 1 x(Si,T1) 1
check 0 0 1 0 0 0
obi 4 4 x(Fe,M1) 1 x(Mg,M2) 2 x(Al,T1) 1 x(Si,T1) 1
make 2 phl 2/3 ann 1/3
DQF -10.73 0 0
check 1/3 0 0 0 1

% =====
% Biotite: KFMASHTO: White et al (2007)

% x(bi) = Fe/(Fe + Mg)
% y(bi) = x(Al,M1)
% Q(bi) = 3(x - x(Fe,M2))
% f(bi) = x(Fe3,M1)
% t(bi) = x(Ti,M1)

bi 6
x(bi) 0.6372
y(bi) 0.4761
f(bi) 0.02182
t(bi) 0.1184
Q(bi) 0.1160

% -----
p(phl) 2 1 1 5 -1 f -1 t -1 x -1 y -2/3 Q
2 0 1 1 x 0 3 1 f 1 t 1 y
p(ann) 1 1 0 2 -1/3 Q 1 x
p(obi) 2 1 0 1 1 Q
2 0 1 -1 x 0 3 1 f 1 t 1 y
p(east) 1 1 0 1 1 y
p(tbi) 1 1 0 1 1 t
p(fbi) 1 1 0 1 1 f

% -----
sf
W(phl,ann) 9 0 0
W(phl,obi) 3 0 0
W(phl,east) 10 0 0
W(phl,tbi) 0 0 0
W(phl,fbi) 0 0 0
W(ann,obi) 6 0 0
W(ann,east) -1 0 0
W(ann,tbi) 10 0 0
W(ann,fbi) 8 0 0
W(obi,east) 10 0 0
W(obi,tbi) 0 0 0
W(obi,fbi) 0 0 0
W(east,tbi) 0 0 0
W(east,fbi) 0 0 0
W(tbi,fbi) 0 0 0

% -----
11
xMgM1 2 1 1 5 -1 f -1 t -1 x -1 y -2/3 Q
2 0 1 1 x 0 3 1 f 1 t 1 y
xFeM1 2 1 0 2 1 x 2/3 Q
2 0 1 -1 x 0 3 1 f 1 t 1 y
xAlM1 1 1 0 1 1 y
xFe3M1 1 1 0 1 1 f
xTiM1 1 1 0 1 1 t
xMgM2 1 1 1 2 1/3 Q -1 x
xFeM2 1 1 0 2 -1/3 Q 1 x
xSiT 1 1 1/2 2 -1/2 f -1/2 y
xAlT 1 1 1/2 2 1/2 f 1/2 y
xOHV 1 1 1 1 -1 t
xOV 1 1 0 1 1 t

% -----
phl 4 5 xMgM1 1 xMgM2 2 xSiT 1 xAlT 1 xOHV 2
check 0 0 0 0 0 0
ann 4 5 xFeM1 1 xFeM2 2 xSiT 1 xAlT 1 xOHV 2
check 1 0 0 0 0 0
DQF -3 0 0
obi 4 5 xFeM1 1 xMgM2 2 xSiT 1 xAlT 1 xOHV 2
check 1/3 0 0 0 1
make 2 phl 2/3 ann 1/3
DQF -10.73 0 0
east 1 4 xAlM1 1 xMgM2 2 xAlT 2 xOHV 2
check 0 1 0 0 0 0
tbi 4 5 xTiM1 1 xMgM2 2 xSiT 1 xAlT 1 xOV 2
check 0 0 0 1 0 0
make 3 phl 1 br -1 ru 1
DQF 78 0 0 % 60 with +/- at least 30
fbi 1 4 xFe3M1 1 xMgM2 2 xAlT 2 xOHV 2
check 0 0 1 0 0 0
make 3 east 1 cor -1/2 hem 1/2
DQF 13 0 0 % 13 with +/- at least 10

```

```

%=====
% Biotite: KFMASH: White et al (2007)

% bulk Fe/(Fe + Mg)
% x(Al,M1)
% 3(x - x(Fe,M2))

bi 4      % order-disorder model

x(bi) 0.4      % bulk Fe/(Fe + Mg)
y(bi) 0.25     % x(Al,M1)
Q(bi) 0.14     % 3(x - x(Fe,M2))

% -----

p(phl) 2 2      1 1 -1 x      1 1 -1 y
          1      0 1 -2/3 Q

p(ann) 1 1      0 2 1 x      -1/3 Q

p(east) 1 1      0 1 1 y

p(obi) 2 2      0 1 -1 x      0 1 1 y
          1      0 1 1 Q

% -----

sf

W(phl,ann) 9 0 0
W(phl,east) 10 0 0
W(phl,obi) 3 0 0
W(ann,east) -1 0 0
W(ann,obi) 6 0 0
W(east,obi) 10 0 0

% -----

7      % no of site fractions

x(Al,M1) 1 1      0 1 1 y

x(Fe,M1) 2 2      0 1 1 x      1 1 -1 y
          1      0 1 2/3 Q

x(Mg,M1) 2 2      1 1 -1 x      1 1 -1 y
          1      0 1 -2/3 Q

x(Fe,M2) 1 1      0 2 1 x      -1/3 Q

x(Mg,M2) 1 1      1 2 -1 x      1/3 Q

x(Al,T1) 1 1      1/2 1 1/2 y

x(Si,T1) 1 1      1/2 1 -1/2 y

% -----

phl 4 4      x(Mg,M1) 1      x(Mg,M2) 2      x(Al,T1) 1      x(Si,T1) 1
      check 0 0 0

ann 4 4      x(Fe,M1) 1      x(Fe,M2) 2      x(Al,T1) 1      x(Si,T1) 1
      check 1 0 0
      DQF -3 0 0

east 1 3      x(Al,M1) 1      x(Mg,M2) 2      x(Al,T1) 2
      check 0 1 0

obi 4 4      x(Fe,M1) 1      x(Mg,M2) 2      x(Al,T1) 1      x(Si,T1) 1

      make 2      phl 2/3 ann 1/3
      DQF -10.73 0 0
      check 1/3 0 1

% =====
% Fluid: CH

% x(co2) = CO2/(H2O + CO2)

fluid 2
x(CO2) 0.2

% -----

p(H2O) 1 1      1 1 -1 x
p(CO2) 1 1      0 1 1 x

% -----

ideal
% -----
2
x(H2O) 1 1      1 1 -1 x
x(CO2) 1 1      0 1 1 x

% -----

H2O 1 1      x(H2O) 1
CO2 1 1      x(CO2) 1

```

```

% =====
% Muscovite: NKFMASH: Coggon & Holland (2002)

% fe(mu) = Fe/(Fe+Mg)
% y(mu) = X(Al,M2A)
% na(mu) = X(Na,A)

mu 4
fe(mu) 0.45
y(mu) 0.96
na(mu) 0.11

% -----

p(mu) 1 1      0 2 1 y -1 na
p(po) 1 1      0 1 1 na
p(cel) 1 2      1 1 -1 fe      1 1 -1 y
p(fcel) 1 2      0 1 1 fe      1 1 -1 y

% -----

asf
W(mu,po) 10.12 0.0034 0.353
W(mu,cel) 0.00 0.0000 0.200
W(mu,fcel) 0.00 0.0000 0.200
W(po,cel) 52.00 0.0000 0.000
W(po,fcel) 52.00 0.0000 0.000
W(cel,fcel) 0.00 0.0000 0.000

mu 0.63 0.0 0.0
po 0.37 0.0 0.0
cel 0.63 0.0 0.0
fcel 0.63 0.0 0.0

% -----

7
x(K,A) 1 1      1 1 -1 na
x(Na,A) 1 1      0 1 1 na
x(Al,M2A) 1 1      0 1 1 y
x(Fe,M2A) 1 2      0 1 1 fe      1 1 -1 y
x(Mg,M2A) 1 2      1 1 -1 fe      1 1 -1 y
x(Si,T1) 1 1      1 1 -1/2 y
x(Al,T1) 1 1      0 1 1/2 y

% -----

mu 4 4      x(K,A) 1      x(Al,M2A) 1      x(Al,T1) 1      x(Si,T1) 1
      check 0 1 0
po 4 4      x(Na,A) 1      x(Al,M2A) 1      x(Al,T1) 1      x(Si,T1) 1
      check 0 1 1
cel 1 3      x(K,A) 1      x(Mg,M2A) 1      x(Si,T1) 2
      check 0 0 0
fcel 1 3      x(K,A) 1      x(Fe,M2A) 1      x(Si,T1) 2
      check 1 0 0

% =====
% Muscovite: NKFMASH: combination of Holland & Powell (1990) and Holland (2009)

% x(mu) = Fe/(Fe+Mg)
% y(mu) = x(Al,M2A) =1 in celadonite
% n(mu) Na/(Na+K)

mu 4
x(mu) 0.55
y(mu) 0.93
n(mu) 0.06

% -----

p(mu) 1 1      0 2 1 y -1 n
p(cel) 1 2      1 1 -1 x      1 1 -1 y
p(fcel) 1 2      0 1 1 x      1 1 -1 y
p(po) 1 1      0 1 1 n

% -----

sf
W(mu,cel) 0 0 0
W(mu,fcel) 0 0 0
W(mu,po) 12 0 0.4
W(cel,fcel) 0 0 0
W(cel,po) 14 0 0.2
W(fcel,po) 14 0 0.2

% -----

7
x(K,A) 1 1      1 1 -1 n
x(Na,A) 1 1      0 1 1 n
x(Al,M2A) 1 1      0 1 1 y
x(Mg,M2A) 1 2      1 1 -1 x      1 1 -1 y
x(Fe,M2A) 1 2      0 1 1 x      1 1 -1 y
x(Al,T1) 1 1      0 1 1/2 y
x(Si,T1) 1 1      1 1 -1/2 y

% -----

mu 4 4      x(K,A) 1      x(Al,M2A) 1      x(Al,T1) 1      x(Si,T1) 1
cel 1 3      x(K,A) 1      x(Mg,M2A) 1      x(Si,T1) 2
fcel 1 3      x(K,A) 1      x(Fe,M2A) 1      x(Si,T1) 2
po 4 4      x(Na,A) 1      x(Al,M2A) 1      x(Al,T1) 1      x(Si,T1) 1

```



```

% =====
% Paragonite: NKASH

%  $k(pa) = K/(Na + K)$ 

pa 2
k(pa) 0.12
% -----
p(pa) 1 1 1 1 -1 k
p(mu) 1 1 0 1 1 k
% -----
ideal
% -----
2
x(Na,A) 1 1 1 1 -1 k
x(K,A) 1 1 0 1 1 k
% -----
pa 1 1 x(Na,A) 1
mu 1 1 x(K,A) 1
DQF 14 0 0

% =====
% Pure Phases

ob and sill ru ky q H2O
*
```

Summary

