**P–T Constraints on the Metamorphic Evolution of the Paleoarchean Kromberg Type-Section, Barberton Greenstone Belt, South Africa**

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Low-grade metabasites and hydrothermally altered ultramafic rocks form most of the Paleoarchean Barberton Greenstone Belt (BGB) of South Africa. However, P–T constraints are sparse and the nature of the greenschist-facies metamorphism is poorly characterized. This study provides new P–T estimates and descriptions of the petrological characteristics of altered mafic-ultramafic rocks across the Kromberg type-section, Onverwacht Group, BGB, from both surface samples and fresh drill core. Application of a chlorite thermodynamic multi-equilibrium calculation and pseudosection modelling, in conjunction with petrographic observations, indicates a wide range in metamorphic conditions from sub-greenschist to the uppermost greenschist facies across the type-section. A central fuchsite-bearing zone containing strong mylonitic fabrics, referred to as the Kromberg Section Mylonites (KSM), records at least two metamorphic events: a high-T, low-P (420°C, 2·9 kbar) metamorphism, and a lower-T event (T = 240–350°C, P = 2·9 ± 0·15 kbar) related to retrograde metamorphism associated with extensional quartz-carbonate veins. Pillow lava metabasalts directly beneath the KSM record the same HT–LP-type metamorphic conditions, whereas metabasalts 90 m and 125 m above the KSM record temperatures of 250–340°C. Lower in the Kromberg section, metamorphic conditions vary from 215–321°C (at 2·9 kbar, mid to lower parts) to very low-grade conditions of 140–209°C in the lowermost Kromberg. Thus, an inverted metamorphic field gradient is documented beneath the KSM. Petrological analysis of the fuchsite-bearing rocks of the KSM indicates that they contain listvenite, a hydrothermal alteration product typically found in mafic-ultramafic tectonic mélangé zones of ophiolite terranes. Together with the new metamorphic P–T constraints this means that these rocks are not a product of Archean atmospheric weathering. Rather, the inverted metamorphic field gradient in the Kromberg type-section suggests that the KSM represents a Paleoarchean thrust zone. Cr-spinel compositions in the ultramafic protolith to the KSM listvenites suggest a supra-subduction zone setting. A new geodynamic model is therefore proposed in which the mafic-ultramafic oceanic rocks of the Kromberg type-section were obducted as a thrust nappe pile in a regional transpressional tectonic regime between 3227 and 3230 Ma.

**KEY WORDS:** Barberton Greenstone Belt; Archean inverted metamorphic profile; Archean listvenite; chlorite–mica thermodynamics

**INTRODUCTION**

The Paleoarchean Barberton Greenstone Belt (BGB) of South Africa (Fig. 1) comprises some of the world’s best preserved c. 3·2–3·5 Ga volcanic and sedimentary rocks.
The central parts (Fig. 1) consist mostly of low-grade mafic–ultramafic rock units preserved as the Onverwacht Group of the Barberton Supergroup (Brandl et al., 2006; Lowe & Byerly, 2007). These low-grade rocks may provide an important window into understanding Paleoarchean hydrothermal and crustal processes. However, there are no metamorphic pressure constraints and very few temperature estimates available in the literature for the pervasive greenschist-facies metamorphism in the BGB, particularly in a stratigraphic context. Furthermore, a number of different low-grade metamorphic types, for example seafloor, regional, contact and burial metamorphism, have been proposed for the Onverwacht Group; but distinguishing between these has proven to be controversial (Viljoen & Viljoen, 1969a; Anhaeusser, 1973; de Wit et al., 1987a, 1987b; Cloete, 1999; Hofmann & Harris, 2008). This is in part due to the general challenge of deriving P–T constraints from low-grade metabasalts, in comparison with metapelitic rocks (e.g. Frey et al., 1991; Spear, 1993). For example, on the basis of empirical chlorite geothermometry and bulk-rock considerations, Xie et al. (1997) proposed a regional metamorphic temperature estimate of ~320°C for the greenstone supracrustal rocks of the BGB. However, various workers have cautioned against the use of empirical chlorite geothermometry and, for a number of crystal-chemical reasons, have questioned its reliability (e.g. de Caritat et al., 1993; Jiang et al., 1994).

Our study focuses on deriving metamorphic P–T constraints for the mafic–ultramafic rocks of the Kromberg Formation (traditionally the Kromberg type-section), which occurs in the upper part of the Onverwacht Group on the SE limb of the Onverwacht Antiformal Fold (OAF; Figs 1 and 2). The central part of the Kromberg type-section includes a 150 m wide zone consisting of strongly banded Cr-mica (fuchsite)–carbonate–chlorite–quartz rocks (Fig. 2). The petrogenetic origin of these banded fuchsitic rocks has historically been at the centre of much debate, with contrasting interpretations ranging from very low-temperature Archean atmospheric weathering of komatiitic flows, to formation in early extensional ‘glide planes’ (Viljoen & Viljoen, 1969a; 1969b; Anhaeusser, 1973; de Wit, 1982, 1986a, 1986b; de Wit et al., 1982, 1983, 1987a, 1987b, 1992, 2011; Lowe & Byerly, 1986). A detailed petrological analysis of these rocks in terms of
what they are, how they formed, and their geodynamic setting is therefore important for understanding the geological evolution of the BGB.

In conjunction with petrographic analysis, we present chlorite, white mica and chromium-spinel data systematically across the metabasaltic and banded fuchsitic rocks of the Kromberg type-section. The sample set includes both surface samples and new drill core collected during the Barberton Scientific Drilling Program (BSDP; Grosch et al., 2009a, 2009b). P–T constraints are derived using a chlorite–quartz–H$_2$O thermodynamic modelling approach, a chlorite–mica–quartz–H$_2$O multi-equilibrium...
calculation, a new geothermobarometer that considers hydration in white mica and pseudosection modelling using THERMOCALC v.3.3i. The results are compared with traditional and recently developed empirical chlorite geothermometers. On the basis of the new petrological and field data, we aim to provide constraints on the geodynamic setting and metamorphic evolution of the rocks in the Kromberg type-section. In addition, we propose a new model for the petrogenetic origin of these banded fuchsitic rocks.

REGIONAL GEOLOGY
Stratigraphy and structural geology
The c. 3.5–3.1 Ga Barberton Greenstone Belt forms part of the easternmost margin of the Kaapvaal Craton as a NE-SW-trending tectonometamorphic belt between northeasternmost Mpumalanga and Swaziland (Fig. 1b). The overall structural geometry of the BGB comprises a NE-SW-trending series of antiforms and synforms, with vertical to sub-vertical bedding and cleavage surfaces (e.g. Brandl et al., 2006). The BGB consists of two components: a low-grade, volcano-sedimentary sequence of supracrustal greenstones that is juxtaposed against a high-grade, tonalite–trondhjemite–granitoid (TTG) gneiss terrane. These two crustal components are separated by highly sheared rocks along the NE-SW-trending Kromberg type-section on the SE limb of the OAF (see Fig. 1b). These shear zones with quartz-carbonate-talc-fuchsite rocks near the central and basal parts of the Kromberg type-section overlies a sedimentary succession informally referred to as the Kromberg section overlies a sedimentary succession (sensu stricto, see Grosch et al., 2011, and Figs 1 and 2).

Some workers have challenged a continuous stratigraphy for the BGB and have argued that banded fuchsite-carbonate–quartz rocks near the central and basal parts of the Kromberg type-section on the SE limb of the OAF (see Fig. 1b) are shear zones with ‘quartz-carbonate-talc-fuchsite schists/gneisses’ or ‘flaser-banded tectonites’ (Fripp et al., 1980; de Wit, 1982, 1986a, 1986b; de Wit et al., 1982, 1983, 1987a, 1987b, 1992, 2011; Armstrong et al., 1990; de Ronde & de Wit, 1994). Moreover, these shear zones have been interpreted as early extensional ‘overthrust glide planes’ that were subsequently overprinted in cataclastic D2 thrust zones along which the stratigraphy has been duplicated into ‘lithotectonic complexes’ (e.g. de Wit, 1982, 1986a, 1986b; de Wit et al., 1982, 1983, 1992, 2011). The uppermost of these proposed shear zones in the central Kromberg type-section is investigated in the present study (Fig. 2). The lowermost shear zone has been argued to represent a major thrust that separates the Kromberg type-section from the underlying Hoogogenoeg Formation (see inferred thrust in Figs 1 and 2). The timing of deformation within these proposed shear zones has not been directly dated, but some workers have argued, on the basis of field relationships, for early thrusting at 3.45 Ga during intra-oceanic ophiolite obduction (de Wit et al., 1987b, 1992, 1999).

On the other hand, proponents of a continuous stratigraphy for the BGB have argued that these banded fuchsite–carbonate–quartz rocks of the Kromberg do not represent a major D1 structural break in the Onverwacht Group stratigraphy but rather, very low-temperature (60°C) chemical weathering zones on the tops of komatiitic flows (Lowe & Byerly, 1986, 1999a, 1999b, 2007). This interpretation is similar to that of early workers, who also argued for an intact stratigraphy and that the fuchsite–carbonate–quartz rocks represent altered silicified tuffs (Anhaeusser, 1973; Viljoen & Viljoen, 1969a, 1996b).
Metamorphic constraints in the BGB

High-grade metamorphic conditions have been reported from within and in the immediate vicinity of the Stolzburg TTG granitoid–gneiss terrane (Fig. 1) with \( T = 550-700^\circ C \) and \( P = 6-15 \) kbar (e.g. Dziggel et al., 2002; Stevens et al., 2002; Diener et al., 2005; Moyen et al., 2006). These high-grade conditions are generally interpreted to be related to a 3.23 Ga (D2) major deformation phase in the BGB. These high-grade \( P-T \) conditions are contrasted with the observed low-grade conditions for the BGB greenschist sequence.

At least four types of low-grade greenschist-facies metamorphism have been proposed for the BGB: contact metamorphism related to early intrusive activity (Anhaeusser, 1969, 1973; Viljoen & Viljoen, 1969a, 1969b), regional low-grade metamorphism (Xie et al., 1997; Tice et al., 2004), seafloor hydrothermal alteration (de Wit et al., 1983, 1987a, 1987b; Lopez-Martinez et al., 1992; de Ronde & de Wit, 1994) and burial metamorphism (Cloete, 1991, 1999). Some of these proposed metamorphic types are based largely on mineral assemblages observed in the field and constraints on the timing of these proposed low-grade events are lacking or very sparse (e.g. Lopez-Martinez et al., 1992). A regional metamorphic temperature estimate of \( \sim 320^\circ C \) has been derived by applying traditional empirical chlorite thermometry relations to a range of basalts, dacites and ultramafic rocks from the upper Onverwacht Group (Cloete, 1991, 1999; Xie et al., 1997). Regional metamorphic temperatures of \( \sim 35^\circ C \) have also been argued for by Tice et al. (2004) based on Raman spectroscopy of organic carbon in various Kromberg chert horizons. Hofmann & Harris (2006) have argued on the basis of oxygen isotope thermometry of silicified basalts that seafloor metamorphism at less than \( 150^\circ C \) is recorded across the entire Onverwacht Group beneath capping sedimentary chert horizons.

LOCAL GEOLOGY

The study area is located in the Kromberg area, on the southeastern limb of the OAF (Fig. 1) where the Kromberg Formation is exposed in its type-section along the Komati River in the Songimvelo Nature Reserve. Here, the Kromberg Formation forms a synclinal structure, separated from the uppermost Hooggenoeg Formation by a thrust fault (Fig. 1; de Wit, 1986a; de Wit et al., 1987b). The most detailed lithological description for the Kromberg type-section has been provided by Furnes et al. (2011) and is illustrated in Fig. 2 (see also corresponding section line A–A′ on the map in Fig. 1). A fuchsite–carbonate–quartz zone in the central Kromberg type-section has been interpreted to represent a shear zone (Furnes et al., 2011; de Wit et al., 2011) or chemically weathered komatitites (Lowe & Byerly, 1986). In a recently proposed and controversial stratigraphy model, the former Kromberg Formation (type-section) has been subdivided into an upper Mendon Complex and a lower Kromberg Complex (de Wit et al., 2011; Furnes et al., 2011). However, it is not clear how the newly proposed Mendon Complex relates to the Mendon Formation of Lowe & Byerly (1999a, 1999b, 2007) further north in the Onverwacht Group and in the general stratigraphy of the BGB.

A strong banded fabric is observed in the fuchsite–carbonate–quartz zone that overlies proto-mylonitic to mylonitic fabrics in metapyroxenite in the central Kromberg type-section (Grosch et al., 2009a, 2009b). For simplicity, we will therefore refer to the Kromberg type-section in two parts: an upper Kromberg sequence A, above the fuchsite–carbonate–quartz zone, and a lower Kromberg sequence B. The variably sheared mafic–ultramafic rocks, including the banded fuchsitic zone, will be referred to in this study as the Kromberg Section Mylonites (KSM, Fig. 2). The KSM is \( \sim 150 \) m thick in outcrop and is here further subdivided into two parts: a lower, composite, variably mylonitic, dominantly metapyroxenitic section \( \sim 85 \) m thick (KS2), and an upper, \( 65 \) m thick banded fuchsite–quartz–carbonate unit (KSI; see Fig. 2 and also Grosch et al., 2009a, 2009b).

The upper strongly banded KSI zone consists of distinct bright emerald green, foliated mafic–ultramafic microthion bands (Cr-mica, Cr-chlorite) and reddish brown carbonate–quartz bands (Fig. 3a). In the central part of KSI, the carbonate–quartz veins appear to represent tension gashes that developed during intense shearing. Although the strong banding in the rock records intense shearing related to compressional deformation, the widespread occurrence of extensional cross-fibrous carbonate–quartz veins at high angles to the ultramafic microthion bands indicates a major component of extensional deformation (Fig. 3b). This demonstrates a major extensional phase synchronous with or post-dating a main phase of shearing. The lower part (KS2) consists of foliated, mylonitic to protomylonitic metapyroxenite with sections of deformed pillow lava screens and metagabbroic blocks (Fig. 3c). A number of anastomosing shear splays occur in this composite metapyroxenitic zone (Fig. 3d) with foliation and shearing appearing to be moderately more intense closer to the transition zone between KSI and KS2. In the BSDP (Grosch et al., 2009a, 2009b), the lower part of the Kromberg sequence A and the KSM were intercepted in the drill core KDI (Figs 1 and 2).

LITHOLOGY OF THE DRILL CORE

A summary of the lithological log data for KDI (total length 261 m, drill at 45°) is shown in Fig. 4 (also see Grosch et al., 2009a). Towards the end of the KDI, the banded mylonitic zone (KSI) occurs as a 20 m thick zone...
of carbonate–quartz bands or veins in a matrix of dark green chlorite microlithon bands with sulphides (Fig. 4). Examples of drill core sections from the upper and lower parts of KSI are shown in Fig. 3e and f. The upper banded part of the KSM displays a complex central zone (Fig. 4) with evidence for both ductile and brittle deformation (Fig. 4 and petrographic descriptions below). The ductile deformation is recorded by stylolites and partly folded dynamically recrystallized carbonate–quartz veins and early mylonitic domains. In contrast, the brittle deformation is recorded by 25 cm thick extensively fractured composite quartz veins, angular black chert breccia, dolomite and calcite breccia fragments. The outer banded flanks of KSI display dilation with extensive cross-fibrous extensional quartz–carbonate (calcite) veins between altered foliated mafic–ultramafic bands. Both the upper
Fig. 4. Drill core lithological data for KD1. The banded mylonites in KSM occur at the end of KD1 and are shown at a higher resolution in the sketch on the right. The position of the drill core samples are indicated (T-samples).
and lower banded parts of KSI incorporate ‘xenoliths’ or foreign angular blocks of highly altered gabbro (upper section) and silicified pillow lava screens (lower section). The upper and the lower parts have an overall appearance of previously deformed mylonitic rocks with in situ brecciation fabrics and with subsequent retrograde fluid infiltration recorded by extensional veins. In the deeper parts of KSI the mafic–ultramafic microlithon bands found between extensional carbonate–quartz bands preserve a foliation. In places, they record a mylonitic fabric that becomes more pronounced and is better preserved with depth.

**PETROGRAPHY**

**Banded mylonite (KSI surface samples)**

In thin section, the bright green mafic–ultramafic bands or microlithons in the banded mylonite consist of fine-grained fuchsite ± chlorite overprinting a previously formed mylonitic fabric or foliation preserving an augen-type texture between carbonate–quartz bands or veins (Fig. 5a and b). The accessory minerals include epidote, rutile, Cr-spinel, magnetite and sulphide (pyrite). The assemblage fuchsite + carbonate + quartz ± sulphide (±Cr-spinel), corresponds to that of listvenite (Halls & Zhoa, 1993, and references therein). In some samples, altered asymmetric σ-style (winged) pyroxene and possibly olivine porphyroclasts are preserved as carbonated pseudomorphs or replaced by fine-grained fuchsite and/or by chlorite (Fig. 5a and b). The textures preserved as σ-style porphyroclast pseudomorphs indicate a sinistral shear sense (Fig. 5a and b) and in places are carbonized (Fig. 5a). Extensional cross-fibrous growth textures of carbonate are observed in carbonate–quartz bands that typically follow the orientation of the foliation and augen-texture, with fine-grained carbonate forming extensional fringes along the margins (Fig. 5c). The cross-fibrous carbonate–quartz veins are in turn cross-cut by second generation carbonate–quartz veins, indicating at least two vein generations (Fig. 5d). Chlorite occurs in rare microdomains in the matrix of the mafic–ultramafic microlithon bands and is also preserved at the border of extensional carbonate–quartz veins where it displays anomalous brown birefringence (Fig. 5e). In the ultramafic microlithon bands fuchsitic mica is the dominant sheet silicate and chlorite occurs in sparse microdomains in textural contact with the mica (Fig. 5e).

**Banded mylonite (KSI drill core samples)**

On the outer flanks of KSI, quartz–carbonate extensional veins occur parallel to the mafic–ultramafic microlithon bands. An overall porphyroclastic texture is displayed in the ultramafic microlithon bands by brecciated Cr-spinel, silicified ferromagnesian grains, magnetite and pyrite with disseminated Ti-bearing phases, such as rutile. Chlorite together with disaggregated Cr-spinel and pyrite defines a foliation adjacent to the carbonate–quartz bands. Samples collected from the uppermost parts of the banded mylonite contain completely altered (silicified) pyroxene and olivine crystals with a poikilitic appearance in a matrix of fine-grained colourless to pale green Cr-chlorite. In the central part of KSI (central complex zone in Fig. 4), the mylonitic microlithon bands are disrupted and in situ brittle deformation appears to have overprinted the early mylonitic banding and dynamically recrystallized carbonate–quartz veins. In this complex central zone, the carbonate–quartz veins are in places folded and brecciated, pointing to possible brittle failure during shearing (Fig. 5f). Deformation and brecciation of formerly deformed quartz–carbonate veins resulted in angular quartz clasts with pronounced undulose extinction (Fig. 5f). The occurrence of chlorite in the microlithon bands is similar to that described in surface samples, except that chlorite is the dominant alteration phase in the microlithons and fuchsitic mica is absent (Fig. 5g). Shear fabrics and foliation development appear to intensify with depth. Intense augen-type textures similar to those preserved in fuchsite-bearing surface samples (Fig. 5a and b) are observed near the base of the KSI in the banded mylonite zone, which are now completely overprinted and chloritized. Immediately below KSI in the sheared metapyroxenitic zone, radial growth of elongate pyroxene crystals is pseudomorphed by quartz. In the lower parts of KS2 more massive, variably deformed and intensely carbonated metapyroxenite occurs with a hydrothermally altered pillow lava metabasalt screen. Petrographic analysis of samples throughout the banded KSM zone indicated no serpentine minerals.
Fig. 5. Photomicrographs of selected samples from across the Kromberg type-section. (a, b) Varibly developed mylonitic fabrics in listvenite surface samples. (a) Former ultramafic (olivine?) σ-style porphyroclast now carbonatized in a matrix of fuchsitic mica overprinting a previous ultramafic mylonitic fabric. (b) Former winged ultramafic (pyroxene?) porphyroclast indicating sinistral shear sense and completely silicified and altered to fine-grained mica with rare chlorite. The former mylonitic to protomylonitic ultramafic augen-type deformation fabric is completed overprinted by the fuchsitic mica. (c) Quartz–carbonate vein with extensional cross-fibrous growth texture and fibrous carbonate margins. (d) Two carbonate vein generations in KSM samples with cross-cutting relationship. (e) Occurrence of rare chlorite associated with abundant fuchsite and quartz in an ultramafic microlithon in listvenite sample KP1-2. (f) Brittle deformation superimposed on ductile deformation indicated by deformed, brecciated quartz–carbonate veins now cataclastic clasts with undulose extinction recording in situ brecciation during deformation (in drill core sample T32 in KSI). (g) Drill core sample T37 in KSI, indicating chlorite observed along quartz–carbonate veins and in the central parts of ultramafic microlithon bands. (h) High degree of alteration observed in P9 metabasalt samples below KSM. (i) Complete alteration observed in sample PI-155 from the mid to lower Kromberg type-section (THERMOCALC sample). (j) Low degree of alteration observed in Pl-l63 metabasalt sample from the lowermost part of the Kromberg type-section in study area.
Metabasalts of the Kromberg above and below the KSM

Although completely altered, igneous textures are still preserved in the metabasalts, and range from equigranular to porphyritic, amygdoloidal and sub-ophitic. In almost all samples, the igneous phases are completely replaced by metamorphic minerals. The metamorphic minerals are typically chlorite, actinolite, epidote, albite, quartz, titanite, sericite and magnetite. Carbonate and sulphides are also present in some of the rocks. No hornblende has been observed in any of the samples.

The proportion and occurrence of these greenschist-facies minerals varies considerably depending on the metamorphic grade. Unlike the KSM samples, no deformation fabric or preferred orientation of metamorphic phyllosilicate minerals is observed. Rather, the metamorphic minerals occur in restricted microdomains or alteration veins. Chlorite and quartz are present in all samples, typically in the groundmass and in alteration microdomains. Actinolite is texturally associated with groundmass chlorite in most samples and typically occurs as fine nematoblasts fringing the grain boundaries of relic pyroxene grains. In the most advanced stages of alteration (uralitization) of pyroxene, actinolite completely replaces clinopyroxene with a higher proportion of chlorite + epidote + quartz present. Metamorphic titanite and/or rutile occur in monomineralic aggregates or as rim overgrowths replacing minor ilmenite. Clinozoisite/epidote shows more Fe-rich rims and occurs as single, stubby, euhedral, prismatic crystals of variable size, or as granular aggregates in textural equilibrium with chlorite and quartz, or in places with actinolite, quartz and albite.

The intensity of greenschist-facies metamorphic alteration in the Kromberg metabasalts varies with distance below the shear zone. Metabasalts (P9-samples) directly below the shear zone display advanced degrees of chloritization and alteration of pyroxene to actinolite (Fig. 5h). Pyroxene grains are completely pseudomorphed by actinolite and chlorite (Fig. 5h) and there is extensive development of large metamorphic microdomains consisting of chlorite, epidote, actinolite and quartz, with the former igneous textures completely overprinted (Fig. 5h). No hornblende is present in these samples and these metabasalts lack deformation fabrics. Metamorphic
epidote forms clusters that display a radial, elongate growth morphology.

In metabasalts above the shear zone, hydrous alteration minerals are not as abundant as in the metabasalts directly beneath the shear zone (i.e. P9-30 and P9-25). Metabasalts of the mid to lower Kromberg also show chlorite + epidote + actinolite + albite present as alteration minerals, but generally a lower abundance of actinolite development in comparison with the P9 pillow basalts. Sample PI-55 is completely altered with no igneous phases present (Fig. 5i). A metabasalt (PI-161) from the lowermost Kromberg displays a very low degree of alteration, indicated by sericitization of plagioclase and a low proportion of chlorite and epidote in the groundmass. Uralitization of pyroxene is limited, with the relic pyroxene cores still being preserved (Fig. 5j).

**MINERAL CHEMISTRY**

Electron microprobe analysis of the metamorphic mineral phases chlorite and fuchsite was carried out to establish the range of mineral compositions for calculation of P–T conditions. Relic Cr-spinel was also analyzed as it is the only unaltered mineral that could be used to establish the nature of the protolith. Cr-spinel analysis has a long history of use as a petrogenetic indicator and in the interpretation of the tectonic setting of ophiolitic rocks (Irvine, 1965; Dick & Bullen, 1984; Matveev & Balhaus, 2002), as well as of Archean rock sequences (Cotterill, 1969; Chadwick & Crewe, 1986; Stowe, 1994; Byerly, 1999; Kusky & Li, 2010).

Chemical analyses of the minerals were obtained using a JEOLJXA-8600 electron microprobe, equipped with four wavelength-dispersive spectrometers (WDS), at the Department of Geological Sciences, University of Cape Town, and an ARL electron microprobe equipped with a WDS at the Department of Earth Science, University of Bergen. Running conditions were 20 nA and 15 nA beam current for each instrument respectively, and 15 kV accelerating voltage for both instruments. Counting time for the elements determined ranged from 10 to 60 s at both peak and background. Mineral abbreviations used below are after Kretz (1983) and Bucher & Frey (1994).

**Chlorite**

All of the chlorite grains analyzed in this study have a sum of Ca + Na + K below 0.08 a.p.f.u. (Table 1). Significant contamination with smectite or other clay minerals can thus be ruled out. Chemical formulae were calculated based on 14 oxygen atoms (see Table 1 legend). Site occupancy was calculated according to Vidal et al. (2005, 2006). All chlorites in the Kromberg type-section in surface and drill core samples are tri-octahedral in nature (Fig. 6a and b). Their composition can be expressed as a linear combination of clinochlore, daphnite, 14A amesite and sudoite, and therefore fulfill the compositional requirements for applying the thermodynamic model of Vidal et al. (2001, 2005, 2006) to estimate metamorphic temperature. In the banded mylonite KSM samples, the chlorite analyzed along carbonate–quartz veins has an intermediate Fe–Mg composition (Fig. 6a and b). Chlorite in the ultramafic microlithon bands away from the margins of these extensional veins is Cr-bearing, reflecting the protolith composition (Fig. 6c and d). The Cr-bearing chlorite has Cr contents in the range of 0.05–0.21 a.p.f.u. (Fig. 6c). Chlorite compositions in the massive pillow lava metabasites of Kromberg sequence A and B display generally intermediate XMg [Mg/(Mg + Fe)] values ranging between 0.38 and 0.58 (Fig. 6a). The chlorite composition in the banded mylonite zone (KSM) has generally higher XMg values, ranging between 0.52 and 0.75 (Fig. 6b). Surface sample KP1-2 from the banded mylonitic zone (KSM) contains Cr-bearing chlorite with a higher 14A amesite content in comparison with the other chlorites (Fig. 6b).

**White mica**

Mica structural formulae are presented in Table 2 and have been calculated on the basis of 11 oxygens, assuming Fe_{total} = Fe^{2+}. Mica compositions correspond to that of muscovite (Fig. 6c) with low celadonite and phengite components (Fig. 6c). The micas are also chromium bearing (fuchsitic) with Cr^3+ contents mostly ranging between 0.05 and 0.15 a.p.f.u. (Fig. 6f). The low Si content is also shown in Fig. 6f, indicating that the studied micas most probably formed under low-pressure conditions (e.g. see Velde, 1965).

**Cr-spinel**

Spinels were analysed in the banded mylonites and their structural formulae were calculated on the basis of 32 oxygens (Table 3). The compositions of the studied spinels are compared with a limited dataset available for spinel compositions in the BGB (Fig. 7a). The spinels have XCr [(Cr/(Cr + Al)] ratios between 0.69 and 0.75 a.p.f.u. and XMg [Mg/(Mg + Fe^2+)] ratios between 0.25 and 0.61. The Cr-numbers [XCr × 100 = 100 × Cr/(Cr + Al)] of the studied spinels (69–75) are different from those of spinel in basaltic komatiites and komatiites in the Onverwacht Suite (Byerly, 1999), with relatively low Mg-numbers (Fig. 7a). The low Mg-number and high Cr-numbers of the Kromberg spinels are very similar to those that occur in mantle peridotites in supra-subduction zone settings (e.g. Dick & Bullen, 1984; Kusky & Li, 2010). The Kromberg Cr-spinel compositions contrast with those in Barberton komatiites that show a boninitic and oceanic plateau association (Fig. 7a).
Table 1: Representative chlorite compositions from the Kromberg type-section

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<th>Sample:</th>
<th>Above the KSM</th>
<th>In the KSM</th>
<th>Drill core</th>
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<td>b.d.l.</td>
<td>b.d.l.</td>
</tr>
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</tr>
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<td>b.d.l.</td>
<td>b.d.l.</td>
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<td>0.09</td>
</tr>
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<td>b.d.l.</td>
<td>b.d.l.</td>
</tr>
<tr>
<td>K₂O</td>
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<td>0.03</td>
<td>0.02</td>
</tr>
<tr>
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<td>86.71</td>
<td>87.47</td>
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</table>

Structural formula calculated on the basis of 14 O

| Si          | 2.77          | 2.70       | 2.69       | 2.78       | 2.79 | 2.83 | 2.66 | 2.58 | 2.59 |
| A1⁺        | 1.23          | 1.30       | 1.31       | 1.22       | 1.21 | 1.17 | 1.34 | 1.42 | 1.41 |
| A1³⁺       | 1.02          | 1.03       | 1.04       | 1.02       | 0.97 | 0.98 | 1.62 | 1.31 | 1.47 |
| Mg          | 2.12          | 2.24       | 2.23       | 2.49       | 2.65 | 2.58 | 1.94 | 2.11 | 2.07 |
| Fe₂⁺       | 1.46          | 1.78       | 1.91       | 1.13       | 0.99 | 1.00 | 2.10 | 1.44 | 1.98 |
| Fe³⁺       | 0.97          | 0.71       | 0.63       | 0.96       | 0.99 | 1.00 | 0.21 | 0.71 | 0.24 |
| Sum other elem. | 0.06   | 0.02       | 0.02       | 0.02       | 0.02 | 0.03 | 0.05 | 0.14 | 0.08 |
| Oct. sum   | 5.62          | 5.78       | 5.82       | 5.62       | 5.62 | 5.59 | 5.88 | 5.78 | 5.88 |
| X Mg        | 0.99          | 0.56       | 0.64       | 0.69       | 0.73 | 0.72 | 0.51 | 0.48 | 0.59 |
| XFe³⁺      | 40.29         | 25.4        | 25.46      | 50.12      | 50.12| 50.12| 22.8 | 50.12| 50.12| 32.8 | 32.8 | 32.8 | 50.12| 50.12|

(continued)
### Table 1: Continued

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<td>FeO</td>
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<td>MnO</td>
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<td>0.45</td>
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<td>Na₂O</td>
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</tr>
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<td>K₂O</td>
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<tr>
<td>Total</td>
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<td>91.43</td>
</tr>
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</table>

Structural formula calculated on the basis of 14 O:

- **Si**: 2.76 2.70 2.70 2.74 2.69 2.68 2.68 2.68 2.77 2.73 2.69 2.80 2.78 2.69 2.92 2.94 2.98
- **Al³⁺**: 1.24 1.30 1.30 1.26 1.31 1.31 1.31 1.31 1.23 1.27 1.31 1.24 1.27 1.31 1.20 1.22 1.31 1.08 1.06 1.14
- **Al⁴⁺**: 1.18 1.14 1.24 1.06 1.01 0.98 1.05 1.16 1.22 1.08 1.07 1.18 1.12 1.12 1.16 1.15 0.97 1.18 0.96 1.06 0.96
- **Mg**: 2.31 2.49 2.44 1.75 1.81 1.92 2.03 2.05 2.08 1.98 2.07 2.04 1.86 1.98 1.99 2.01 2.25 2.18 2.11 2.02 2.25
- **Fe³⁺**: 1.27 1.65 1.92 1.89 2.34 2.40 1.45 2.08 2.32 1.59 1.77 2.07 1.79 1.80 2.08 1.40 1.35 1.88 1.47 1.38 1.48
- **Fe²⁺**: 0.79 0.51 0.26 0.85 0.62 0.97 0.51 0.28 0.93 0.79 0.51 0.84 0.77 0.65 0.94 0.98 0.50 0.98 1.00 0.99
- **Sum other elem.**: 0.08 0.05 0.04 0.11 0.05 0.04 0.09 0.02 0.01 0.02 0.01 0.01 0.01 0.01 0.01 0.01 0.01 0.01 0.06 0.05 0.04 0.06
- **Oct. sum**: 5.63 5.83 5.90 5.67 5.84 5.89 5.82 5.90 5.60 5.71 5.81 5.63 5.69 5.80 5.56 5.63 5.81 5.57 5.50 5.46
- **X:Mg**: 0.65 0.60 0.56 0.48 0.44 0.38 0.58 0.50 0.47 0.55 0.54 0.50 0.51 0.52 0.49 0.59 0.62 0.54 0.59 0.69 0.69
- **XFe³⁺**: 0.38 0.24 0.12 0.31 0.21 0.19 0.40 0.20 0.11 0.31 0.37 0.31 0.32 0.32 0.32 0.32 0.28 0.31 0.32 0.32 0.32

Chlorite chemical formulae were calculated on the basis of 14 oxygens and ignoring H₂O. T°C calculated at convergence of equilibria up to XFe³⁺ = 0.6. Uncertainties on calculated chlorite-quartz-H₂O temperature estimates are ±30°C. Temperature estimates using the empirical thermometers of Hillier & Velde (1991) and Inoue et al. (2009) are also reported for each chlorite analysis. b.d.l., below detection limit.
As a first-order attempt to estimate metamorphic $P^T$-conditions for the Kromberg rocks, a general pseudosection was calculated for a typical pillow lava metabasite from the mid to lower Kromberg (sample P1-155) using THERMOCALC tc331 (Powell & Holland, 1988), the internally consistent dataset of Holland & Powell (1998) and the solution models that are cited in the caption of Fig. 8. Sample P1-155 was used here as it shows a homogeneous...
Table 2: Representative white mica compositions in listvenite samples from the study area

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<tr>
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<td>0.85</td>
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<td>1.51</td>
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<td>0.74</td>
<td>0.99</td>
<td>0.72</td>
<td>0.97</td>
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<tr>
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<td>0.03</td>
<td>0.04</td>
<td>0.04</td>
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<td>0.07</td>
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<td>b.d.l.</td>
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<td>0.56</td>
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<td>0.00</td>
<td>0.12</td>
<td>0.09</td>
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<td>0.26</td>
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<td>0.13</td>
<td>0.15</td>
<td>0.07</td>
<td>0.11</td>
<td>0.14</td>
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<td>0.08</td>
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<td>0.28</td>
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<tr>
<td>Total</td>
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<td>98.47</td>
<td>98.41</td>
<td>98.25</td>
<td>98.29</td>
<td>96.51</td>
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<td>98.19</td>
<td>98.94</td>
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</table>

The chemical formulae were calculated on the basis of 11 oxygens, ignoring H₂O and with Fe₂⁺ = Fe²⁺. Mole fractions of muscovite, pyrophyllite and paragonite end-members are indicated. The temperature estimates from mica compositions are after the method of Dubaqué et al. (2010). b.d.l., below detection limit.
alteration texture (equilibrium texture) and is completely altered with the absence of igneous mineral relics (see Fig. 5i). The XRF whole-rock analysis of sample PI-155 shows that K₂O and MnO constitute less than 1% of the bulk-rock (see Fig. 8 caption). A P–T pseudosection was therefore constructed in the chemical system NCFMASHTO + qz + H₂O (Fig. 8a).

The P–T pseudosection (Fig. 8a) is characterized by a series of mineral assemblage fields with steep boundaries. Three divariant fields (ep chl di lws pmp ab ttn), (act hhb ep pl ab ttn) and (act hhb ep pl ab ttn mag) appear at 250°C, 4.5 kbar and 280°C, 5.3 kbar; 450°C, 2.5 kbar and 485°C, 5 kbar; and 450°C, 2.7 kbar and 500°C, 4.5 kbar; respectively. The disappearance of pmp and lws occurs above 250°C, 3.7 kbar and 310°C, 7 kbar. Actinolite is stable over a wide range of pressure and temperature (>270°C), indicating greenschist-facies conditions above this temperature. Hornblende becomes stable above ~450°C, 2.7 kbar. Magnetite-in reactions appear between 440°C, 2 kbar and 500°C, 4.5 kbar. The plagioclase stability field starts at 430°C, 2 kbar and 485°C, 5 kbar. Diopside is a stable phase at low-temperature conditions over a wide pressure range. The diopside stability region over these P–T conditions was defined by Banno (1998) as the pumpellyite–diopside region. No metamorphic or relic diopside is present in sample PI-155 (see Fig. 5i).

The equilibrium assemblage of sample PI-155 (act ep chl ab ttn) appears in a quadviriant field (see white P–T field in Fig. 8a) in the region between 270 and 432°C at 2 kbar, which narrows to higher pressure where it terminates at 442–488°C at 7 kbar. Within this field, the P–T conditions can be further constrained using the epidote mineral composition (see epidote composition given in Table 4) with the site distribution f(ep) = Re⁸⁺(M₁ & M₃)/(Fe³⁺(M₁ & M₃) + Al(M₁ & M₃)), Q(ep) = 1/2(Fe³⁺(M₃) – Fe³⁺(M₁)) (Patier et al., 1991; Holland & Powell, 1998), as well as four reactions between the end-members, which are cited in the figure caption. Epidote composition isolets plotted in the P–T stability field (dashed lines in Fig. 8a) for the equilibrium

### Table 3: Representative chromium spinel compositions from the banded mylonites

<table>
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<tr>
<th>Sample:</th>
<th>KP21-8</th>
<th>KP21-12</th>
<th>KP21-15</th>
<th>KP21-1</th>
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<th>T37-2</th>
<th>T37-7</th>
<th>T37-24</th>
<th>T46-4</th>
<th>T46-6</th>
<th>T46-10</th>
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<th>T44-9</th>
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<th>T44-16</th>
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</thead>
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<td>0.11</td>
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<td>0.06</td>
<td>0.49</td>
<td>0.08</td>
<td>0.16</td>
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<td>0.06</td>
<td>0.08</td>
<td>0.07</td>
<td>0.09</td>
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<td>0.41</td>
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<td>0.65</td>
<td>0.52</td>
<td>0.48</td>
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<td>0.46</td>
<td>0.95</td>
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<td>50.25</td>
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<td>49.39</td>
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<td>6.34</td>
<td>5.95</td>
<td>6.5</td>
<td>5.79</td>
<td>5.37</td>
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<td>6.26</td>
<td>6.13</td>
<td>7.43</td>
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<td>0.33</td>
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<tr>
<td>MgO</td>
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<td>10.77</td>
<td>8.64</td>
<td>9.71</td>
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The structural formulae were calculated on the basis of 32 oxygens. b.d.l., below detection limit.
assemblage intersect with the end-member reactions and indicate metamorphic conditions of $T = 310\text{--}360^\circ \text{C}$ and $P < 2.9 \ \text{kbar}$ (see dark grey area in Fig. 8a).

From the petrographic observations calcite is present in some of the pillow lava metabasalts above and below the KSM, indicating that the study area was affected by a hydrothermal fluid containing some amount of CO$_2$. Therefore as a second step, a $P-a$(CO$_2$) pseudosection was drawn at a fixed temperature ($330^\circ \text{C}$) for metabasalt sample P1-155 in the mid to lower Kromberg section using THERMOCALC tc331 (Powell & Holland, 1988) to further determine the pressure conditions and the fluid activity of CO$_2$. The $P-a$(CO$_2$) pseudosection was calculated in the system NCFMASHTO-CO$_2$ and using the bulk-rock and mineral solution models that are cited in the caption of Fig. 8. The $P-a$(CO$_2$) pseudosection (Fig. 8b) indicates that magnetite is stable at lower pressure conditions ($< 1.6 \ \text{kbar}$) over a wide range of CO$_2$ activity, whereas it becomes a stable phase at higher pressures ($1.6\text{--}6 \ \text{kbar}$) only at $a$(CO$_2$) $> 0.26$. The observed equilibrium assemblage of sample P1-155 (act ep chl ab ttn H$_2$O) is stable at lower pressure conditions ($1.1\text{--}3.1 \ \text{kbar}$) and $a$(CO$_2$) $< 0.25$. The $X_Mg$ of chlorite involved in this assemblage is calculated to decrease with increasing temperature (dashed lines in white highlighted $P-T$ field in Fig. 8b; Table 4). The observed $X_Mg$ of chlorite ($0.49\text{--}0.52$; see grey dashed line in Fig. 8b and chlorite composition in Table 4) in sample P1-155 constrains the activity of CO$_2$.
In summary, the THERMOCALC pseudosections indicate low-grade $P$–$T$ conditions of $T = 310–360 ^\circ C$ at $P < 3$ kbar, with low $a(\text{CO}_2)$ fluid conditions of less than 0.15 for the mid to lower Kromberg. The pseudosections indicate that the upper thermal stability limit of chlorite for the assemblage in PI-155 near the greenschist–amphibolite-facies transition is around 440 ± 30°C (Fig. 8b). Above this temperature, the hornblende stability field occurs at pressures above ~2.7 kbar, whereas the epidote–amphibolite-facies transition occurs below ~2.7 kbar. However, it is necessary to point out that the thermodynamic solid solution model for chlorite in THERMOCALC tC331 (Powell & Holland, 1988) cannot account for all natural chlorites, because it assumes no vacancies and full octahedral occupancy in the chlorite crystal structure, whereas low-temperature natural chlorites show octahedral vacancies (see Cathelineau & Nieva, 1985; Vidal et al., 2001, 2005, 2006). For this reason, further thermodynamic modeling and analysis is required for the altered mafic–ultramafic rocks of the Kromberg.

**Metamorphic $P$–$T$ estimates based on the composition of chlorite and mica**

$\text{Chl}–\text{Qtz}–\text{H}_2\text{O}$ equilibrium

Cathelineau & Nieva (1985) showed that the amount of Al$^{IV}$ increases and the amount of vacancies decreases in the chlorite structure with increasing temperature. Vidal et al. (2001, 2005, 2006) showed that these trends observed in chlorite compositions can be explained and modelled thermodynamically. The substitutions in chlorite can be modelled with five end-members: clinohlore (Clin); Fe- and Mg-end-members of amesite composition (Fe-ames and Mg-ames); daphnite (Daph); Mg-sudoite (Sud). Using these end-members, four reactions (two independent) can be written for the chlorite-quartz-water assemblage:

$$2\text{Clin} + 3\text{Sud} = 4\text{Mg-Ames} + 7\text{Qtz} + 4\text{H}_2\text{O} \quad (1)$$

$$4\text{Clin} + 5\text{Fe-Ames} = 4\text{Daph} + 5\text{Mg-Ames} \quad (2)$$

The pressure estimate of $T = 310–360 ^\circ C$ and $P < 3$ kbar is based on the epidote isopleths in Fig. 8a. The pressure estimate in the $P-a(\text{CO}_2)$ pseudosection is identical to that derived from the epidote isopleths in Fig. 8a.

Using these end-member reactions and indicate metamorphic conditions of $T = 310–360 ^\circ C$ and $P < 3$ kbar (dark grey area). Mineral abbreviations are after Kretz (1983). (b) $P-a(\text{CO}_2)$ pseudosection for the assemblage of PI-155 near the greenschist–amphibolite-facies transition is around 440 ± 30°C (Fig. 8b). Above this temperature, the hornblende stability field occurs at pressures above ~2.7 kbar, whereas the epidote–amphibolite-facies transition occurs below ~2.7 kbar. However, it is necessary to point out that the thermodynamic solid solution model for chlorite in THERMOCALC tC331 (Powell & Holland, 1988) cannot account for all natural chlorites, because it assumes no vacancies and full octahedral occupancy in the chlorite crystal structure, whereas low-temperature natural chlorites show octahedral vacancies (see Cathelineau & Nieva, 1985; Vidal et al., 2001, 2005, 2006). For this reason, further thermodynamic modeling and analysis is required for the altered mafic–ultramafic rocks of the Kromberg.
The temperature locations of the equilibria \(1\)–\(4\) depend on the activity of the Clin, Daph, Sud and amesite end-members as well as the activity of water. At fixed water activity and pressure, the increase in sudoite components (increase in \(\text{Si}\)) leads to a shift in the equilibria towards lower temperatures, consistent with the numerous empirical thermometers based on the amount of \(\text{Al}^{3+}\) in chlorite (e.g. Cathelineau & Nieva, 1985; Cathelineau, 1988; Hillier & Velde, 1991; Inoue et al., 2009). In theory, the pressure could be estimated from the location of the point where equilibria \(1\)–\(4\) intersect in the \(P^\cdot T\) space. However, these equilibria are near vertical, so that the pressure estimate is poor and not reliable. Vidal et al. (2005, 2006) suggested that a simultaneous estimation of \(\text{Fe}^{3+}\) in chlorite and equilibrium temperature for the \(\text{Chl}^\cdot \text{Qtz}^\cdot \text{H}_2\text{O}\) assemblage can be made using a criterion based on the convergence of equilibria \(1\)–\(4\) at a given pressure, which is achieved for a minimal \(\Delta \text{Fe}^{3+} = (\text{Fe}^{3+})_\text{Chl} - (\text{Fe}^{3+})_\text{tot}\) of chlorite. Following this approach, \(\Delta \text{Fe}^{3+}\) is increased and the structural formula of chlorite is recalculated until convergence of \(1\)–\(4\) is achieved. Although this method has been validated quantitatively by comparing estimated values of \(\text{Fe}^{3+}\) with XANES measurements (Munoz et al., 2006; Vidal et al., 2006), and also Mossbauer measurements on natural chlorites (Tarantola et al., 2009), it provides minimum \(\Delta \text{Fe}^{3+}\) and maximum \(T\) estimates only. Depending on the chlorite composition, a further increase of \(\text{Fe}^{3+}\) is sometimes possible without losing convergence of \(1\)–\(4\), which are all shifted at \(LT\).

In the present study, we determined the whole range of possible \(\text{Fe}^{3+}\) contents and temperature conditions for which convergence of \(1\)–\(4\) could be achieved for all chlorite compositions (Fig. 9a) from the Kromberg type-section. The calculation was made at a pressure of 2 kbar for all samples and with \(\Delta \text{Fe}^{3+}\) varying from 0.05 to 0.6 in 0.02 steps. The highest value possible for chlorite is \(\Delta \text{Fe}^{3+} = 0.6\) (Munoz et al., 2006; and references therein), and most estimated \(\Delta \text{Fe}^{3+}\) values in the present study are lower than this value. The presence of carbonate in the studied samples suggests that the activity of water can be less than unity, as a result of \(\text{H}_2\text{O}^\cdot \text{CO}_2\) mixing. However, as discussed above, the activity of \(\text{CO}_2\) in the fluid phase is likely to be less than 0.15, so that the activity of water should be close to unity. Moreover, given that equilibrium (2) does not involve water, varying \(\text{aH}_2\text{O}\) has no influence on its temperature location at 2 kbar and for a given \(\Delta \text{Fe}^{3+}\). For these reasons, all calculations were performed at \(\text{aH}_2\text{O} = 1\). Except for sample PI-661 (see below), convergence was considered to be achieved when the temperature difference between all equilibria \(1\)–\(4\) was less than 30°C. This value was adopted to take into account the cumulative uncertainties stemming from errors and

### Table 4: Mineral compositions and end-member activities of sample PI-155 that are used in the pseudosection modelling (Fig. 8)

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### End-member activities

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The activities of the end-members were calculated using the AX program (unpublished program). b.d.l., below detection limit.
Fig. 9. Chl–Qtz–H_{2}O equilibrium temperatures (Vidal et al., 2006) calculated for α(H_{2}O) = 1. The temperatures are plotted as a function of calculated $X_{\text{Fe}^{3+}} = \frac{\text{Fe}^{3+}}{\text{Fe}_{\text{total}}}$ (a, b), $X_{\text{Mg}} = \frac{\text{Fe}^{2+}}{\text{Fe}^{2+} + \text{Mg}}$ (c), $X_{\text{Al}_{2}} = \frac{\text{Al}_{\text{t}}}{2}$ (d), the temperatures derived with the empirical thermometer of Hillier & Velde (1991) (e) and Inoue et al. (2009) (f). In (a), the lines show for each chlorite analysis the range of possible temperature and the symbols show the average temperatures reported in (b)–(f) and in Figs 10 and 11.
The mean chlorite temperatures for all samples at 2 kbar are shown in relation to modelled XRFe3+ content, X(Mg2+)/(Mg2+ + Fe2+) and XAlIV in Fig. 9b and c. The XMg values vary systematically with modelled XRFe3+ values and temperature estimates, as higher Fe3+ values also correspond to lower Fe2+ contents and thus higher Mg-number (Fig. 9c). However, drill core T-samples in KS1 clearly show the highest XMg values, typically between 0.65 and 0.87, reflecting their ultramafic protolith, in comparison with chlorite compositions in typically between 0.45–0.50. The temperature limit of about 420°C is also made when using the T vs AlIV equations of Cathelineau & Nieva (1985) or Cathelineau (1988). In comparison, a much better agreement is observed with the temperatures derived with the empirical equation proposed by Inoue et al. (2009), which also highlights two groups of HT and LT chlorites (Fig. 9f).

The two chlorite generations, HT (375–420°C) and LT (240–350°C), occur in drill core samples (e.g. T39, T44, T37) from within the KSM (KSl) and in pillow lava samples directly below the KSM (P9-30, P9-25, P9-23). Chlorite temperature estimates are shown in relation to textural information in back-scattered electron images for drill core samples T41 and T44 (Fig. 10a and b). In sample T41, a low-temperature chlorite occurs along a quartz-carbonate vein with metamorphic temperatures between 250 and 326°C (Fig. 10a), whereas in sample T44 the HT chlorite group is recorded along a similar extensional carbonate-quartz vein with temperatures as high as 414°C (Fig. 10b). This indicates multiple fluid infiltration events in the KSM recorded by the HT and LT chlorite groups along extensional carbonate-quartz veins.

An attempt was made to reconstruct a thermal profile in the drill core across KSI from the outer banded zone into its central complex fluidized and veined zone (sample T37 to T59 in Fig. 11). In general, the two chlorite groups, HT and LT, are recorded in the outer banded zones (samples T37-T44 in upper KSI), whereas closer to the central complex zone only lower temperatures of 280–350°C are recorded (e.g. T46). Metamorphic temperature estimates could not be calculated for samples T45, T47 and T48 further down in the central complex zone as chlorite compositions do not satisfy the compositional criteria of Vidal et al. (2005, 2006). The distribution of T estimates in the drill core suggests lower T fluid infiltration from the central complex zone overprinting earlier higher temperature fluid infiltration events in the outer banded zones. Electron microprobe analyses from listvenite surface sample KPI-2 (KSM) indicate the presence of trace amounts of chlorite still preserved within mafic-ultramafic microdomains in a matrix consisting of fuchsite. Chlorite temperature estimates for sample KPI-2, for example, indicate T = 290–335°C (see Tables 1 and 5).

In contrast to the drill core samples, the metabasalts of Kromberg sequence A, P2-166 and P2-182, at distances of 90 and 125 m above the KSM, respectively, record low metamorphic conditions of T = 263–336°C and 257–334°C (Figs 9 and 11). In the mid to lower parts of Kromberg sequence B, pillow lava metabasalts generally record variable and much lower grade greenschist-facies conditions. Sample P1-155, for example, records metamorphic conditions of T = 240–30°C, and similarly sample P1-152 records chlorite temperature estimates as low as T = 215–321°C (Figs 9 and 11). A metabasalt sample in the lowermost parts of the Kromberg sequence (P1-161) is also made when using the T vs AlIV equations of Cathelineau & Nieva (1985) or Cathelineau (1988). In comparison, a much better agreement is observed with the temperatures derived with the empirical equation proposed by Inoue et al. (2009), which also highlights two groups of HT and LT chlorites (Fig. 9f).
records the lowest metamorphic grade with chlorite temperature estimates ranging between 140 and 209°C (Figs 9 and 11). Similar low temperature estimates (170–250°C) were also obtained using the empirical thermometer of Inoue et al. (2009), typically applied to very low-grade, Si-rich chlorites as in the case of sample P1-161. These low temperatures estimated for the mid to lower parts of Kromberg sequence B are consistent with petrographic observations; that is, the very low abundance of epidote (one or two grains), which typically forms well above 200°C (Cho et al., 1986; Liou et al., 1985), and very low proportion of chlorite (<2%) and the absence of actinolite in this sample. Given that no zeolite (or prehnite) could be identified in thin section, it is possible that this sample may have been affected by a low-temperature, CO2-containing fluid.

Metamorphic P–T conditions based on chlorite–mica–quartz–H2O equilibria

Chlorite–phengite–quartz–water equilibrium has been recently applied to various low-temperature metapelites (e.g. Trotet et al., 2001a, 2001b; Le Hebel et al., 2002; Parra et al., 2002b; Augier et al., 2005; Rimmelé et al., 2005; Willner, 2005; Vidal et al., 2006; Yamato et al., 2007). With five chlorite and four dioctahedral mica end-members (Mg- and Fe-celadonite, muscovite, phyllosilicate), 64 reactions (four independent) can be written in the system SiO2–Al2O3–MgO–FeO–K2O for an assemblage comprising mica, chlorite, quartz and water (Parra et al., 2002a, 2002b). As previously described, the temperature and (Fe3+)mica estimation is then adjusted to locate the Fe–Mg exchange reaction between chlorite and mica at a similar temperature. Pressure is then calculated with the 64 reactions using the INTERSX software included in the TWEEQ package (Berman, 1991). In practice, the intersection between the 64 equilibria is not always perfect and gives scattered results owing to the accumulated uncertainties in the calculation of the position of the reactions (e.g. uncertainties in the thermodynamic data, standard state properties, solid solution models).

The method described above was used to constrain average P–T conditions and to test for chemical equilibrium between chlorite and mica in the selected pairs for sample KPI-2 in the KSM (Table 5). The P–T estimates are shown by boxes with a size that corresponds to the P–T scatter and uncertainty (Fig. 12a and b). Some chlorite–mica pairs in selected microdomains show a wide scatter in equilibria and a large P–T box owing to a poor intersection between all equilibria (Fig. 12b). In this case, P–T boxes are large but low pressures of 3.5–1.3 kbar are still indicated. Other chlorite–mica pairs, particularly near the margins of carbonate–quartz veins, give a very good intersection between all 64 equilibria with P–T conditions estimated to be $T = 325 \pm 30°C$ and $P = 2.9 \pm 0.15$ kbar (Fig. 12a).

Temperature estimates based on dioctahedral mica–Qtz–H2O equilibrium

LT dioctahedral micas show a decrease in their interlayer content (IC, excluding interlayer water) and an increase in their Si content with decreasing T. This has been observed in LT and HP phengite (Bishop & Bird, 1987;
Cathelineau, 1988; Agard et al., 2001; Battaglia, 2004; Vidal & Dubacq, 2009; Dubacq et al., 2010, 2011), where the vacant sites of LT and LP illite are partially hydrated (Loucks, 1991; Drits & McCarty, 2007; Vidal & Dubacq, 2009; Vidal et al., 2010; Dubacq et al., 2010). These compositional variations can be modelled as a function of \( T \) and \( P \) using the recently proposed formalism of Dubacq et al. (2010) that involves dehydrated micas and hydrated pyrophyllite-like thermodynamic end-members. The equilibrium conditions of quartz + water + K-bearing mica of fixed 2:1 composition are represented by a divariant \( P^T \) line along which the interlayer water content varies. The \( P^T \) location of this line can be calculated from the condition of equilibrium of the following equilibria:

\[
\begin{align*}
3 \text{ Celadonite} + 2 \text{ Pyrophyllite} &\leftrightarrow 2 \text{ Muscovite} + \text{ Biotite} + 11 \text{ Quartz} + 2 \text{ H}_2\text{O} & (5) \\
\text{Pyrophyllite} \cdot \text{H}_2\text{O} &\leftrightarrow \text{Pyrophyllite} + \text{H}_2\text{O} & (6)
\end{align*}
\]

where celadonite, pyrophyllite, pyrophyllite.H\(_2\)O, muscovite and biotite are solid solution components of the mica phase. Dubacq et al. (2010) and Vidal et al. (2010) showed that the mica–quartz–water equilibrium curves could be used as a thermometer in low-pressure rocks.

In the present study, the temperature, pressure and water content for the convergence of (5)–(7) was determined for the compositions of micas occurring in KSM surface samples KPI-2, KP21 and Al7. The phengite hydration isopleths are shown in \( P^T \)-space in Fig. 13 for these three samples. For pressure conditions of around 3 kbar or less (as determined from the chl–mica multi-equilibrium
Table 5: Mineral compositions for chlorite–mica pairs in listvenite K-samples and calculated P–T conditions with uncertainties indicated.

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Structural formula calculated on the basis of 14 O

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<td>293</td>
<td>336</td>
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<tr>
<td>T(°C) Inoue (2009)</td>
<td>308</td>
<td>351</td>
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<td>T(°C) H &amp; V (1992)</td>
<td>384</td>
<td>339</td>
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<td>T(°C) no equilibrium</td>
<td>324</td>
<td>±30</td>
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<tr>
<td>P (bar)</td>
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<td>1 ±150</td>
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</table>

XFe3+ calculated after the method of Vidal et al. (2006); b.d.l., below detection limit.

calculation above), metamorphic temperatures of between 250 and 425°C are indicated by the steep isopleths for KSM listvenite surface samples KP2l, KA17 and KPI-2, respectively (Fig. 13a–c). At a pressure of 3 kbar (as determined by the chl–mica–quartz–H2O equilibrium calculation), mica compositions in microlithon bands in samples KP2l and KA17 record metamorphic temperatures of between 300 and 365°C, whereas much higher temperature metamorphic conditions are recorded by mica in sample KPI-2 up to about 420±30°C (Fig. 13d). The range and distribution in metamorphic temperature estimates derived from the Phg–Qtz–H2O equilibria are in good agreement with the temperature estimates derived from Chl–Qtz–H2O thermodynamic modelling in the drill core KSM samples.

Pressure estimates based on Si-in-mica geobarometry

The celadonite content of phengite is mainly dependent on pressure and has been proposed as a geobarometer (Ve1de, 1965, 1967; Massonne & Schreyer, 1987; Massonne & Szpurka, 1997; Parra et al., 2002a). Using the Si content in micas, applying the classic geobarometer of Ve1de (1965) and temperature estimated from the Chl–Qtz–H2O equilibrium, low pressures of less than 3 kbar are indicated for all mica compositions in the KSM banded fuchsite-bearing mylonites (see K-samples in Fig. 6f). The Si isopleths of the micas were also used to estimate the pressure according to the geobarometer of Massonne & Szpurka (1997). The intersection between the wide temperature range (up to 420±30°C) calculated from chlorite thermodynamic modelling, and Si-isopleths in mica using the geobarometer of Massonne & Szpurka (1997), indicate low pressure conditions P<3 kbar (Fig. 12c). Given that this Si-in-mica geobarometer was calibrated for high-pressure rocks, the uncertainty on this low pressure estimate is probably large. However, this low-pressure estimate is compatible with low pressure estimates derived from both the chlorite–mica multi-equilibrium calculation, and the THERMOCALC P–T grid (see Figs 12 and 8 respectively).

DISCUSSION AND INTERPRETATION

P–T constraints across the Kromberg type-section

The Kromberg type-section, including the central KSM, records a wide range in metamorphic conditions from sub-greenschist- to uppermost greenschist-facies conditions (Figs 11 and 14). The KSM records two chloride temperature populations, with an HT group of T = 422–375°C and an LT chloride group of T = 240–350°C that are related to alteration along extensional quartz–carbonate veins. The pressure recorded in the KSM is 2.9±0.15 kbar as calculated from chlorite–mica–quartz equilibria and is related to the LT event. Underformed, extensively altered pillow lavas directly below the KSM also recorded these two metamorphic events. Although these metabasalts record high temperatures near the upper greenschist–amphibolite-facies transition, no hornblende is found, but rather actinolite + epidote. This indicates that the pressure conditions were probably less than 3 kbar (see P–T grid in Fig. 8a with facies transition at around T = 440±50°C, also Maruyama et al., 1983). Metamorphic temperatures recorded in undeformed metabasalts above the shear zone in Kromberg sequence A are T = 250–340°C. These low-grade metabasalts show no deformation fabrics, and
probably also experienced pressures lower than $\sim 3$ kbar. Pillow lava metabasalts in the mid to lower Kromberg sequence record metamorphic conditions of $T = 215$–$321^\circ$C. Sample PI-155 from this part of the Kromberg sequence B (about 1.5 km beneath the KSM) shows a pressure condition of less than 2–9 kbar and low fluid $\text{CO}_2$ conditions of less than 0.15 (Fig. 8a and b). Much lower grade conditions of between 140 and 209$^\circ$C are estimated in the lowermost part of the Kromberg type-section for sample PI-161 (Figs 11 and 14).

A first-order observation is that the metamorphic field gradient is inverted, decreasing from conditions near the uppermost greenschist–amphibolite-facies boundary to metamorphic conditions as low as 140–209$^\circ$C in the lower Kromberg, over a distance of about 1.2 km. The new $P$–$T$ data across the Kromberg type-section are not compatible with current regional metamorphic profiles for the Onverwacht Group as a whole, which argue for either low-temperature ($<150^\circ$C) seafloor metamorphism for the entire Onverwacht Group (Hofmann & Harris, 2008) or a gradual increase in metamorphic grade with depth regionally in the Onverwacht Group (Cloete, 1991, 1999).

The new inverted metamorphic field gradient reported here indicates much more internal variation in metamorphic grade for the Onverwacht Group and that more than one origin for the low-grade metamorphism can be distinguished.

### Petrogenesis of the fuchsite-bearing rocks in KSM

The assemblage fuchsite–carbonate–quartz $\pm$ chlorite $\pm$ sulphide (with relic Cr-spinel) in the rocks of the KSM corresponds to that of listvenite and rocks of the listvenite series (for a review see Halls & Zhao, 1995). Listvenite is typically reported from shear zones in dismembered ophiolite terranes (e.g. Kishida & Kerrich, 1987; Ash, 2001; Johnson et al., 2004; Tsikouras et al., 2006; Nasir et al., 2007; Plessart et al., 2009) and occasionally also from greenstone belts; for example, in Burkina Faso (Béziat et al., 1998) and the Archean greenstone belts of Ontario and Quebec in the Canadian Shield (e.g. Kishida & Kerrich, 1987; Auclair et al., 1993). The observation of mylonitic...
Textures together with the $P-T$ conditions recorded in the highly altered mafic–ultramafic rocks of KSI do not support previous interpretations that they are low-temperature ($\leq 60^\circ$C) chemical weathering products of komatiitic flow tops (Lowe & Byerly, 1986) or low-temperature alteration of volcanic tuffs (Viljoen & Viljoen, 1969a, 1969b; Anhaeusser, 1973). The origin of the fuchsitic rocks as a weathering product is also highly unlikely because: (1) thermodynamic modelling indicates greenschist-facies metamorphic temperatures, up to $420 \pm 30^\circ$C, that are not compatible with low-$T$ atmospheric weathering; (2) chlorite–mica pairs are in chemical equilibrium at conditions of $T = 325 \pm 30^\circ$C and $P = 2.9 \pm 0.15$ kbar, conditions that cannot be explained by weathering.

In addition, atmospheric surface weathering of ultramafic rocks has been reported to result in Ni–Fe–Cr laterites in ophiolite terranes rather than listvenites (e.g. the Semail ophiolite, Oman; Stanger, 1985; Sharhan & Nasir, 1996; Nasir et al., 2007).

We propose a petrogenetic model in which the listvenitic rocks of KSI formed as a result of deformation and multi-stage fluid infiltration along a basal thrust zone in the Kromberg type-section. The rocks of the listvenite series develop between the serpentinite and fuchsite-bearing (true) listvenite, depending on the extent of fluid–rock interaction. For example, a typical serpentinite protolith may undergo the following alteration and mineralogical changes (after Halls & Zhao, 1995): (1)
serpentine → (2) serpentine–carbonate (breunnerite) → (3) serpentine–talc–carbonate rocks → (4) chlorite–quartz–carbonate → (5) chlorite–carbonate → (6) fuchsite listvenite (end product). As observed in the drill core and surface samples, the mylonitic rocks of KSI include a spectrum of hydrothermal alteration assemblages typical of the more advanced stages (stages 4–6) of listvenitic alteration (see carbonate and/or silica pseudomorphs after pyroxene and petrographic descriptions in Fig. 5a–f).

### Geodynamic model and metamorphic evolution

The KSM contains mafic–ultramafic rocks with abundant carbonate veining that show some similarities to rocks formed in oceanic detachment faults in modern mid-ocean ridge settings, or off-axis extensional detachment faults along transform boundaries (see Dick et al., 1991; Karson & Lawrence, 1997; Kelley et al., 2001; Karson et al., 2006). However, the P–T conditions of ~3 kbar recorded in and around the KSM are too high for a purely oceanic extensional setting (e.g. Alt & Honnorez, 1984; Bach et al., 2001). In addition, the listvenite rocks of the KSM most probably do not reflect typical oceanic mid-ocean ridge type alteration (Miyashiro et al., 1971; Alt & Honnorez, 1984; Bach et al., 2001), and differ from the talc-bearing schists observed in modern-day detachment fault scarps in oceanic core complexes (e.g. Karson et al., 2006). For these reasons, a purely oceanic (core-complex detachment model or mid-oceanic ridge) geodynamic setting for the Kromberg type-section is unlikely.

The presence of an inverted metamorphic field gradient in the Kromberg type-section is more compatible with
development of the KSM as a basal thrust zone that formed during tectonic obduction of oceanic crust in which Kromberg sequence A was thrust over Kromberg sequence B (e.g. as observed in metamorphic ‘soles’ beneath obducted ophiolites such as the Newfoundland ophiolite; Williams & Smyth, 1973; Malpas, 1979; Jamieson, 1980, 1986). On the basis of field observations, these banded fuchsitic rocks were previously interpreted to represent complex extensional ‘glide planes’ along which sub-horizontal overthrusting occurred during intra-oceanic obduction at 3455–3460 Ma (de Wit, 1982, 1986a, 1986b; de Wit et al., 1982, 1987a, 1987b, 2011). This intriguing mid-Archean ‘hovercraft tectonics’ model has, however, remained highly controversial because of a lack of detailed petrological data on the fuchsite-bearing rocks, together with the absence of field and geochronological data to support intra-oceanic shear zone development and early ophiolite obduction at 3455 Ma in the BGB (e.g. see Lowe & Byerly, 1986, 1999a, 1999b, 2007; de Wit et al., 1992; Kamo & Davis, 1994; de Wit, 1998; Hamilton, 1998; Van Kranendonk et al., 2009). In contrast, there are field data and numerous U–Pb age constraints for major deformation and crustal accretion in the BGB at a much later stage between 3227 and 3230 Ma, owing to oblique subduction, thrusting and transpressional tectonics (e.g. Kamo & Davis, 1994; Lowe & Byerly, 1999a, 1999b, 2007; Lowe et al., 1999; de Ronde & Kamo, 2000; Moyen et al., 2006; Schoene et al., 2008).

Although the timing of shearing in the KSM is not directly constrained in the present study, detrital zircon ages of 3334 ± 3 Ma from the ‘Footbridge’ chert in the upper part of Kromberg sequence A (see black chert horizon in Fig. 2 and Byerly et al., 1996) provide a maximum age constraint for the underlying Kromberg shear zone. A more likely geodynamic origin for the Kromberg type-section is thus a thrust nappe duplex that developed during a regional crustal accretion event at c. 3230 Ma. In addition to the inverted metamorphic field gradient observed, the calculated pressure estimate of 2.9 ± 0.4 kbar also supports a thrust and obduction model. The occurrence of listvenite in the KSM is similar to that reported in basal tectonic mélanges zones of dismembered Neoproterozoic ophiolite terranes of the Arabian Shield that were obducted during a transpressional tectonic regime (e.g. Johnson et al., 2004; Tsikouras et al., 2006; Nasir et al., 2007; Plissart et al., 2009). Furthermore, Cr-spinel compositions in KSM banded mylonites record high Cr-numbers, which indicate that the Cr-spinels formed from high degrees of partial melting from a depleted mantle source, typical of fore-arc tectonic settings (e.g. Dick & Bullen, 1984; Kusky & Li, 2010). The KSM Cr-spinel compositions are distinct from those of high-temperature BGB komatiites, but are very similar to peridotitic spinel compositions of Archean ophiolites and oceanic rock sequences, as well as Neoproterozoic and Cenozoic ophiolites, which are argued to have formed in a supra-subduction zone tectonic setting (Dick & Bullen, 1984; Stowe, 1994; Stern et al., 2004; Kusky & Li, 2010).

The Kromberg type-section B beneath the KSM is c. 1.4 km thick and therefore we interpret it to represent a relatively thin, dismembered sequence of oceanic, mafic–ultramafic rocks. In this context, we prefer a geodynamic model in which the Kromberg type-section represents sequential tectonic stacking of relatively thin, shear zone-bounded allochthonous slices of Paleoarchean oceanic crust and mantle obducted at c. 3230 Ma over the underlying clastic sedimentary rocks of the c. 3432 Ma Noisy Formation (see Figs 1 and 2). Thin oceanic crust in the early to mid-Archean, formed as a result of elevated mantle geothermal gradients, has previously been proposed (e.g. see Arndt, 1983). The upper parts of the KSM (KSI) display late extensional features in the form of quartz–carbonate veins, whereas KS2 deeper in the section displays only ultramafic mylonitic fabrics. This suggests that the major transpressional tectonic event at 3227–3230 Ma may have involved upper crustal level extension in the KSM with fluid infiltration, and simultaneous lower crustal shearing and compression, a tectonic mechanism that has been proposed in major transpressional settings (see Carson et al., 1997). Alternatively, thrusting during obduction may have been followed by extensional detachment or retrograde fluid infiltration subsequent to obduction of the oceanic sequence. In summary, we prefer a tectonic model for the Kromberg type-section in which multi-stage retrograde metamorphism occurred in a syn- and/or post-obduction regime that most probably involved tectonic burial of an ophiolitic-type thrust nappe pile. Extensive retrograde fluid alteration appears to have been focused locally along the KSM shear zone. Although previously the low-grade upper Onverwacht Group has been thought to have escaped the major 3.2 Ga tectono-thermal episode, we argue to the contrary that the Kromberg type-section records obduction and tectono-thermal metamorphism at that time.

CONCLUSIONS

Metamorphic constraints determined here indicate a more variable and complex metamorphic history for the Kromberg type-section rocks than previously indicated. A wide range in metamorphic conditions spanning sub-greenschist- to uppermost greenschist-facies conditions is reported. Two metamorphic events are recorded in the Kromberg Section Mylonites (KSM): (1) a high-T (420 ± 30°C), P < 3 kbar event; (2) a lower-T event (240–350°C), P = 2.9 ± 0.4 kbar, related to late-stage fluid infiltration during extension and recorded by multiple carbonate–quartz veins. The earliest metamorphic event related to deformation and mylonite development in the KSM could not be constrained in this study owing to subsequent...
retrograde overprinting by the HT and LT fluid infiltration events. Pillow lavas directly below the mylonite zone record the same metamorphic conditions. Metabasalts at a distance of 90 and 125 m above the KSM record much lower temperature conditions of 250–340°C. In the lower Kromberg sequence beneath the KSM, metamorphic conditions vary from 215–321°C (at \( P < 2.9 \text{ kbar} \)) to low-grade, sub-greenschist-facies conditions of between 140 and 209°C in the lowermost Kromberg sequence. Thermodynamical modelling of fluid conditions in the mid to lower Kromberg section indicates a low CO₂ activity of \(< 0.15\). Petrographic observations and the high-temperature upper greenschist-facies conditions determined by thermodynamical modelling on chlorite and fuchsitic micas of the KSM indicate that these rocks cannot be the result of low-temperature conditions \(\sim 60°C\) chemical weathering of komatitic flow tops under Archean atmospheric conditions as previously proposed. Rather, the field observations, metamorphic conditions, relic protolith phases and alteration mineral assemblages of the KSI mylonites indicate that they are listvenites and rocks of the listvenite series.

The presence of an inverted metamorphic field gradient (see Fig. 11) suggests thrust repetition in the Kromberg type-section. In this new geodynamic model, the KSM represents a basal thrust zone along which Kromberg sequence A was thrust over sequence B. Cr-spinel-bearing rocks in the KSM with high Cr-numbers are similar to ultramafic rocks formed in late Archean, Neoproterozoic and Cenozoic supra-subduction zone settings. We therefore prefer a tectonic model in which the oceanic rocks of the Kromberg type-section were obducted as a series of allochthonous thrust nappes over the underlying clastic sedimentary rocks of the c. 3432 Ma Noisy formation. The occurrence of listvenites in the KSM further supports a thrust model, given that listvenite formation is observed in basal thrust zones in dismembered ophiolite terranes (e.g. the Cretaceous Semail ophiolite, Oman). Although no absolute ages are available to date, thrusting and emplacement of relatively thin oceanic crust most probably occurred during a major transpressional event in the BGB, during oblique subduction and ocean closure between 3227 and 3230 Ma.

Traditionally, the low-grade volcanic rocks of the Onverwacht Suite in the BGB are interpreted to be part of a continuous volcanic stratigraphy recording regional low-grade metamorphism with progressive increase in metamorphic grade with depth. Although Paleoproterozoic low-temperature seafloor metamorphism may be preserved in some parts of the Onverwacht Group (e.g. Lopez-Martinez et al., 1992; Hofmann & Harris, 2008), the distribution of the metamorphic conditions with depth, preserved as an inverted metamorphic field gradient, in the Kromberg type-section does not support a simple regional greenschist-facies metamorphic profile for the Onverwacht Group of the BGB. Rather, the 150 m wide KSM represents a basal thrust zone that marks an important tectonothermal discontinuity in the stratigraphy of the upper Onverwacht Group along which there was extensive deformation, followed by high-\( T \), low-\( P \) metamorphism and significant fluid infiltration.

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**REFERENCES**


