1. Introduction

Fluid flow in the deeper subsurface is often largely influenced by faults, which can act as highly permeable pathways for mineralising fluids. In an extensional tectonic regime, substantial quantities of fluids are expelled during normal faulting (Muir-Wood & King, 1993). In a region of compressional deformation, arrays of extension fractures develop by hydraulic fracturing adjacent to reverse faults prior to failure (Sibson, 1994). Failure of the reverse faults allows fluids from the overpressured reservoirs to migrate upwards. Typical examples of such fault-valve action are the mesozonal gold mineralisation hosted in steep reverse faults and brittle-ductile shear zones (Sibson et al., 1988, Sibson 1994; Cox et al., 1991; Robert et al., 1995; Groves et al. 1998; McCuaig & Kerrich, 1998).

The study area is located at Bierghes, situated in the Senne valley (Fig. 1). The intrusive body found at this locality, can be classified as a dacitic metaporphy (André, 1983). In the quarry in this sill (Fig. 1), new outcrops with a multistage mineralised quartz-carbonate vein system became exposed due to recent excavations (Figs 2A and B). In the Bierghes sill, several deformation features have been identified and described (Corin & Ronchesne, 1936a, b; Corin et al., 1963; Schippers, 1979; André & Deutsch, 1985; Debacker, 2001). According to Schippers (1979) and André & Deutsch (1985) the quartz-carbonate veins (Figs 2B to D) formed during Givetian strike-slip deformation.

The southern margin of the Lower Palaeozoic of the Brabant Massif (Belgium) several polysulphide mineralisations associated with faults, have been identified. A multistage mineralised quartz-carbonate vein system occurs in the Bierghes sill. Macroscopic observations, geochemical and mineralogical investigations indicate a narrow alteration envelope around the veins, characterised by carbonitisation, sericitisation, chloritisation and silicification. The veins dominantly consist of dolomite and later generations of quartz. The opaque minerals are pyrite, galena and chalcopyrite. Vein development, alteration and the polysulphide mineralisation in the Bierghes sill can be considered contemporaneous with deformation and occurred after cleavage development. The mineralising fluids have a low-salinity H$_2$O-CO$_2$-X-NaCl-KCl composition, and a δ$^18$O, δ$^13$C and δ$^34$S isotopic composition of respectively +5.5 to +7.5 ‰ V-SMOW, –11.8 to –10.5‰ V-PDB and – 2.8 ‰ to +1.8 ‰ CDT. The mineralisation at Bierghes is interpreted as a mesozonal orogenic mineralisation, which formed by the circulation of fluids with a metamorphic signature. The location of the Bierghes mineralisation in the same high strain zone as the mineralisation at the Marcq area, indicates an extensive circulation of metamorphic fluids along shear zones at the northern margin of a supposed granitoid basement block, during the late Silurian to early Devonian deformation of the Anglo-Brabant Fold belt.

The southern margin of the Lower Palaeozoic Anglo-Brabant Fold belt is characterised by a complex and particular deformation history due to the presence of an inferred rigid granitoid basement block at depth (Everaerts et al., 1996; Sintubin, 1999; Sintubin & Everaerts, 2002). Different faults, vein systems and associated polysulphide mineralisation have been identified at this margin of the fold belt (Fig. 1). The vein system in the Marcq area (Fig. 1) has been identified as a mesozonal orogenic mineralisation formed during the tectonometamorphic deformation of the Anglo-Brabant fold belt (Piessens et al., 2002). The migration of metamorphic fluids has been dated at 416 ± 3 Ma (Dewaele et al., 2002). However, the faults at the southern margin of the Anglo-Brabant Fold belt have been active during different geological time periods (Bergerat & Vandycke, 1994; Vandycke et al., 1991; Vandycke, 2002), even up to today (Oudenaarde earthquake 1938).
The aim of this study is to characterise and to determine the origin of the mineralising fluid in the Bierghes sill. Since this magmatic body falls in the high-strain zone north of the supposed granitoid crustal block at depth (Sintubin & Everaerts, 2002), the fluid characteristics are compared to those of the polysulphide mineralisation in the Marcq area. This allows us to obtain a more general view of fluid migration along the southern margin of the Anglo-Brabant Fold belt.

2. Geological setting

The Lower Palaeozoic rocks in central Belgium, northern France and the Netherlands form the southeastern part of the Anglo-Brabant Fold belt (Van Grootel et al., 1997; Verniers et al., 2002). This Caledonian slate belt is moulded around the Cadomian Midlands microcraton, forming the core of the East Avalonia microcontinent (Blundell et al., 1992).

This Anglo-Brabant Fold belt (Fig. 1) consists of folded, faulted and weakly metamorphosed rocks. Its sedimentary and tectonometamorphic evolution can be divided in three megasequences (Verniers et al., 2002), related to the palaeogeographic evolution of the East Avalonian microcontinent. During the first megasequence (Cambrian until early Ordovician), Avalonia is still connected to Gondwana. More specific, the Anglo-Brabant area has been identified as an extensional rift basin at the margin of Gondwana (e.g. Nance et al., 2002; Sintubin & Everaerts, 2002). During the Middle Ordovician, Avalonia disconnected from Gondwana and migrated northward as an isolated continental fragment (second megasequence). Finally, Avalonia encountered Baltica and collided with Laurentia during the Middle Silurian (Verniers et al., 2002). The oldest sediments in the Anglo-Brabant fold belt are of Early Cambrian age and occur in the central part of this fold belt (Fig. 1). Middle Cambrian to Silurian rocks flank this Cambrian core to the north and south. The Anglo-Brabant Fold belt in Belgium is characterised by a dominance of turbidity sediments. These sediments consist of claystones and siltstones, which were subjected to lower diagenetic up to greenschist metamorphic conditions (Van Grootel et al., 1997).

Several volcanic and intrusive rocks are distinguished along the southern margin of the Anglo-Brabant Fold belt (Fig. 1). The rocks have a dominantly dacitic composition, with some rhyolites and rare andesitic and basaltic rocks (Hertogen & Verhaeren, 1999; Verniers et al., 2002). The intrusive rocks of the Bierghes sill can be classified as dacitic. The rocks found at the Quenast plug have a similar composition, while the intrusive rocks of the Lessines sill are more andesitic (André, 1983). Based on composition and

Fig. 1: Subcrop map of the Brabant Massif (after De Vos et al., 1993), with indication of the study area. B = Bierghes, C = Court St.-Etienne, D = Dongelberg, H = Hulste, I = Idegem, K = Kuurne, L = Lessines, M = Marcq area, S = Schendelbeke and Q = Quenast.
distribution, the intrusive rocks are related to volcanic rocks, which are supposed to have formed as subaerial and submarine ignimbrites (Hertogen & Verhaeren, 1999). The magmatism has been related to a short-lived subduction of oceanic lithosphere from probably a small oceanic basin within the Avalonia microcontinent (Debacker, 2001; Verniers et al., 2002) beneath the Avalonian microcontinent during Early Ashgill, i.e. the end of the second megasequence (André et al, 1986; Van Grootel et al., 1997; Verniers et al., 2002).

Fig. 2: A. Schematic plan of the quarry at Bierghes (after Schippers, 1979). Shaded area indicates zone with carbonate-quartz veins. Study area is indicated with X. B. Photograph of study area, dots indicate mineralised vein. C & D. Example of carbonate-quartz veins
Deformed Lower Palaeozoic basement metasediments are discordantly concealed by Givetian sediments, while the latest deformed sediments are of Ludlovian age (Debacker, 2001, Verniers et al., 2002). At present, there is only evidence for one major tectonometamorphic deformation of the Anglo-Brabant fold belt in Belgium, occurring during the Late Silurian to early Middle Devonian (André et al., 1981; Van Grootel et al., 1997; Debacker, 2001; Verniers et al., 2002; Dewaele et al., 2002). Since the accretion of Eastern Avalonia to Baltica occurred during the Late Ordovician to Early Silurian (Cocks et al., 1997; Verniers et al., 2002), the deformation in the Anglo-Brabant Fold belt can be considered as the result of a post-accretionary intracontinental accommodation within the East Avalonia microcontinent due to the closure of the Rheic Ocean (Pickering & Smith, 1995; Cocks et al., 1997; Rey et al., 1997). The Anglo-Brabant Fold belt in Belgium is recently considered as a compressional wedge system, where material is pushed out during the tectonometamorphic deformation phase (Sintubin & Everaerts, 2002). At the southwestern margin of this belt a low-density body is identified at depth, recently interpreted to be a Precambrian crustal block of granitoid composition (Everaerts et al., 1996; Sintubin & Everaerts, 2002). This fault-bounded basement block controlled the deformation and a NW-SE trending high-strain zone developed along its northeastern edge (Sintubin, 1999; Sintubin & Everaerts, 2002). The low-angle reverse shear zone, identified in the Marcq area (Fig. 1), can be considered as the result of this thrusting (Debacker, 1999).

The study area is the Bierghes sill (Figs 1 and 2). According to André (1983), this intrusion crystallised at the same time, from the same magma as the Quenast plug, which has been dated by U-Pb at 433 +/- 10 Ma (André & Deutsch, 1984). The Bierghes sill is however characterised by a widespread set of fracture/foliation planes with different orientation. Detailed observations in the quarry made it possible to distinguish a cleavage with an orientation of N20W41N and a fracture directed N60W50N (Fig. 3). The fracture planes seem to cut the foliation (Fig. 3). Fractures and foliation are better developed in the central part of the quarry, where also the dolomite-quartz veins are present (Corin et al., 1963; Schippers, 1979). These veins are clearly oriented parallel to the fracture planes (Fig. 2B). According to Legrand (1968), this zone forms part of the Oudenaarde-Bierghes fault zone, which has been identified to form part of the Nieuwpoort-Asquempont fault zone (De Vos et al. 1993; Sintubin, 1999; Debacker, 2001). André & Deutsch (1985) interpreted the veins as syn-tectonic; they formed during strike-slip faulting in the late Givetian (~373 ± 11 Ma).

3. Methodology

Different samples have been taken from the quartz-carbonate vein system in the Bierghes sill for petrographic, mineralogical and geochemical investigation. The combination of different geochemical techniques allows to study the fluid-rock interaction. Fluid inclusion analysis has been carried out to identify fluid remnants trapped in mineral phases. Stable isotope composition was investigated to determine the origin of the fluids.

3.1 Mineral chemistry

Thin sections were prepared to determine the mineralogy of the vein minerals, the alteration zone and the host-rock. Examination of polished sections led to the identification of the ore minerals. The mineralogy of the fine-grained crystals (e.g. sericite, chlorite and other phyllosilicates) in the altered and unaltered rocks was determined by X-ray diffraction. The measurements were conducted on a Phillips PW3710 diffractometer with Cu-tube. The major element geochemistry of the host rock, and of the different alteration zones and vein types, has been measured by Atomic Absorption spectrometry (AAS).

Chlorite geothermometry was carried out on two polished thin sections with chlorite, present in two different stages of vein development. The Si, Al,
Fe, Mn, Mg, Ti content of the chlorites was measured with a Cameca France SX 50 electron microprobe using the standards fayalite (Si), sapphire (Al), hematite (Fe), olivine (Mg), rhodonite (Mn) and rutile (Ti). From the Al in the tetrahedral position and eventually the Fe/(Fe+Mg) ratio, the formation temperature of the chlorites can be calculated with the empirical models of Cathelineau (1988), Hillier & Velde (1991) and Jowett (1991). The thermodynamic model of Walshe (1986) uses the complete chemical composition of the chlorites to calculate their precipitation temperature at given pressure.

3.2 Fluid inclusion microthermometry

Doubly polished sections (300 µm thick) were prepared for the study of fluid inclusions in the gangue minerals. A detailed description of the sample preparation technique and of the measurement procedure has been given by Muchez et al. (1994). Microthermometric analyses of fluid inclusions were carried out on a Fluid Inc. adapted U.S.G.S. gas-flow heating-freezing system, mounted on an Olympus BX60 microscope. Reproducibility was within 0.2°C for the melting temperatures and < 3°C for the total homogenisation temperature (Thom).

Since no Raman analysis have been carried out to determine the additional gas components, the maximum salinity and density of the fluid inclusions with an H2O(-CO2(X))-NaCl-KCl composition, with X the additional gas components, has been estimated with the program Flincor of Brown (1989), using the Equation Of State (EOS) of Brown & Lamb (1989) for an H2O-CO2-NaCl system.

3.3 Stable isotope chemistry

The isotopic composition of the dolomite, present in the alteration zones and veins (δ18O and δ13C), and of sulphides (δ34S) was measured to determine the origin of the mineralising fluids. The different dolomite generations and sulphides were selected based on their occurrence in the paragenetic sequence and sampled by diamond micro-drilling. Carbonate powders were reacted with 100% phosphoric acid (density > 1.9 g/cm3, Wachter and Hayes, 1985) at 75°C in an online carbonate preparation line (Carbo-Kiel – single sample acid bath) connected to a Finnigan Mat 252 mass spectrometer. All values are reported in per mil relative to V-PDB.

The oxygen isotopic composition of the dolomites was corrected using the fractionation factors given by Rosenbaum & Sheppard (1986). Reproducibility is better than 0.04‰. Oxygen and carbon stable isotopic analyses were carried out at the University of Erlangen, Germany. δ34S analysis of sulphides were analysed by standard techniques at the SURRC centre in East Kilbride, Scotland. The SO2 gas was liberated by combusting the sulphides with excess CuO at 1075°C, in vacuo (Robinson and Kusakabe, 1975). Liberated gases were analysed on a VG Isotech SIRA II mass spectrometer. Reproducibility is better than 0.2‰ (1σ). Data are reported in δ34S notation as per mil (‰) variations from the Canyon Diablo Troilite (CDT) standard.

4. Petrography and mineral paragenesis

The primary -unaltered- magmatic rocks dominantly consist of feldspar (especially albite), amphibole, pyroxene and quartz. Ilmenite, apatite, zircon and magnetite are present as accessory minerals. Epidote and chlorite are the most important secondary minerals (e.g. André, 1983). The original magmatic texture has however significantly been altered by relatively low-temperature secondary processes. Hydrothermal activity linked to the injection of dacitic magmas within water-saturated Ordovician sediments (André & Deutsch, 1986), metamorphism associated with the Early Devonian deformation phase (André et al, 1981) and fluid flow causing a Late Givetian to Late Devonian Sr resetting (André & Deutsch, 1985) could have resulted in an intense sericitic alteration of feldspars, epidotisation, chloritisation and saussuritisation.

Since the orientation of the mineralised veins is identical to that of the fractures, their development is thought to be related to this fracturing. The isolated veins seem to form lenses, which have a thickness up to a metre. The veins are found in a NW-SE oriented zone (N60W50N) transecting the quarry (Fig. 2A). The quartz-dolomite veins are associated with an alteration characterised by dolomitisation, silicification and sericitisation. A polysulphide mineralisation occurs in the veins.

A detailed scheme of the mineral paragenesis of the multistage mineralised quartz-carbonate vein system is presented in Fig. 4, indicating five main stages in its evolution. In an initial step, bands of green-coloured altered rocks are formed (=stage 1). This alteration stage consists of dolomite, quartz, muscovite and chlorite (Plate 1A). This stage is followed by the formation of red-coloured altered rocks, consisting of dolomite, quartz, muscovite and hematite (stage 2). The latter mineral is responsible for the red colouring of the rocks (Plate 1B). The alteration is followed by vein generation 1 (=stage 3), which dominantly consists of dolomite (Plate 1C). Traces of muscovite are present, but chloride is absent. Some quartz occurs as a late vein-filling phase. This vein generation also contains pyrite, chalcopyrite and galena (Plate 1D). The pyrites are often characterised by cataclastic deformation. The next vein generation 2 consists of dolomite and quartz (=stage 4). The last generation 3 formed perpendicular to the former vein generation and only contains quartz (=stage 5).

The alteration stages are characterised by carbonitisation, chloritisation, silicification and sericitisation compared to the host rocks. This is...
illustrated by the geochemical analysis (Table 1), which shows a higher concentration of MnO, K2O and CaO in the green-coloured alteration, compared to the host rocks. The Al2O3, MgO and TiO2 content remains similar, while there is a decrease of Na2O, Fe2O3 and SiO2. Additionally, the red-coloured alteration is characterised by the highest amount of Fe2O3, MnO, MgO and CaO; but the lowest concentration of SiO2, TiO2, Al2O3, K2O and Na2O. While sericite is omnipresent in the alteration stages and in the first vein generation, chlorite can only be identified in the green-coloured alteration. This observation has been confirmed by X-ray diffraction.

Some remains of strongly altered feldspars can still be identified in the green-coloured alteration.

5. Microthermometry

Doubly-polished sections have been investigated from vein generation 1, 2 and 3. The fluid inclusions in vein generation 1 were studied in dolomite, while fluid inclusions in vein generation 2 and 3 were measured in quartz. Due to the complex, progressive deformation history of the earliest generations of veins,
it was impossible to determine the chronology of fluid inclusions in vein generations 1 and 2. Only in vein generation 3, it was possible to distinguish between primary and secondary inclusions.

The fluid inclusions in vein generation 1 are two-phase (L + V) aqueous-gaseous inclusions at room temperature (Plate 1E). Some isolated three phase inclusions have been found. The melting temperature of CO$_2$ (Tm$_{cis}$) is only measurable in the largest inclusions, since the formation of clathrate consumed the majority of CO$_2$. Tm$_{cis}$ is -57.5°C (Fig. 5A). Since this temperature is below -56.6 °C, it indicates the presence of one or several gasses, CH$_4$, N$_2$, H$_2$S, ..., in addition to CO$_2$. In a few inclusions the eutectic melting temperature of the aqueous phase could be measured and is around -22°C, indicative for the presence of KCl in addition to NaCl (Shepherd et al., 1985). The ice melting temperatures (Tm$_{ice}$) vary between -3.8°C and -5.3°C (Fig. 5B). Clathrate dissociation temperatures (Tm$_{clath}$) occur between 7.6°C and 8.4°C (Fig. 5C). The homogenisation of the CO$_2$ (Tm$_{hom}$) can be measured in few inclusions and varies between 24.8 and 28.0 °C. Decrepitation occurs at a temperature of 240°C, and total homogenisation temperatures vary between 237 °C and 363°C (Fig. 5D). The two-phase aqueous-gaseous inclusions have an H$_2$O-CO$_2$-(X)-NaCl-KCl composition, with X the additional gas components. The calculated maximum salinity and density for all measured inclusions respectively vary between 6.0 and 8.2 eq. wt% NaCl, and 0.67 and 0.88 g/cm$^3$.

<table>
<thead>
<tr>
<th>Sample</th>
<th>SiO$_2$</th>
<th>TiO$_2$</th>
<th>Al$_2$O$_3$</th>
<th>FeO$_3$</th>
<th>MnO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na$_2$O</th>
<th>K$_2$O</th>
<th>LOI</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>33m from vein</td>
<td>62.67</td>
<td>0.71</td>
<td>14.53</td>
<td>6.93</td>
<td>0.03</td>
<td>5.27</td>
<td>0.52</td>
<td>1.90</td>
<td>2.91</td>
<td>4.05</td>
<td>98.97</td>
</tr>
<tr>
<td>22m from vein</td>
<td>64.32</td>
<td>0.72</td>
<td>14.33</td>
<td>6.09</td>
<td>0.03</td>
<td>3.52</td>
<td>1.36</td>
<td>3.06</td>
<td>2.74</td>
<td>3.64</td>
<td>99.36</td>
</tr>
<tr>
<td>11m from vein</td>
<td>62.66</td>
<td>0.71</td>
<td>14.68</td>
<td>5.85</td>
<td>0.08</td>
<td>4.10</td>
<td>1.34</td>
<td>2.36</td>
<td>3.02</td>
<td>4.60</td>
<td>98.92</td>
</tr>
<tr>
<td>11m from vein</td>
<td>63.10</td>
<td>0.70</td>
<td>14.13</td>
<td>5.72</td>
<td>0.02</td>
<td>4.01</td>
<td>1.19</td>
<td>2.38</td>
<td>2.99</td>
<td>5.00</td>
<td>98.71</td>
</tr>
<tr>
<td>wallrock adjacent vein</td>
<td>64.47</td>
<td>0.60</td>
<td>13.40</td>
<td>5.29</td>
<td>0.02</td>
<td>5.65</td>
<td>0.93</td>
<td>0.84</td>
<td>3.25</td>
<td>5.27</td>
<td>99.20</td>
</tr>
<tr>
<td>green-coloured alteration</td>
<td>50.55</td>
<td>0.67</td>
<td>14.42</td>
<td>5.14</td>
<td>0.17</td>
<td>5.38</td>
<td>6.34</td>
<td>0.25</td>
<td>4.82</td>
<td>12.14</td>
<td>99.59</td>
</tr>
<tr>
<td>red-coloured alteration</td>
<td>29.75</td>
<td>0.07</td>
<td>3.48</td>
<td>7.44</td>
<td>1.37</td>
<td>9.37</td>
<td>16.26</td>
<td>0.01</td>
<td>1.22</td>
<td>31.30</td>
<td>99.68</td>
</tr>
<tr>
<td>red-coloured alteration</td>
<td>28.02</td>
<td>0.05</td>
<td>3.68</td>
<td>9.25</td>
<td>1.35</td>
<td>9.35</td>
<td>15.90</td>
<td>0.01</td>
<td>1.20</td>
<td>31.30</td>
<td>99.53</td>
</tr>
</tbody>
</table>

**Table 1.** Major element geochemistry of the host rock, and of the different alteration zones at Bierghes.

![Fig. 5. Microthermometric data of aqueous-gaseous and aqueous fluid inclusions in dolomite or quartz associated with the Bierghes vein system. VG1 = vein generation 1, VG2 = vein generation 2, VG3(prim) = primary inclusions in vein generation 3 and VG3(sec) = secondary inclusions in vein generation 3. Melting temperature of CO$_2$ (Tm$_{cis}$); Melting temperature of ice (Tm$_{ice}$); Melting temperature of clathrate (Tm$_{clath}$); Total homogenisation temperature (Tm$_{hom}$).](image-url)
PLATE 1

A. Photograph of green-coloured alteration. Note the preferred orientation of the muscovite fibers. Muscovite (Musc); dolomite (Dol). Crossed nicols. Scale bar is 135 µm.

B. Photomicrograph of red-coloured alteration. Note presence of muscovite (Musc); hematite (Hem); dolomite (Dol). Crossed nicols. Scale bar is 65 µm.

C. Photomicrograph of vein generation 1, dominantly consisting of dolomite. Crossed nicols. Scale bar is 65 µm.

D. Photomicrograph showing detail of sulphide mineralisation at Bierges. Cp = chalcopyrite, Py = pyrite. Scale bar is 65 µm.

E. Detail of a trail showing two-phase aqueous fluid inclusions in dolomite. Scale bar is 35 µm.

F. Detail of the trail showing two-phase aqueous-gaseous fluid inclusions in vein generation 3. Scale bar is 35 µm.

G. Photomicrograph of chlorite type 1, occurring in the green-coloured alteration zone. Crossed nicols. Scale bar is 65 µm.

H. Photomicrograph of chlorite type 2, at the contact of the green-coloured and red-coloured alteration zones. Crossed nicols. Scale bar is 65 µm.
The fluid inclusions of vein generation 2 are two-phase (L + V) aqueous-gaseous inclusions at room temperature in quartz. Also, some isolated three phase inclusions have been found. $T_m^{\text{CO}_2}$ was measurable in a limited number of inclusions and varies between $-56.6$ °C and $-57.9$ °C (Fig. 5A). Three temperatures are below $-56.6$ °C, indicating the presence of an extra gas phase in addition to CO$_2$. The eutectic melting temperature of the aqueous phase is ~ $-22$ °C, indicative for the presence of KCl in addition to NaCl in the aqueous phase (Shepherd et al., 1985). $T_m^{\text{Ice}}$ varies between $-2.5$° and $-5.5$°C (Fig. 5B). $T_m^{\text{Clath}}$ values are between $5.8$° and $7.3$°C (Fig. 5C). $T_{\text{Tot}}$ of the inclusions occurs between $235$ °C and $265$ °C (Fig. 5D). Like the inclusions in the first vein generation, the fluid inclusions in the second vein generation have an H$_2$O-CO$_2$-(X)-NaCl-KCl composition. The calculated maximum salinity and density for all measured inclusions respectively vary between 4.5 and 8.5 eq. wt% NaCl, and between 0.83 and 0.88 g/cm$^3$.

The fluid inclusions of vein generation 3 are two-phase (L + V) aqueous-gaseous inclusions at room temperature (Plate 1F). In this vein generation, a chronology of different orientations of fluids was observable. However, only a small decrease in homogenisation temperatures can be observed in the secondary fluid inclusions (Fig. 5D). No traces of CO$_2$ or another gas phase have been found. The temperature of eutectic melting of the aqueous phase could be measured and lies around ~ $-21.0$ °C, indicative for the H$_2$O-NaCl system. $T_m^{\text{water}}$ varies between $-3.6$° and $-4.4$°C (Fig. 5B). $T_{\text{tot}}$ values are between $158$ °C and $235$ °C, with an average of 200 °C (Fig. 5D). The salinity and density has been calculated for all measured inclusions and the values vary respectively between 5.7 and 6.9 eq. wt% NaCl, and between 0.88 and 0.95 g/cm$^3$.

### Table 2A.
Geochemical composition (weight%) of the chlorites found in the green alteration and the border of the green-red alteration.

<table>
<thead>
<tr>
<th>Analysis</th>
<th>SiO$_2$</th>
<th>TiO$_2$</th>
<th>Al$_2$O$_3$</th>
<th>FeO</th>
<th>MnO</th>
<th>MgO</th>
<th>H$_2$O</th>
</tr>
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<tbody>
<tr>
<td><strong>Green alteration</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>25.31</td>
<td>0.087</td>
<td>18.59</td>
<td>20.56</td>
<td>0.164</td>
<td>17.92</td>
<td>17.77</td>
</tr>
<tr>
<td>2</td>
<td>26.00</td>
<td>0.050</td>
<td>19.18</td>
<td>20.99</td>
<td>bd</td>
<td>18.17</td>
<td>15.60</td>
</tr>
<tr>
<td>3</td>
<td>26.44</td>
<td>0.058</td>
<td>19.20</td>
<td>20.92</td>
<td>0.097</td>
<td>18.10</td>
<td>15.18</td>
</tr>
<tr>
<td>4</td>
<td>26.13</td>
<td>0.045</td>
<td>19.20</td>
<td>20.99</td>
<td>0.134</td>
<td>18.10</td>
<td>15.40</td>
</tr>
<tr>
<td>5</td>
<td>25.91</td>
<td>bd</td>
<td>19.20</td>
<td>21.12</td>
<td>0.115</td>
<td>17.80</td>
<td>15.84</td>
</tr>
<tr>
<td>6</td>
<td>25.91</td>
<td>bd</td>
<td>19.22</td>
<td>21.00</td>
<td>0.125</td>
<td>18.18</td>
<td>15.53</td>
</tr>
</tbody>
</table>

| **Border green-red alteration** | | | | | | | |
| 1 | 28.87 | 0.063 | 19.83 | 12.51 | bd | 24.64 | 14.10 |
| 2 | 28.18 | 0.052 | 19.95 | 12.73 | 0.096 | 24.55 | 14.45 |
| 3 | 27.62 | bd | 20.28 | 13.57 | bd | 24.08 | 14.44 |
| 4 | 28.23 | 0.315 | 23.91 | 8.32 | bd | 14.02 | 15.21 |
| 5 | 27.99 | bd | 19.94 | 13.11 | 0.093 | 24.09 | 14.74 |

<table>
<thead>
<tr>
<th>Analysis</th>
<th>Si</th>
<th>Ti</th>
<th>Al(tetr)</th>
<th>Al(oct)</th>
<th>Fe</th>
<th>Mn</th>
<th>Mg</th>
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</thead>
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<tr>
<td><strong>Green alteration</strong></td>
<td></td>
<td></td>
<td></td>
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<td></td>
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<tr>
<td>1</td>
<td>2.79</td>
<td>0.0075</td>
<td>1.21</td>
<td>1.21</td>
<td>1.89</td>
<td>0.014</td>
<td>2.88</td>
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<tr>
<td>2</td>
<td>2.79</td>
<td>0.0040</td>
<td>1.21</td>
<td>1.21</td>
<td>1.88</td>
<td>bd</td>
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<tr>
<td>3</td>
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<td>0.0048</td>
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<td>1.22</td>
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<td>0.012</td>
<td>2.88</td>
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<tr>
<td>4</td>
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<td>0.0032</td>
<td>1.20</td>
<td>1.22</td>
<td>1.87</td>
<td>0.012</td>
<td>2.88</td>
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<tr>
<td>5</td>
<td>2.78</td>
<td>bd</td>
<td>1.21</td>
<td>1.23</td>
<td>1.90</td>
<td>0.010</td>
<td>2.85</td>
</tr>
<tr>
<td>6</td>
<td>2.78</td>
<td>bd</td>
<td>1.21</td>
<td>1.21</td>
<td>1.88</td>
<td>0.011</td>
<td>2.90</td>
</tr>
</tbody>
</table>

| **Border green-red alteration** | | | | | | | |
| 1 | 2.88 | 0.0450 | 1.11 | 1.23 | 1.04 | bd | 3.37 |
| 2 | 2.83 | 0.0380 | 1.16 | 1.21 | 1.07 | 0.077 | 3.68 |
| 3 | 2.79 | bd | 1.20 | 1.21 | 1.14 | bd | 3.63 |
| 4 | 3.65 | 0.0220 | 0.35 | 2.34 | 0.66 | bd | 1.99 |
| 5 | 2.83 | bd | 1.16 | 1.21 | 1.11 | 0.007 | 3.63 |
| 6 | 2.84 | bd | 1.15 | 1.19 | 1.11 | bd | 3.68 |

### 6. Chlorite geochemistry

Two types of chlorite have been recognised at Bierghes. The first type occurs in the green-coloured alteration, while the second chlorite occurs at the contact between the green and the red-coloured alteration. The first type of chlorite has a blue-greenish anisotropy colour (Plate 1G), while the second type of chlorite has a distinct light green colour (Plate 1H). According to the chlorite classification diagram of Hey (1954), the first chlorite has a composition between ripidolite and pycnochlorite, while the second type of chlorite has a composition dominantly corresponding to clinochlore (Table 2 and Figure 6). This means that the latter type of chlorite is Mg-richer. The composition of the chlorite, found in the green-coloured alteration of the Bierghes vein system, has a different composition from the chlorites found associated with veins in the Marcq area (Dewaele et al., 2001), the Bolland borehole (Zhang et al., 1997) and the Stavelot Massif (Schroyen, 2000; Schroyen & Muchez, 2000). The chlorites in these latter areas have a ripidolite to daphnite composition (Fig. 6).

From the Al in the tetrahedral position and eventually the Fe/(Fe+Mg) ratio, the formation temperature of the chlorites has been calculated with the empirical models of Cathelineau (1988), Hillier &
Velde (1991) and Jowett (1991) and with the thermodynamic model of Walshe (1986). The different chlorite geothermometers show no distinct variation between the average values of the two types of chlorite (Table 3). The geothermometer of Cathelineau (1988) and Jowett (1991) show a similar temperature (~320°C), while Hillier & Velde (1991) results in a lower temperature of 265°C. The geothermometer of Walshe indicates a temperature of 280°C for 100 MPa and of 300°C for 200 MPa (Table 3).

The different chemical composition of the chlorites present in the Bierghes vein system, compared to the veins in the Marcq area, Bolland borehole and the Stavelot Massif likely reflect a different equilibration temperature (Dewaele et al., 2001, Zhang et al., 1997, Schroyen, 2000; Schroyen & Muchez, 2000; Dewaele et al., 2001). However, buffering of the chemical composition of the fluids by the surrounding rocks can not be excluded. Chlorite geothermometry should always be supported by additional temperature indicators (De Caritat et al., 1993).

7. Stable isotope analyses

Ten dolomite samples have been selected for stable isotope analyses ($\delta^{18}O$, $\delta^{13}C$). Samples have been taken from the two alteration zones and from the different vein generations. The isotopic composition of the dolomites and of five sulphides ($\delta^{34}S$) was measured to determine the origin of the mineralising fluids. The $\delta^{18}O$ values and $\delta^{13}C$ values vary respectively between +13 ‰ and +14.5 ‰ V-SMOW and between –12.4 ‰ and –14.3 ‰ V-PDB (Table 4). Pyrite has $\delta^{34}S$ values between –2.8 ‰ and +2.9 ‰ CDT, while the investigated galena sample has a value of –5.6 ‰ CDT (Table 4).

The formation temperature of the sulphides has been estimated by using coexisting sulphide pairs that precipitated at equilibrium conditions (Ohimoto & Rye, 1979). Coexisting pyrite (~2.8 ‰ CDT) and galena (~5.6 ‰ CDT) have a $\Delta_{\text{py-gn}}$ of 2.8, indicating a precipitation temperature of 300-350°C. This calculated temperature based on isotopic fractionation between coexisting sulphide minerals is in good agreement with the temperature obtained by fluid inclusion microthermometry. It should be noticed that if the precipitation temperature changes with 50°C, the $\delta^{18}O$ and $\delta^{13}C$ values of the fluids change by respectively ~ 2 ‰ and ~ 0.8 ‰. $\delta^{34}S$ values are less effected by temperature fractionation (Ohimoto & Goldhaber, 1997). The calculated oxygen ($\delta^{18}O$) and carbon ($\delta^{13}C$) isotopic
composition of the ambient fluid fall respectively, between +5.5 ‰ and +7.5 ‰ V-SMOW and between – 10.5 ‰ and –11.8 ‰ V-PDB. The sulphur (δ\(^{34}\)S) isotopic composition varies between – 2.8 ‰ and +2.9 ‰ CDT.

### 8. Discussion

#### 8.1 Structural setting

In the multistage mineralised quartz-dolomite vein system at Bierghes, the occurrence of the sulphides is restricted to the vein generation 1. The fracture-parallel orientation of the mineralised veins and the deformation of the sulphides demonstrate that the mineralisation took place during a progressive deformation in which the fracture planes were reactivated as fluid pathways. Based on Rb-Sr radiometric dating, André & Deutsch (1985) interpreted the veins to be syn-tectonic during Late Givetian (373 ± 11 Ma) shear strike-slip faulting. The authors interpreted this age as an \(^{87}\)Sr resetting of the intrusive rocks by hydrothermal circulation along the shear zone at temperatures below 250°C (André & Deutsch, 1985). Resetting of Rb-Sr whole rock ages has been proven by the formation of secondary hydrated minerals by water-rock interaction from Rb-rich minerals (biotite, K-feldspar) and acid ash-flow tuffs in rocks affected by low-grade metamorphism (Evans, 1991; Evans et al., 1995). Since the first vein generation in the quartz-dolomite vein system has homogenisation temperatures up to 350°C and temperatures of 250°C are only found in vein generation 2 and 3, the age obtained by André & Deutsch (1985) is considered as a minimum age of the mineralisation in the Bierghes sill. Since

### Table 4. Results of stable isotope analyses of dolomite (δ\(^{18}\)O, δ\(^{13}\)C), and sulphides (δ\(^{34}\)S) in veins at Bierghes vein system. V-SMOW: Vienna-standard mean ocean water, V-PDB: Vienna Peedee Belemnite, CDT: Canyon Diablo meteorite troilite.

<table>
<thead>
<tr>
<th>Composition</th>
<th>δ(^{18})O</th>
<th>δ(^{13})C</th>
<th>δ(^{34})S</th>
</tr>
</thead>
<tbody>
<tr>
<td>green alteration</td>
<td>-12.7</td>
<td>13.1</td>
<td>1.3</td>
</tr>
<tr>
<td>green alteration</td>
<td>-12.4</td>
<td>13.3</td>
<td>1.3</td>
</tr>
<tr>
<td>red alteration</td>
<td>-15.3</td>
<td>13.4</td>
<td>1.3</td>
</tr>
<tr>
<td>green alteration</td>
<td>-13.8</td>
<td>13.1</td>
<td>1.3</td>
</tr>
<tr>
<td>red alteration</td>
<td>-14.2</td>
<td>13.1</td>
<td>1.3</td>
</tr>
<tr>
<td>vein generation 1</td>
<td>-12.3</td>
<td>13.6</td>
<td>1.3</td>
</tr>
<tr>
<td>vein generation 1</td>
<td>-13.8</td>
<td>13.2</td>
<td>1.3</td>
</tr>
<tr>
<td>vein generation 2</td>
<td>-13.8</td>
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<td>vein generation 2-3</td>
<td>-14.6</td>
<td>14.5</td>
<td>1.3</td>
</tr>
<tr>
<td>polema</td>
<td>-2.8</td>
<td>0.7</td>
<td>1.8</td>
</tr>
<tr>
<td>pyrite</td>
<td>2.9</td>
<td>-0.3</td>
<td>-5.6</td>
</tr>
<tr>
<td>pyrite</td>
<td>1.8</td>
<td>-0.3</td>
<td>-5.6</td>
</tr>
</tbody>
</table>

The alteration around the veins at Bierghes, developed along thin NE-dipping zones, parallel to the fracture planes. Such a well-defined pattern reflects a strong tectonic control on the migration of the fluids. Geochemical and XRD analysis showed an alteration pattern, characterised by carbonitisation, sericitisation, silicification and chloritisation features. Mineralisation with this kind of alteration pattern in magmatic rocks is often recognised in mesozonal orogenic mineralisation. It is caused by the alteration of mostly mafic and ultramafic rocks by a CO\(_2\)-rich fluid (Ash & Arksey, 1989; Kishida & Kerrich, 1987; Leitch et al., 1991). Carbonate minerals which replace these rocks form by hydrolysis of iron, magnesium, calcium and manganese silicates to carbonates in which the wall rocks provide the bivalent metal cations (Kerrich, 1989). At Bierghes, André & Deutsch (1985) and Schippers (1979) interpreted the carbonates as the results of the dynamometamorphic breakdown of epidote. Mesozonal orogenic deposits occur worldwide, from the Archean to the Tertiary (Goldfarb et al., 2001). Typical Phanerozoic orogenic deposits are associated with convergent plate margins, in close proximity to major translithospheric structures or transpressional-transensional shear zones (Bierlein & Crowe, 2000). Mesozonal orogenic deposits are almost entirely structurally controlled, and often formed at or above the brittle-ductile transition in greenschist terranes.

#### 8.2 Fluid temperatures

The hydrothermal system of the multistage mineralised quartz-carbonate vein system found at Bierghes is characterised by three groups of fluids. CO\(_2\)-rich fluids have been identified in vein generation 1 and 2. In vein generation 1 the homogenisation temperature falls between 237°C and 363°C, while in vein generation 2 T\(_{\text{SN}}\) varies between 235°C and 265°C. Detailed observation of the T\(_{\text{SN}}\) distribution plot indicates two clusters for vein generation 1 (Fig. 5D). The cluster between 341 and 363°C could correspond to the unaltered fluids of vein generation 1 and forms a minimum range of trapping temperatures for vein generation 1. The cluster of lower temperatures between 237° and 277°C could correspond to fluids of vein generation 1, reset by
the fluids in vein generation 2 (T<sub>min</sub> between 235° and 265°C). Fluids from vein generation 2 could also partly have reset other minerals, e.g. chlorite, which could explain the lower temperatures obtained by the chlorite geothermometers.

The fluids associated with vein generation 3 can be considered as a late, relatively low temperature fluid (minimum temperature between 158° and 235°C), with a different composition as the fluids from vein generation 1 and 2.

### 8.3 Fluid composition and origin

Fluids with an H<sub>2</sub>O-CO<sub>2</sub>-NaCl-KCl composition have also been identified in the vein system in the Marcq area (Dewaele et al., 2001; Piessens et al., 2002), which has been identified as a typical mesothermal orogenic fluid formed by the circulation of metamorphic fluids (Piessens et al., 2002). This fluid composition is characteristic for the Phanerozoic polysulphide and gold mineralisation in the Caledonides of the British Isles (Shepherd et al., 1991; Bottrel et al., 1988; Duller et al., 1997; Ixer et al., 1997; Mason, 1997; Steed & Morris, 1997; Lowry et al., 1997; Wilkinson et al. 1999). However, the mineralisation in these areas are often associated with granitic intrusions. The main emplacement of the Bierghes sill is restricted to the Ashgillian (Late Ordovician). It precedes the deformation event in the Anglo-Brabant Fold belt, occurring during the Late Silurian to Early Devonian, by at least 30 Ma (André et al., 1981; Van Grootel et al., 1997; Debaker, 2001; Verniers et al. 2002; Dewaele et al., 2002). Magmatic fluids can therefore be excluded as the dominant source or transport medium for the mineralisation. The alternative explanation, involving mainly metamorphic fluids, is in complete agreement with the tectonometamorphic framework.

The metamorphic origin of the fluid is supported by the isotopic investigations. The δ<sup>34</sup>S composition of the dolomitising fluids at Bierghes (between +5.5 ‰ and +7.5 ‰ V-SMOW), is similar to the isotopic composition of the fluid causing the mineralisation in the Marcq area (+3 to +8 ‰ V-SMOW), which has been identified as a metamorphic fluid (Piessens et al., 2002). The δ<sup>13</sup>C values of the dolomite are negative, between −10.5 ‰ and −11.8 ‰ VPDB, which can be explained by the diagenesis/metamorphism of organic material-rich sediments (Ohmoto & Goldhaber, 1994). The δ<sup>34</sup>S range between −2.8 ‰ and +2.9 ‰ CDT, falls in the range of I-type magmatic rocks (-5 to 5 ‰ CDT). However, the isotopic signature of sulphides is strongly depending on the δ<sup>34</sup>S composition of sulphides and sulphates in the country rocks (Ohmoto & Goldhaber, 1997; Ohmoto, 1986). This indicates that the δ<sup>34</sup>S values of the Bierghes sulphides certainly do not exclude a metamorphic origin. Based on the geotectonic setting, the similarity with the Marcq area and the lack of clear evidence for a magmatic origin, the mineralising fluid at Bierghes is interpreted to have a metamorphic origin. It should be stated that only fluids from vein generation 1 and 2 are likely to have a metamorphic origin. Fluids from vein generation 3 can be anything, notably surface waters (e.g. marine waters), that infiltrated at depth.

This study has demonstrated a spatial and temporal relationship between alteration, polysulphide mineralisation and deformation in the Anglo-Brabant fold belt in Belgium. The vein system in the Bierghes sill is the second location identified, in addition to the Marcq area, where a migration of metamorphic fluids, accompanied by progressive deformation along a high-strain zone, is proposed. Both areas are located in the same fault zone, located at the northeastern edge of a low-density body, identified at depth (Sintubin, 1999; Sintubin & Everaerts, 2002). This granitoid basement block controlled the deformation (Sintubin, 1999; Sintubin & Everaerts, 2002) and associated faults acted as conduits for the release of metamorphic fluids during the Lower Palaeozoic deformation of the Anglo-Brabant Fold belt in Belgium.

### 9. Conclusion

The alteration and polysulphide mineralisation in the Lower Palaeozoic magmatic rocks in the Bierghes sill are spatially associated with NE dipping fracture planes. The zone of alteration and mineralisation comprises several small lenses, up to one metre thick. The total thickness of the mineralised zone is 50 metres. Mineralisation and alteration occurred during progressive deformation, but after cleavage development.

The mineralising fluids have an H<sub>2</sub>O-CO<sub>2</sub>-(X)-NaCl-KCl composition. The δ<sup>18</sup>O, δ<sup>34</sup>S isotopic composition of the fluids is respectively between +5.5 ‰ to 7.5 ‰ V-SMOW, −10.5 to −11.5 ‰ V-SMOW and −2.8 ‰ to +2.9 ‰ CDT. These fluids caused the precipitation of the sulphide minerals and the carbonitisation, sericitisation, silicification and chloritisation of the host rock.

The vein system in the Bierghes sill has been identified as a mesozonal orogenic mineralisation. The mineralising fluids most likely had a metamorphic origin, based on the geotectonic and geochemical characteristics. The similarity between the vein system in the Bierghes sill and the polysulphide mineralisation in the Marcq area, reflects a related origin.

The occurrence of the veins at Bierghes in the same geotectonic context as those in the Marcq area, indicates an important migration of metamorphic fluids in the high-strain zone at the northern margin of the supposed granitoid basement block, during the Lower Palaeozoic deformation of the Anglo-Brabant Fold belt, Belgium.
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