Aulis Kinnunen

A PALAEOPROTEROZOIC HIGH-SULPHIDATION EPITHERMAL GOLD DEPOSIT AT ORIVESI, SOUTHERN FINLAND
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Abstract

The metamorphosed Palaeoproterozoic Orivesi gold deposit in southern Finland is located within the Tampere Schist Belt, which belongs to the Svecofennian domain. The Orivesi mine, run by Outokumpu Mining Oy, was in production from 1994 to 2003, during which time a total of approximately 1.7 million tons of ore was extracted, with an Au content of 9.31 g/t, implying a total output 13.115 tons of gold in concentrate.

The hydrothermal alteration halo can be divided successively into chlorite-dominant, sericite-dominant and quartz-dominant rocks from the outer zone inwards. The host rocks of the ore are quartz rocks with andalusite-rich quartz rocks. Topaz-bearing rocks also occur in the inner part of the alteration halo. In addition to Au, the elements Ag, Te, Bi, Sb, S, As, Se, Cu, Zn, Pb, Sn and Mo are enriched to varying degrees within the alteration halo.

The main ore minerals include base metal sulphides, sulphosalts and tellurides. Pyrite is the most common sulphide. The sulphosalts are represented by tetrahedrite, bournonite, boulangerite and meneghinite. The most common gold, gold-silver and silver tellurides are calaverite, montbrayite, petzite, kostovite, sylvanite and hessite. Other known tellurides include tellurobismuthite, altaite, melonite, frohbergite, tsumoite, tetradyrite and rucklidgeite.

Gold occurs mostly in fine-grained native grains containing an average of 5% Ag. The native gold is usually of very small grain size, generally < 20 µm. Most of the gold grains in the deposit occur as intergrowths with tellurides.

The adjacent hypabyssal intrusion is an obvious source of both hydrothermal fluids and metals. A comb quartz layering has been discovered in the transition zone between the intrusion and the alteration halo.

The Orivesi deposit is thought to belong to the high-sulphidation epithermal type. Soon after its formation the deposit encountered deformation and metamorphism that amounted to lower amphibolite facies conditions. The subsequent retrograde metamorphism caused the reappearance of some hydrothermal minerals typical of high-sulphidation epithermal deposits.

Keywords: comb quartz layering, deformation, epithermal, Finland, gold deposit, high alumina minerals, high sulphidation, hydrothermal alteration, hypabyssal intrusion, metamorphic rocks, Orivesi, Palaeoproterozoic, Precambrian, sulphides, sulphosalts, Svecofennian, tellurides
Preface

The Orivesi Kutemajärvi gold deposit, located some 30 km NE of the city of Tampere in southern Finland, was mined from autumn 1994 to the end of 2003. I worked at the mine for the first time for two months in early spring 1996, and gained an initial impression that the deposit was both very interesting and enigmatic. Over a year later, in autumn 1997, a possibility opened up to work there as the mine geologist. This meant that a lot of detailed material came into my hands in the course of my everyday work. One very valuable source of information was the geological mapping of the drifts, totalling ca. 5 km, and the logging of diamond drill cores (ca. 20 km) also opened my eyes to understand the details of the deposit, a perspective that was widened by prospecting in the nearby surroundings.

The stimulus for this present work was a desire to document the deposit, which developed into the most notable gold mine to operate in Finland at that time. A second aim was to display the characteristics of the deposit to such an extent that the data could be utilized elsewhere in gold exploration. The data, accumulated by professional work, paved the way to an understanding of the nature and genetic type of the deposit, which were not so well known in the context of an Archean and Proterozoic terrain representing age groups that are common in the Finnish bedrock. I puzzled over the data and my observations at length, piecing the facts together, shaping my thinking on the subject and increasing my knowledge in gold deposits via the literature. Those ideas, presented below, entered my consciousness gradually and took shape bit by bit while I was immersed in my work.
Acknowledgements

I am highly indebted to Mr. Tuomo Mäkelä, former President of Outokumpu Mining Oy, for permission to use the data that have accumulated concerning the Orivesi deposit, and to the late Prof. Tuomo T. Alapieti, University of Oulu, for accepting the topic of my doctoral thesis and for his kind guidance. After his sudden and premature death, Docent Aulis Kärki took over his role as an encouraging supervisor of this research. The critical comments made by the official examiners of the thesis, Prof. Pekka Nurmi and Dr. Anders Hallberg, have been highly appreciated. I am obliged to the staff of the Orivesi mine, especially to Heikki Saarnio, M.Sc., who has provided many stimulating ideas, and Aatto Ratia who gathered many samples that were of great help. The summer trainees Saku Vuori, Jukka Nieminen and Matti Talikka in chronological order, assisted with the geological mapping both in the nearby area and especially underground, and their keen eyes observed many valuable details which were in turn of great help for creating the overall picture.

Dr. Irmeli Mänttäri of the Geological Survey of Finland kindly performed the age determinations and thus had a decisive influence on the formulation of the schema for the metamorphic episodes.

I am very much obliged to my colleagues Arto Peltola, M.Sc., of the University of Turku and Dr. Kari Kojonen of the Geological Survey of Finland for their help in mineralogical issues, and to Aimo Hattula, M.Sc. (Tech.), who kindly helped me in geophysical matters.

I am also indebted to the numerous visitors to whom I have had the honour of explaining the deposit, and who have offered many useful ideas. Furthermore, the preparation of this thesis has benefited considerably from detailed discussions with many colleagues not mentioned here by name.

Mrs. Kristiina Karjalainen has drawn the maps, figures and diagrams and put the skilful finishing touches to them, and Mr. Malcolm Hicks, M.A., has corrected the language of the manuscript.

I am also greatly indebted to K.H. Renlunds Stiftelse for crucial financial help.

Finally, I would like to express my warmest thanks to my family for their encouraging attitude and patience.
List of figures

Fig. 1. Lithological map of the central Tampere Schist Belt with a simplified structural interpretation. .................................................. 26
Fig. 2. The commencement of the building of the inclined drift within the unaltered intermediate volcanic rocks (left) and the shells of the open pit on Pipe 1 (right). ........................................................... 34
Fig. 3. Classification and variation in composition of the Pukala porphyry, protolith suite and unaltered volcanic rocks. ............................................. 35
Fig. 4. The Pukala porphyry (a), a microphotograph of the porphyry, crossed nics (b), and inclusions in the porphyry of the southern margin (c)........... 39
Fig. 5. The transition zone. Rhyolitic fragments in a matrix rich in sericite and chlorite. ................................................................. 41
Fig. 6. Comb quartz layering. .......................................................... 42
Fig. 7. Geological map of the Orivesi Deposit and its surroundings. .............. 47
Fig. 8. Detailed geological map of the pipes and the proximal area on the ground level (z = +15). .......................................................... 48
Fig. 9. Geological map of the alteration halo around the Orivesi Deposit, level z = +500. ................................................................. 49
Fig. 10. Detailed geological map of the pipes and the proximal area on the level z = +500. ................................................................. 50
Fig. 11. Total magnetic image of the ground survey data on a 10 m × 50 m grid. ... 51
Fig. 12. Schematic vertical section depicting the intensity of hydrothermal alteration from the unaltered protolith to the mineralized zone. ................. 52
Fig. 13. Microphotographs of variants of topaz-dominant and topaz-bearing rocks. 53
Fig. 14. Bluish kaolinitized fine-grained topaz rock fragments in quartz-sericite schist and quartz-dominant rock. ........................................... 54
Fig. 15. The breccia lying at the margin of the unit containing topaz-bearing and topaz-dominant rocks. The fragments consist mainly of sericite-dominant, quartz-dominant and topaz-dominant rocks. ......................... 55
Fig. 16. Variants of the volcanioclastic structures showing strong lineation. ........ 57
Fig. 17. Microphotographs of typical rocks within the ore pipes and in their proximity. 58
Fig. 18. Three-dimensional model of the deposit, viewed from the north-east. .... 60
Fig. 19. Microphotographs of tellurides and of the occurrence of gold. ............. 74
Fig. 20. Distribution of Au, Te and Bi between fragments and the matrix within Pipe 3. 75
Fig. 21. Grain size distribution and fineness of gold, and modes of gold occurrence. 76
Fig. 22. Quartz-tourmaline veins on the wall of a drift within Pipe 2 (a), and a microphotograph of a quartz-tourmaline vein located within a topaz-bearing schist (b). ................................................................. 79
Fig. 23. Au, Te and Bi content of samples containing Au ≥2 ppm in the entire deposit. 83
Fig. 24. Gold, Te and Bi in Pipe 1 at various levels in the mine. ....................... 84
Fig. 25. Gold, Te and Bi in Pipe 2 at various levels in the mine. ....................... 85
Fig. 26. Gold, Te and Bi in Pipe 3 at various levels in the mine. ....................... 86
Fig. 27. Gold, Te and Bi in Pipe 5 at various levels in the mine. ....................... 87
Fig. 28. Rectangular profile across the alteration halo illustrating the area covered by the samples chosen in order to review the dispersion of elements. 89
Fig. 29. Distribution of Au, Ag, Te and Bi on a profile across the alteration halo. ... 91
Fig. 30. Distribution of Sb, Mo, W and Sn on a profile across the alteration halo. ... 92
Fig. 31. Distribution of Cu, Zn, Pb and As on a profile across the alteration halo. ... 94
Fig. 32. Distribution of S, Se, U and Th on a profile across the alteration halo. .... 95
Fig. 33. Distribution of Ni, Co, Cr and Li on a profile across the alteration halo. .... 97
Fig. 34. Distribution of Sr, Ba, Sc and Y on a profile across the alteration halo. .... 99
Fig. 35. Distribution of La, P, and V on a profile across the alteration halo. ....... 100
Fig. 36. Fluorine content in relation to gold content in the proximity of the ore pipes. 101
Fig. 37. Type of hydrothermal alteration, expressed in terms of the ratio of aluminium to total alkali (A/CNK). 103
Fig. 38. Detailed map illustrating D3 folding within the alteration halo. 107
Fig. 39. The almost vertical hinge of a fold caused by deformation D3 on the wall of a drift cut by minor quartz veins and segregations. 108
Fig. 40. Two separate lineations, the older almost vertical, while the younger forms an angle of about 25º with it. 109
Fig. 41. Kink folding visible on the wall of a drift. The width of the view is ca. 2 m. 109
Fig. 42. The present post-deformational setting of the alteration halo and the Pukala porphyry, according to magnetic modelling of airborne survey data using plate models. 111
Fig. 43. P-T diagram showing estimated metamorphic conditions for the Orivesi gold deposit at the different stages. 113
Fig. 44. Microphotograph (crossed nicols) displaying two generations of phlogopite in a schist containing quartz, sericite, chlorite and altered andalusite porphyroblasts (a), microphotograph (one nicol) of a late andalusite porphyroblast (b), and a large quantity of late phlogopite and andalusite porphyroblasts in a chlorite-rich layer (c). 114
Fig. 45. Retrograde minerals occurring within the andalusite-muscovite breccia. 118
Fig. 46. Microphotographs of retrograde metamorphosed rocks. 119
Fig. 47. Thucholite nodules, max. ca. 5 mm in diameter, located in andalusite-muscovite breccia (a) and inclusions within the thucholite (b). 119
Fig. 48. U-Pb isochron diagrams for the thucholite samples analysed. 121
Fig. 49. Late tectonic phenomena, a fissure filled with fluor spar and later with kaolinite. 124
## List of tables

Table 1. Annual gold production at the Orivesi mine. .......................... 20  
Table 2. Compositions of the Pukala porphyry. ................................. 43  
Table 3. Selected whole rock compositions for Pipes 2 and 5. ................. 62  
Table 4. Selected whole rock compositions illustrating the variation in the quantity of pyrite within Pipe 1. ................................. 62  
Table 5. Compositions of pyrite. ................................................. 65  
Table 6. Average compositions of sphalerite. ................................. 66  
Table 7. Compositions of galena. ................................................. 67  
Table 8. Average compositions of arsenopyrite. ............................ 68  
Table 9. Average compositions of sulphosalts. .............................. 69  
Table 10. Average compositions of tellurides. ................................ 71  
Table 11. Average microprobe analyses of the most common tellurides in Pipe 4. 72  
Table 12. Average compositions of native gold, electrum, native tellurium and native antimony. ................................. 73  
Table 13. Average tourmaline compositions, in wt %. ........................ 80  
Table 14. Conventional U-Pb age data on thucholite from the Orivesi mine. 120  
Table 15. Nomenclature used for the deposit types related to two contrasting epithermal environments. ................................. 130  
Table 16. Geological evolution and characteristic features of the Orivesi gold deposit and its host rocks. ................................. 163
## Contents

**Abstract**

**Preface**

**Acknowledgements**

**List of figures**

**List of tables**

**Contents**

1 **Introduction**
   1.1 Exploration history .............................................................. 17
   1.2 The development of mining .................................................... 19
   1.3 Mining methods ................................................................. 20
   1.4 Ore processing ...................................................................... 20

2 **Sample sets and analytical methods**

3 **Regional geology of the Tampere Schist Belt**
   3.1 Historical background.......................................................... 25
   3.2 Subsequent investigations ...................................................... 27
   3.3 Characteristic features of the central Tampere Schist Belt......... 27
   3.4 Ore deposits and industrial minerals ...................................... 30

4 **Geology of the Orivesi area**
   4.1 Rocks bordering on the hydrothermally altered domain .......... 36
           4.1.1 Unaltered volcanic rocks .............................................. 36
           4.1.2 The protolith suite ..................................................... 36
           4.1.3 The Pukala porphyry ................................................... 37
                   Comb quartz layering ................................................. 39
   4.2 The hydrothermally altered domain ...................................... 44
           4.2.1 General features ...................................................... 44
           4.2.2 Rock types and mineral assemblages ......................... 45
   4.3 Ore pipes ............................................................................. 55
           4.3.1 General framework .................................................... 55
           4.3.2 Geometry of the pipes ................................................. 59
           4.3.3 Chemical and mineralogical composition of the ore pipes.. 61
   4.4 Ore minerals ........................................................................ 63
           4.4.1 Iron and titanium oxides .............................................. 63
           4.4.2 Sulphides ................................................................... 64
           4.4.3 Sulphosalts ................................................................ 68
4.4.4 Tellurides ................................................................................... 69
4.4.5 Native gold, tellurium, lead, antimony and copper .................. 72
4.4.6 Parameters of gold occurrence within the deposit .................... 75

4.5 Dykes and veins ........................................................................... 76
  4.5.1 Mafic dykes .......................................................................... 76
  4.5.2 Milky quartz veins ............................................................... 77
  4.5.3 Quartz-tourmaline veins ....................................................... 78
  4.5.4 Age relations of dykes and veins ......................................... 79

5 Geochemical features resulting from ore deposition and hydrothermal alteration 81
  5.1 Distribution of Au, Te and Bi within the Orivesi deposit ......... 81
  5.2 Elements characterizing hydrothermal alteration .................. 88
    5.2.1 Gold and silver ................................................................. 89
    5.2.2 Tellurium, bismuth and antimony ................................... 89
    5.2.3 Molybdenum, tungsten and tin ....................................... 90
    5.2.4 Copper, zinc and lead ..................................................... 93
    5.2.5 Arsenic, sulphur and selenium ....................................... 93
    5.2.6 Uranium and thorium ..................................................... 93
    5.2.7 Nickel, cobalt and chromium ........................................ 96
    5.2.8 Lithium, strontium and barium ...................................... 96
    5.2.9 Scandium, yttrium and lanthanum ................................. 98
    5.2.10 Phosphorus and vanadium ......................................... 98
    5.2.11 Fluorine ........................................................................ 100
  5.3 Hydrothermal enrichment and metamorphic remobilization .... 101
  5.4 Alteration style in the proximity of the deposit ...................... 102

6 Deformation and metamorphism 105
  6.1 Deformation in the Tampere Schist Belt .................................. 105
    6.1.1 Deformation in the proximity of the deposit ................... 106
    6.1.2 Structure according to magnetic modelling ...................... 110
  6.2 Metamorphism ......................................................................... 110
    6.2.1 Retrograde metamorphism ............................................. 112
      Porphyroblastic minerals ..................................................... 112
      Andalusite-muscovite breccia ............................................ 115
      Thucholite .......................................................................... 117
      Muscovite veins and other late overprintings ..................... 123
    6.2.2 Successive events of deformation and metamorphism ....... 123
1 Introduction

1.1 Exploration history

The hydrothermally altered domain on the northern border of the central Tampere Schist Belt (TSB) (Fig. 1) has been known for several decades (see Seitsaari 1951; Simonen 1952). After the Second World War economic activity increased greatly due to the shortage of various materials, and the sericite formation located NW of Lake Kutemajärvi also became a target of economic evaluation. A company named Viento Oy made a reservation for a claim in 1946, in accordance with Finnish mining law, to secure the right to investigate a supposed occurrence of kaolin in the area (Ollila et al. 1990).

Within the same year the exploration license was transferred to the company Renlundin Tiili Oy, which applied for and later obtained a mining concession (1947) for the extraction of sericite and kaolin. On the basis of investigations that included diamond drilling (9 holes, each less than 6 m in length) the company estimated that the amount of sericite schist was 2 Mt to a depth of 5 m and 18–20 Mt to a depth of 50 m. A small open sericite pit (10 m * 30 m, 6–7 m deep) was quarried at that time. At the end of 1957, Renlundin Tiili Oy, was amalgamated with Oy Renlund Ab and lost the mining concession because it had failed to begin mining within ten years. (Saikkonen 1981; Ollila et al. 1990).

Following this a new claim on the area was made by a private person for a short time, again on account of the kaolin. At the same time Suomen Mineraali Oy became interested in the sericite, but this was not among the minerals eligible for a claim under Finnish mining law. Suomen Mineraali Oy concluded an exploration agreement with the landowner (Metsähallitus, the Finnish Forest and Park Service) for 25 years at the beginning of 1959. Through a subsequent merger the agreement was transferred to Paraisten Kalkkivuori Oy, which renounced it at the end of 1964. (Grönholm 1992).

Kemira Oy applied for a claim in 1966 which was awarded a year later. This led to the commencement of a detailed geological investigation, after a preliminary inquiry. The study included geological mapping and diamond drilling (3 holes, 428.00 m in total). The efforts of the company were directed towards separating out potassium and aluminium by leaching from concentrated sericite, but the leaching methods tested proved to be uneconomic. The possibility of using topaz as a basis for producing fireproof materials was also under investigation. The final estimate of mineral resources resulting from the investigations carried
out by Kemira Oy amounted to 14 Mt of sericite schist containing 37% sericite, 6% topaz, 8% kaolinite, 1% andalusite and 48% quartz. It was noted in this connection that the rocks were rich in fluorine, and in the next phase the company investigated metals that are known to follow fluorine in hydrothermal fluids, such as Sn, Li, Rb, Au and Ag. The concentrations of precious metals in the analyses were not encouraging for further research, and the company abandoned its claim at the end of 1974. (Saikkonen 1981; Ollila et al. 1990).

A fresh company to enter the scene was Oy Lohja Ab, whose exploration department reserved a claim over the hydrothermally altered area in 1975 that was registered one year later. The company was investigating the possibility of making high-quality fillings of mica minerals for the various branches of industry. This led to detailed geological mapping, trenching for sampling, diamond drilling and geophysical surveys, which indicated that the area containing sericite schist was about 20–24 hectares, with abundant internal variation in quality. The most sericite-rich schist amounted to 4–5 Mt down to a level of 20 m. The recovery rate in this unit was 27% and the sericite concentrate would contain 31.9% Al₂O₃ and 0.16% Fe₂O₃, in addition to which it would be possible to obtain a quartz concentrate with 0.01% Fe₂O₃ as a by-product. Topaz was to be found in considerable amounts both in disseminations and as monomineralic inclusions in the sericite schist, and it would also be possible to produce topaz concentrate as a by-product of sericite. It is worth mentioning, too, that analyses of the topaz rock have shown that it contains 53.64% Al₂O₃, 0.90% Fe₂O₃ and 15–40 ppm Sn. After efforts to concentrate industrial minerals, it had to be admitted, however, that there was no basis for economic exploitation. (Saikkonen 1981).

After this exploration for industrial minerals, base metals and gold became the targets. Gold exploration in the TSB and Pirkkala migmatite belt was initiated by a University of Helsinki project in 1980–1983 (Nurmi et al. 1984), and the Kutemajärvi area at Orivesi was one of the principal target areas selected. The work was carried out in collaboration with Oy Lohja Ab. A total of 72 bedrock samples taken along three N–S profiles over the sericite-quartz schist (Rasilainen et al. 1999) revealed a zoned anomaly pattern with barren sericite schist in the north, Au in the middle and base metals in the south, matching well with the geophysical picture. The concentrations were low, at most 0.2 g/t Au and 283 g/t Cu, but high enough for further work. Two diamond drill holes were made by Oy Lohja Ab in summer 1982 to check the anomalies, and these intersected an ore-grade zone that is now called Pipe 2. An intensive period of diamond drilling followed, totalling 15,045 m, as a result of which a gold mineralization occupying
three almost vertical pipes was located. This contained 631,000 t of ore with a concentration of 9.4 g/t Au, given a 2 g/t cut-off (Ollila et al. 1990). Mineral resources were calculated to a depth of about 275 m.

At the same time, during the late 1980’s, Oy Lohja Ab was carrying out Au exploration in hydrothermally altered areas analogous to Kutemajärvi near the northern border of the Tampere schist belt. The results of a relatively short exploration period did not point to any further mineral resources in the surroundings, and Oy Lohja Ab, today Lohja Rudus Oy Ab, which was acquired by CRH plc in 1999, decided to give up its mining concession. The mining rights were purchased by Outokumpu Finnmines Oy, the predecessor of Outokumpu Mining Oy, in March 1990 (Ollila et al. 1990).

1.2 The development of mining

Pilot open pit mining started in 1990, and 9847 tons of ore had been stoped from the open pit (Pipe 1) by the end of the year and concentrated in the dressing plant attached to the Vammala Ni-Cu-mine about 90 km to the SW. The pilot phase of underground mining was in progress three years later, in 1993–1994, and the decision to open the Orivesi mine was made in 1994. Since the mine at Vammala had been exhausted and had to be closed down at about the same time, its concentrating plant was used for dressing the gold ore, broken rock from Orivesi being transported to Vammala by road. Production increased from 1994 until 1999 and sustained that level for a couple of years before declining during the last years (Table 1).
Table 1. Annual gold production at the Orivesi mine.

<table>
<thead>
<tr>
<th>Year</th>
<th>Feed (t)</th>
<th>Au (g/t)</th>
<th>In concentrate (kg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1990</td>
<td>7926</td>
<td>9.3</td>
<td>60</td>
</tr>
<tr>
<td>1994</td>
<td>63604</td>
<td>8.9</td>
<td>477</td>
</tr>
<tr>
<td>1995</td>
<td>147353</td>
<td>8.4</td>
<td>1050</td>
</tr>
<tr>
<td>1996</td>
<td>155918</td>
<td>10.0</td>
<td>1336</td>
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<tr>
<td>1997</td>
<td>172226</td>
<td>10.8</td>
<td>1551</td>
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<td>1998</td>
<td>190301</td>
<td>9.6</td>
<td>1500</td>
</tr>
<tr>
<td>1999</td>
<td>214362</td>
<td>10.0</td>
<td>1761</td>
</tr>
<tr>
<td>2000</td>
<td>174272</td>
<td>9.4</td>
<td>1351</td>
</tr>
<tr>
<td>2001</td>
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<tr>
<td>Total</td>
<td>1691157</td>
<td>9.31</td>
<td>13115</td>
</tr>
</tbody>
</table>

1.3 Mining methods

The gold deposit consists of five almost vertical pipes or lenses, and the main mining method was that known as bench and fill stoping. The mine was divided vertically into main blocks of height 80 m, with four sub-levels inside each main block. Stoping started in the bottom bench and continued to the next bench after the preceding one had been back-filled. After three benches the pillar between the main blocks was stoped by sub-level caving. The broken rock was hoisted by centre-articulated dumpers.

The inclined drift has reached a level of +720 m. By the end of mining Pipe 2 had been stoped to the +720 level, Pipe 5 to +700 and Pipe 1 and Pipe 3 to +610. The ground level is at +15 m on the mine coordinates.

1.4 Ore processing

Ore dressing, including crushing and milling, took place at the Vammala concentrating plant, which had principally been built for processing nickel ore. The flotation process proved suitable for treating gold ore after a few minor changes. Since the particle size of the gold minerals was very small, gravity separation was not possible, and all the concentrates were produced by flotation.
The ore was crushed in a primary jaw crushe operating at a 120 mm setting and transferred via a feed conveyor to the secondary crushe, a cone crusher run at a 40 mm setting. Screened oversized material was fed to another cone crushe run at 6 mm. After crushing, the ore was taken from the fine ore bin by three drum feeders to a rod mill of the overflow type with manganese steel liners that was operating in an open circuit. The coarse underflow product of the hydrocyclones passed to a rubber-lined ball mill, also of the overflow type, which used 35 mm diameter hardened steel balls for grinding. The cyclone overflow, with a particle size of about 70% -53 µm, passed to a conditioner prior to flotation. The conditioned pulp then passed first to a rougher stage comprising two OK16 (16 m³) flotation cells, followed by a scavenger stage of four 16 m³ cells. The concentrates from the previous stages were fed into the cleaner cells, comprising one to four cleaning steps. Eight cells OK3 (3 m³) could be connected flexibly to match a higher grade of final concentrate. The final concentrate was pumped into an outdoor thickener, where a flocculant was added to aid settlement. The underflow of the thickener was pumped to an automatic pressure filter, and the concentrate, at a moisture content of 8 to 10%, was transported by a conveyer into a covered store. A Courier-30 online analyser was used to provide continuous assays of the key elements in the concentrate stream.

Since some proportions of the ore were rich in tellurium and bismuth, the concentrate after flotation contained commercially excessive amounts of these elements, which would be harmful in a smelter. Tellurium and Bi were therefore leached out of the concentrate by dissolving them in NaCl and H₂SO₄, whereupon the gold remained in the concentrate, from which the chlorides could be washed out carefully and neutralised.
2 Sample sets and analytical methods

Various analyses published earlier were available for use in this research, together with unpublished analyses carried out during the exploration phase before mining. The lithogeochemistry played an essential role in the discovery of the Orivesi gold deposit in 1982, and for that reason large amounts of data on analysed samples exist covering the alteration area. Samples were taken from outcrops by the former Oy Lohja Ab company at the time of discovery, each comprising 3 to 5 pieces of core about 15–20 cm long obtained with a portable drill. The contents of 24 elements were determined by instrumental epithermal neutron activation analysis (IENAA) in the Reactor Laboratory of VTT (Technical Research Centre of Finland). Further information on the method and its detection limits, precision and accuracy is available from Rosenberg et al. (1982) and Lipponen & Rosenberg (1988). Of the elements analysed Au, Ag, Sb, As, Sn, Mo, W, Zn, Ba, Sc, La, U and Th are included in this research.

Another body of material comprises samples collected from diamond drill cores in the proximity of the deposit in the mid-1990’s, including contractual research performed by the Geological Survey of Finland for Outokumpu Mining Oy in 1996. Most of the samples were located within the alteration zone in the proximity of the deposit, within the ore pipes or in their immediate vicinity. These samples were analysed by standard procedures in the chemical laboratory of the Geological Survey of Finland. Gold, Ag, Bi, Se and Te were analysed by GAAS and As, Ba, Co, Cr, Cu, La, Li, Sr, Th, V, Y, Mo, Ni, Pb, S, Sc, Sb and Zn by ICP–AES (Rasilainen 1996; Rasilainen & Nurmi 1999).

By combining these two sets of data it is possible to review the occurrence of certain elements within the alteration halo. The network of samples chosen here covers a stripe approximately 700 m long in a NS direction and 300 m wide in an EW direction running through the alteration halo. Drill core samples located deeper than 150 m are excluded. The combined material includes 157 samples from outcrops and 234 drill core samples. The contents of some elements are below the detection limit in certain samples and have thus been omitted. The numbers of analyses are as follows: Au 391, Ag 346, Te 234, Bi 234, Sb 219, Mo 388, W 157, Sn 116, Cu 230, Zn 308, Pb 199, As 344, S 234, Se 229, U 156, Th 377, Ni 224, Co 200, Cr 234, Li 202, Sr 234, Ba 391, Sc 390, Y 234, La 390, P 234 and V 224.

Most of the subsequent analyses carried out during the mining operations, including whole-rock X-ray fluorescence and fire assay Au-analyses of diamond drill cores, were performed by the Chemical Technology, Mineral Processing and
Geoanalytical Services sections of VTT (referred to for brevity as VTT Geoanalytical Services), which later became a department of the Geological Survey of Finland.

A large number of analyses are for the purpose of reviewing the distribution of Au, Te and Bi within the deposit as a whole and its various parts. After diamond drilling, the exact shape and volume of the deposit was determined by sludge drilling in each pipe and at each mining level. This provided an essential means of locating the ore deposit and estimating its Au, Te and Bi concentrations in situ. Sampling proceeded continuously during drilling, most samples being of length 1.8 or sometimes 1.5 metres, depending on the length of drill rod driven, and the weight of the cuttings per sample was approximately 10 kg. Tellurium, Bi and Au were analysed by FAAS (after dissolving in nitrohydrochloric acid) in the laboratory of the dressing plant.

Petrographic studies were performed on 75 thin sections, 83 polished thin sections inherited from previous researchers (Siivonen 1984; Grönholm 1992), and 54 polished thin sections prepared in the laboratory of the Department of Geology at the University of Turku. Some microprobe analyses and mineral identifications were carried out by the author using the Jeol JXA-8200 equipment at the Institute of Electron Optics University of Oulu. Unless otherwise stated, the photographs were taken by the author.
3 Regional geology of the Tampere Schist Belt

3.1 Historical background

The Tampere Schist Belt is a narrow, discontinuous EW-trending belt approximately 200 km long and about 20 km wide as its maximum. The Orivesi gold deposit is located near its central part, about 25 km NE of the city of Tampere (Fig. 1). The belt is one prominent subarea of the Palaeoproterozoic Svecofennian domain that extends from Russia in the east to Sweden in the west, and it borders on the large granitoid complex of central Finland in the north and migmatitic mica gneisses with associated Ni-prospective ultramafic intrusions, sulphide schists and amphibolites to the south.

Since the time of Jacob Johannes Sederholm (1863–1934) the well-exposed volcanic-sedimentary belt has been considered one of the key areas of the Svecofennian domain (see Sederholm 1987). The initial phase of its geological investigation was historic in the sense that, as a young geologist, Sederholm had found some fossils on the shore of the bay of Aitolahä in Lake Näsijärvi, slightly over five km NE of Tampere, in summer 1890, at the time when he was preparing his doctoral thesis on the bedrock of SW Finland (Sederholm 1891). He noted that the volcanic rocks there were very similar to modern basalts.

Three years later the young researcher was appointed director of the Geological Commission, the predecessor of the Geological Survey of Finland and being in charge of the mapping of glacial deposits in the Tampere Schist Belt, was able to participate in this work himself. He was interested in the notion of uniformitarianism put forward by the founder of modern geology, the Scotsman James Hutton (1726–1797), and wished to test it. Hutton’s views were supported by the English geologist Sir Charles Lyell (1797–1875), who summarized them in the adage “The present is a key to the past.” Sederholm principally investigated sedimentary rocks in the Tampere Schist Belt, and his work (Sederholm 1897) became well-known everywhere in the world as a confirmation of the actualistic principle.
Fig. 1. Lithological map of the central Tampere Schist Belt with a simplified structural interpretation. Modified after Kähkönen (1999). In the inset: a) the Archaean craton; b) the Karelian Domain; c) the Svecofennian Orogen, with the Tampere area shown as a black rectangle; d) Post-Svecofennian Precambrian domains; and e) Palaeozoic rocks.
During their research into sedimentary rocks, Sederholm and his team found many more fossils of the same type as in summer 1890: strange small, dark circles on the surface of the cliffs lining the shore of Lake Näsijärvi. Detailed research proved them to be sacs of some sort that had been immersed in the rock, as it were, and filled with material of the same kind as outside and in the vicinity. The sacs were flattened and elongated, and their walls contained graphite and to some extent phosphorus. The smallest circles were less than one cm in diameter, and the biggest over 30 cm in length. Many years later Sederholm (1913) gave these fossils the name *Corycium enigmaticum* and stated that they were the oldest fossils encountered hitherto.

### 3.2 Subsequent investigations

The Tampere Schist Belt has been the target of many subsequent investigations. Geological mapping of the central part was performed by Seitsaari (1951), Simonen (1952, 1953a), Matisto (1961, 1964, 1977) and Laitakari (1986), and Simonen (1953b) incorporated it into a stratigraphy for south-western Finland. Later on, Kähkönen (1989) examined the geochemistry and petrology of the metavolcanic rocks and specified the stratigraphy (Kähkönen 1999). The metasedimentary rocks, their geochemistry, lithostratigraphy and type of sedimentation, have also been investigated by Simonen and Kouvo (1951), Ojakangas (1986), Rautio (1987) and Kähkönen and Leveinen (1994).

The age of the schist belt (ca. 1.9 Ga) has been discussed by many researchers, at least in passing, and the results of geochronological and associated isotope studies have been published, among others, by Kouvo and Tilton (1966), Kähkönen *et al.* (1989), Huhma *et al.* (1991), Claesson *et al.* (1993), Kähkönen *et al.* (1993) and Vaasjoki *et al.* (1999). The structure, multiphase folding and metamorphism of the belt have been addressed by many authors, e.g. Campbell (1978, 1980), Nironen (1989) and Kilpeläinen (1998).

### 3.3 Characteristic features of the central Tampere Schist Belt

The Tampere Schist Belt is a well-preserved volcanic-sedimentary belt in the Paleoproterozoic Svecofennian terrain, and is thought to have been formed by extensive mantle-derived crustal growth (see Huhma 1986). In its central part it is composed of an EW-striking major synform (Fig. 1.) with subvertical axial planes and subhorizontal fold axis lineations attributable to the main phase of folding.
(Kähkönen 2005). Because of this major structure it has been common practice for researchers to consider the lithology and petrology of the northern limb in relation to those of the southern limb. The northern limb of the synform is dominated by metavolcanic rocks which are mostly pyroclastic, although sills and rocks of lava origin are also known (Kähkönen 1989). The volcanic rocks vary from basaltic to rhyolitic in composition, and the southern limb of the synform is rich in metasedimentary rocks.

Pillow lavas have been observed with certainty only in the Haveri formation at Viljakkala (Mäkelä 1980). This formation is also the lowermost volcanic unit in the northern limb, and is overlain by turbidites (Osara formation) which are coeval with the oldest turbidites on the southern limb (the Myllyniemi formation) and probably lateral counterparts representing the lower sedimentary group of the central Svecofennian terrain (Kähkönen 1999, Kähkönen 2005). Geochemical studies have indicated that the metalavas of the Haveri formation differ substantially from other metavolcanic rocks within the schist belt being mostly low-K/medium-K tholeiitic basalts (Kähkönen & Nironen 1994), and the concentration of TiO₂, ca.1.7%, is also significantly higher than in the adjacent rocks (Mäkelä 1980). Tectonomagmatic affinities suggest that the Haveri metavolcanics are not arc-type rocks but are more reminiscent of marginal basin basalts. The difference between the rocks of the Haveri formation and the adjacent ones marks a change from a restricted, shallow extensional basin to a larger arc-related basin environment (Kähkönen & Nironen 1994).

In the light of lead and neodymium isotope analyses, Huhma and Vaasjoki (1999) suggest that the Haveri metabasalt may have originated from a source which had been enriched in LREE for a considerable length of time. There was no significant amount of older crustal material available but the overall geochemical and isotopic data for the area suggest that the continental crust was forming there before the culmination of the Svecofennian orogeny, although probably not earlier than 1.95 Ga ago (Huhma & Vaasjoki 1999). Geological and geochemical evidence suggests that tholeiitic mafic volcanism, EMORB-like basalts in an extensional setting, were followed by an arc-type magmatism (Kähkönen 2005).

The Koskuenjärvi formation at Orivesi is the lowermost unit on the northern limb of the major synform, and is assumed to be younger than the Haveri formation or the oldest turbidites in the southern limb (the Myllyniemi formation and the Osara formation at Viljakkala) (Kähkönen 2005). The Koskuenjärvi formation has a U-Pb zircon age of 1904 ± 4 Ma (Kähkönen et al. 1989), making it the oldest volcanic unit dated in the Tampere Schist Belt so far and is characterised
by massive to stratified plagioclase-porphyritic high-K dacites and andesites. Felsic rocks are relatively abundant, but mafic volcanics are rare. The volcanic rocks are largely of a pyroclastic origin typical of sub-aerial and shallow-water environments (Kähkönen 1999). The host rock of the Orivesi Au deposit belongs to this formation.

The Koskuenjärvi formation is overlain by the Valkjärvi formation, which is separated from it by a shear zone, and is divided into three volcanic members on the basis of its chemical composition: a subalkaline member, the oldest, a shoshonitic member and a trachytic member, the youngest (Kähkönen 1999).

The volcanic rocks of the subalkaline member are characteristically of pyroclastic origin, being tuffs and lapilli tuffs, and its strata represent sub-aqueous depositional environments distal from the volcanic centres. The shoshonitic member is separated from the lower subalkaline member by thin-bedded silty mudrock or tuffites, 5–10 metres thick, indicating waning volcanic activity. The formation itself consists of massive and fragmental porphyritic lavas together with tuff breccias, lapilli tuffs and fine-grained tuffs or tuffites. The trachytic member has a wide lateral extent relative to its thickness, and evidently originates from a pyroclastic flow deposit. (Kähkönen 1999).

Trachyandesites or basaltic trachyandesites with shoshonitic affinity also occur in the southernmost part of the schist belt. The alkali-rich character agrees with the fairly high P and LREE contents. In view of the Ti vs. Zr diagram and the high Th/Ta ratios, the southern metavolcanics of the Tampere Schist Belt were formed in an arc-type setting. The shoshonitic to high-K character indicates an evolved arc environment (Kähkönen 1994).

The Takamaa formation in the Ylöjärvi area of the northern limb, is dominated by basalts and basaltic andesites, both pyroclastic and lava rocks, and have tectonomagmatic affinities that are transitional between the arc and EMORB or WPB types (Kähkönen 1999). The Takamaa formation, yielding a U-Pb age of 1889 ± 5 Ma, is the youngest volcanic unit in the schist belt dated so far (Kähkönen et al. 1989).

The sedimentary rocks of the belt occur mostly in the southern part and are weathering products of the volcanic rocks. The strata are characterised by greywackes, siltstones and mudstones, which have been transformed to mica schists and slates (Simonen 1953b, Ojakangas 1986). Intraformational conglomerates occur in substantial amounts in places. Mauri arenites occupy a considerable area in the western part of the belt, comprising a unit almost 2.5 km thick and up to 20 km long (Matisto 1968, Kähkönen 1999).
The composition of the metasedimentary rocks shows variation according to stratigraphic position. The variation in the chemical composition of the graywackes and of the mudstones associated with the uppermost graywackes reflects variations in the character of their provenance and in the volcanic activity. The analytical results suggest that the metasedimentary rocks have prominent LREE enrichments and mostly do not display significant Eu depletions. The bulk of the Palaeoproterozoic source rocks of the sediments were fairly well evolved, but intracrustally differentiated rocks were not important components in the source areas. Even the lowermost turbidites are derived from an evolved volcanic arc rather than from an immature arc. (Kähkönen & Leveinen 1994).

It is possible from the published investigations to establish the following frameworks for the development of the Tampere Schist Belt. The development commenced with extensive mantle-derived generation of a Palaeoproterozoic crust, obviously during a relatively short period of time. There is no Archaean crust or any Archaean remnants. The volcanic rocks vary widely in chemical composition, ranging from rhyolites to basalts, from subalkaline to alkaline, from low-K to very high-K or shoshonitic, and from calc-alkaline to tholeiitic. Intermediate rocks with calc-alkaline affinities dominate or are abundant in most areas, however. The volcanic rocks resemble those of present-day volcanic arcs and comparison with the latter suggests that the crust was relatively thick during the eruption of most of the volcanic rocks. On the other hand, some mafic rocks do not provide convincing indications of arc affinities. The basalts of the Haveri formation and the mafic volcanics of the Takamaa formation in Ylöjärvi, the youngest dated volcanic unit, are relatively rich in Ti, and may indicate temporal or local extensional environments. It is also obvious on the grounds of stratigraphical investigations that the Haveri formation belongs to a relatively early stage and the Takamaa formation represents a late stage. It is thus conceivable that the extensional environments were not restricted to one single episode. The U-Pb zircon ages range from 1904 ± 4 Ma to 1889 ± 5 Ma indicating a very short period of volcanic activity even if the Haveri formation is assumed to be older than 1904 Ma.

### 3.4 Ore deposits and industrial minerals

Metavolcanic rocks occur richly on the northern edge of the Tampere schist belt, whereas sedimentary rocks prevail in the southern part. Signs of hydrothermal
alteration are visible in many places along the zone of metavolcanic rocks, and the hydrothermal activity has resulted in economic ore deposits on some occasions.

Two exhausted mines exist in the schist belt in addition to the Orivesi mine. The Haveri mine is situated roughly 35 km north-west of Tampere (Fig. 1), and the deposit was known from 1737 onwards. It was mined on a small scale for its magnetite in the 18th and 19th centuries and reinvestigation in 1935 indicated that it contained economic grades of Au and Cu. This led a company named Vuoksminika Oy to start mining operations in 1942, but the mine was closed down in 1960, possibly because of a change in company strategy, before the ore deposit totally exhausted. An open pit together with underground mining produced 1.5 Mt of ore containing 2.8 g/t Au and 0.37% Cu (Isokangas 1978). Gold production totalled 4129 kg. The pyrrhotite-dominated auriferous and cupriferous disseminated to massive sulphide deposits at Haveri are hosted by basaltic metavolcanic rocks. The Haveri occurrence is correlated with Cyprus-type deposits and was probably generated at the initial-stage of an island arc (Mäkelä 1980).

The Ylöjärvi mine is located about 20 km north-west of the city of Tampere (Fig. 1). The deposit was discovered by the Geological Survey of Finland as a result of exploration activities in 1937. Outokumpu Oy started mining operations in 1942. By the time of its closure in 1966 the mine had produced slightly over 4 Mt ore, yielding 28 322 t Cu, 49.5 t Ag, 270 kg Au, 427 t W and 567 t As by open pit and underground mining combined (Himmi et al. 1979). The host rock of the ore is a tourmaline breccia in which the fragments consist mostly of tuffite, with a minority of plagioclase porphyrite. The fragments are silicified throughout or at the borders. The matrix contains quartz and tourmaline, the latter occurring as needles 2–20 mm long and arranged radially in clusters. Chlorite, apatite and ore minerals also occur in the matrix. Chalcopyrite is the most important ore mineral, and arsenopyrite the most common, while scheelite is not unusual. Other ore minerals detected at Haveri are cubanite, pyrrhotite, pyrite, mackinawite, sphalerite, stannite, galena, molybdenite, ilmenite, magnetite, cassiterite and uraninite including native Bi, Ag and Au (Himmi et al. 1979). The Ylöjärvi Cu-W deposit has been interpreted as analogous to porphyry copper deposits (Perry 1961, Mikkola 1963, Front 1981). According to some researchers (Gaál et al. 1981, Front 1981) the deposit is genetically related to the nearby batholith, which is mainly granodioritic in composition.

The Eräjärvi pegmatite area is located in the eastern part of the Tampere schist belt, and being about 250 km² in size, is one of the most prominent pegmatite
provinces in Finland. According to Lahti (1981), the pegmatites are Palaeoproterozoic in age and seem to be associated genetically with the late orogenic granites occurring in the area. Most of the dykes are found in schists near or at the contacts of granite plutons. In total about 70 rare-element pegmatite dykes enriched in B, Be, Li, Rb, Cs, Nb, Ta, Sn and P are known. According to the classification created by Lahti (2000), the pegmatites are divided into beryl-columbite, beryl-columbite-phosphate and complex pegmatites enriched in Li minerals. The U-Pb age of the best-known and most prominent dyke, the Viitaniemi pegmatite, is 1794 ± 15 Ma (Teertstra et al. 1993). Small-scale mining was carried out in the district up to the mid-1960s, and altogether about 30 dykes, most of them relatively small, have been mined, mainly for feldspar, quartz, beryl and columbite-group minerals.
4 Geology of the Orivesi area

This chapter is concerned with the rocks occurring in the neighbourhood of the Orivesi deposit and their relationship to the central part of the TSB. Primary importance is attached to the rock types within the hydrothermally altered halo, the ore deposit itself, its form and ore minerals. The various dykes and veins encountered in the mine are also considered.

The Koskuenjärvi formation, which is the lowermost unit on the northern limb of the major synform, i.e. in the Orivesi area, is about 1 km wide and extends at least 10 km in an E–W direction. The formation includes mainly massive to stratified plagioclase-porphyritic high-K dacites and andesites and felsic rocks are relatively abundant, whereas mafic volcanics are rare. The volcanic rocks frequently display pyroclastic origin and evidently erupted in sub-aerial to shallow-water environments, because turbidite interbeds are rare or absent. The relict lithological layering is almost vertical and trends in an E–W direction, whereas the tops of the strata lie to the south. (Kähkönen 1999).

The Orivesi gold deposit is located near the western end of the Koskuenjärvi formation (Fig.1). According to observations in the mine, the mining area and its vicinity, the Koskuenjärvi formation can be divided into two members: a southern suite of hydrothermally unaltered intermediate volcanic rocks and a northern suite which includes the alteration halo and is referred to here as a protolith suite (Fig. 7). The former has not encountered hydrothermal alteration at all. The contact between the unaltered and altered volcanic units is very sharp and visible over a long distance (cf. Fig. 2), and does not give any impression of being of a tectonic nature, with sericite, chlorite and pyrite simply appearing within the altered rock. This was noticed numerous times during the logging of diamond drill cores and underground mapping. The transition zone is repeatedly about 15–20 cm thick and contains no signs of major sedimentation or erosion, but is more like a weathering crust. This feature provides additional evidence supporting the sub-aerial nature of the volcanism, because a submarine environment will produce sediments even within a short time interval. This kind of very sharp contact with an abrupt change in mineralogy cannot be interpreted otherwise than that the suite, which has remained unaltered, is younger than the protolith suite and thus erupted after a period of hydrothermal alteration. The new volcanic activity with accompanying earthquakes and movements may well have been the reason for the hydrothermal circulation coming to an end.
The samples available for dating the Koskuenjärvi formation were collected from the rocks representing the protolith suite a couple of kilometres east of the mine and yielded the aforementioned U-Pb zircon age of 1904 ± 4 Ma (Kähkönen et al. 1989), denoting the oldest volcanic unit in the Tampere Schist Belt dated up to the present.

A subvolcanic porphyritic intrusion, the “hypabyssal Pukala porphyry” in Fig. 1, probably comagmatic with the volcanism, is situated north of the protolith suite and the altered domain. The Pukala porphyry is extensive in area, and is named after the lake whose southern shore it occupies.

The chemical composition of the hydrothermally unaltered volcanic rocks, the protolith suite and the Pukala porphyry varies only slightly, and all the rocks are calc-alkaline in nature (Fig. 3). The Pukala porphyry and the protoliths are rhyolitic in composition and the unaltered volcanic rocks dacitic. The samples representing the porphyry were gathered north of the alteration area (coordinates in Table 2), and those representing the protolith are from outcrops west of the border of the alteration halo, just outside the map in Fig. 7, and contain no indication of alteration except for a few tiny grains of pyrite. Six of the analyses, representing the unaltered volcanic rocks, were published by Luukkonen (1994), and the others are from somewhat south of the ore pipes.

![Fig. 2. The commencement of the building of the inclined drift within the unaltered intermediate volcanic rocks (left) and the shells of the open pit on Pipe 1 (right). The photographs (by T. Ahola 1993) demonstrate the colour difference between the southern unaltered volcanic rocks and the hydrothermally altered rocks.](image)

Although the samples analysed from the protolith suite were all felsic in composition, intermediate volcanic lithologies also exist there. As a whole, the protolith suite contains about 45% felsic volcanic rocks and the rest are intermediate volca-
nic rocks, as shown by diamond drilling observations within the Koskuenjärvi for-
motion. Near the Orivesi deposit, within and outside the alteration area, the pro-
portion of felsic volcanic rocks in the protolith suite is considerably greater. The
unaltered intermediate volcanic rock contains only about 10% felsic volcanic
rocks, principally tuffs in origin.

Fig. 3. Classification and variation in composition of the Pukala porphyry, protolith
suite and unaltered volcanic rocks. X-ray fluorescence analyses by VTT Geoanalytical
Services. Oxides in wt %. Total Fe as FeO*. The boundary between the tholeiitic and
calc-alkaline fields in the AFM diagram is after Irvine & Baragar (1971) and the fields in
the Na₂O + K₂O vs. SiO₂ diagram are after Le Bas et al. (1986).
4.1 Rocks bordering on the hydrothermally altered domain

4.1.1 Unaltered volcanic rocks

The hydrothermally unaltered intermediate volcanic rocks near the contact with the altered rocks are mostly lavas in origin. The groundmass between the plagioclase phenocrysts consists of quartz and biotite with muscovite, apatite and carbonate as accessories. Approximately 100 m to the south, towards the tops of the strata, the rock reverts to an intermediate tuff sedimented in shallow water, and contains narrow layers of light-coloured felsic tuff as decimetre-scale intercalations and narrow interlayers of amphibole schist with carbonate stripes. Layers of this kind, principally of calcareous clay, may conceivably have been the source of the mobilised carbonate seen mostly in narrow hair-like veins even near the contact with the altered rocks. The intermediate volcanic rocks also include the mafic volcanic rocks, which are conceivably sub-volcanic dykes containing principally pyroxene and with plagioclase phenocrysts in places, and have been metamorphosed to uralite porphyrites and uralite plagioclase porphyrites. Whole-rock analyses of the unaltered intermediate volcanic rocks are shown in Appendix 3.

4.1.2 The protolith suite

The characteristics of the protolith suite as seen outside the altered domain and particularly within the alteration halo, in spite of the mask caused by pervasive alteration, are discussed below.

The protolith is mainly of pyroclastic origin and rhyolitic in composition. Intermediate volcanic rocks also occur in a minority both within and outside the alteration halo. The proportion of basic rocks is insignificant. In many cases it is impossible to identify the primary nature of the pervasively altered rocks, but their volcaniclastic structure is recognizable almost everywhere, so that only the size of the clasts varies. The pyroclastic rocks are composed mainly of air-fall deposits, tuff breccias, crystal tuffs and tuffs, while some signs of pumice and ignimbrites also occur. Most clasts suggest a penecontemporaneous felsic to intermediate volcanic source, but none show any evidence of derivation from hydrothermally altered rocks. Some of the clasts are both rounded and subangular, favouring a resedimented rather than primary pyroclastic origin. Autobrecciation of rhyolitic magma is obviously a common phenomenon, and some features may also be inter-
interpreted as representing hyaloclastic breccias and peperites, i.e. the intrusion of rhyolitic magma into sediments. Pelitic meta-sediments are rare.

The existence of primary breccias would have been an important prerequisite for focusing the hydrothermal fluid flow and alteration. Some porous sedimentary or pumiceous and brecciated volcaniclastic units might be relatively permeable, and would have provided a strong lithological control over the fluid flow. In addition to primary porosity, the networks between grains and the fluid flow along fractures as well as the alteration processes themselves may also have led to changes in rock properties with time, due to fluid-rock interaction, giving rise to changes in mineral assemblage and grain size, and corrosion by acid fluids. A direct connection has been noticed between the size of the clasts and the intensity of fluid flow, i.e. the intensity of alteration. The clasts are overwhelmingly larger in the area of the ore pipes than in the surroundings.

4.1.3 The Pukala porphyry

The Pukala hypabyssal intrusion lies immediately north of the alteration halo and is separated from it by a narrow transition zone (Fig. 7). The EW-striking intrusion is ca. 18 km long and 1–2 km wide, bordering on the protolith suite. On the whole the porphyry is a quite homogeneous, light-coloured rock (Fig. 4). The country rocks on the north side of the intrusion are mainly 1880–1890 Ma old (Korsman et al. 1997) and comprise granite, granodiorite, and gabbro of the Central Finland Granitoid Complex (CFGC).

The porphyry near the alteration halo was referred to earlier as “soda granite”, representing a marginal modification of the CFGC. The most typical rock consists mainly of sodic oligoclase and quartz with only small amounts of dark minerals. The porphyritic texture is due to phenocrysts of plagioclase, and in places also quartz phenocrysts or accumulations. Potash feldspar is locally present. The foliation depends on local factors, but is usually distinctive. (Seitsaari 1951).

On the whole, the phenocrysts within the Pukala porphyry vary in quantity and quality, yielding slightly different mineralogical compositions. The intrusion consists mainly of porphyritic granodiorite and trondhjemite, with some granitic, tonalitic and aplitic parts. The plagioclase and K-feldspar phenocrysts in the granodiorite and trondhjemite are 1–5 mm in diameter and usually sericitized and saussuritized, while the fine-grained (0.01–0.2 mm) matrix is mainly quartz with some plagioclase and biotite. The only abundant mafic mineral is biotite (3–8%), which appears in elongated aggregates parallel to the schistosity. The granodiorite variant
contains 40–45% quartz, about 25% plagioclase and approximately 20 vol.% alkali-feldspar. Alkali-feldspar is rare in the trondhjemite variant, but the proportion of quartz is about 60 vol.% and that of plagioclase 30 vol.% The accessories in both types are chlorite, muscovite, epidote, apatite, fluorite, titanite, allanite and zircon.

The gradual contacts between the rock types suggest either a single magma pulse or two pulses closely related in time. The granodiorite forms mainly the interior of the intrusion and the trondhjemite is commonly found near the edges, e.g. at the southern edge north of the hydrothermal alteration halo. The trondhjemite, which has fewer phenocrysts and a matrix of a slightly smaller grain size, crystallised earlier than the granodiorite. Thermobarometry has given crystallization temperatures of 760–790°C (Talikka 2003). The zircon U-Pb age for the granodiorite is 1896 ± 3 Ma (Talikka & Mänttäri 2005).

A number of mafic and intermediate xenoliths of the country rock are found in the south-western parts of the intrusion (Fig. 4c), and there is usually a broad zone along the southern border of the porphyry in particular that contains xenoliths of this kind, which are most likely derived from the adjoining metavolcanic rocks. The xenoliths are composed mainly of plagioclase porphyrite, basic volcanics and mica schists, and are oriented in an east-west direction. They are sometimes angular in shape, but can in many locations be roundish, and vary in diameter from a few centimetres to several metres.

Seitsaari (1951) noted that the composition of the plagioclase in the xenoliths was identical to that in their immediate surroundings, where the anorthite percentage is comparatively high (An25-30). In the pure “soda granite” with only a few xenoliths the plagioclase is distinctly poorer in anorthite (An12-17). Furthermore, the surroundings of the xenoliths are richer in hornblende and biotite than the rock located outward from it. The “soda granite” containing abundant schist inclusions is more mafic than usual throughout. These features suggest that there was a lot of contamination due to partial melting of the xenoliths.

The geochemical data suggest that the Pukala intrusion is a peraluminous volcanic-arc granitoid (Talikka 2003, Talikka & Mänttäri 2005). The Pukala porphyry intruded into subhorizontal volcanic rocks at subvolcanic depths during the early stages of the orogeny, 10–20 Ma before the main regional metamorphic peak. Judging from the stratigraphy observed today, the thickness of the volcanic rocks above the Pukala intrusion at the time of emplacement was approximately 1.5–2.5 km.
During the Svecofennian orogeny the Pukala intrusion was submerged gently to the west and tilted steeply (80–90°) to the south along with the metavolcanic rocks of the northern limb of the major synform. The geophysical and diamond drilling data indicate that the depth of the intrusion is probably at least 1.5–2 km in the west and slightly less in the east.

Comb quartz layering has been discovered at the transition zone between the Pukala porphyry and the alteration halo, the only instance of this encountered in Finland so far.

![Fig. 4.](image)

**Comb quartz layering**

In the north, between the alteration halo and the subvolcanic Pukala porphyry, there is a transition zone nearly 30 m wide where the porphyry turns into altered rock (Fig. 5). Within this zone comb quartz layering occurs in a zone ca. three
metres wide on the southern edge of the porphyry. It has been possible to identify the layering only in one location, in an exploration trench, and thus nothing is known of its continuity in the direction of the porphyry contact.

The comb quartz layers (Fig. 6) are curved in shape, with no notable ptymatic folding, and contain mainly recrystallized quartz of grain size 0.2–0.5 mm although primary, optically homogenous grains of up to 5 mm are preserved in some places. The grains are usually recrystallized either totally or at least on the margins. The few primary quartz crystals are bluish in colour and are turbid under the microscope, conceivably on account of being rich in fluid inclusions, whereas the recrystallized quartz seems to have been partly purified of fluid inclusions.

The thickness of the quartz layers varies from 1 mm to almost 1 cm and is mostly about 5 mm, while the interlayered porphyric aplite also varies greatly in thickness. A single layer can be only 0.5 mm thick. The phenocrysts of oligoclase (An25) are in places about 1.5 mm long, but mostly 1 mm or less, and occur frequently in clusters of a couple of crystals, forming nucleation centres. Quartz and in places K-feldspar also occur as phenocrysts within porphyritic aplite. The matrix consists of fine-grained quartz and plagioclase, grain size < 0.1 mm. The micrographic textures that are so common in the interlayered aplite are missing here, conceivably destroyed in the metamorphic recrystallization, or else these did not exist primarily. There is only a minor amount of K-feldspar, but it is this that is usually the host for intergrowths. The accessory minerals of aplite are biotite, sericite, chlorite, rutile, apatite, tourmaline, zircon and monazite.
Fig. 5. The transition zone. Rhyolitic fragments in a matrix rich in sericite and chlorite, a) ca. 5 m, b) 8 m, and c) 13 m south of the comb quartz layering on the side of the alteration halo. The diameter of the coin is 22 mm.
The intrusions containing comb quartz layers are generally of similar compositions. In terms of major elements they are characterised by high SiO$_2$ (> 74%), Na$_2$O (> 3%) and K$_2$O (> 4%) and low total Fe (FeO < 2%), TiO$_2$ (< 0.2%), CaO (<0.9%) and MgO (< 0.8 %) relative to most granites, and they have high differentiation indices. Although rich in alkalis, they tend to be metaluminous to slightly peraluminous (i.e. corundum normative) rather than peralkaline. (Kirkhan & Sinclair 1988).

Analyses of samples of the sub-volcanic porphyry at Orivesi covering about one square kilometre north of the alteration halo are presented in Table 2. The two samples located nearest to the transition zone (columns 1 and 2) are trondhjemitic in composition and the others are granodioritic. The analyses in Table 2 published by Talikka (2003) represent the entire porphyry.
Table 2. Compositions of the Pukala porphyry. X-ray fluorescence analyses performed by VTT Geoanalytical Services.

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Columns 1 and 2, trondhjemite
Columns 3 to 8, granodiorite
TR, trondhjemite composition as an average of six samples, after Talikka (2003)
GRDR, granodiorite composition as an average of seven samples, after Talikka (2003)
The trondhjemite and granodiorite varieties are nearly similar in composition to those intrusions described by Kirkhan and Sinclair (1988), except that the K content at Orivesi is lower and the Ca content higher, a difference that can be anticipated in view of the mineral composition.

4.2 The hydrothermally altered domain

4.2.1 General features

A picture of hydrothermal alteration can be created in three dimensions on the grounds of knowledge accumulated in connection with the advanced mining operations and associated diamond drilling. The author has been able to compile a geological map not only for the ground level (Fig. 7) but also at the level of +500 (Fig. 9). The data for the ore pipes and their close proximity are naturally more detailed (Fig. 8 and Fig. 10). The total magnetic image of the ground survey data over the hydrothermally altered domain completes the picture given by the geological map (Fig. 11).

The signs of alteration appear gradually as one approaches the domain of intensive alteration from either the east or the west. The increase in chlorite and sericite with pyrite is accompanied by initial expressions of silification through the appearance of quartz augens at some locations. These spots contain fine-grained quartz and display no signs of primary phenocrysts. Thus the border with the hydrothermally altered realm is gradual and its precise location as drawn on the map is a matter of taste. The border had been a winding one originally, even before deformation. The degree of alteration is naturally highest in the proximity of the ore pipes, and more intensive underground, approximately from the level +200 to +550 m, than at the present ground level.

Only at a short distance north of the ore pipes is there a rock type unit of a kind that may conceivably have originally been composed of uralite-plagioclase porphyrite, or locally plagioclase porphyrite, an intermediate-basic rock of lava origin. The pyroxene phenocrysts have undoubtedly been altered to uralite and finally to biotite, indicating fairly pervasive alteration, and the plagioclase has been totally altered to sericite. Quartz, chlorite and andalusite occur in variable amounts. This lithology, extending approximately 700 m in an E–W direction and downwards at least to a depth of 800 m, being thin and barely perceptible at ground level but a unit 30–40 m wide at a greater depth, contains porphyroblasts of magnetite and differs from the rocks in the vicinity by virtue of its slightly higher
magnetic susceptibility (Fig. 11). The intensity of the positive anomaly is 300–400 nT. An unusually extensive, coherent and impermeable protolith like this could have had a far-reaching effect in focusing hydrothermal fluids.

4.2.2 Rock types and mineral assemblages

The lithologies within the alteration halo may be divided into the following groups according to the prevailing minerals:

(1) chlorite-dominant
(2) sericite-dominant
(3) topaz-bearing and topaz-dominant
(4) andalusite-dominant
(5) quartz-dominant

The mineral assemblage concerning main minerals is not complicated, but the minerals vary greatly in abundance. Chlorite, sericite, quartz, andalusite and pyrite are the most prominent minerals in the chlorite-dominant rocks, while sericite-dominant rocks contain sericite, quartz, chlorite, andalusite and pyrite. The main constituents of the topaz-bearing and topaz-dominant rocks are quartz, topaz, sericite and kaolinite. The andalusite-dominant rocks contain quartz, andalusite, sericite, tellurides, gold and pyrite, and the quartz-dominant rocks quartz, sericite, andalusite, tellurides, gold and pyrite.

The primary zonal structure within the alteration halo (Fig. 12) is still recognizable in spite of deformation and metamorphism, and an overwhelming proportion of the alteration halo is occupied by chlorite-dominant and sericite-dominant rocks. The outer, distal zone consists of chlorite-dominant rocks, but sericite increases gradually inwards to become dominant. Andalusite emerges increasingly in the sericite-dominant rocks on the way to the proximal zone, and is encountered in abundance in the proximity of the alteration halo. At the same time the amount of quartz also increases, and finally the majority of the sericite is replaced by quartz and andalusite. Quartz-dominant rocks prevail in the core of the alteration halo, enveloped by andalusite-quartz schist, and andalusite-dominant rocks occur in a minority near or within it. Topaz-bearing lithologies are met with in the inner zone of the alteration halo, including a minority of topaz-dominant rocks.

The chlorite-dominant rocks are chlorite schists that always contain varying amounts of sericite, quartz, pyrite, and in places also andalusite, in addition to chlorite. Biotite is common in porphyroblasts. A volcaniclastic structure is com-
mon, and the clasts are most frequently of lapilli size. An almost monomineralic chlorite schist is very rare, and in that event the chlorite is accompanied by andalusite porphyroblasts. This is probably a pelitic sediment in origin.

The usual accompaniments of sericite in the sericite-dominant rocks are quartz, andalusite, chlorite and biotite in various combinations, so that quartz-sericite and quartz-andalusite-sericite schists are very common. Chlorite schists occur in connection with sericite schists in some cases, reflecting the variation in the primary composition of the protolith. An exact estimate of the interrelation between the amounts of sericite and chlorite is difficult to obtain from hand specimens in the case of some variants due to the very fine grain size and the greenish colour of the sericite. Schists containing almost as much chlorite as sericite usually contain quartz and considerable amounts of andalusite. One exception is the unit south of the porphyry (Fig. 7 and Fig. 9), which contains almost equal amounts of sericite, chlorite and quartz, but andalusite is unusually rare. Sericite also dominates over chlorite in many places there as well, and quartz is usually more common than either sericite or chlorite. This unit, which is classified among the quartz-dominant rocks, is transitional between the sericite-dominant and quartz-dominant categories but differs from those of the latter category that are situated in the proximity of the deposit. A volcaniclastic structure is present everywhere in the sericite-dominant lithologies as well, and the clasts are mostly of lapilli size.

The quartz-dominant rocks, which are the host rocks of the ore, range from almost monomineralic quartz rock to a sericite-quartz schist (cf. Fig. 17) and an andalusite-quartz schist. Andalusite-dominant rocks occur in close connection with them, but in a minority, and a transition in the abundances of quartz and andalusite exists between these end members with respect to all the combinations.

Andalusite is fairly common, and is present in various lithologies within the altered domain. In addition to markedly altered andalusite, fresh grains of andalusite also occur, mostly bluish violet in colour. The chlorite-dominant and sericite-dominant lithologies contain andalusite porphyroblasts, locally up to matchbox size. The growth of porphyroblasts seems to have remained imperfect. The grains are rich in inclusions and do not show any signs of alteration to sericite. Observations indicate that various generations of andalusite occur, created by chronologically different metamorphic events.
Fig. 7. Geological map of the Orivesi Deposit and its surroundings. Ground level \((z = +15)\).
Fig. 8. Detailed geological map of the pipes and the proximal area on the ground level (z = +15).
Fig. 9. Geological map of the alteration halo around the Orivesi Deposit, level z = +500.
Fig. 10. Detailed geological map of the pipes and the proximal area on the level $z = +500$. 
Fig. 11. Total magnetic image of the ground survey data on a 10 m × 50 m grid. The magnetic highs are denoted by red and pink colours, and the lows by blue ones. The magnetic maxima are about 400 nT.
The occurrence of topaz is mainly restricted to the inner part of the alteration halo. A small proportion of the topaz-bearing rocks are almost monomineralic, and the rest are topaz-bearing quartz rocks (Fig. 13). As the amount of topaz decreases, that of quartz increases, resulting in quartz-rich topaz-bearing rocks which locally have only an anomalous gold content. The topaz in the topaz-bearing quartz rocks occurs in kaolinitized spot-like fragments of a size varying from 5 mm to a few decimetres. These fragments are mostly roundish, but can also be angular in places (Fig. 14), and they contain tiny cracks and splits which are usually filled with quartz. The grain size of the topaz is very fine and the fragments had originally been almost monomineralic. Topaz fragments are similarly not totally unknown in sericite-dominant rocks at contacts with topaz-bearing rocks. In this case the fragmentation involves fragments that vary in size and lithology (Fig. 15) and is unlike the primary volcaniclastic structure which is so common throughout the alteration halo. It is impossible to avoid the conclusion that the fragmentation in both the topaz-bearing rocks and those in close proximity to them was an early phenomenon.
Fig. 13. Microphotographs of variants of topaz-dominant and topaz-bearing rocks: a) quartz-topaz rock, b) almost totally altered andalusite porphyroblasts in topaz rock containing minor sericite, c) topaz-quartz rock, and d) a topaz rock fragment in a quartz-sericite schist.
Fig. 14. Bluish kaolinitized fine-grained topaz rock fragments in quartz-sericite schist (upper) and quartz-dominant rock (lower) within a diamond drill core of diameter 42 mm in both photographs.
Fig. 15. The breccia lying at the margin of the unit containing topaz-bearing and topaz-dominant rocks. The fragments consist mainly of sericite-dominant, quartz-dominant and topaz-dominant rocks. The diameter of the coin is 22 mm.

4.3 Ore pipes

4.3.1 General framework

The deposit consists of five pipes, of which Pipe 5 is a downward continuation of Pipe 4, cut off from it only by a basic dyke with a minor fault relevant to the dyke intrusion. Besides the common features, each pipe has its own particular characteristics. The volcaniclastic structure is equally prominent almost everywhere, but the amount of matrix varies between the clasts to some extent. Another common feature is that the pyroclasts are markedly deformed, with a length at least double the primary length (Fig. 16). This also means that the vertical lengths of the pipes are at least double what they were before the deformation and metamorphism.
The host rock of the ore bodies varies slightly being mostly quartz rock, quartz schist or sericite-quartz schist in Pipes 1 and 2 (Fig. 17a,b,c) and andalusite-quartz schist with a minority of quartz-andalusite schist in Pipes 3 and 5. In the most quartz-rich rocks, the quartz rock and quartz schist, the volcaniclastic structure is not always noticeable. The matrix in Pipe 1 consists mostly of pyrite (Fig. 16a), although sericite and andalusite also occur. For that reason the feed mined from Pipe 1 contained 3.4% S, whereas that from Pipe 5 contained only 0.3% S and that from Pipes 2 and 3 remained more or less at the same level. Apart from the pyrite-rich matrix, Pipe 1 contains semi-massive to massive pyrite lenses and fragments (see map in Fig. 10).

The matrix in Pipe 2 is composed mainly of a thin sericite film with a few andalusite grains between tightly packed clasts, whereas the host rock in Pipes 3 and 5 contains more andalusite and the matrix is rich in andalusite and sericite. The clasts are more loosely packed than in Pipe 2 and the amount of matrix also increases noticeably.

Even though sulphur was not analysed systematically, pyrite, the main sulphide mineral, seems to occur more abundantly as a gangue in the upper parts of the pipes. Sulphur decreases as a whole with increasing depth, a trend that becomes more obvious below the level +500 and also applies to the wall rocks. The occurrence of pyrite within the wall rocks is more abundant in the proximity of the ore pipes than in the more distal zones, and the amount of pyrite in the vicinity of Pipes 2, 3 and 5 is greater than within these pipes.
Fig. 16. Variants of the volcaniclastic structures showing strong lineation: a) quartz-rich clasts and a matrix consisting mostly of pyrite, in Pipe 1, sample size ca. 5 cm × 7.5 cm, b) quartz-rich clasts and a matrix consisting mainly of a thin sericite film which still coats most of the clasts after blasting, in Pipe 2, length of the hammer ca. 30 cm, c) the bluish andalusite-rich matrix between clasts containing quartz and andalusite in Pipe 5, view 30 cm × 40 cm, and d) occasionally abundant amounts of matrix containing andalusite and sericite in Pipe 5, length of the hammer ca. 30 cm.
Fig. 17. Microphotographs of typical rocks within the ore pipes and in their proximity: a) quartz rock, b) quartz schist, c) sericite-quartz schist illustrating a volcanoclastic structure with a minor matrix of sericite, d) andalusite-dominant rock with minor quartz and sericite, e) sericite-quartz schist with minor chlorite and opaque pyrite, and f) quartz-sericite schist with chlorite and opaque pyrite. Crossed nicols.
4.3.2 Geometry of the pipes

The information on gold content accumulated via diamond drilling and sludge drilling was combined by the author to obtain a 3-dimensional model of the ore pipes (Fig. 18). The pipes have fairly sharp boundaries, so that a different cut-off, say 3 or 4 g/t instead of 2 g/t, would not have much effect on their indicated form or volume. In the case of Pipes 1, 3 and 4 mining commenced from an open pit, while Pipe 2 appears at a level of +50, although anomalous Au concentrations do occur at ground level (+15).

The best-known part of the pipe system is nearly vertical, migrating to NE with increasing depth, and the hydrothermal channel continues at least to a depth of 1200 m, following approximately the same direction. This continuation has been traced by diamond drilling, but the gold content is lower and the metal is present only sporadically.
Fig. 18. Three-dimensional model of the deposit, viewed from the north-east. The contours of the pipes are shown at a cut-off of 2 g/t Au at each level.
4.3.3 Chemical and mineralogical composition of the ore pipes

The chemical compositions and mineral assemblages of the host rocks in the pipes resemble each other greatly, displaying only minor variation. The purest quartz rock is the host in Pipe 2, the analyses showing that this may contain over 95% SiO₂ (Table 3). The average grain size of quartz is < 0.1 mm, except in the richly occurring narrow, vertical hair-like veins, where it is an order of magnitude greater, as much as 1.5 mm. Sericite is scarce and very fine-grained, as are the ore minerals. Rutile is a common accessory mineral.

Pipe 1 contains a large amount of pyrite, mainly in the matrix between the clasts, but also in semi-massive and massive lens-like structures. This means that the silica content (Table 4) is lower than in pipe 2. In addition to quartz, the host rock in Pipe 1 contains varying amounts of sericite and andalusite. Rutile is also common as an accessory mineral.

When the amount of sericite increases, the result is a sericite-quartz rock which contains about 15–30% sericite by volume. The grain size of quartz is still fine, averaging < 0.1 mm, and varying amounts of chlorite always occur in addition to sericite. The scales of sericite are < 0.2 mm and frequently slightly greenish in colour, the pleochroism varying, with a spectrum X = pale green or yellowish green, Y = Z = pale bluish green or bluish lilac.

It is also possible for andalusite to increase in quantity in the quartz rock instead of sericite, leading finally to andalusite-quartz schist, or less frequently to quartz-andalusite schist. These occur in Pipes 3 and 5. The amount of quartz in the andalusite-quartz rock is mostly about 70% by volume, with a grain size of about 0.1 mm. Andalusite occurs in porphyroblasts of diameter 0.5–1 mm, locally 2–3 mm. Most of the andalusite grains have been altered to sericite, to such an extent that only a fraction of the primary grain is left.
Table 3. Selected whole rock compositions for Pipes 2 and 5. X-ray fluorescence analyses by VTT Geoanalytical Services.

<table>
<thead>
<tr>
<th></th>
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<tr>
<td>SiO₂</td>
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<td>94.1</td>
<td>96.9</td>
<td>96.3</td>
<td>83.2</td>
<td>81.3</td>
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<tr>
<td>TiO₂</td>
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<td>0.30</td>
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<td>0.21</td>
<td>0.593</td>
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<tr>
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<td>14.2</td>
<td>20.4</td>
<td>11.6</td>
<td>11.3</td>
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<tr>
<td>FeO</td>
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<td>0.36</td>
<td>0.06</td>
<td>0.05</td>
<td>0.21</td>
<td>0.10</td>
<td>0.06</td>
<td>0.06</td>
<td>0.08</td>
</tr>
<tr>
<td>MnO</td>
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<td>0.015</td>
<td>0.005</td>
<td>0.005</td>
<td>0.005</td>
<td>0.005</td>
<td>0.007</td>
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<td>0.00</td>
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<td>0.00</td>
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<td>0.076</td>
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<td>0.068</td>
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<td>0.059</td>
<td>0.019</td>
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<tr>
<td>Na₂O</td>
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<td>0.00</td>
<td>0.02</td>
<td>0.20</td>
<td>0.27</td>
<td>0.07</td>
<td>0.09</td>
<td>0.04</td>
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<tr>
<td>K₂O</td>
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<td>3.51</td>
<td>0.665</td>
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<td>0.486</td>
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<tr>
<td>ZrO₂</td>
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<td>0.027</td>
<td>0.012</td>
<td>0.021</td>
<td>0.014</td>
<td>0.015</td>
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<td>0.015</td>
<td>0.014</td>
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<td>P₂O₅</td>
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<td>0.113</td>
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<td>0.077</td>
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<td>99.9</td>
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<td>100.0</td>
<td>99.9</td>
<td>99.9</td>
<td>99.9</td>
</tr>
</tbody>
</table>

Samples nos. 1–6 from Pipe 2, and nos. 7–9 from Pipe 5. Abbreviations: QzS quartz schist; SerQzS sericite-quartz schist; QzAdlS quartz-andalusite schist; AdlQzS andalusite-quartz schist.

Table 4. Selected whole rock compositions illustrating the variation in the quantity of pyrite within Pipe 1. Main components analysed by X-ray fluorescence, S by Leco and base metals by the AAS method, all at VTT Geoanalytical Services.

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<tr>
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<td>SiO₂</td>
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<td>TiO₂</td>
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<td>Al₂O₃</td>
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<td>FeO</td>
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<td>MnO</td>
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</tr>
<tr>
<td>MgO</td>
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<td>0.00</td>
</tr>
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<td>CaO</td>
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<td>0.03</td>
</tr>
<tr>
<td>Na₂O</td>
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<td>0.15</td>
<td>0.00</td>
</tr>
<tr>
<td>K₂O</td>
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<td>1.90</td>
<td>0.529</td>
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<td>ZrO₂</td>
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<td>0.008</td>
<td>0.003</td>
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<tr>
<td>P₂O₅</td>
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<td>0.105</td>
<td>0.015</td>
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<td>40.64</td>
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<tr>
<td>ppm</td>
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<td>2228</td>
<td>278</td>
</tr>
<tr>
<td>Cu</td>
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<td>127</td>
<td>341</td>
</tr>
<tr>
<td>Co</td>
<td>8175</td>
<td>1295</td>
<td>33</td>
</tr>
<tr>
<td>Zn</td>
<td>48</td>
<td>486</td>
<td>124</td>
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</tbody>
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Abbreviations: SerQzS sericite-quartz schist; SemPyr semi-massive pyrite; MasPyr massive pyrite.
4.4 Ore minerals

The most common sulphides within and in the vicinity of the Orivesi gold deposit are pyrite, pyrrhotite, sphalerite, chalcopyrite, galena and arsenopyrite, while the sulphosalts identified are tetrahedrite, boulangerite, bournonite and meneghinite. The most important group of minerals is that of the tellurides, which occur in many species. The extremely quartz-rich host rocks are ordinarily almost free of sulphides, and tellurides are more common than sulphides or sulphosalts in places. An illustrative view of the metal concentrations and the abundance of ore minerals can be obtained from the annual feed to the dressing plant, the average composition of which in 1999 was as follows:

<p>| | |</p>
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<tbody>
<tr>
<td>S</td>
<td>1.8 %</td>
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<td>Fe</td>
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<tr>
<td>Cu</td>
<td>0.012</td>
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<tr>
<td>Ni</td>
<td>&lt; 0.001</td>
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<tr>
<td>As</td>
<td>&lt; 0.03</td>
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<tr>
<td>Te</td>
<td>0.02</td>
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<tr>
<td>Bi</td>
<td>0.011</td>
</tr>
<tr>
<td>Zn</td>
<td>0.014</td>
</tr>
<tr>
<td>Pb</td>
<td>&lt; 0.05</td>
</tr>
<tr>
<td>Au</td>
<td>9.3 g/t</td>
</tr>
<tr>
<td>Ag</td>
<td>2.3</td>
</tr>
</tbody>
</table>

Pyrite is the prevalent mineral within the alteration halo, whereas topaz-bearing and topaz-dominant rocks are exceptional and contain little in the way of ore minerals. Pyrite is minor in abundance in these latter, with chalcopyrite and pyrrhotite more common, and some arsenopyrite can be noted as well. The chalcopyrite is observed to contain hessite and tellurobismuthite inclusions, while the pyrrhotite contains galena.

4.4.1 Iron and titanium oxides

Magnetite is not common within the alteration zone, as it mostly occurs in porphyroblasts of maximum diameter 5 to 7 mm within a special slightly magnetic unit (cf. Fig. 11). Elsewhere magnetite porphyroblasts occur only occasionally, in chlorite-rich rocks. The type of occurrence suggests that magnetite is not a primary hydrothermal mineral but was created by a late metamorphic event. Hematite has
also been identified (Ollila et al. 1990), and is usually a product of the weathering of pyrite.

Rutile is a common oxide in the hydrothermally altered zone, with the majority of it occurring in the interstices of other minerals, and also as inclusions in most minerals. In many places the grains are organised into trails according to the prevailing lineation. A minor portion of the rutile occurs in the form of grain aggregates, the shape and habit of which suggest that they are pseudomorphs after ilmenomagnetite occurring primarily in the protolith. Analyses show that the rutile contains 0.05 wt % Ta₂O₅, 0.41 wt % Nb₂O₅ and 0.10 wt % FeO (Luukkonen 1994).

### 4.4.2 Sulphides

Pyrite is overwhelmingly the most common sulphide within the hydrothermally altered rocks, occurring in the least amounts in the quartz-dominant host rocks and the topaz-dominant and topaz-bearing quartz rocks. The pyrite mostly occurs in disseminations, but also in stripes following the foliation, dots, aggregates of a few grains and mobilized veins and veinlets. It occurs in the matrix of volcanic clasts in Pipe 1, with a considerable proportion in semi-massive to massive bodies as well. The grains of disseminated pyrite are mostly subhedral to euhedral and usually of size 0.1–2 mm. Not only pyrrhotite but also arsenopyrite and chalcopyrite occur with pyrite and as inclusions in it. The semi-massive to massive pyrite in Pipe 1 contains inclusions of chalcopyrite, galena, arsenopyrite and telluride.

According to some microprobe analyses (Table 5) compatible elements such as Co and Ni are not abundant in pyrite, their concentrations being only slightly elevated. It is obvious that variations in pyrite composition occur between different parts of the deposit and the alteration halo.
Pyrrhotite is much less common than pyrite, but increases in amount slightly with depth in the entire deposit, and especially in Pipe 1. Pyrrhotite seems to occur mostly in association with deformed pyrite stripes, as a result of the metamorphic mobilization of pyrite. The anhedral grains, of diameter < 2 mm, are typically free of ore mineral inclusions. Sphalerite and chalcopyrite are common in association with pyrrhotite, and pentlandite has also been identified in connection with it (Sotka 1994). According to Luukkonen (1994), the pyrrhotite contains 0.31% As on average and is almost invariably hexagonal.

Sphalerite is also an uncommon mineral and is seen with the naked eye only on a few occasions. Judging by the varying colour, it varies considerably in composition. It commonly occurs as small grains (diameter <10 µm) in accumulations with chalcopyrite, pyrrhotite and pyrite, while sphalerite inclusions in pyrite are common. The compositions of sphalerite in various petrographic associations are presented in Table 6, the figures in which point to a high Fe content (max. 10.22 wt %) with some arsenic also present.

Table 5. Compositions of pyrite. Results in wt %, n.d. = not determined, n = number of analyses.

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<th>II 6</th>
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<td>n.d.</td>
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<td>47.14</td>
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<tr>
<td>Co</td>
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<td>0.04</td>
</tr>
<tr>
<td>Ni</td>
<td>0.00</td>
<td>0.03</td>
<td>0.01</td>
</tr>
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<td>Cu</td>
<td>0.01</td>
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<td>Zn</td>
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<td>n.d.</td>
</tr>
<tr>
<td>Pb</td>
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<td>n.d.</td>
</tr>
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<td>As</td>
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<tr>
<td>Sn</td>
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<td>n.d.</td>
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<td>n.d.</td>
</tr>
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<td>Sb</td>
<td>n.d.</td>
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<td>0.01</td>
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<td>Ag</td>
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<td>n.d.</td>
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<tr>
<td>Tot</td>
<td>99.89</td>
<td>100.84</td>
<td>100.85</td>
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</table>

1 Semi-massive – massive pyrite in Pipe 1; II Pyrite after Kojonen et al. (1996); III Pyrite after Luukkonen (1994).
Table 6. Average compositions of sphalerite, after Luukkonen (1994). Results in wt %, n.d. = not detected, and n = number of measurement points.

<table>
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<td>Cu</td>
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<tr>
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<td>Cd</td>
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<td>0.06</td>
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<tr>
<td>As</td>
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<td>0.11</td>
<td>0.16</td>
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</tr>
<tr>
<td>Tot</td>
<td>99.49</td>
<td>100.03</td>
<td>99.93</td>
</tr>
</tbody>
</table>

1 Sphalerite in various petrographic associations with pyrite; II Sphalerite inclusions in pyrite; III Fe-rich sphalerite inclusions in pyrite.

Chalcopyrite occurs occasionally and is seen mostly within and in the vicinity of Pipe 1. The striking feature with regard to its occurrence is its mobilization, as it has clearly been driven into small hair-like fractures by deformation and metamorphism in many places to form tiny veinlets and small aggregates.

Chalcopyrite usually occurs in anhedral grains (<1.5 mm) or in association with sphalerite, but it is also found intergrown with bournonite and sphalerite, and with tetrahedrite and native antimony. It may occasionally contain cubanite exsolution lamellae, or less frequently small irregular inclusions. Microprobe analyses have shown that the chalcopyrite contains 0.45% Zn and 0.26% As (Luukkonen 1994). It should also be mentioned that bornite has been identified here (Ollila et al. 1990).

Galena is seen with the naked eye less frequently than chalcopyrite and sphalerite, as its grain size varies from a few microns to 1 mm. It is found with sphalerite and pyrite, and is intergrown with chalcopyrite in certain instances (Luukkonen 1994). Electron microprobe analyses of galena (Table 7) indicate that it may contain small amounts of Bi, Ag, Sb, Au and Hg. Its Se content may be extremely high, and Te also increases with Se. The composition seems to vary from habitual galena via selenian galena to clausthalite. It is generally agreed that a galena-clausthalite solid solution series exists in nature, as suggested by Coleman (1959).
Table 7. Compositions of galena with increasing selenium content. Results in wt %, n = number of analyses, and n.d. = not determined.

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<td>S</td>
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<td>&quot;</td>
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<td>Cu</td>
<td>&quot;</td>
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<td>&quot;</td>
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<td>0.10</td>
<td>0.03</td>
<td>0.04</td>
<td>0.06</td>
</tr>
<tr>
<td>Te</td>
<td>0.12</td>
<td>0.18</td>
<td>0.19</td>
<td>0.47</td>
<td>5.91</td>
</tr>
<tr>
<td>Au</td>
<td>0.20</td>
<td>0.14</td>
<td>0.00</td>
<td>0.26</td>
<td>0.26</td>
</tr>
<tr>
<td>Hg</td>
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<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
<td>n.d.</td>
</tr>
<tr>
<td>Tot</td>
<td>100.66</td>
<td>100.49</td>
<td>100.37</td>
<td>101.04</td>
<td>102.20</td>
</tr>
</tbody>
</table>

1 Galena (Luukkonen 1994); II Galena (Kojonen et al. 1996); III Selenian galena Pb(S,Se) (Kojonen et al. 1996); IV Selenian galena (Kojonen et al. 1996); V Clausthalite PbSe (Kojonen et al. 1996).

Arsenopyrite occurs mainly in small individual euhedral crystals within silicates, and it is thus rarely observed with the naked eye. It may contain inclusions of galena, sphalerite, tellurides and native gold, and may occur sporadically as small subhedral inclusions in pyrite and pyrrhotite. Luukkonen (1994) noted that certain grains proved to be concentrically zoned, with their central parts richer in arsenic than the fringes. The average compositions of arsenopyrite, as presented in Table 8, show that it contains on the whole only small amounts of compatible elements, the concentrations of Co, Ni, Sb and Bi being low.
Table 8. Average compositions of arsenopyrite. Results in wt %, n = number of analyses, and n.d. = not determined.

<table>
<thead>
<tr>
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</tr>
</thead>
<tbody>
<tr>
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<td>21.99</td>
<td>21.74</td>
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<tr>
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<td>43.74</td>
<td>40.72</td>
<td>41.63</td>
</tr>
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<td>Fe</td>
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<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td>Co</td>
<td>0.05</td>
<td>0.03</td>
<td>0.03</td>
</tr>
<tr>
<td>Cu</td>
<td>0.03</td>
<td>0.02</td>
<td>0.02</td>
</tr>
<tr>
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<td>n.d.</td>
<td>n.d.</td>
</tr>
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<td>&quot;</td>
<td>&quot;</td>
</tr>
<tr>
<td>Se</td>
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<td>&quot;</td>
<td>&quot;</td>
</tr>
<tr>
<td>Ag</td>
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<td>&quot;</td>
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</tr>
<tr>
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<td>n.d.</td>
<td>0.01</td>
</tr>
<tr>
<td>Tot</td>
<td>101.14</td>
<td>100.14</td>
<td>100.30</td>
</tr>
</tbody>
</table>

1 Arsenopyrite (Kojonen et al. 1996); II Arsenopyrite in various associations with other minerals (Luukkonen 1994); III Arsenopyrite inclusions in pyrite and pyrrhotite (Luukkonen 1994).

4.4.3 Sulphosalts

The sulphosalts identified in connection with the deposit are represented by tetrahedrite, bornonite, boulangerite and meneghinite (Luukkonen 1994, Kojonen et al. 1996, Kojonen et al. 1999). The quantity of sulphosalts is minor, however, and fine-grained minerals are only identifiable with difficulty other than in the laboratory, although the observations lend support to the opinion that tetrahedrite is the most plentiful sulphosalt. The compositions of the sulphosalts are displayed in Table 9. Tetrahedrite occurs mostly in association with chalcopyrite and has been shown in microprobe analyses to contain a certain amount of Zn, As, Bi and Ag.

Bornonite is observed in association with galena, chalcopyrite and sphalerite, and may contain inclusions of exsolved sphalerite (Luukkonen 1994). The compositions of bornonite indicate that it contains small amounts of the compatible elements As, Bi, Se and Te.

Boulangerite grains are commonly very small (< 10 μm), and their occurrence in association with galena and altaite is readily understandable. The microprobe analysis (Table 9, column V) indicates that the boulangerite contains approximately 55 wt % Pb, and about 0.5 wt % As and Se as impurities.
Meneghinite is another sulphosalt that has been identified and analysed here, the samples coming close to the ideal composition deducible from its chemical formula (Table 9, column VI). Arsenic, Bi and Se are present as impurities.

In spite of their scarce occurrence, sulphosalts are important as hosts to antimony within and in the vicinity of the deposit. Aurostibite (AuSb₂) is the only antimonide to have been identified here up to the present time (Luukkonen 1994, Kojonen et al. 1999).

Table 9. Average compositions of sulphosalts. Results in wt %, n = number of analyses, n.d. = not detected, and open space in table = not determined.

<table>
<thead>
<tr>
<th></th>
<th>I 7</th>
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<th>III 16</th>
<th>IV 3</th>
<th>V 5</th>
<th>VI 8</th>
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<td>19.23</td>
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<td>0.02</td>
<td>0.01</td>
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<tr>
<td>Cu</td>
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<td>1.55</td>
</tr>
<tr>
<td>Zn</td>
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<td>0.02</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pb</td>
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<td>40.77</td>
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<tr>
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<tr>
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<tr>
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</tr>
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<tr>
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<td>99.35</td>
<td>98.79</td>
<td>99.51</td>
<td>99.69</td>
</tr>
</tbody>
</table>

I Tetrahedrite, (Cu,Fe)₁₂Sb₄S₁₃ (Kojonen et al. 1996); II Tetrahedrite in association with chalcopyrite (Luukkonen (1994); III Bournonite, PbCuSbS₃ (Kojonen et al. 1996); IV Bournonite in association with chalcopyrite and sphalerite (Luukkonen (1994); V Boulangerite, Pb₅Sb₄S₁₁ (Kojonen et al. 1996); VI Meneghinite, Pb₁₃CuSb₇S₂₄ (Kojonen et al. 1996).

4.4.4 Tellurides

The deposit contains various tellurides within all the ore pipes. Mineralogical investigations (Luukkonen 1994, Sotka 1994, Huhtelin 1995, Kojonen et al. 1996) suggest that the most common gold, gold-silver and silver tellurides are calaverite (AuTe₂), montbrayite ((Au,Sb)₅Te₇), kostovite (CuAu₄Te₇), petzite (Ag₅AuTe₇), sylvanite ((Au,Ag)₂Te₄) and hessite (AgTe₂), of which calaverite is overwhelm-
ingly the most frequently observed. Petzite and hessite are more common than monbrayite, kostovite and sylvanite, which are fairly rare.

Other known tellurides include tellurobismuthite (Bi₂Te₃), altaite (PbTe), melonite (NiTe₂), frohbergite (FeTe₂), tsumoite (BiTe), tetradymite (Bi₂Te₂S) and rucklidgeite ((Bi,Pb)₃Te₄), while minor ones that have been identified include joseite (Bi₄(S,Te)₃) (Kojonen et al. 1996) and volynskite (AgBiTe₂), and tellurantimony, Sb₂Te₃, has also been identified alongside tellurobismuthite (Sotka 1994). It should be mentioned that tiny crystals of krennerite (AuTe₂) and coloradoite (HgTe) have been identified occasionally (Grönholm 1992, Luukkonen 1994).

The tellurides are mostly of a grain size of 10–50 µm, occasionally 50–100 µm, although tellurobismuthite, the most common Te mineral, mostly forms grains in the size range 100–300 µm and even grain aggregates some millimetres in diameter. Tellurides occur very often as aggregates of two or more species, sometimes as inclusions in base metal sulphides, particularly galena and sphalerite, or with native gold (see Fig. 19). The largest observed aggregates have been several centimetres long and up to 10 mm wide. The compositions of the tellurides in the present material are indicated in Table 10.

The mutual relationships in manner of occurrence and abundance between the tellurides vary slightly in the mineralogical data, with the abundance of calaverite, for example, varying in different parts of the deposit. In many cases calaverite occurs with native gold or contains inclusions of native gold. Petzite is observed to occur together with altaite and is intergrown with tellurobismuthite and frohbergite. Tellurobismuthite is associated with hessite, native gold and galena, and also occurs as inclusions in frohbergite and altaite and in grain aggregates with calaverite. Altaite is usually intergrown with native gold, and it has been found as an inclusion in sphalerite. Both altaite and hessite are common inclusions in galena. Frohbergite occurs in various associations with tellurobismuthite, calaverite, native gold and petzite, but most commonly together with altaite.
Table 10. Average compositions of tellurides. Results in wt %, n = number of analyses, and n.d. = no determination.

<table>
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<tr>
<th></th>
<th>I</th>
<th>II</th>
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<th>V</th>
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<th>VIII</th>
<th>IX</th>
<th>X</th>
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<td>n.d.</td>
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<td>0.04</td>
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<td>0.02</td>
<td>n.d.</td>
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<td>n.d.</td>
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<td>n.d.</td>
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<td>0.02</td>
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<td>61.15</td>
<td>0.08</td>
<td>0.00</td>
<td>0.23</td>
<td>0.11</td>
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<td>0.03</td>
<td>0.06</td>
<td>0.04</td>
<td>0.01</td>
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<td>98.60</td>
<td>100.96</td>
<td>99.91</td>
<td>98.75</td>
</tr>
</tbody>
</table>

I Calaverite, AuTe₂ (Kojonen et al. 1996); II Calaverite (Sotka 1994); III Montbrayite, (Au,Sb)₂Te₃ (Sotka 1994); IV Petzite, Ag₃AuTe₂ (Kojonen et al. 1996); V Kostovite, CuAu₄Te₄ (Kojonen et al. 1996); VI Sylvanite, (Au,Ag)₂Te₄ (Kojonen et al. 1996); VII Hessite, Ag₂Te (Kojonen et al. 1996); VIII Tellurobismuthite, Bi₂Te₃ (Kojonen et al. 1996); IX Tellurobismuthite (Sotka 1994); X Altaite, PbTe (Kojonen et al. 1996); XI Altaite (Sotka 1994); XII Melonite, NiTe₂ (Kojonen et al. 1996); XIII Frohbergite, FeTe (Kojonen et al. 1996); XIV Frohbergite (Sotka 1994); XV Tsumoite, BiTe (Kojonen et al. 1996); XVI Tetradymite, Bi₂Te₂S (Kojonen et al. 1996); XVII Rucklidgeite, (Bi,Pb)₃Te₄ (Kojonen et al. 1996).
The total amounts and mutual ratios of the tellurides vary considerably within the deposit. Pipe 4, for example, is somewhat exceptional as its gold content is fairly high and tellurides are also very common and occur richly. At the same time the amount of sulphides is fairly low and the S content is only 0.1–0.2 wt %. The main sulphides are pyrite and chalcopyrite. The amount of tellurobismuthite locally exceeds 2 wt %, while the total Te content is about 1.9 wt % (Huhtelin 1995). Calaverite, altaite and melonite are also common tellurides, whereas frohbergite and sylvanite are less common. The calaverite in Pipe 4 contains a significant proportion of the gold, at least 1/3 and perhaps even 2/3 of that in the samples investigated and analysed being bound to calaverite (Huhtelin 1995). Conversely, the calaverite and petzite in Pipe 1 contain 15–20% of the gold, the rest being native. The compositions of some of the most common tellurides in Pipe 4 are presented in Table 11.

Table 11. Average microprobe analyses of the most common tellurides in Pipe 4. Results in wt %, n = number of measurements. Analysed by Huhtelin (1995).

<table>
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<th></th>
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<th>II 15</th>
<th>III 3</th>
<th>IV 4</th>
<th>V 4</th>
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<td>0.00</td>
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<td>0.10</td>
<td>61.71</td>
<td>0.07</td>
<td>0.07</td>
</tr>
<tr>
<td>Bi</td>
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<td>49.66</td>
<td>0.13</td>
<td>0.24</td>
<td>0.06</td>
</tr>
<tr>
<td>Ag</td>
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<td>0.01</td>
<td>0.05</td>
<td>0.02</td>
<td>0.01</td>
</tr>
<tr>
<td>Sb</td>
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<td>0.01</td>
<td>0.02</td>
</tr>
<tr>
<td>Te</td>
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<td>49.07</td>
<td>38.69</td>
<td>81.31</td>
<td>81.89</td>
</tr>
<tr>
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<td>0.16</td>
<td>0.55</td>
</tr>
<tr>
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<td>100.00</td>
<td>101.36</td>
<td>100.55</td>
<td>99.94</td>
</tr>
</tbody>
</table>

1. Calaverite (AuTe₂); II Tellurobismuthite (Bi₂Te₃); III Altaite (PbTe); IV Melonite (NiTe₂); V Frohbergite (FeTe₂).

4.4.5 Native gold, tellurium, lead, antimony and copper

Native gold particles occur either as single grains or as groups of grains and are unevenly distributed. The grain size is usually very small, in general < 20 µm and frequently 5–10 µm (Fig. 19). Grains of size 1–2 µm also occur, and some even less than 1 µm. The grains found in connection with telluride aggregates are usually the largest, about 40–60 µm in diameter. If aggregates comprise several species of tellurides, native gold is regularly present. Gold inclusions occur most often in altaite or tellurobismuthite, but also in calaverite and frohbergite. Conversely,
no observations exist of petzite or hessite that contains native gold inclusions. Apart from intergrowths with tellurides, native gold commonly occurs in its host rocks as inclusions in quartz grains and at quartz grain boundaries, and can also be detected as inclusions in sulphides. At least pyrite and arsenopyrite have been observed to possess gold inclusions.

Native gold contains mostly 5–6 wt % Ag (Table 12), with minor amounts of Pb, Bi, Te and Fe as impurities. The occurrence of electrum is sporadic due to a scarcity of silver throughout the deposit. It is obvious that a notable proportion of the silver is bound to gold-silver and silver tellurides (petzite and hessite). According to Luukkonen (1994) the native gold and electrum, like the tellurobismuthite, altaite and hessite, contain minor amounts of Hg as an impurity.

Native tellurium occurs in association with tellurides and can also be found as small inclusions (<20 µm) in pyrite. The composition of native tellurium is shown in Table 12 (column VI).

Native lead has been identified in association with galena. The grains can be up to 100 µm in diameter and contain mainly Fe, Au and Te as impurities, as well as minor amounts of Hg and Cu (Luukkonen 1994). The native antimony grains contain Bi and Pb as major impurities (Table 12, column VII). Small native copper grains have also been observed (Grönholm 1992).

**Table 12. Average compositions of native gold, electrum, native tellurium and native antimony. Results in wt %, n = number of determinations, and open space in table = not analysed.**

<table>
<thead>
<tr>
<th></th>
<th>I</th>
<th>II</th>
<th>III</th>
<th>IV</th>
<th>V</th>
<th>VI</th>
<th>VII</th>
</tr>
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<tr>
<td>n</td>
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<td>4</td>
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<td>4</td>
<td>4</td>
</tr>
<tr>
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<td>0.01</td>
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<tr>
<td>Fe</td>
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<td>0.01</td>
<td>0.69</td>
<td>0.00</td>
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</tr>
<tr>
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<td>0.02</td>
<td>0.01</td>
<td>0.03</td>
<td>0.00</td>
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</tr>
<tr>
<td>Co</td>
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<td>0.02</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cu</td>
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<td>0.03</td>
<td></td>
<td></td>
<td></td>
<td>0.01</td>
<td>0.07</td>
</tr>
<tr>
<td>Zn</td>
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<td>0.02</td>
<td></td>
<td></td>
<td></td>
<td>0.06</td>
</tr>
<tr>
<td>Pb</td>
<td>0.10</td>
<td>0.09</td>
<td>0.05</td>
<td>0.00</td>
<td>0.00</td>
<td>0.06</td>
<td>1.61</td>
</tr>
<tr>
<td>Se</td>
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<td>0.02</td>
<td></td>
<td></td>
<td></td>
<td>0.07</td>
<td></td>
</tr>
<tr>
<td>Ag</td>
<td>6.62</td>
<td>5.93</td>
<td>5.10</td>
<td>26.59</td>
<td>24.77</td>
<td>0.27</td>
<td>0.00</td>
</tr>
<tr>
<td>Au</td>
<td>91.63</td>
<td>91.74</td>
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<td>70.93</td>
<td>75.09</td>
<td>0.29</td>
<td>0.13</td>
</tr>
<tr>
<td>Hg</td>
<td></td>
<td></td>
<td></td>
<td>0.05</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bi</td>
<td>0.34</td>
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<td>0.15</td>
<td>0.02</td>
<td>0.16</td>
<td>0.04</td>
<td>3.24</td>
</tr>
<tr>
<td>Sb</td>
<td>0.02</td>
<td>0.00</td>
<td>0.02</td>
<td>0.16</td>
<td>0.60</td>
<td>92.26</td>
<td></td>
</tr>
<tr>
<td>Te</td>
<td>0.02</td>
<td>0.04</td>
<td>0.16</td>
<td>0.00</td>
<td>0.02</td>
<td>98.59</td>
<td>0.03</td>
</tr>
<tr>
<td>Tot</td>
<td>99.58</td>
<td>98.88</td>
<td>100.08</td>
<td>97.83</td>
<td>100.84</td>
<td>100.93</td>
<td>97.79</td>
</tr>
</tbody>
</table>

*Native gold (Kojonen et al. 1996); II Native gold (Sotka 1994); III Native gold (Huhtelin 1995); IV Electrum (Kojonen et al. 1996); V Electrum (Sotka 1994); VI Native tellurium (Luukkonen 1994); VII Native antimony (Kojonen et al. 1996).*
Fig. 19. Microphotographs of tellurides and of the occurrence of gold: a) altaite (white) and hessite (grey), length of the hessite grain 0.25 mm, b) a gold grain (10 µm) embayed in brown chalcopyrite, white calaverite and frohbergite, c) a gold grain (ca. 40 µm) encircled in white altaite and minor grey hessite, d) a gold grain (12 µm) intergrown with petzite (light brown) and tellurobismuthite, and e) gold grains of size < 10µm. Original photographs by K. Kojonen.
4.4.6 Parameters of gold occurrence within the deposit

A volcaniclastic structure is ubiquitous within the deposit and in its vicinity, with the clasts or fragments varying greatly in size, from centimetre to decimetre scale. The proportion of matrix also varies, from ca. 10 to 40% by volume. The distribution of gold between matrix and clasts has been under investigation with regard to Pipe 5 below the level +235, where the volcaniclastic structure is most easily distinguished and the amount of matrix is considerable. Fragments contain about 80–90% quartz, 5–10% sericite and 5% andalusite, while the matrix contains approximately 40% quartz, 50% andalusite, 10% sericite and tellurides to the extent that they are visible with the naked eye. According to some analyses the Au content of the clasts is low, only about 5% of total gold, the rest being situated in the surrounding matrix (Fig. 20). The situation is very much the same with tellurium and bismuth. It is conceivable that the proportion of gold in the fragments increases as the proportion of matrix decreases. In Pipe 2, for example, the matrix is a thin film composed mainly of sericite between quartz-rich clasts, but the pipe had a conspicuous gold content, even extremely high in some stopes.

Fig. 20. Distribution of Au, Te and Bi between fragments and the matrix within Pipe 5. Unpublished data from H. Saarnio.

Some estimates of the frequencies of different mineralogical modes of gold occurrence have been made concerning the deposit as a whole (Fig. 21). It seems to be that almost half of the gold grains are intergrown with various tellurides, while approximately one quarter of the gold is bound to gold tellurides. The remainder consists of native grains, either as inclusions in quartz or located between grains of quartz and other silicates. A small minority of the native gold occurs as inclusions in sulphides and arsenides.
4.5 Dykes and veins

The dykes and veins occurring in the area whose age relations are best visible in the mine can be classified into two groups: mafic dykes and quartz veins. The former group includes two categories, older and younger mafic dykes, while the quartz veins, which are younger than the quartz segregations and veins related to volcanism, can be divided into planar, nearly vertical milky quartz veins, quartz-tourmaline veins that are also nearly vertical and sub-horizontal quartz veins.

4.5.1 Mafic dykes

Older mafic dykes occur frequently at various levels in the mine and vary widely in thickness. Some dykes are about two metres wide and others a couple of centimetres wide. The strike is approximately east-west and the dip vertical. Some dykes, when occurring along the drift, have been visible over a distance of 20
metres. The dykes consist of fine-grained plagioclase (andesine), quartz and amphibole porphyroblasts, which are sometimes in the form of needles at least 3–4 mm long and contain quartz and plagioclase inclusions. The amphibole porphyroblasts are usually unoriented, but the needles cross the schistosity. Accessories are epidote, biotite and sometimes carbonate. The narrow dykes contain sulphides throughout, but those that are several decimetres in thickness contain sulphides only on the endocontact, as if the sulphides had migrated from the side of the altered rocks. The main sulphide is pyrite, with pyrrhotite in the minority. Chalcopyrite occurs occasionally, as does magnetite.

The timing of the intrusion of the older mafic dykes is puzzling. There is no age determination available, and an estimate has to be made on geological grounds. The absence of sericite, and in a few cases the presence of carbonate, may indicate that the dykes have not encountered hydrothermal alteration, so that their intrusion obviously took place after the alteration. The mobilization of sulphides into the dykes and the amphibole porphyroblasts together indicate that the dykes have been involved in deformation and metamorphism, at least in the latest phases. It is conceivable that the dykes may belong to the age group of EW-striking metamorphosed dykes that are commonly composed of uralite-porphyrites and are approximately 1880 Ma old as known in southern Finland (Pesonen & Neuvonen 1981).

The younger class of mafic dykes is rare in the mine but more common in the surroundings. These are usually equipped with chilled margins, show no signs of sulphide mobilization and are undisturbed in every respect.

A very extensive diabase dyke swarm occurs in the Tampere – Heinola – Lappeenranta area of southern Finland, comprising unmetamorphosed Subjotunia diabases formed 1650–1600 Ma ago. The dykes cut the Svecokarelian metamorphic complex in the direction N 50º W with sharp contacts, and show well-developed chilled margins. According to field observations, they are older than the rapakivi of SE Finland, but in any case the swarm may be geotectonically related to the rapakivi in view of the fact that the largest dykes lie close to rapakivi intrusions (Laitakari 1987). There is only minor doubt as to whether the younger basic dykes belong to this group.

4.5.2 Milky quartz veins

Milky, planar, almost vertical quartz veins are very common, occurring most frequently in connection with Pipe 2, which is also the richest in quartz, so that the
mobilization of quartz would seem natural. The veins occur many times in swarms running quite close to each other in a NE–SW direction and are several metres in length and seldom of a thickness greater than a decimetre. The veins are barren of ore minerals and indicate no concentration of gold. Sulphides are not enriched even at the contacts, where conversely apatite has sometimes been noted.

4.5.3 Quartz-tourmaline veins

Quartz-tourmaline veins (Fig. 22) occur mostly with milky quartz veins in connection with Pipe 2, usually within the ore body but occasionally outside it. The shape is mostly that of a proper vein, sometimes spot-like or cloud-like, but it may also resemble a segregation. The contacts with the quartz rock seem not to be very sharp, but are for the most part gradual. The relative abundances of tourmaline and quartz vary, sometimes being equal, although quartz is usually more abundant. Tourmaline occurs in crystal aggregates which are spot-like or elongated in shape. It is nearly black in colour in the hand specimen, with a grain size of about 0.2 mm, and the longest prisms are about 1 mm. Its colour and pleochroism suggest that the composition varies slightly. The average grain size of the quartz in the veins is 0.2 mm, and the colour is bluish or greyish and does not differ greatly from that of the quartz in the quartz rock. Minor amounts of sericite occur in the interstices of the quartz grains. No special concentration of gold has been observed in the veins.
In addition to the black tourmaline in veins, a very uncommon green one has been noticed in the mine. It is conceivable that this tourmaline, which does not occur in veins together with quartz, may be of primary origin and belong to the hydrothermal alteration assemblage. The composition of this tourmaline, identified as dravite, is different from that of the black tourmaline in the quartz-tourmaline veins (M. Poutiainen, unpublished data). The average tourmaline compositions are presented in Table 13.

### 4.5.4 Age relations of dykes and veins

All the rock types are intersected by irregular but mostly sub-horizontal quartz veins in which the grains have a light rosy tint and the vein sometimes contains sulphide minerals. These veins are not common, and the most important point concerning them is that they represent the youngest generation of quartz veins. The age relations between the dykes and veins are indistinct, however. The quartz-tourmaline veins favour the same locations as the older group of mafic dykes, and it has been noted that fragments of basic dykes are situated in quartz-tourmaline veins, which suggests that the latter are younger than the mafic dykes.
Table 13. Average tourmaline compositions, in wt %. Ku 117 is green tourmaline (dravite), Ku 263 black tourmaline in a quartz-tourmaline vein, and n is the number of determinations.

<table>
<thead>
<tr>
<th></th>
<th>Ku 117</th>
<th>Ku 263</th>
</tr>
</thead>
<tbody>
<tr>
<td>n</td>
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<td>8</td>
</tr>
<tr>
<td>SiO₂</td>
<td>37.53</td>
<td>35.86</td>
</tr>
<tr>
<td>TiO₂</td>
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<td>0.346</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>35.46</td>
<td>32.22</td>
</tr>
<tr>
<td>Cr₂O₃</td>
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<td>0.00</td>
</tr>
<tr>
<td>FeO</td>
<td>0.97</td>
<td>4.86</td>
</tr>
<tr>
<td>MnO</td>
<td>0.09</td>
<td>0.07</td>
</tr>
<tr>
<td>MgO</td>
<td>9.47</td>
<td>8.48</td>
</tr>
<tr>
<td>CaO</td>
<td>0.245</td>
<td>0.613</td>
</tr>
<tr>
<td>Na₂O</td>
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</tr>
<tr>
<td>K₂O</td>
<td>0.01</td>
<td>0.008</td>
</tr>
<tr>
<td>ZnO</td>
<td>0.065</td>
<td>0.13</td>
</tr>
<tr>
<td>SnO</td>
<td>0.00</td>
<td>0.052</td>
</tr>
<tr>
<td>F</td>
<td>0.67</td>
<td>0.503</td>
</tr>
<tr>
<td>B₂O₃</td>
<td>11.12</td>
<td>11.58</td>
</tr>
<tr>
<td>Total</td>
<td>97.23</td>
<td>96.53</td>
</tr>
</tbody>
</table>

Milky quartz veins and quartz-tourmaline veins also occur at the same locations, with the latter even crossing the former, especially in connection with Pipe 2, where the longitudinal direction of the quartz-tourmaline veins deviates by 10–15 degrees in an eastern direction from that of the milky quartz veins. Thus the quartz-tourmaline veins seem to cut across the milky quartz veins and are younger, although the age difference is not necessarily a conspicuous one. It is also possible that there exist two generations of quartz-tourmaline veins, with a minor age difference. The contact relations between the older mafic dykes and the milky quartz veins have not been determined, but it is probable that the latter are younger than the older group of mafic dykes and, together with the quartz-tourmaline veins, are older than the younger mafic dykes. The youngest sub-horizontal quartz veins formed almost contemporaneously with the younger mafic dykes or soon after them.
5 Geochemical features resulting from ore deposition and hydrothermal alteration

The compositions of the hydrothermal fluids are critical parameters controlling the nature of a hydrothermal ore, and it is these that reflect the process that leads to the generation of the fluids. The interaction between the fluids and the wall rock results in certain alteration assemblages, but it is generally agreed that there are a lot of differences between the elements enriched in different types of gold deposits and in alteration halo selvages.

Alteration assemblages in Archaean orogenic lode gold deposits, for instance, generally show enrichment in K, CO₂ and S, and a metal association of Au, Ag ± As, B, Bi, Sb, Te and W is displayed by most deposits (Hagemann & Cassidy, 2000). In Proterozoic lode gold deposits the consistent association of pathfinder elements is Ag, Bi, Sb and more importantly As in conjunction with gold (Partington & Williams 2000). Most newly recognized gold deposits related to reduced granitic intrusions display a consistent and striking Au-Bi-Te-As (W, Mo, Sb) association (Thompson & Newberry 2000), while it may be mentioned that a Fe-Cu-Au-REE association is typical of iron oxide Cu-Au deposits, another newly recognized type, commonly with anomalous Ag, As, Co, F, Mo, Nb, Ni, P and U (Kerrich et al. 2000).

The metals and elements that are enriched in epithermal low-sulphidation deposits are Au, Ag, Zn, Pb, Cu, As, Sb, Se, Ba, Mn, Hg and Tl, while in high-sulphidation deposits they are Cu, Au, Ag, Bi, Te, Sb, As, Mo, Sn, Pb and Hg (Hedenquist et al. 1996; Hedenquist et al. 2000). It is significant that epithermal gold deposits related to alkaline magmatism are characterized by telluride-rich mineralization in which hydrothermal quartz and acid styles of alteration are poorly developed or absent (Jensen & Barton 2000). Each deposit is unique, however, regardless of the ore type and displays a particular combination of metals, most frequently from among those mentioned above. Various data sets are available for studying the Orivesi deposit and determining the elements enriched in it.

5.1 Distribution of Au, Te and Bi within the Orivesi deposit

Tellurium and bismuth are highly characteristic of the deposit. Previous researchers have noted that these are greatly enriched within the ore and show a close correlation with each other and with Au (Rasilainen & Nurmi 1999).
Analyses are available for the purpose of reviewing the distribution of these elements within the deposit, since in addition to diamond drilling, the exact shape, volume and element content of the deposit was determined by sludge drilling in each pipe and at each mining level, yielding numerous samples, analyses of which provide means of estimating Au, Te and Bi in situ in the entire deposit and at various levels in the separate pipes.

The samples containing ≥ 2 ppm Au were chosen as a basis for reviewing the entire deposit, and the data also suggest that the gold really does have a very good positive correlation with Te and also with Bi (Fig. 23). The highest values of Te and Bi are about 1%, with those of Bi exceeding those of Te on only a very few occasions, since the maximum Bi content is generally lower than that of Te or locally equal to it. The average Te content of 2439 samples was 196 ppm and Bi content 99 ppm (Au 11.3 ppm). The Te/Bi ratio was almost exactly 2:1.

Certain mining levels were chosen for the review of the distribution of Au, Te and Bi, according to depth within the deposit and the availability of sufficiently large numbers of analyses (Fig. 24, 25, 26 and 27). Data on samples containing Au > 0.1 ppm were included. The separate pipes varied in both shape and volume and the tellurium content was highest in Pipe 5 (Fig. 27), as also noted in the hand specimens. A further feature typical of this pipe was that the Bi content was also frequently very high. The cross-sectional area of the pipe was greatest between the levels +250 and +500, and the gold content was high at the same place. Pipe 4, lying above Pipe 5, was very small in cross-section, but extremely rich in Te and Bi, and the gold content was also high in some places. Unfortunately, no systematic data are available for characterizing Pipe 4.

The gold content of Pipe 2 (Fig. 25) was approximately equal to that of Pipe 5, and the cross-sectional area was greatest between the levels +200 and +500. Tellurium and Bi contents were slightly lower than in Pipe 5, however. The major part of Pipe 1 (Fig. 24) was located between the levels +100 and +250, an interval that was excluded from this review on account of the lack of sufficient analyses. The gold content was considerably lower than in Pipe 2, and the Te and Bi contents slightly lower. The best part of Pipe 3 (Fig. 26) extended from +250 to +350, but Au generally decreased with depth, and the Te content was lower than in the other pipes and the Bi content conspicuously low, especially at the deeper levels.
Fig. 23. Au, Te and Bi content of samples containing Au ≥ 2 ppm in the entire deposit.
As a whole the data indicate that Au content decreases downwards in Pipes 1, 3 and 5, peak values become more infrequent with depth and contents are generally reduced. The trends in Te and Bi are more or less the same, although not quite so striking in Pipe 2. The inevitable conclusion is that the trends in Te and Bi follow the decrease in gold with depth.

Fig. 24. Gold, Te and Bi in Pipe 1 at various levels in the mine.
Fig. 25. Gold, Te and Bi in Pipe 2 at various levels in the mine.
Fig. 26. Gold, Te and Bi in Pipe 3 at various levels in the mine.
Fig. 27. Gold, Te and Bi in Pipe 5 at various levels in the mine.
5.2 Elements characterizing hydrothermal alteration

The following review of the dispersion of elements within the alteration halo is based on two data sets: analyses of samples taken from outcrops by the former Oy Lohja Ab company at the time of discovery of the Orivesi gold deposit in 1982, covering the elements Au, Ag, Sb, As, Sn, Mo, W, Zn, Ba, Sc, La, U and Th as far as this review is concerned, and samples collected from among the diamond drill cores obtained in the proximity of the deposit in connection with contractual research performed by the Geological Survey of Finland for Outokumpu Mining Oy in 1996. Most of the latter samples were located within the alteration halo in the proximity of the deposit, within the ore pipes or in their immediate vicinity. The aim of the investigation had been to identify the best pathfinder elements for Au in the vicinity of the gold deposit (Rasilainen 1996; Rasilainen & Nurmi 1999). The analyses of Au, Ag, Te, Bi, Sb, Mo, Cu, Zn, Pb, As, S, Se, Th, Ni, Co, Cr, Li, Sr, Ba, Sc, Y, La, P and V are considered here.

The network of samples chosen for this review covers a stripe that is 700 m long in a N–S direction and 300 m wide in an E–W direction (coordinates 6838200–6838900 N and 2508400–2508700 E), and runs from unaltered volcanic rocks in the south through the alteration halo and ore pipes to the northern porphyry (Fig. 28).

Unfortunately, the assortment of elements analysed is not quite the same in the two sets of material. When an element such as Au is being analysed in the two overlapping sets of data, the picture of its dispersal is nevertheless complete, covering the whole profile, although otherwise it would remain slightly incomplete. In spite of the differing coverage of the two sets of samples, the results of the analyses reveal the general levels of the contents of the elements in the alteration halo as compared with the porphyry and the slightly younger unaltered volcanic rocks, and display trends towards either enrichment or depletion. It is significant, and essential, that there are also a lot of samples from within the ore pipes, which indicate the contents in the ore itself.

The elements under examination are displayed in groups mainly according to their chemical properties e.g. those forming precious metals or base metals and radioactive elements, or elements that follow each other in ore-forming processes.
5.2.1 Gold and silver

The enrichment of gold (Fig. 29) is prominent in the proximity of the ore pipes, and offers a good spatial reference for the other elements. Also, the whole alteration halo is anomalous with respect to gold and sharply delimited in the south. The enrichment coefficient is approximately $10^4$ relative to the average gold content of igneous rocks 0.005 ppm (Rankama & Sahama 1949), or its crustal abundance 0.004 ppm (Govett 1983).

The amount of silver, and hence its enrichment, is appreciably lower, as the highest concentrations, in the proximity of the ore pipes, are slightly over 10 ppm. The average silver content of igneous rocks is 0.1 ppm (Goldschmidt 1937a; Rankama & Sahama 1949) and the crustal abundance is estimated to be 0.08 ppm (Hamaguchi & Kuroda 1959).

5.2.2 Tellurium, bismuth and antimony

The average tellurium content, including all rock types, is 0.0018 ppm (Goldschmidt 1937b) and the natural range reported for Te is 0.001–0.005 ppm in
maggmatic rocks (Hall & Pelchat 1997), which means that the enrichment within the ore pipes is much higher than for gold, and the entire alteration area is highly anomalous (Fig. 29).

The abundance of bismuth in crustal rocks is 0.025 ppm (Cox 1989), and according to determinations of Goldschmidt (1937b), its average content in igneous rocks is 0.2 ppm. The highest bismuth concentrations indicate a remarkable enrichment coefficient both within the pipes and in the alteration halo (Fig. 29).

Antimony is also enriched in the alteration halo, but not to such an extent as tellurium and bismuth (Fig. 30). The abundance of antimony in the most common rock types is 0.1 to 0.3 ppm (Govett 1983) and the average content in igneous rocks is 1 ppm (Goldschmidt 1937b).

The trend in the enrichment of Te, Bi and Sb coincides well with the dispersal of gold within the ore pipes, in their proximity and in the alteration halo.

5.2.3 Molybdenum, tungsten and tin

The distribution of molybdenum is interesting. Since igneous rocks contain an average of 2.5 ppm Mo (Sandell & Goldich 1943), hardly any enrichment can be said to exist in the proximity of the deposit, but the distal alteration halo is markedly anomalous in that molybdenite is fairly rare in its occurrence there (Fig. 30).

The W content displays neither conspicuous local enrichment nor depletion throughout the section from the unaltered volcanic rocks to the porphyry (Fig. 30), but the results should be regarded only as semi-quantitative due to possible contamination caused by the sampling and preparation methods. Nevertheless, the tungsten content is fairly high. Scheelite has sometimes been identified and is the main carrier of W. The abundance of tungsten in crustal rocks is 1.0 ppm (Govett 1983) and 1.5 ppm in silicic and intermediate igneous rocks (Sandell 1946).

The crustal abundance of Sn is thought to be 2.5 ppm (Govett 1983). The proximity of the deposit, and especially the vicinity of Pipe 1, is slightly enriched in tin (Fig 30). Cassiterite has been identified (Ollila et al. 1990) and the topaz rock contains 15–40 ppm Sn (Saikkonen 1981). Tin can be a participant in sulphides and sulphosalts.
Fig. 29. Distribution of Au, Ag, Te and Bi on a profile across the alteration halo.
Fig. 30. Distribution of Sb, Mo, W and Sn on a profile across the alteration halo.
5.2.4 Copper, zinc and lead

The average crustal abundance of copper is 60 ppm (Govett 1983). Most of the highest figures in the Orivesi alteration halo are situated within the ore pipes or in their vicinity (Fig. 31). The copper content is low by comparison with many other gold deposits of the same type. Igneous rocks contain 80 ppm Zn on average (Sandell & Goldich 1943). Zinc is enriched most around Pipes 1 and 2 (Fig. 31). The distribution of lead resembles that of zinc in many features (Fig. 31) and lead has an estimated average crustal abundance of 16 ppm (Govett 1983).

The averages for all the samples under review here are 113 ppm Cu, 327 ppm Zn and 83 ppm Pb, so that lead may be said to be enriched by a factor of about five, zinc four and copper only two in relation to their crustal abundances or average contents in igneous rocks.

5.2.5 Arsenic, sulphur and selenium

The average As content of igneous rocks is 5 ppm (Goldschmidt & Peters 1934) and the crustal abundance is 2 ppm (Govett 1983). An enrichment of As is visible almost throughout the alteration halo, but it is striking in the vicinity of Pipe 1 (Fig. 31). The peak sulphur content is also reached around Pipe 1, where a level of ten percent is attained locally (Fig. 32).

The abundance of selenium in crustal rocks is 0.09 ppm, and the S : Se ratio in igneous rocks is 6,000 : 1 (Goldschmidt 1937b). Selenium is enriched in the alteration halo (Fig. 32), and the trend in its distribution is very similar to that for sulphur and arsenic.

5.2.6 Uranium and thorium

Granitic rocks contain 4 ppm U on average (Senftle & Keevil 1947), while the thorium content of igneous rocks is 11.5 ppm (Goldschmidt 1937b), and that of granite in particular 13.45 ppm (Senftle & Keevil 1947). The Th : U ratio varies from 3.4 in acidic igneous rocks to 4.0 in intermediate ones (Senftle & Keevil 1947). It should also be mentioned that the crustal abundance of uranium is 3 ppm and that of thorium 10 ppm (Govett 1983). The thorium and uranium contents in the alteration halo are of this order, and there is no major enrichment or depletion, although a few occasional samples may contain amounts above this level (Fig. 32).
The main carrier of thorium and uranium in the alteration halo is zircon, and thucholite, on the rare occasions when it occurs, also contains brannerite inclusions and hence uranium and thorium.

Fig. 31. Distribution of Cu, Zn, Pb and As on a profile across the alteration halo.
Fig. 32. Distribution of S, Se, U and Th on a profile across the alteration halo.
5.2.7 Nickel, cobalt and chromium

The abundance of nickel and cobalt varies greatly in different types of igneous rocks, from ultramafic to felsic. The average cobalt content of igneous rocks is 23 ppm and that of nickel 80 ppm, yielding a Co : Ni ratio of 0.29 (Sandel & Goldich 1943). The chromium content also varies greatly. Some granites contain 2 ppm Cr (Goldschmidt 1937b) or in some regions 2 to 6.8 ppm (Sahama 1945), and felsic rocks generally have 25 ppm on average (Govett 1983). The chromium content of the primary protolith is not known, but it is conceivable there has been neither enrichment nor depletion within the alteration halo. Conversely, cobalt, and more especially nickel, have been depleted during the alteration process (Fig. 33).

5.2.8 Lithium, strontium and barium

The data regarding the alkaline metals are limited to lithium (Fig. 33) and the information that is available on the alkaline earths strontium and barium (Fig. 34). The abundance of lithium in igneous rocks is 65 ppm (Strock 1936; Rankama & Sahama 1949), and the estimated crustal abundance 30 ppm (Govett 1983). The lithium content of the alteration halo is low, which is readily understandable on mineralogical grounds, due to the scarcity of dark minerals capable of adopting it.

The abundance of strontium and barium in igneous rocks is 150 ppm and 250 ppm, respectively (Rankama & Sahama 1949), and higher estimates have also been put forward. According to Govett (1983), intermediate rocks contain 650 ppm Ba and 800 ppm Sr, while felsic rocks contain 830 ppm Ba and 300 ppm Sr, the crustal abundances being 430 ppm and 350 ppm respectively. The very low strontium content of the alteration halo is due to the destruction of K-feldspar and plagioclase. Barium was principally bound to K-feldspar, and its fluctuation within the alteration halo is due to the variation in the amount of sericite in the whole-rock samples.
Fig. 33. Distribution of Ni, Co, Cr and Li on a profile across the alteration halo.
5.2.9 Scandium, yttrium and lanthanum

The rare-earth elements are represented by scandium, yttrium (Fig. 34) and lanthanum (Fig. 35). The abundance of scandium in crustal rocks is 26 ppm (Cox 1989), while practically all the scandium in igneous rocks is contained in pyroxenes, amphiboles and biotite (Oftedal 1943). Pyroxene does not occur at all within the alteration halo and even amphibole is very rare, so that the main host of scandium has to be biotite. The quartz-rich rocks within the alteration halo are naturally poor in scandium.

Yttrium and lanthanum, like the other lanthanides, have a pronounced affinity for phosphorus and also for fluorine. The abundance of yttrium in crustal rocks is 31 ppm and that of lanthanum 35 ppm (Cotton 2006). Yttrium and lanthanum fluctuate in abundance between different kinds of igneous rocks, being highest in syenites and granites. According to Goldschmidt (1937b), for example, the average lanthanum content of granite is 43 ppm. Yttrium and lanthanum do not show any enrichment or depletion on a major scale in the proximity of the ore pipes or elsewhere within the alteration halo.

5.2.10 Phosphorus and vanadium

Data on samples analysed for phosphorus and vanadium are also available (Fig. 35). Phosphorus is one of the most abundant trace elements in the upper lithosphere. Estimates of its crustal abundance suggest 1000 ppm (Govett 1983; Cox 1989). Apatite is one of the most common accessory constituents of igneous rocks, and it is evident that by far the greatest part of the phosphorus in igneous rocks, about 95%, is present in apatite (Rankama & Sahama 1949). The phosphorus content of the alteration halo is by no means high, but apatite occurs richly in places, obviously a result of accumulation by retrograde metamorphic fluids.

Vanadium is also one of the more abundant trace elements. Its abundance in igneous rocks varies, being highest in basic rocks and lowest in felsic rocks. The crustal abundance is estimated to be 190 ppm (Cox 1989). The vanadium content of the alteration halo is considerably lower than for crustal rocks on average, and thus the occurrence of roscoelite, the vanadian equivalent of muscovite, is in no way evident. In fact, roscoelite belongs to the assemblage that is present in low-sulphidation systems (Ronacher et al. 2002).
Fig. 34. Distribution of Sr, Ba, Sc and Y on a profile across the alteration halo.
Concerning the halogens, only a minor body of data on fluorine and bromine is available, a set of 67 analyses published by Luukkonen (1994). The bromine content in the proximity of the ore pipes is < 0.5 ppm, mostly < 0.1 ppm, and

**5.2.11 Fluorine**

Concerning the halogens, only a minor body of data on fluorine and bromine is available, a set of 67 analyses published by Luukkonen (1994). The bromine content in the proximity of the ore pipes is < 0.5 ppm, mostly < 0.1 ppm, and
displays no dramatic variation. The average abundance of bromine in crustal rocks is 3.0 ppm (Cox 1989).

Due to its common occurrence in volcanic exhalations, fluorine is a very common element in alteration halos, its main carriers being topaz and fluorite. No data are available on the fluorine content of sericite. Fine-grained topaz is a common mineral within the proximal alteration halo at Orivesi, as is fluorite, which usually occurs in narrow, hair-like late veins. The fluorine content in the proximity of the ore pipes is about 0.25% (Fig. 36), which denotes an enrichment of about 4 to 5-fold relative to igneous rocks, where the average fluorine content is 600 ppm (Wasserstein 1947), and relative to its crustal abundance of 500 ppm (Govett 1983).

**Fig. 36. Fluorine content in relation to gold content in the proximity of the ore pipes. Analyses according to Luukkonen (1994).**

### 5.3 Hydrothermal enrichment and metamorphic remobilization

The elements reviewed here fall into various groups in relation to their behaviour in the alteration process. Some elements such as Au and also Ag, Te, Bi, Sb, As, Se, Zn, Pb and S are considerably enriched, while others such as Cu, Sn and Mo are enriched slightly or only to some extent. In the case of Cr, U, Th, P, Ba and La it is hard to see either enrichment or depletion, as the trend is indistinct. Indisputably depleted elements also exist, such as Ni, Co, Li, Sr, Sc, Y and V.
Silver and Sn are enigmatic. They are enriched in the proximity of the ore pipes and to some extent elsewhere in the alteration halo, but the unaltered volcanic rocks in the south also demonstrate high contents by comparison with the altered rocks. Some other elements similarly display migration towards unaltered volcanic rocks. Gold, Sb, As, Zn, Cu and Mo show anomalous contents within the unaltered volcanic rocks to some extent. On the other hand, the line of demarcation between altered and unaltered volcanic rocks is very sharp within many metals (e.g. Te, Bi).

The cause of the inconsistency probably lies in the metamorphic events in many phases. Metamorphic remobilization is a natural reason for the migration of elements, and there are observations of narrow quartz-sulphide veins located near the contact between the altered and unaltered volcanic rocks in the vicinity of Pipe 1, for example, that have been detected both in diamond drill cores and in underground geological mapping. Without any doubt these veins extend for some tens of metres into the unaltered volcanic rocks. One of the quartz-sulphide veins intersected by a diamond drill hole has been analysed, for instance, and contains not only SiO₂ 61.00%, FeO 20.1% and S 13.05% but also Cu 4.38%, Ni 102 ppm, Co 54 ppm, Zn 4118 ppm, Pb 3275 ppm, Ag 56 ppm, Bi 48 ppm, Te 2 ppm and Au 11.5 ppm. In the light of this single analysis many of the metals seem to be easily mobilized. It is also within the bounds of possibility that the magma pulse, being younger than the protolith and including intermediate unaltered volcanic rocks, could have become enriched through the assimilation of altered volcanic rocks, or may even have been primarily enriched in some elements.

5.4 Alteration style in the proximity of the deposit

The primary hydrothermal minerals indicate the composition of fluids and the host rock, whereas the subsequent metamorphism will alter the minerals to other species corresponding to the prevailing temperature and pressure. Whole-rock chemistry is useful for the investigating the style of hydrothermal alteration. Whole-rock analyses of samples located in the proximity of the deposit have been published by Luukkonen (1994), and the author has gathered additional samples, mainly from altered rocks (Appendix 1) and the protolith (Appendix 2), to complete the picture of alteration.

If the altered rocks in different locations display striking similarities in whole-rock chemistry, there is every reason to suggest a common hydrothermal origin. One common characteristic of the altered rocks containing high-alumina minerals
with a relatively normal aluminium content, for example, is a high ratio of aluminium to total alkali. This ratio is defined as \( \frac{Al_2O_3}{(CaO + Na_2O + K_2O)} (A/CNK) \) and has been used as an alteration index (Hashimoto 1977).

It has been noted that this ratio is consistently greater than 3 for altered rocks containing high-alumina minerals in the proximity of a mineralization. Acid-sulphate (high-sulphidation) hydrothermal alteration is responsible for producing an alteration chemistry of this type, generally caused by low-pH (< 5) fluids that were probably driven by a magma body at depth. Chemical changes that resulted from the acid-sulphate alteration must have included a depletion of alkalis and an addition of silica. The whole-rock aluminium content is not highly anomalous, and it does not appear that aluminium has been added or removed from the altered rocks. (Ririe 1990).

In the case of the Orivesi deposit the ratio A/CNK is 3 or more in the proximity of the ore pipes (Fig. 37), whereas the protoliths and unaltered volcanic rocks exhibit quite a different ratio, about 1.5 to 2. Aluminium may be slightly enriched in the altered rocks in relation to the protoliths. Some samples exhibit an extremely high \( Al_2O_3 \) content, however. This topic will be discussed later in the section on deformation and metamorphism.

![Fig. 37. Type of hydrothermal alteration, expressed in terms of the ratio of aluminium to total alkali (A/CNK). The five exceptionally aluminium-rich samples represent retrograde metamorphosed rocks. The whole rock analyses published by Luukkonen (1994) are included.](image)
6 Deformation and metamorphism

6.1 Deformation in the Tampere Schist Belt

The Palaeoproterozoic tectonic evolution of the Fennoscandian Shield is complex and includes several orogenic events. The development within the Svecofennian domain is delineated mainly by a north-south accretion stage at 1.92–1.87 Ga that led to a collage of several accretionary units and almost simultaneous amalgamation of several microcontinents and island arcs (Lahtinen et al. 2005). Deformation due to N–S compression continued until 1.87 Ga (Kilpeläinen 1998, Nironen 1999), i.e. to the end of the accretionary stage. Thereafter the change in plate motions resulted in a continental extension stage at 1.86–1.84 Ga followed in turn by continent-continent collision at 1.84–1.80 Ga. Docking of Sarmatia with Fennoscandia and oblique collision from the south-east occurred at ca 1.84 Ga. Collision of Amazonia with Fennoscandia from north-west at 1.82–1.80 Ga created deformational impacts mainly on the northern and central parts of the western edge of the Fennoscandian shield. (Lahtinen et al. 2005).

While focusing on the minor details of the Svecofennian orogeny, it is worth noting that the structure and tectonic events that resulted from the multiphase folding and deformation history of south-western Finland, including the TSB, have been topics of a great deal of research (e.g. Nironen 1989, Kilpeläinen 1998), and that general agreement exists regarding the main events concerned with deformation and metamorphism. The TSB formed in a developing island arc environment where the lowest formation, the Haveri formation, represented an early rifting state. Subduction, resulting in the formation of the present TSB, started at least 1905 Ma ago (Lahtinen 1994), but the subduction stage was evidently fairly short, 15–20 Ma, because the last formations related to it were deposited 1889 Ma ago (Kähkönen et al. 1989).

The broad outlines of the tectonic and metamorphic history of the TSB and the Pirkanmaa belt south of it are in harmony with the events in south-western Finland, and according to several meritorious investigations (e.g. Nironen 1989, Kilpeläinen 1998, Väisänen 2002) are briefly as follows. D1 structures are visible in the Pirkanmaa belt and are represented by the horizontally recumbent folds, but deformational and metamorphic structures from the D1 stage are largely lacking in the TSB. The original lithological layering was deformed by D2 folds, which are the regionally dominant tectonic feature. The F2 axial surfaces generally strike E–W and dip subvertically, while the F2 fold axes are almost horizontal and the folds
are mostly upright and isoclinal. The wavelength is extremely variable, mostly > 1 km. The penetrative $S_2$ schistosity is axial planar to the $F_2$ folds. $D_3$ structures are found throughout the TSB, but are well developed only within the $F_2$ synforms. The axial surfaces of the $F_3$ folds are subvertical and range from E–W to NE–SW in strike. The fold axes plunge steeply to the east. The $F_3$ folds are open to tight, mostly asymmetric, and the wavelength remains less than remarkable. The $S_3$ crenulation cleavage is axial planar to the $F_3$ folds and overprints the penetrative $S_2$ schistosity. $D_4$ structures are rarely observed, and the effect of $D_4$ deformation has been minor. Earlier tectonic imprints are deformed by $F_4$ kink folds in narrow zones, and small-scale extensional fracturing is associated with this $F_4$ kink folding.

The $D_1$ structures can be explained by north-south compression, and both the $D_2$ and $D_3$ structures could also have been produced in a north-south stress field. The subhorizontal $D_1$ structures are estimated to be over 1889 Ma in age, and even as much as 1920 Ma according to recent investigations (the metamorphic event in the Vammala Migmatite Belt, Rutland et al. 2004). The $D_2$ synforms and antiforms with horizontal axes and vertical axial planes developed around 1889–1878 Ma. The last folds to significantly rework the structure of the area developed during the prolonged $D_3$ period, in the interval 1878–1830 Ma. (Kilpeläinen 1998).

### 6.1.1 Deformation in the proximity of the deposit

The mining area and its surroundings offer only a tiny shutter for examining the geological structure. There are no unambiguous signs of the $D_1$ deformation phase in the surroundings of the mine, where the oldest noticeable tectonic feature is a very weak, possibly diagenetic foliation parallel to the bedding. The $S_2$ schistosity, produced by the $D_2$ deformation phase, is the dominant structure in the mining area, striking almost E–W and being subvertical, mostly dipping 85–88º to the north. The major synform of the TSB was created by the $D_2$ deformation phase.

The $D_3$ deformation phase is very important in the mining area. The gliding of layers in relation to each other, resulting in fragmentation of the more competent layers, particularly within unaltered volcanic rocks, or small-scale, almost vertical faulting across the layering at a low angle are minor effects belonging to this deformation phase. The folding is visible in outcrops, especially in the eastern part of the alteration halo (Fig. 38), and the existence of the folds is also observed underground (Fig. 39). Deformation $D_3$ created two subvertical planes of schistosity. The more obscure of these deviates from the EW-trending $S_2$
schistosity by 10 to 15º to the SE, and the other, which is more distinct, deviates 10 to 15º to the NE. The very strong lineation, dipping steeply (85–87º) to the NE or ENE is a result of this folding event. A younger lineation is also observed on occasions, dipping steeply to the NW (Fig. 40). The amplitude of the F₃ folds is approximately 30–50 m at its greatest.

D₃ is interpreted as representing a prolonged phase of deformation, and for this reason it is possible that a tiny change in the direction of the strain field could have brought about a fresh divergent cleavage and associated lineation. As a whole, it is probable that the folding events, especially F₃, pressed the pipes closer together, which has made mining operations easier and more economical, through a saving in drifting.

The F₄ kink folds belonging to the D₄ deformation occur in the form of narrow zones (Fig. 41). The proportional transition of layers in kink folding sometimes crosses the ductile-brittle transition line, resulting in open fissures into which a piece of paper or even a knife could be pushed. The axial plane is often gently dipping. The most prominent kink folding has developed in the chlorite-dominant rocks.

Fig. 38. Detailed map illustrating D₃ folding within the alteration halo. Modified after Sorjonen-Ward (1998).
Fig. 39. The almost vertical hinge of a fold caused by deformation $D_3$ on the wall of a drift cut by minor quartz veins and segregations. On the right, large numbers of fluorite aggregates on a shear plane. The length of the hammer is ca. 30 cm.
Fig. 40. Two separate lineations, the older almost vertical, while the younger forms an angle of about 25º with it. The wall of the drift trends almost east-west, and the view, about 15 cm in width, is from the north.

Fig. 41. Kink folding visible on the wall of a drift. The width of the view is ca. 2 m.
6.1.2 Structure according to magnetic modelling

Magnetic modelling makes it possible to view the present post-deformational setting of the alteration halo, including the ore pipes, in relation to the Pukala porphyry. That reported here (Fig. 42) was carried out by Juntti (1986) on the basis of aeromagnetic low-altitude measurements on the profile 2508800 E, which crosses the central part of the alteration halo ca. 250 m east of the ore pipes in a NS direction (see Fig. 11). One feature of great help from the modelling point of view is the slightly magnetized unit north of the ore pipes within the alteration halo. This unit, plate 8 at Fig. 42, reaches a length of 700 m in an E–W direction.

Both the alteration halo and the Pukala porphyry are deep-reaching. The deposit is located on the northern limb of a major syncline where the tops of the strata point south. These observations support the view that all the volcanic formations have turned into an upright position. Consistent with this, about it may be assumed that the porphyry has also turned, the roof now facing towards the south, so that it is lying edgeways in its present position. The hydrothermal fluids usually seem to be ascending from the intrusion when seeking the conduit of least resistance and the alteration halo has also moved to its present position, flattened and bent against the porphyry like a long blade of grass bends underfoot.

The structural model given by the magnetic modelling is compatible with the real situation observed during mining. The roots of the alteration halo indeed reach deep. The deepest level in the alteration funnel for which diamond drilling data are available, with occasional evidence of gold, is approximately 1200 m. The alteration halo also flares slightly upwards in spite of tectonic re-shaping, thus corresponding to the alteration funnel in recent deposits.

6.2 Metamorphism

Combining the results gained by various researchers, it may be concluded that metamorphism took place in the TSB under conditions of 470–600°C and 1.5–4 kb, varying slightly in different parts of the belt (Campbell 1978, Mäkelä 1980, Törnroos 1982, Kilpeläinen et al. 1994). It should be mentioned that the central area of the Tampere schist belt represents the least metamorphosed part of the Svecofennian domain in Finland, metamorphosed under greenschist facies conditions at its lowest.
Metamorphism in the TSB is thought to have culminated 1.88 Ga ago and to have been followed by a strong thermal pulse at approximately 1.83 Ga, which caused extensive K-granitic migmatization and is thought to be a reflection of the metamorphism that culminated 1.84–1.83 Ga ago in south-westernmost Finland (Kilpeläinen 1998, Väisänen 2002).

The metamorphic events concerning the deposit and its immediate vicinity have been investigated via fluid inclusions (Poutiainen & Grönholm 1996, Poutiainen et al. 1999), employing samples mainly collected from D2 and D3 quartz veins, but also including some younger veins in the host rocks of the ore and representing surface exposures, drill cores and underground stopes. Studies of fluid inclusion properties within the host rock proper were prevented by the small grain size of the quartz and its intense foliation and sparse fluid inclusion content.
(only a few microns in size). Most of the inclusions studied are considered to be secondary and are related to deformation and re-crystallization processes. It is also likely that earliest fluid inclusions underwent some post-entrapment modifications (i.e. volume changes and leakage) during deformation.

The research revealed three principal types of fluids and five sets of entrapment conditions (Fig. 43). The earliest observed fluid is represented by the homogeneously trapped (constant phase ratios) low-salinity (< 3 wt % NaCl equiv.) mixed (< 30 mole % CO₂) fluid inclusions in the apatite crystals of the phlogopite-chlorite schist. Since the primary H₂O–CO₂ fluid inclusions in apatite indicate the conditions under which the enclosing host was formed, this is considered to be the nearest approximation of the fluid present during the peak conditions of metamorphism of the TSB rocks. The pseudo-secondary CH₄-bearing H₂O–CO₂ fluid inclusions in sulphide-quartz veins, recording conditions of post-peak metamorphism, imply entrapment conditions of 320⁰ to 380⁰C and 2.0 to 2.8 kb, while secondary CO₂–CH₄ ± N₂ gaseous fluid inclusions within re-crystallized quartz grains and along healed micro-fractures indicate a stage at 270⁰ to 330⁰C and 0.7 to 1.6 kb. Later thermal events (<380⁰C at 1 kb) associated with ductile-brittle deformation gave rise to residual mineralization and remobilization. (Poutiainen et al. 1999).

6.2.1 Retrograde metamorphism

Porphyroblastic minerals

Judging by the conditions of entrapment of the fluid inclusions, the deposit has encountered retrograde metamorphism, which was such a prominent and eventful phenomenon that it is necessary to review it more closely. According to fluid inclusion measurements the temperature was approximately 270–380⁰C and the pressure 0.5–2.8 kb during this episode (rectangles 2, 3, 4 and 5 in Fig. 43).

The late metamorphic events can be discerned by the occurrence of some post-peak porphyroblastic minerals (Fig. 44). One of these is phlogopite, which is very common throughout the alteration halo. It occurs in two generations, the older form in scales oriented according to the cleavage, and the younger one in unoriented porphyroblasts a few millimetres in length. Phlogopite also occurs occasionally in narrow, almost monomineralic vein-like pods. The composition varies, and microscopically the grains vary from almost colourless to very pale brown, or sometimes a much darker brown with an increase in iron, upon which
the composition approaches that of biotite. The phlogopite contains a large number of inclusions undoubtedly of zircon in most cases, with pleochroic halos.

Andalusite also occurs in apparently late porphyroblasts mostly in chlorite-dominant rocks, where it forms grains that may be several centimetres in length and rich in inclusions.

Garnet is not a common mineral, but it does occur sometimes as porphyroblasts, mostly 3 to 5 mm in diameter, in chlorite-dominant rocks. The composition corresponds to that of spessartite (MgO 1.00, FeO 10.49, MnO 22.78, CaO 5.17, TiO₂ 0.54, Al₂O₃ 20.66, SiO₂ 36.66 wt %).

Fig. 43. P-T diagram showing estimated metamorphic conditions for the Orivesi gold deposit at the different stages. Rectangles: 1) P-T range for the primary H₂O–CO₂ fluid inclusions in the apatite of the phlogopite-chlorite schist, 2) estimated conditions for the pseudo-secondary / secondary H₂O–CO₂–CH₄ fluid inclusions related to sulphide-quartz veins, 3) estimated conditions for the secondary CO₂–CH₄ fluid inclusions related to Au-Te-bearing quartz veins, 4) residual mineralization and 5) remobilization (Puotiainen & Grönholm 1996, Puotiainen et al. 1999). Rectangle TSB: P-T range for the metamorphism of the Tampere Schist Belt (Campbell 1978, Mäkelä 1980, Törnroos 1982, Kilpeläinen et al. 1994). The stability field for a variety of aluminosilicates is also shown: Ka, kaolinite; Py, pyrophyllite; Ad, andalusite; Ky, kyanite; Si, sillimanite. The petrogenetic grid is based on data in Spear & Cheny 1989 (for Py, Ad, Ky, and Si) and Hemley et al. 1980 (for Ka).
Fig. 44. Microphotograph (crossed nicols) displaying two generations of phlogopite in a schist containing quartz, sericite, chlorite and altered andalusite porphyroblasts (a), microphotograph (one nicol) of a late andalusite porphyroblast (b), and a large quantity of late phlogopite and andalusite porphyroblasts in a chlorite-rich layer (c), where the height of the view is 55 mm.
Andalusite-muscovite breccia

A specific retrograde type of rock occurs within a restricted sphere in the mine. The occurrence is limited mainly to the vicinity of Pipe 5 between levels +256 and +490 and the vicinity of Pipe 3 between levels +200 to +310. The occurrence is more important in association with Pipe 5 and occasionally forms a minor part of it. The retrograde rock also occurs in places south of Pipe 5, in a zone which is practically barren. At its largest it occupies a few percent of the total cross-section of Pipe 5. This rock, which has been called andalusite-muscovite breccia ever since the first encounter, during mining operations, consists of andalusite-rich blocks a decimetre or two in diameter at their largest and an intervening matrix of mainly muscovite and pyrophyllite (Fig. 45). The light mica is not so fine-grained as the sericite elsewhere, sometimes forming scales several millimetres in diameter. One cannot avoid the impression that a highly andalusite-rich rock has been brecciated and has encountered a mineralogical transformation. The brecciation is located only in a limited zone and is separated from the unbrecciated wall rock by a muscovite-rich seam. Besides muscovite and pyrophyllite, the matrix also contains minerals such as diaspore, gem topaz, apatite, lazulite and thucholite. The rock bodies themselves, on the horizontal level, are mostly NE-trending elongated lens-shaped and sub-vertical, mostly 1 to 1.5 m thick and up to 10 m in length. The vertical dimension is considerable. The bodies are totally without any cleavage, but it is possible that small-scale shearing has occurred along the contacts afterwards.

The aluminium content of the andalusite-muscovite breccia, according to the whole-rock chemistry, is very high, as much as ca. 50 wt % Al₂O₃, and it is not certain how such an extremely high content could have been achieved. The rocks that have suffered alkali depletion do not appear to be enriched or depleted in aluminium, however, except for areas of very high fluid flow, and thus any aluminium transport must have been over relatively short distances (Ririe 1990). Although the rocks of epithermal high-sulphidation deposits contain abundant high-alumina minerals, the aluminium content of these rocks is not unusually high, and it does not appear that there has been any significant transport of aluminium either into or out of the rocks that were altered. This is the situation concerning primary hydrothermal alteration when the fluids are moderately to extremely acid. In the case of retrograde metamorphism the fluids do not need to be primarily hydrothermal nor extremely acid in order to leach aluminium minerals, and the ability to transport aluminium is even less prominent.
The high aluminium content of the andalusite-muscovite breccia is the result of the mineral assemblage, as all the main constituents are aluminium minerals, so that only the disappearance of quartz by reaction with other minerals and by removal of the extra quartz in fluids is required in order to increase the aluminium content. Of the most common high-alumina minerals, andalusite contains approximately 61% Al₂O₃, pyrophyllite at least 27% and muscovite 31–38% (Deer et al. 1992), so that a mixture consisting of one third of each would contain 40 to 42% Al₂O₃. It is possible, however, that the andalusite-dominant rocks containing various amounts of quartz might originally have been markedly rich in aluminium in some places.

The following reactions with quartz are possible (Ririe 1990). Andalusite may react with quartz at temperatures below 350ºC to form pyrophyllite:

\[
\text{Al}_2\text{SiO}_5 + 3\text{SiO}_2 + \text{H}_2\text{O} = \text{Al}_2\text{Si}_4\text{O}_{10}(\text{OH})_2,
\]

or alternatively, andalusite and quartz may react to form sericite:

\[
3\text{Al}_2\text{SiO}_5 + 3\text{SiO}_2 + 3\text{H}_2\text{O} + 2\text{K}^+ = 2\text{KAl}_3\text{Si}_3\text{O}_{10}(\text{OH})_2 + 2\text{H}^+.
\]

The microscopic observations lend support to the view that the above reactions really were of principal significance in the formation of the mineral assemblage in the andalusite-muscovite breccia (Fig. 46). Both pyrophyllite and muscovite occur in abundance and the quartz was exhausted or possibly the extra silica was removed in a fluid.

A proportion of the andalusite within a typical andalusite-muscovite breccia will sometimes have altered to diaspore probably through a reaction with water:

\[
\text{Al}_2\text{SiO}_5 + \text{H}_2\text{O} = 2\text{HAIO}_2 + \text{SiO}_2,
\]

whereupon the resulting quartz may have reacted further with andalusite to produce muscovite or pyrophyllite.

The grain size of the andalusite is mainly about 1 mm, but with single grains of up to 3 mm. Totally altered andalusite grains consist of a fine-grained sericite mat with slender prismatic diaspore grains lying on it (Fig. 46 a). The diaspore varies in size, but the largest single prisms can reach over almost the whole andalusite pseudomorph. Where andalusite has been pervasively altered, muscovite and pyrophyllite occur in the matrix. The muscovite is simply coarser than the sericite, its scales being about 0.2 mm, but there are also scales of muscovite several millimetres in diameter. Pyrophyllite occurs in connection with
sericite/muscovite, and occasionally the grains are arranged radially, forming aggregates (Fig. 45 b).

The andalusite-muscovite breccia sometimes contains rich amounts of apatite in particular spots, occurring in rounded anhedral grains of diameter up to 5 mm. According to XRD studies this is mostly fluorapatite in composition, although carbonate-apatite has also been identified (A. Peltola, unpublished data). It is worth mentioning that fluorapatite has also been identified in thin, greenish apatite-rich veins. It is not entirely unreasonable to suppose that apatite occurs in different generations and is of different compositions.

Bluish or greenish euhedral gem topaz (Fig. 45 c), sometimes reaching a length of several centimetres, also occurs in a muscovite/sericite and pyrophyllite-rich matrix, or merely in the matrix as such.

Lazulite has also been identified in connection with the andalusite-muscovite breccia, often in thin joints a couple of centimetres wide, but also in large grains or aggregates of grains (Fig. 45 d). The colour is usually blue, but zoning frequently occurs, so that the cores of the centimetre-scale grains are blue and the margins green. The occurrence of green lazulite is unusual in nature, and efforts have been made by many visiting researchers to find out what has created the green colour. It has been attributed to Fe(II)–Fe(III) colour centres and variation in Fe(II) / Fe(III) or small quantities of the transition metals Co, Ni and Cr (cf. Lahti 2002), but no noticeable variation in chemical composition has been observed between the blue and green lazulite in electron microprobe analyses.

**Thucholite**

Much attention has been aroused by the presence of black nodules or balls, mostly sited in the mass of pyrophyllite. These can be several millimetres in diameter as in Fig. 47, or even more than one centimetre and have been observed to contain amorphous carbon with inclusions consisting of U and Ti (J. Jääskeläinen, unpublished data). Brannerite, \((U, Ca, Fe, Y, Th)_3 Ti_5 O_{16}\), has previously been identified by microprobe analysis as occurring in Pipe 4 (Huhtelin 1995) and it was assumed that the inclusions in the carbon nodules could be brannerite, although it was later confirmed that there may be other species as well that are not easy to identify due to their occurrence in very minor amounts. Anyway, the inclusions evidently contain Ti, U, Th and Pb (K. Kojonen, unpublished data), as the average of 14 microprobe determinations is \(TiO_2\) 26.8 wt %, \(U_2O_3\) 42.3 wt %, and \(ThO_2\) 4.4 wt %.
Fig. 45. Retrograde minerals occurring within the andalusite-muscovite breccia: a) dark spots consist almost exclusively of andalusite, the matrix being of muscovite and pyrophyllite (width of the view ca. 35 cm), b) radially arranged pyrophyllite grains 6 to 8 mm in length, c) bright sky-blue gem topaz ca. 25 mm in length, and d) large amounts of lazulite aggregates and a vein ca. 12 mm in diameter.

Pieces of thucholite grains from both the andalusite-muscovite breccia and the muscovite-rich veins/fracture fills were selected for conventional U-Pb dating. The resulting data are presented in Table 14. All the determinations were carried out by I. Mänttäri at the laboratory of the Geological Survey of Finland. The radiogenic $^{208}\text{Pb}/^{206}\text{Pb}$ ratios of the thucholites bifurcate according to the sample site, with the $A,B,C$ and $D$ fractions from the andalusite-muscovite breccia having clearly lower ratios than those from the muscovite-rich vein/fracture fill ($E,F$). Tucholite fraction $A$ is slightly reversely discordant and does not deserve any further discussion, but the analyses of the $B,C$ and $D$ fractions give an approximate age of 1.5 Ga and the fractions $E$ and $F$, with higher thorogenic lead, indicate an even younger age of ca. 1.0 Ga (Fig. 48). These results are not inconsistent with the field evidence for a multi-phase retrograde metamorphic history, and indicate first
separation and concentration of U from Th and secondly late events in the thucholite U-Pb system.

Fig. 46. Microphotographs of retrograde metamorphosed rocks (one nicol): a) andalusite altered partially to diaspore and muscovite, b) andalusite altered totally to muscovite with special pleochroism, and c) a fine-grained mass consisting mainly of pyrophyllite.

Fig. 47. Thucholite nodules, max. ca. 5 mm in diameter, located in andalusite-muscovite breccia (a) and inclusions within the thucholite (b).
Table 14. Conventional U-Pb age data on thucholite from the Orivesi mine. Determinations by I. Mänttäri.

<table>
<thead>
<tr>
<th>Sample information</th>
<th>Sample U Pb</th>
<th>206Pb/ 204Pb</th>
<th>208Pb/ 206Pb</th>
<th>ISOTOPIC RATIOS</th>
<th>APPARENT AGES/ Ma ± 2σ</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mineral and fraction analysed</td>
<td>weight/ mg</td>
<td>% ppm measured</td>
<td>radio-genic</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Thucholite/brannerite</td>
<td></td>
<td></td>
<td></td>
<td>206Pb/238U±2σ%</td>
<td>207Pb/235U±2σ%</td>
</tr>
<tr>
<td>A) piece of grains</td>
<td>1.03</td>
<td>0.98</td>
<td>2205</td>
<td>51484</td>
<td>0.0003</td>
</tr>
<tr>
<td>B) 10% HF leached pieces of grains</td>
<td>0.38</td>
<td>1.60</td>
<td>3275</td>
<td>23320</td>
<td>0.0006</td>
</tr>
<tr>
<td>C) larger piece of grains</td>
<td>0.40</td>
<td>1.36</td>
<td>1609</td>
<td>29686</td>
<td>0.0006</td>
</tr>
<tr>
<td>D) small pieces of grains</td>
<td>0.11</td>
<td>2.95</td>
<td>3318</td>
<td>63080</td>
<td>0.0002</td>
</tr>
<tr>
<td>E) small pieces of grains, muscovite rich vein/ fracture fill</td>
<td>0.13</td>
<td>0.25</td>
<td>234</td>
<td>10481</td>
<td>0.05</td>
</tr>
<tr>
<td>F) small pieces of 5%HF-leached, muscovite rich vein/fracture fill</td>
<td>0.17</td>
<td>0.19</td>
<td>246</td>
<td>16311</td>
<td>0.05</td>
</tr>
</tbody>
</table>

1 Isotopic ratios corrected for fractionation, blank (50 pg), and age related common lead (Stacey & Kramers 1975: 206Pb/204Pb±0.2, 207Pb/204Pb±0.1, 208Pb/204Pb±0.2).

Analytical methods: The selected pieces of thucholite grains were dissolved in hot HNO₃ in Savillex beakers. The extraction of U and Pb for conventional isotopic age determinations follows mainly the procedure described by Krogh (1973). 235U-208Pb-spiked and unspiked ratios were measured using a VG Sector 54 thermal ionization multicollector mass spectrometer. The measured lead and uranium isotopic ratios were normalized to the accepted ratios of SRM 981 and U500 standards. The U-Pb age calculations were done using the PbDat-program (Ludvig 1991) and the fitting of the discordia lines using the Isoplot/Ex 3 program (Ludvig 2003).
Fig. 48. U-Pb isochron diagrams for the thucholite samples analysed, compiled by I. Mänttäri from the data in Table 14.

Thucholite is common species in ore deposits of various kinds and of all ages and it has a firm connection with hydrocarbons, uranium and minerals containing uranium, including thorium. The hydrocarbons in the ore fields of north-east Wales (Lower Jurassic), for example, occur as masses of bitumen hosting uraninite inclusions (Eakin 1989). Hydrocarbons also occur in close association with gold-bearing quartz pebble conglomerates in the Witwatersrand region of South Africa. The uranium was precipitated from the meteoric waters by reduction around carbon to form uraninite and by interaction with Ti-bearing minerals to form brannerite (Phillips & Law 2000).

Numerous occurrences of uraniferous hydrocarbons have been described in the Proterozoic rocks of Scandinavia, commonly associated with areas of mineralization and faulting and brecciation in various ore deposits (Dons 1956, Mänttäri 1995). All the uraniferous thucholite samples contain inclusions of uraninite or pitchblende, and it seems likely that the uraniferous hydrocarbons of Scandinavia, although found in a variety of geological environments, all represent alteration products of pre-existing Precambrian uraninite or pitchblende brought about by younger hydrocarbons (Welin 1966).
Thucholite also occurs in the Boliden deposit in Sweden (Grip & Ödman 1944), where it was initially identified by Aminoff (1943) and called titano-thucholite because of the considerable proportion of titanium in it (37.26% TiO₂). The thucholite of Boliden occurs mainly in an andalusite rock in the deeper levels of the mine. In the quartz-free type of andalusite rock the thucholite appears in fissures and stringers that are sometimes filled with coarse sericite. It also occurs in veins or narrow gash veins of quartz in the quartz-bearing andalusite rock. The presence of rutile probably accounts for the TiO₂ content (Grip & Ödman 1944).

Diamond drill holes at Boliden have also released natural gas, and there is a geographical connection between thucholite and gas, in that thucholite has been found in the vicinity of the gas-emitting drill holes. The main components of the gas were methane and nitrogen, the next richest component being helium.

Radioactive radiation causes a polymerization of the hydrocarbons and a solid substance is precipitated from the gas. According to Erland & Ödman (1944), there have been both radioactive minerals and a gas rich in hydrocarbons at Boliden, and the hydrocarbons may have come into contact with the radioactive minerals, so that a substance rich in carbon must then have been precipitated. Thus the thucholite must have been formed by methane and other hydrocarbons being polymerised under the influence of radioactive radiation from U-Th-minerals. Furthermore, the natural gas would seem to have been of a fossil nature, and the helium content a product of the radioactive disintegration of minerals (Erland & Ödman 1944).

Comparison of the Orivesi deposit with that at Boliden is highly interesting, as they have many features in common. Inflammable natural gas has also been noticed in the Orivesi mine on some occasions, and analyses suggest that it contained mainly methane and hydrogen sulphide. It is pure guesswork, however, as to the contributions made by methanogens to recent gas discharge.

It is conceivable that the metamorphic fluids within the Orivesi gold deposit, containing slight amounts of U and Th, could have induced the occasional alteration of rutile to brannerite, so that radioactive radiation from the brannerite could have caused polymerization of a hydrocarbon (methane) and a solid substance, amorphic carbon, precipitated from the gas, around the brannerite at a later stage. It is possible that rutile initially formed mixed grains on some occasions, e.g. with monazite, galena and uraninite. As stated above, the pseudo-secondary fluid inclusions recording conditions of post-peak
metamorphism, and particularly the secondary gaseous fluid inclusions, contained methane.

**Muscovite veins and other late overprintings**

Veins about a couple of centimetres wide which are filled with coarse-grained muscovite scales have been encountered in diamond drill cores outside the circle of the ore pipes, about 250 m NW, and elsewhere. The host rock is quartz-andalusite, and the veins contain over 95% muscovite and less than 5% transparent, clear, colourless topaz, a little fine-grained kaolinite and millimetre-sized thucholite nodules. The muscovite veins are an additional expression of a retrograde episode. Age determinations show the thucholite to be substantially younger than that in the andalusite-muscovite breccia. The coarse-grained muscovite offered a suitable target for investigating the composition of the mineral more accurately, but no special features were noted other than an unusually high Ba content, ca. 1500 ppm (A. Peltola, unpublished data). For the sake of comparison, the fine-grained sericite occurring in the vicinity was also analysed, and was similarly found to contain ca. 1500 ppm Ba. Other frequently occurring retrograde minerals are kaolinite and fluorite (Fig. 49). The richness of the kaolinite varies greatly. In some places the walls of the drift were covered in white dust after blasting, but mostly the amount of kaolinite was more minor and bound to the degree of kaolinitization of the sericite and the richness of the kaolinitized topaz. Fluorite occurs in thin veins filling late fractures, and also in rather small aggregates. It is also worth mentioning as a curiosity that gypsum has been identified in thin veins a couple of millimetres wide (A. Peltola, unpublished data).

Some researchers (Siivonen 1984, Luodes 1989) have observed a wide regional event of potassium metasomatism subsequent to hydrothermal alteration and connected with the regional shearing. It is quite obvious that these are one and the same phenomenon, the late metamorphic event that created potash feldspar and sericite in the surroundings and caused large amounts of retrograde minerals to accumulate within the alteration halo.

**6.2.2 Successive events of deformation and metamorphism**

It is logical to suggest that deformation phase D₂ represents the culmination at Orivesi, and that the temperature conceivably remained fairly high for a prolonged
time during the subsequent deformation phase $D_3$. This view is supported by the relatively late titanite U-Pb age of the granodiorite variety of the Pukala intrusion, $1851 \pm 5$ Ma (Talikka & Mänttäri 2005).

Fig. 49. Late tectonic phenomena, a fissure filled with fluorspar and later with kaolinite. The material on the right-hand edge is fine-grained kaolinitized topaz. The height of the sample is ca. 42 mm.

The retrograde metamorphism is somewhat complicated, as major mineralogical changes occurred in that phase wherever the overall chemical composition allowed it and the borders of stability fields were exceeded. The age of the brannerite/thucholite, 1.5 Ga, is astonishingly low, and hence that of the surrounding pyrophyllite and andalusite-muscovite breccia, assuming that they formed contemporaneously. An extensional geotectonic setting during rapakivi magmatism prevailed in Fennoscandia (Baltica) at that time as indicated by the extensive subparallel swarms of mainly NW-trending diabase dykes (Haapala 1988). The Fennoscandian rapakivi granites and associated rocks, including diabases, range in age from 1.65 Ga to 1.5 Ga, while the Wiborg batholith and its satellites in SE Finland represent the age group 1.60 to 1.65 Ga (Lukkari & Haapala 2000).

The Häme diabase swarm extends from the Wiborg batholith to the Orivesi area, and the number of known major dykes is about 200. These dykes are
unmetamorphic and intersect the Svecofennian metamorphic complex with sharp contacts. Most of the dykes consist of olivine diabase. (Laitakari 1987).

The younger thucholite age (1.0 Ga) is conceivably bound to the 1.25 Ga dyke activity (Vaasa and Satakunta dolerites) or rather to Grenvillian-age collisional events in the interval 1.10–1.00 Ga (Pesonen et al. 2003).

The rapakivi magmatic activity lasted a very long time, and it is evident from this that some of the late retrograde metamorphic events (at least the event depicted by rectangle 2 in Fig. 43) are associated with the diabase and rapakivi magmatism. It is also evident that the kink folding of deformation phase D₄ is associated with the extensional geotectonic setting. Evidence of a multi-phase event is also visible mineralogically, but it is hard to ascertain what phenomena, veins, fractures, fissures or late minerals, were coeval with the younger thucholite.
7 Traits of epithermal gold deposits

7.1 Definition and significance

Numerous schemes have been proposed for the classification of gold deposits in the course of time, based on a variety of criteria. Epithermal gold deposits are one of the early classes to be recognised and have been of considerable significance. The prime evidence for this is that Kerrich et al. (2000), when identifying gold deposits on a world scale, distinguished arc-related epithermal gold-silver deposits as one of the six classes.

Epithermal deposits have been exploited since antiquity. The gold and silver veins of central Europe have been worked for almost 2000 years, and those of Peru and Mexico provided gold and silver for the New World as early as the 16th century. Epithermal deposits account for 18% of the world’s present gold production and a major part of its silver production. Of the currently known epithermal deposits, more than 30 exceed 100 t Au, and the largest ones (Porgera and Ladolam in Papua New Guinea and Cripple Creeck in Colorado) exceed 500 t Au. (Marcoux & Milési 2000).

Epithermal gold deposits form in shallow parts of magma-related hydrothermal systems, commonly in subaerial volcanic arcs. A rigorous, unambiguous definition of the term epithermal is difficult to achieve. Lindgren (1922) first defined and later specified (Lindgren 1933) an epithermal environment as being shallow in depth, typically hosting deposits of Au and Ag and of base metals. Epithermal systems have also been exploited for a wide range of metals and minerals, including Hg, Sb, S, kaolinite, alunite and silica. Lindgren’s estimate for the maximum depth of formation, based on geological reconstruction, was 3,000 feet (about 1,000 m), and his estimate for the upper pressure limit was 100 atmospheres, which corresponds to that of low-salinity, low-gas water that is boiling at the depth of ~ 1000 m. Lindgren deduced a temperature range of 50º to 200ºC from the perceived stability limits of various minerals and similarities in vein textures with hot spring deposits formed at less than 100ºC.

Subsequent research (e.g. Hedenquist et al. 2000, Cooke & Simmons 2000) has revealed that deposits with textures and mineral assemblages characteristic of an epithermal environment have minerals and fluid inclusions that point to a maximum temperature of about 300ºC, although most deposits form in a temperature range of about 160º to 270ºC. The maximum temperature at a given depth under hydrostatic pressure is limited by the vapour pressure of boiling water.
As there is abundant evidence that boiling is common within epithermal ore zones, this temperature interval corresponds to a depth range below the palaeowater table of about 50 to 700 m. Few deposits with epithermal characteristics have formed below a depth of 1,000 m (Hedenquist et al. 1996, Sillitoe 1999). Lindgren (1933) concluded that ore deposition occurs because focused, rapidly ascending fluids quickly change composition within a kilometre of the surface as they rise. According to subsequent research, this change is caused by boiling, the process that most favours the precipitation of bisulphide-complexed metals such as gold.Boiling and concomitant rapid cooling also result in many related features such as gangue-mineral deposition of quartz with a colloform texture, adularia and bladed calcite and the formation of steam-heated waters that create advanced argillic and argillic alteration blankets and halos. In addition, sharp depressurization follows hydraulic fracturing, and this also focuses the flow of vigorously boiling fluid (Hedenquist et al. 2000).

### 7.2 End-member styles of hydrothermal systems

Two contrasting styles of hydrothermal systems exist within an epithermal environment and are well known from the study of active examples (Henley & Ellis 1983), and two epithermal deposit styles of contrasting alteration and ore assemblages form within these distinctly different systems and in somewhat contrasting volcanic settings.

At one extreme are geothermal systems with a near-neutral pH and reduced deep fluid. The fluid is in equilibrium with the altered host rocks owing to its relatively slow ascent, resulting in a rock-dominated system, and the liquid is typically of low salinity, less than 1 to 2 wt % NaCl equivalent, and may be gas-rich, with CO$_2$ and H$_2$S the dominant gases. When this liquid discharges at the surface, boiling is common, and neutral-pH springs deposit silica sinter. Steam-heated water also occurs in this environment, formed by the condensation of vapour into the groundwater. Vapour that condenses within the vadose zone above the water table forms blankets of sulphate-rich water, but where the vapour condenses on the margins of the system below the water table, the water is CO$_2$-rich. Surface features associated with the steam-heated zone include steaming ground, mud volcanoes and collapse craters in clay-altered ground.

Geothermal systems typically occur some distance from a volcanic edifice, although they can also occur in areas without contemporaneous volcanic activity or in volcanic rocks without any distinct connection with volcanic activity. In most
cases the geothermal system is driven by an intrusion located as much as 5 to 6 km below the surface. Although systems with relatively saline water also occur, the high density of the deep liquid prevents discharge at the surface. (Hedenquist et al. 1996, Hedenquist et al. 2000).

At the opposite extreme, volcanic-hydrothermal systems occur in a location proximal to volcanic vents that focus the discharge of magmatic vapours at the surface. Their principal surface expressions are high-temperature fumaroles and related condensates of extremely acid water. The acidic, oxidised fluid is far from being in equilibrium with the host rocks, reflecting its magmatic affiliation (Giggenbach 1992a). Strong structural control causes the rapid fluid ascent that is responsible for its reactive nature and fluid-dominated character (Giggenbach 1992b). The parent intrusions may be very shallow, even erupting to the surface (Hedenquist et al. 2000).

These volcanic-hydrothermal systems have distinctly different characteristics from their geothermal counterparts, although the two styles can coexist almost side by side. In some cases there is a transition downward from a geothermal to a volcanic-hydrothermal environment at a depth of 1 to 2 km, where hypogene acid fluid rises along fractures into an overlying neutral-pH geothermal system (Reyes et al. 1993).

7.3 High-sulphidation and low-sulphidation styles of epithermal deposits

Two distinctly different chemical environments have also been recognised for epithermal deposits, and a variety of terms have been used to describe them (Table 15). The ore deposit characteristics are strongly influenced by the contrasting compositions of the deep fluid, either neutral and reduced or acidic and oxidised. The various terms that have been used in the past usually reflected either the different processes or the resulting products of these two end-member types of fluid.

The terms low-sulphidation and high-sulphidation are currently used to reflect the two end-member sulphidation states as deduced from sulphide mineral assemblages (Hedenquist et al. 2000). These terms were first suggested on the basis of the oxidation state of the sulphur in the fluid (Hedenquist 1987), but they are now used to refer to the sulphidation state of the sulphide assemblage (Hedenquist et al. 2000). This mineralogical feature reflects the intrinsic nature of the ore fluid, both its origin and the degree of fluid-rock interaction, rock
dominated by low-sulphidation systems and fluid dominated by high-sulphidation systems (Heald et al. 1987, Hedenquist 1987, Giggenbach 1992b). It should be emphasized that the terms low and high-sulphidation do not necessarily refer to low and high concentrations of sulphide minerals, but rather reflect the oxidation potential and sulphur fugacity of the fluid that deposited the sulphides.

Table 15. Nomenclature used for the deposit types related to two contrasting epithermal environments. Modified after Hedenquist et al. (2000).

<table>
<thead>
<tr>
<th>Geothermal</th>
<th>Volcanic-hydrothermal</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dominated by neutral pH and</td>
<td>Dominated by early acidic and</td>
<td></td>
</tr>
<tr>
<td>reduced hypogene fluid</td>
<td>oxidized hypogene fluid</td>
<td></td>
</tr>
<tr>
<td>Au-quartz veins in andesite and</td>
<td>Au-alunite</td>
<td>Lindgren 1922, 1933</td>
</tr>
<tr>
<td>rhyolite; Ag-Au, Ag, Au-Te, and</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Au-Se veins; base metal veins</td>
<td></td>
<td></td>
</tr>
<tr>
<td>with Au, Ag; cinnabar, stibnite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>veins</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Enargite Au</td>
<td></td>
<td>Ashley 1982</td>
</tr>
<tr>
<td>Hot spring</td>
<td></td>
<td>Giles &amp; Nelson 1982</td>
</tr>
<tr>
<td>Adularia-sericite</td>
<td>Acid sulphate</td>
<td>Held et al. 1987</td>
</tr>
<tr>
<td>Low sulphur</td>
<td>High sulphur</td>
<td>Bonham 1986</td>
</tr>
<tr>
<td>Low-sulphidation</td>
<td>High-sulphidation</td>
<td>Hedenquist 1987</td>
</tr>
<tr>
<td>Adularia-sericite</td>
<td>Alunite-kaolinite</td>
<td>Berger &amp; Henley 1989</td>
</tr>
</tbody>
</table>

In nature, however, just as a transition exists between the end-members, geothermal and volcanic hydrothermal systems, so there are also deposits recognised with features intermediate between the low-sulfidation and high-sulfidation styles (Sillitoe 1993, Hedenquist et al. 2000). The sulphide assemblage of end-member low-sulphidation epithermal deposits is typically pyrite-pyrrhotite-arsenopyrite and Fe-rich sphalerite, while the high-sulphidation deposits are characterised by enargite-luzonite-covellite assemblages with ubiquitous pyrite. Hedenquist et al. (2000) subdivide the low-sulphidation category into end-member low-sulphidation deposits and those with assemblages indicating an intermediate sulphidation state. This classification is able to emphasize the possibility that an intermediate sulphidation state and end-member low-sulphidation deposits may form in different tectonic settings and have different magmatic affiliations (John et al. 1999).

The hypogene acid fluid that is generated in a volcanic-hydrothermal environment leaches the rock, creating a core of residual, commonly vuggy silica
that re-crystallizes to quartz. The silicic zones invariably form the core of an advanced argillic alteration halo, and the silicic core may serve as an aquifer for a subsequent ore fluid (White 1991, Arribas 1995). This ore fluid, when present, is distinct in composition from the early-stage fluid responsible for leaching. It is less acidic and less oxidised, and is also relatively saline (Hedenquist et al. 1998). Sulphide minerals deposited during this later stage include enargite and pyrite (Hedenquist et al. 2000).

The sulphidation state terminology is based on the occurrence of ore minerals, whereas barren prospects – those that experienced only the early stage of leaching – do not contain diagnostic high-sulphidation ore minerals such as enargite. Barren prospects with alteration zone conduits are common in a volcanic-hydrothermal environment, but not in a geothermal environment, the barren alteration being similar to the silicic and quartz-alunite lithocaps that form over degassing intrusions (Sillitoe 1995, Hedenquist et al. 1998, Sillitoe 2000). In these situations the high-sulphidation ore fluid either did not form at depth or did not ascend to epithermal depths, thus leaving an alteration assemblage typical of a lithocap, the term used for a barren system of this alteration style (Hedenquist et al. 2000).

The low-salinity liquid responsible for the formation of low-sulphidation ore veins and disseminations is similar to the water encountered when drilling beneath geothermal hot springs. The low-sulphidation state minerals that form this reduced, neutral-pH water are in equilibrium with the host rock alteration minerals, while the salinity of the fluids that form intermediate sulphidation state deposits is somewhat higher, and the sulphide assemblage indicates a sulphidation state that has not been fully equilibrated with the host rock (Hedenquist et al. 2000).

The two end-member styles of epithermal deposits, low-sulphidation and high-sulphidation, can also be clearly distinguished on the basis of their hypogene alteration mineralogy. Quartz-adularia-carbonate veins with sericitic or clay halos commonly host low-sulphidation ore, while a leached silicic core zone with quartz ± alunite ± pyrophyllite ± dickite halos will host high-sulphidation ores. The silicate alteration mineralogy of intermediate sulphidation state deposits is broadly similar to that of end-member low-sulphidation deposits, indicating that they also form from a near-neutral pH ore fluid. The distinctions include an abundance of rhodochrosite and anhydrite in intermediate sulphidation state deposits versus chalcedony and adularia in end-member low-sulphidation deposits (Hedenquist et al. 2000).
7.4 Geodynamic setting

As active geothermal systems show, epithermal gold mineralization is essentially related to volcanic-plutonic arcs, associated with subduction zones, both island arcs and intra-continental arcs (Heald et al. 1987). Epithermal deposits are thus seen to line the Pacific ring of fire and are also found in recent to ancient subduction zones. The deposits are emplaced at shallow crustal levels, which explains their predominant occurrence in recent (Tertiary and Quaternary) orogenic settings. Epithermal mineralization is commonly associated with a back-arc setting located as much as 100 km from the active volcanic front (Hedenquist & Lowenstern 1994). In certain arcs, e.g. in Japan and New Zealand, the magmatic and associated hydrothermal activity induced by subduction tends to move away from the trench with time. In all volcanic centres hydrothermal activity and gold-bearing mineralization occurs approximately 0.5 Ma after the initial volcanic phase (Marcoux & Milési 2000).

The gold deposits are predominantly hosted by volcanic rocks and contemporaneous volcano-sedimentary successions, and locally by basement rocks. Although the associated igneous rocks are commonly calcalkaline subaerial volcanic rocks of andesitic to rhyolitic composition, alkaline and shoshonitic rocks can be found in mature intra-oceanic arcs, e.g. in Papua New Guinea (Moyle et al. 1990) and Fiji (Ahmad et al. 1987), and in continental settings (Jébrak 1986).

7.5 Structural setting

Epithermal mineralization is controlled by major structural constraints which must be defined on several scales. Many authors have stressed the importance of large-scale structural constraints such as the intersection of tectonic plates in the localization of regions rich in deposits, such as the Andes, the Philippines and New Guinea. Crustal faults serve as major drainage pathways which subsequently enable pluton emplacement and hydrothermal circulation (Marcoux & Milési 2000). On a district scale, major fault displacement (tens to hundreds of kilometres) is a first-order metallotect which controls the distribution of mineral occurrences e.g. in Romania (Milési et al. 1994).

The faults, sometimes ancient, are reactivated during eruptive phases, but the mineralization is generally hosted by smaller-order –2 and –3 scale faults. Statistically, the most common mineralized fractures are gash veins, dilational jogs and Riedel shears associated with movement along major faults (Marcoux &
Milési 2000), but occasionally mineralization can occur where strike-slip faults develop into normal faults (Genna et al. 1996).

Mineralizations can also be hosted by volcano-tectonic structures (cauldrons and calderas) related to the volcanic activity. The epithermal deposits in New Zealand, Indonesia and Japan are located in various types of volcano-tectonic depression: grabens or pull-apart structures caused by shearing. (Marcoux & Milési 2000).

There is an example in Romania of the emplacement of a large underlying batholith determined by structural control. All the mineralized structures and volcanic products are restricted to the surface projection of the batholith, which appears to have been the controlling factor on a district scale in terms of both thermal drive and the source of material (Milesi et al. 1994).

7.6 The link with volcanic activity

The spatial relationship of an epithermal deposit to volcanic centres and/or hypovolcanic intrusion is almost systematic. On a district scale (2 – 20 km), a spatial relationship between deposits and volcano-tectonic collapse structures such as calderas is very common but not a governing rule (Marcoux & Milési 2000). Likewise the nesting of two or three calderas is not an uncommon pattern e.g. Rodalquilar in Spain (Rytuba 1994).

Mineralized veins generally occur along the caldera rims, which are preferential evacuation sites for hydrothermal fluids, and less commonly along transverse fractures in order to precipitate metals. Hydrothermal circulation can occur as much as 1–3 Ma after the collapse of a caldera (Marcoux & Milési 2000). On the scale of individual volcanic deposits (domes, flows, breccia pipes, dykes, sills), the mineralized fluids preferentially circulate through naturally permeable structures, whether tectonic or volcanic in origin. Tectonic permeable structures include faults, feeder faults and tectonic boundary structures, while the volcanic permeable structures comprise phreatomagmatic breccia, explosion breccia and late intra-centre faults, for instance. Volcanic vents (diatremes) and their surface fillings (maars) can serve as important catchment sites for hydrothermal fluids, and there exist examples of diatremes hosting first-class deposits, such as Acupan in the Philippines (Cook & Bloom 1990) and Cripple Creek in Colorado, USA (Sillitoe & Bonham 1984), which have produced 200 and 700 t Au, respectively.

The pore space within pyroclastics and volcanioclastics is also very important for the circulation of fluids, in that it can facilitate circulation or form rheological
barriers. The reaction of fluids with country rocks, particularly carbonates, influences the course of the mineralization process, and the distribution of the deposits is partly related to the morphology of the volcanic centres. (Marcoux & Milési 2000).

7.7 Genetic model

A model for the genesis of volcanic-plutonic mineralizations has been proposed, refined and revised by many authors in past years (Henley 1985, Heald et al. 1987, Sillitoe 1989, Hedenquist & Lawenstern 1994, Marcoux 1995, Marcoux & Milési 2000). The starting point for volcanic-plutonic mineralization is the generation of a rising silicic magma following deep crustal melting. The magma provides not only most of the essential components, such as fluids, gases and metals, but also the energy required by the hydrothermal system (Cathles 1981).

The silicic magma crystallizes at various depths (up to 3–5 km) liberating fluids rich in exsolved heavy metals. These metals are transported mainly by water vapour, but also gases rising to the surface through fracture systems and permeable layers. The fluids, which are initially hot (> 600°C), are dominated by brine (commonly > 40% NaCl equivalent) in equilibrium with low-density vapour. The salts dissolved in the brine crystallize out along the roof of the batholith or stock in veinlet and fracture systems, forming deposits of the porphyry copper type, which are commonly also rich in molybdenum and low-grade gold (generally < 1 g/t). Hydrothermal alteration in these deposits generates characteristic concentric halos around them, and the mineralogy reflects the high-temperature oxidizing conditions of their formation. The low-density vapour composed of volatiles (H₂O, CO₂, SO₂, H₂S, etc.) escapes from the magma and gradually condenses to an acid fluid of lower temperature and salinity. These fluids generate deposits of varying shape and characteristics along a column approximately 2 km high between the zone of plutonic rock crystallization and the surface. The blocking of fluids at depth is common and can occur at different levels. In the immediate vicinity of the pluton this gives rise to mineralizations in fractures (vein systems) or lithologically favourable layers (e.g. skarns in a carbonate environment). The fluids that reach the surface or the shallow subsurface give rise to two types of epithermal mineralization, high-sulphidation and low-sulphidation. (Marcoux & Milési 2000).

This genetic model explains the spatial relationship discovered between porphyry copper and epithermal, particularly high-sulphidation types of deposit in certain districts (Cook & Bloom 1990, Moyle et al. 1990, Hedenquist et al. 1996).
It also provides for two types of epithermal mineralization having the same fluid source, with the final type depending on the degree of fluid/rock interaction during the rise of the fluid to the surface.

7.8 High-sulphidation epithermal mineralization

High-sulphidation epithermal systems constitute the first major type of epithermal deposits and are mined for gold and commonly copper, but not for lead or zinc. They are generally formed in a relatively proximal position with respect to a hypovolcanic intrusion and are considered to be apical.

Being commonly associated with rhyodacitic volcanic activity, high-sulphidation epithermal systems mark the arrival of a hot solution (200 – 300°C) rich in \( \text{SO}_2, \text{CO}_2 \) and \( \text{HCl} \) that is saline (2 – 30% NaCl equivalent) and oxidising at the surface (Hedenquist 1995). In contrast to low-sulphidation epithermal systems, there is little or no interaction between acid epithermal fluids and the country rocks. Oxidizing, acidic waters form via the disproportionation of magmatic \( \text{SO}_2(g) \), which generates abundant sulphuric acid and minor \( \text{H}_2\text{S} \) (Cooke & Simmons 2000). The arrival of these vapours at the surface and their absorption by groundwater gives rise to acid fluids (pH < 2), which can become concentrated to form acid crater lakes and which intensely leach the surface rocks surrounding the hydrothermal pipes. Aluminium passes into solution below pH 2, and all aluminium silicates are dissolved. The only remaining surface expression is a highly characteristic vuggy facies composed of a whitish, porous, alunite-rich siliceous residue. This facies constitutes an ideal site for subsequent gold and sulphide deposition, once the mixing with meteoric water has lowered the pH sufficiently for precipitation. Hypogene leaching leaves a silica residue that recrystallizes to quartz, and this zone may also be silicified at later stages, typically before and/or during the introduction of a high-sulphidation ore (Hedenquist et al. 2000).

The highly permeable leached zones create different rheological conditions. The mineralized fluids circulate in large volumes, so that the resultant deposits are essentially disseminated ore replacing or impregnating the leached rocks. Structural control appears to be minor and is replaced by essentially lithological control through porosity and permeability. Occasionally vein structures exist, especially at depth and in the lateral zones, with a brecciated texture in places, but such structures are unusual for this type of deposit.
The acid conditions have an influence on certain characteristics of the deposits. The low pH prevents the deposition of carbonates, adularia and silica, and thus no sinter is found in this type of mineralization. Silica caps locally overlie the mineralization at the palaeo-groundwater levels, however (Marcoux & Milési 2000).

The characteristic alteration zoning, both lateral and vertical, points to a decrease in temperature and rise in pH with increasing distance from the hydrothermal conduits. Three zones can generally be distinguished: an innermost siliceous zone characterised by a vuggy silica due to maximum acid alteration (pH < 2, T 200 – 250ºC), an outer acid clay-rich zone of advanced argillic alteration, also formed during initial acid leaching (pH 2 – 3) and comprising a halo with primary quartz-alunite (passing to andalusite or diaspore very locally at depth), surrounded by a halo with quartz, dickite, pyrophyllite and then kaolinite farther out, reflecting a rise in pH, and finally an enveloping clay-rich (pH ~ 5) zone indicating argillic alteration in association with a typical high-sulphidation ore body (Steven & Ratté 1960). The diagnostic minerals are illite, illite-smectite or smectite, depending on the palaeotemperature (Stoffregen 1987, Hedenquist et al. 1996, Marcoux & Milési 2000).

The silicic core is the principal host to the high-sulphidation ore, although portions of the advanced argillic zone can also contain ore, particularly where pyrophyllite dominates over the silicic zones (White 1991). Patches of advanced argillic assemblages (quartz-alunite) may be contained within the silicic core, most probably caused by permeability variations that result in some zones being incompletely leached.

The prevailing hydrothermal alteration conditions are expressed by minerals that are stable under acid conditions (Henley & Ellis 1983, Reyes 1990), some of which are also temperature-sensitive. The advanced argillic alteration assemblage formed during initial leaching comprises minerals, of which the following occur under acidic conditions, according to increasing pH: alunite, kaolinite, dickite, pyrophyllite, diaspore, zunyite, topaz, anatase, rutile, quartz and pyrite, the last-mentioned being transitional to neutral conditions. The temperature range for acid-stable minerals such as dickite, pyrophyllite, diaspore, zunyite, topaz and rutile is approximately 200º – 300ºC, while alunite and pyrite are stable throughout the epithermal temperature range and kaolinite in the range 100ºC to more than 200ºC (Hedenquist et al. 1996).

The ore mineralogy also reflects the acidic nature of the environment. The deposits commonly contain Cu-As minerals, particularly enargite, luzonite and
tennantite, indicators of high sulphur activity (Hedenquist 1995), and the association of quartz, pyrite and enargite-luzonite in variable amounts is ubiquitous, commonly accompanied by very fine native gold and silver minerals farther away. Common, albeit minor minerals are tellurides, covellite, tetrahedrite, chalcopyrite, sphalerite and galena. Uncommon or rare ore minerals are electrum, selenides, pyrargyrite, arsenopyrite, cinnabar and stibnite. Most of the mineralization is generally focused on the central siliceous zone, where the most acid conditions prevail, but tennantite, the other copper sulphides and the rare silver sulphides occur preferentially in the peripheral clayey zone or in quartz breccia structures. (Hedenquist et al. 1996, Marcoux & Milési 2000).

7.9 Low-sulphidation epithermal mineralization

Low-sulphidation epithermal systems are associated with rhyolitic to andesitic environments. Most low-sulphidation epithermal deposits are characterized by lower temperatures (190–250°C) and salinities (0–3 % NaCl equivalent), suggesting that they are more distal in relation to the subvolcanic intrusion. Sulphides precipitate from reduced, almost neutral fluids dominated by meteoric water but containing an indisputable magmatic component (CO₂, SO₂ and HCl) (Marcoux & Milési 2000).

The final chemical characteristics of these initially acid fluids result from intense equilibration with the host rocks through interaction and dilution due to the powerful influence of meteoric water (Giggenbach 1992b). Boiling at shallow depths releases a vapour rich in CO₂ and H₂S which condenses in the vadose zone to form steam-heated acid sulphate water through oxidation upon contact with meteoric water. These fluids are identical to those encountered in bore holes in the geothermal areas of New Zealand, Italy and the United States (Henley & Ellis 1983).

Where preserved, the surface expression of low-sulphidation epithermal systems is spectacular, consisting of hot springs and geysers, both with neutral pH commonly associated with siliceous sinter or carbonate travertine (Vikre 1985, White et al. 1989). Yellowstone in Wyoming, USA, and the Taupo Volcanic Zone in New Zealand are two famous examples of active low-sulphidation epithermal systems (Browne & Ellis 1970, Henley & Ellis 1983). The sinter is locally gold-bearing in the immediate vicinity of the hydrothermal vents, commonly within volcanic explosion craters (Krupp & Seward 1987).
Low-sulphidation epithermal systems give rise to polymorphic, discordant vein deposits, commonly with polyphase brecciated and banded textures. The filling indicates several successive stages of hydraulic fracturing accompanied by mineralization. The depositional process results from boiling of the hydrothermal fluids, which releases the dissolved metals. CO$_2$ is released first, depositing carbonates and base metals, followed by H$_2$S, which carries most of the gold in bisulphide complexes. Although gold is deposited in trace amounts throughout the process, it is mainly deposited and concentrated at the end of the final phase, during the final degassing of H$_2$S. (Marcoux & Milési 2000).

Low-sulphidation epithermal deposits are generally polymetallic (Pb, Zn and very locally Cu) and are major carriers of gold and/or silver. Their sulphide content is highly variable, ranging from a few percent to almost 50% in the sulphide-rich zones of some deposits, but end-member low-sulphidation deposits contain minor base metal (Zn-Pb) sulphides, in contrast to intermediate sulphidation state deposits (Hedenquist et al. 2000).

In contrast to high-sulphidation epithermal deposits, which are characterized by alteration under acid pH conditions, low-sulphidation epithermal deposits contain minerals that are stable under neutral pH conditions (Reyes 1990). The alteration paragenesis associated with mineralization also shows characteristic vertical and lateral zoning due to the decrease in temperature towards the surface and away from the up-flow channels. The zonation is generally composed of a typically potassic centre with adularia and illite grading to smectite in the meteoric alteration zones and an outer propylitic zone. Wall-rock alteration assemblages include illite, chlorite, albite, epidote, zeolites and pyrite in addition to quartz, adularia and calcite (Hedenquist et al. 2000, Marcoux & Milési 2000). The gangue contains quartz, adularia and manganese minerals, rhodochrosite and rhodonite. The bladed calcite is a consequence of boiling (Simmons & Christenson 1994).

The ore contains characteristic sulphides: pyrite is ubiquitous and tellurides, silver sulphides such as acanthite, polybasite or pyrargyrite, electrum and native gold are common. Sphalerite and galena are much more abundant than in the high-sulphidation epithermal deposits. Conversely, chalcopyrite is rare, the fluids being too cold to transport much copper. (Marcoux & Milési 2000).

### 7.10 Origin of acid fluid in an epithermal environment

Acid water generates the advanced argillic alteration assemblages that are very typical of high-sulphidation epithermal deposits. There are three principal sources
of natural acidity: hypogene magmatic condensates, steam-heated oxidation and supergene oxidation. The first is responsible for the advanced argillic alteration of high-sulphidation deposits and sometimes of barren lithocaps, whereas the latter two can create blankets of advanced argillic alteration over both high-sulphidation and low-sulphidation deposits (Sillitoe 1993).

The proximal volcanic setting of high-sulphidation deposits accounts for the presence of hypogene acidic species, including HCl, SO₂ and HF, in decreasing abundance (Hedenquist 1995). Dissociation of the dominant acidic species, HCl and H₂SO₄ (equations a and c respectively) occurs at temperatures less than 300º to 350ºC, subsequent to absorption of the high temperature magmatic vapour by groundwater and disproportionation of the SO₂ and its reaction with water (equation b).

(a) \[ \text{HCl} = \text{H}^+ + \text{Cl}^- \]

(b) \[ 4 \text{SO}_2 + 4 \text{H}_2\text{O} = 3 \text{H}_2\text{SO}_4 + \text{H}_2\text{S} \]

(c) \[ \text{H}_2\text{SO}_4 = \text{H}^+ + \text{HSO}_4^- \]

This process results in the development of hypogene acid water containing hydrochloric acid and sulphuric acid, with a pH of about 1, sufficiently acid to leach most components, including aluminium, from the rock. This leaching leaves a siliceous residue that soon recrystallizes to quartz, sometimes with a vuggy texture, and also forms the advanced argillic alteration halo. The creation of acid conditions is dependent on the absorption of magmatic vapours by groundwater. (Hedenquist et al. 2000).

7.11 Complexation and transport of gold

7.11.1 Complexes

It is generally accepted that when simple aqueous ions, e.g. Zn²⁺, Pb²⁺ and Au⁺, predominate in a hydrothermal solution, ore mineral solubilities, i.e. the upper limits to the amounts of dissolved metal that a hydrothermal fluid can transport, are too low for the transport of significant quantities of ore metals (Czamanske 1959, Barnes 1979). The solubilities of several important ore metals in simple aqueous fluids are usually much less than 1 ppm under all conditions considered (Johnson et al. 1992), and ore metal ions must form complexes in order to attain
the solubilities necessary for ore deposit formation. A complex is simply a combination of one or more metal ions (including protons, H⁺) with one or more anionic or neutral species, referred to as ligands. Complexes may be divided into two general classes: coordination complexes and ion pairs. Coordination complexes are formed by metals and ligands capable of bonding with a high degree of covalency, so that their bonds are highly directional in character, because they depend on the overlap of atomic orbitals, and for that reason they have a definite structure that can be identified by spectroscopic methods. These complexes tend to be formed by the transition metal ions and As, Sb, Bi, Sn, Pb, Tl, Se, Ga, and Al. Ion pairs comprise a single cation and a single anion, held together solely by electrostatic attraction, i.e. ionic bonding. As a result, the bonds are non-directional and the geometry of complexes of this type is not well-defined. The ligands can be in direct contact with the metal ion, in which case they form contact or inner-sphere ion pairs, or they may be separated by water molecules, in which case they form solvent-separated or outer-sphere ion pairs (Crerar et al. 1985). The alkalis, alkaline earths, rare earth elements, Sc, Y and actinides (U, Th) tend to form ion pairs rather than coordination complexes, owing to a relatively large difference in electronegativity between the cation and anion, leading towards ionic bonding, but there is actually a continuum between the extremes of purely ionic and purely covalent bonding. (Wood & Samson 1998).

7.11.2 Relative natural abundance of ligands

Inorganic ligands of potential importance in natural hydrothermal fluids include fluoride (F⁻), chloride (Cl⁻), bromide (Br⁻), iodide (I⁻), bisulphide (HS⁻), polysulphide (S₂O₇²⁻), sulphite (SO₃²⁻), bisulphite (HSO₄⁻), sulphate (SO₄²⁻), bicarbonate (HCO₃⁻), carbonate (CO₃²⁻), ammonia (NH₃), hydroxide (OH⁻), hydrogen phosphate (HPO₄²⁻), dihydrogen phosphate (H₂PO₄⁻), phosphate (PO₄³⁻), thioarsonites (e.g. As₃S₆³⁻), thioantimonites (e.g. Sb₃S₆³⁻), telluride (H₇Te⁻, Te²⁻, Te₂²⁻) and cyanide (CN⁻). Organic ligands that exist in some hydrothermal fluids include carboxylic acid anions such as acetate (CH₃COO⁻), propionate (CH₃CH₂COO⁻), oxalate (C₂O₄²⁻) and malonate (OOCCH₂COO⁻), thiols (organic compounds containing an –SH group), amino acids and porphyrins. (Wood & Samson 1998).

Chloride is the most abundant ligand in the majority of hydrothermal fluids, with a concentration ranging from <0.03 to >7 moles/kg H₂O in natural hydrothermal fluids (Barnes 1979). As many metals form reasonably strong
complexes with this ligand, chloride complexation is often the dominant form of ore metal transport in mineralizing fluids. Bromide and iodide are generally several orders of magnitude less concentrated than chloride in hydrothermal fluids, and so, with rare exceptions, complexes with these ligands are far less important (Gammons & Yu 1997). The concentration of fluoride is often restricted to relatively low values, but fluoride complexes may be important for certain hard metal ions, e.g. Be$^{2+}$ (Wood 1992).

Whether sulphide or sulphate predominates in a hydrothermal system is mainly a function of oxygen fugacity, with sulphate predominating under oxidizing and sulphide under reducing conditions. Sulphate forms relatively weak complexes with most ore metals and hence rarely competes with chloride. On the other hand, bisulphide can form very strong complexes with soft metal ions such as Au$^+$ and Ag$^+$, and is thus probably the next most important ligand after chloride in most hydrothermal solutions. (Wood & Samson 1998).

The concentration of total carbonate and total ammonia can be relatively high, and these ligands can be important complexers of ore metals in certain limited environments (Barnes 1979). As hydroxide is always present in aqueous solutions, as a result of self-dissociation of water, it may be an important form of transport for some metals at higher pH.

7.11.3 Pearson’s HSAB principle

All metal ions may be considered to be Lewis acids, because they accept electrons, and all ligands may be considered to be Lewis bases, because they donate electrons. Therefore all metal-ligand complexes are Lewis acid-base combinations. Hard acids are small, highly charged metal ions, generally with d$^0$ electron configurations. As a result, the electron clouds of these metals are not easily deformed and they tend to prefer ionic or electrostatic bonding. Soft acids are large metal ions of low charge, typically with d$^{10}$ electron configurations, and consequently have easily deformed electron clouds. Soft acids therefore prefer to bond covalently. Pearson’s HSAB (hard-soft-acid-base) principle states that, in a competitive situation, hard acids tend to complex with hard bases and soft acids with soft bases (Pearson 1963). In other words, given the choice, metal ions will complex with those ligands that tend to bond in the same way. Metals that tend to bond covalently will generally preferentially form complexes with ligands that bond covalently, and similarly for ionic bonding. Where coordination complexes and ion pairs are concerned, it should be evident that ion pairs are formed in
interactions between hard metal ions and hard ligands, whereas coordination complexes are more likely to be formed in soft-soft interactions. The hard metals include \( \text{W}^{6+}, \text{W}^{4+} \) and \( \text{Mo}^{6+}, \text{Mo}^{5+} \) and \( \text{Mo}^{4+} \), for instance, and the soft metals \( \text{Au}^+, \text{Ag}^+, \text{Cu}^+ \), \( \text{Hg}^{2+}, \text{Cd}^{2+}, \text{Pt}^{2+} \) and \( \text{Pd}^{2+} \), in order of decreasing softness. Borderline elements are arsenic, antimony, iron, manganese, zinc and lead.

A competitive situation is important from the point of view of Pearson's principle, because given no choice, a soft metal ion may be forced to form a complex with a hard or borderline ligand. The soft \( \text{Au}^+ \) ion, for example, would prefer to complex with the soft \( \text{HS}^- \) ion, rather than the harder \( \text{Cl}^- \) ion, but in a situation where the \( \text{HS}^- \) concentration is very low and the \( \text{Cl}^- \) is very high it is forced to form a complex with \( \text{Cl}^- \), which is preferable to an even harder ligand.

### 7.11.4 Gold

The \( \text{Au}^+ \) ion is a prime example of a soft metal ion that should show a strong preference for soft ligands such as bisulphide. This is borne out by considerable experimental evidence that has emerged to suggest that gold is transported predominantly in bisulphide complexes in mildly oxidising to reducing hydrothermal solutions over a wide range of pH and temperature (see Shenberger & Barnes 1989, Pan & Wood 1994, Benning & Seward 1996). It is now very well established that the predominant \( \text{Au(I)} \)-bisulphide complexes are \( \text{AuHS}^0 \) at acidic pH, and \( \text{Au(HS)}_2^- \) at weakly acidic to weakly basic pH (Shenberger & Barnes 1989, Benning & Seward 1996). The experimental equilibrium constants of Benning & Seward (1996) for the following two reactions,

\[
\text{Au(s)} + \text{H}_2\text{S}^0 = \text{AuHS}^0 + \frac{1}{2} \text{H}_2(\text{g}) \quad \text{and} \\
\text{Au(s)} + \text{H}_2\text{S}^0 + \text{HS}^- = \text{Au(HS)}_2^- + \frac{1}{2} \text{H}_2(\text{g}),
\]

cover a wide range of temperature (150° – 500°C) and pressure (500 – 1500 bars).

There has been considerable debate about the relative importance of \( \text{Au-Cl} \) complexes in the hydrothermal transport of gold (see Henley 1973, Gammons & Williams-Jones 1995, Gammons \textit{et al.} 1997), but there is no complete, experimentally-based set of data for \( \text{Au(I)} \)-chloride complexes covering a wide range of pressure and temperature. Most studies have focused on the equilibrium constants for the reaction:

\[
\text{Au(s)} + 2\text{Cl}^- + \text{H}^+ = \text{AuCl}_2^- + \frac{1}{2} \text{H}_2(\text{g}),
\]
values for which have been measured by Nikolaeva et al. (1972), Gammons &

Also, Gammons et al. (1997) made an experimental study of the
disproportionation reaction

\[ 3\text{AuCl}_2^- = 2\text{Au}(s) + \text{AuCl}_4^- + 2\text{Cl}^- \]

at 100° to 200°C. Less thermodynamic data are available for the Au(I)-chloride
complex, AuCl°, but Sverjensky et al. (1997) provide estimates of the equilibrium
constant for the reaction:

\[ \text{Au}(s) + \text{Cl}^- + \text{H}^+ + \frac{1}{2} \text{O}_2 = \text{AuCl}^° + \frac{1}{2} \text{H}_2\text{O} \]

Au(I)-chloride complexes are considered to be more important than Au(III)-
chloride complexes for Au transport in most hydrothermal solutions (Seward
1983), and it has been well established that Au(III)-chloride complexes such as
AuCl°, AuCl₂, AuCl₃, AuCl₄, etc., predominate at lower temperatures and Au(I)-chloride complexes at higher
temperatures (Pan & Wood 1991, Gammons & Williams-Jones 1995, Gammons et
al. 1997). Thus there is a possibility of Au transport in the form of AuCl°, in low-
temperature, highly oxidized, acidic fluids, but Au(I)-chloride complexes should
predominate over Au(III)-chloride complexes under less oxidizing conditions,
especially at higher temperatures.

Thus the picture of gold speciation which emerges is the following: Au(III)-
chloride complexes will predominate under conditions of high chloride activity,
low temperature, low pH and high oxygen fugacity, i.e. near the prevailing surface
conditions, while Au(I)-chloride complexes will be important in the presence of
high chloride activity, high temperature, low pH and moderately high oxygen
fugacity. Au(I)-bisulphide complexes will predominate over a wide temperature
and pH range under mildly oxidizing to reducing conditions, i.e. embracing most
hydrothermal Au-transporting environments. Under comparatively oxidizing,
near-neutral to basic conditions, and in the absence of bisulphide or chloride,
Au(I)-hydroxide (Vlassopoulos & Wood 1990) or mixed Au(III)-hydroxychloride
(Farges et al. 1993, Sharps et al. 1993) complexes may dominate the Au
speciation.

Other ligands that may have limited roles in gold transport include
thiosulphate and sulphite (Webster 1986), which are probably limited to low-
temperature, near-surface environments, polysulphide (Berndt et al. 1994),
telluride, thioarsenic, and thioantimony species such as AsS₂⁻ or SbS₂⁻ (Nekrasov


7.11.5 Silver

Silver occurs together with gold in many gold deposits such as epithermal low-sulphidation Au-Ag deposits, where it may even be of greater economic importance than the gold. The Ag$^+$ ion is a soft ion, although not as soft as Au$^+$, and consequently complexes of Ag$^+$ with chloride would be expected to be more important than the corresponding Au(I)-complexes and those with bisulphide less important. There have been several studies of the stability of Ag(I)-chloride complexes ($\text{AgCl}^0$, $\text{AgCl}^-$, $\text{AgCl}^-$) under hydrothermal conditions (e.g. Seward 1976, Zotov et al. 1987, Levin 1991, Gammons & Williams-Jones 1995) and these indicate that Ag-chloride complexes are responsible for silver transport in a wide variety of such environments.

The thermodynamics of Ag(I)-bisulphide complexes have been determined by several researchers. Gammons & Barnes (1989) suggest that $\text{AgHSO}_3^-$ and $\text{Ag(HS)}_2^-$ are the predominant Ag-bisulphide complexes, the former in acidic sulphide solutions and the latter in near-neutral to slightly basic solutions.

7.11.6 Copper

Epithermal deposits, especially of the high-sulphidation type, may contain economically important amounts of copper. Although Cu may occur in aqueous solutions in both the Cu(I) and Cu(II) oxidation states, Cu(I) is generally expected to be predominant in most hydrothermal solutions (Crerar & Barnes 1976). The Cu$^+$ ion is a moderately soft cation, softer than most transition metals, but harder than Ag$^+$ and Au$^+$, and thus it forms comparatively stable chloride and bisulphide complexes, and hydroxide complexes may also need to be taken into account in basic solutions that are poor in other ligands. The mass transfer of Cu(I) in most hydrothermal solutions will be accomplished via chloride complexation, and the most important chloride complex is likely to be $\text{CuCl}_2^-$. Bisulphide or polysulphide complexes may play a role in certain limited environments. (Wood & Samson 1998).
Tellurium occurs in trace amounts in a wide range of geological environments, but epithermal ore deposits are the most common Te-bearing deposits. The behaviour of Te in hydrothermal fluids is poorly known, making it difficult to understand the formation of deposits of this type, especially ones that are extremely rich in Te, such as that at Orivesi.

There are three main groups of known aqueous Te species, based on their valency state: telluride (Te$^{2-}$), tellurite (Te$^{4+}$) and tellurate (Te$^{6+}$). The predominance of individual Te species depends on the redox state and acidity of the hydrothermal fluids. The more reduced deposits, of the low-sulphidation type, are formed from reduced, near neutral pH fluids, and here the telluride species ($H_2Te(aq)$, $HTe^-$ and $Te_2^{2-}$) are likely to dominate. Tellurium may be transported in other species and complexes, such as precious metal complexes (e.g. $AgHTe(aq)$, $AgTe^-$, $Ag_2Te(aq)$ or $AuHTe(aq)$, $Au(HTe)_2^-$ and $AuTe_2^-$), although the low concentrations of Au and Ag in hydrothermal solutions are not able to enhance Te transport sufficiently. In more oxidized and acidic epithermal (high-sulphidation) environments the tellurite species ($H_3TeO_3^+$, $H_2TeO_3(aq)$ and $HTeO_3^-$) are predicted to dominate. There are a number possibilities for tellurite-bearing complexes, of which perhaps the most important are those with alkali metals (e.g. $NaHTeO_3(aq)$, $NaTeO_3^-$ and potassium analogues) and halide elements (e.g. Te-chlorides). (McPhail 1995).

The tellurium concentrations in hydrothermal fluids at 300°C, calculated for redox and acid conditions that are likely in ore deposition, range from 0.005 to 5 ppb (McPhail 1995), but the solubility of hessite, native tellurium and chlorargyrite in 0.1 to 0.3 m HCl solutions at 300°C has also been measured, giving Te concentrations ranging from 1 to several 1000’s of ppm (McPhail & Bloom 1993).

Deposits can be diverse in nature and there are many possible physicochemical conditions and transport and precipitation processes which may be involved. Perhaps the most likely possibility is that Te could be transported in a gas phase (e.g. $H_2Te(g)$ or $Te_2(g)$). Indications exist that Te in low-sulphidation deposits fractionates strongly to the vapour phase over a wide range of temperature and fluid composition (Cooke et al. 1996, Cooke & McPhail 2001). Gas-phase transport of tellurium is also possible under oxidizing conditions (McPhail 1995).
7.12 Fluid inclusions

The generally corrosive environment in which ore formation occurs limits the material available for fluid inclusion studies in the case of high-sulphidation epithermal deposits. Most data come from quartz crystals filling vugs, although some data have also been obtained with respect to enargite, alunite, barite, pyrite, sphalerite, diasopre, siderite and hübnerite. Abundant trails of secondary fluid inclusions can make the identification of primary fluid inclusions difficult (So et al. 1998). In rare cases, primary fluid inclusions have been documented as occurring within the growth zones of euhedral quartz crystals, although their relationship to ore genesis is unclear (Ruggieri et al. 1997). Some authors have stated that primary fluid inclusions are present, but even they do not explain the criteria used to classify them (Mancano & Campbell 1995, Jannas et al. 1999).

High-sulphidation deposits mostly contain two-phase (liquid + vapour) fluid inclusions and both liquid-rich and vapour-rich inclusions are recognised (Cook & Simmons 2000). Multiphase brine inclusions that contain halite daughter crystals in addition to liquid and vapour ± other daughter phases at room temperature have occasionally been documented in some deposits (Arribas et al. 1995, Ruggieri et al. 1997), and three-phase fluid inclusions that contain visible CO₂ have also been encountered (So et al. 1998).

Microthermometric studies have revealed that some high-sulphidation deposits contain only dilute (<1 to ~ 5 wt % NaCl equivalent), moderate temperature (T_h = 100° – 350°C) two-phase fluid inclusions, while others have fluid inclusions with similar homogenization temperature ranges, but with much greater ranges in salinity (<1 to 46 wt % NaCl equivalent; Arribas 1995). Based on fluid inclusion geobarometry, estimates of palaeodepths below the water table for high-sulphidation deposits range from a minimum of 60 m to 900 m at the deepest.

7.13 Stable isotopes

Over the last couple of decades, from the 1980’s on, oxygen and deuterium isotope studies have significantly advanced our understanding of the formation of high-sulphidation deposits (see Rye et al. 1992). A variety of hydrous minerals that are suitable for analysis occur in different locations and at different paragenetic stages within high-sulphidation deposits (alunite, kaolinite, illite), and oxygen and deuterium isotope analysis of the hydrous phases and quartz has allowed the fluid evolution and water sources of individual deposits to be characterized in detail.
In general, alunite has oxygen and deuterium isotope compositions consistent with deposition from water predominantly of magmatic origin. The isotope compositions also suggest that the water that precipitated the hypogene clays, silica and ore mineralization had a greater component of meteoric water than that which precipitated the alunite. (Cooke & Simmons 2000).

As the most common minerals used in isotope studies are pyrite, enargite, galena, sphalerite, alunite and barite, sulphur isotope data obtained from sulphide, sulphosalts and sulphate minerals and native sulphur have been used to trace sulphur sources and estimate fluid compositions and for geothermometry. The source of sulphur is generally interpreted as being magmatic (Arribas 1995).

Sulphide minerals have distinctly negative $\delta^{34}S$ values in many deposits. Negative sulphur isotope compositions for sulphides and sulphosalts are consistent with sulphide precipitation under oxidising conditions, with redox-controlled isotopic fractionation driving the enrichment of $^{32}S$ (see Heithersay & Walshe 1995). Co-precipitating sulphates become progressively enriched in $^{34}S$, resulting in a trend towards $\delta^{34}S$ values in excess of 30 per mil (Cooke & Simmons 2000).

### 7.14 Precipitation

A given metal can be deposited for a variety of reasons, and different metals in the same solution can be deposited by different mechanisms. Gold deposition in epithermal environments results from the chemical reactions that take place within the fluids and adjacent wall rock. Boiling and mixing are the two principal ore deposition processes that occur in geothermal systems (Giggenbach & Stewart 1982), together with vapour condensation near the surface.

The debate over the genesis of high-sulphidation deposits has arisen from uncertainties concerning the chemical compositions of their mineralizing water. Arribas (1995) synthesized a two-stage genetic model for high-sulphidation deposits, in which magmatic gases cause intense acid leaching and development of advanced argillic alteration assemblages in the first stage and fluid exploits the secondary porosity and permeability created by the acid-leaching event and precipitates gold and copper minerals in the second stage. Arribas (1995) also proposed two possible origins for the mineralizing water in second stage: condensation or dissolution of magmatic gases in a heated groundwater convection cell to produce an acidic, low-salinity mineralizing water, or the ascent of mineralizing magmatic brines to the region of advanced argillic alteration, where they mix with shallow groundwater and precipitate a high-sulphidation ore.
The chemical reactions that control gold deposition in high-sulphidation epithermal deposits where gold is transported as \( \text{AuHS}_{(aq)} \) in low-salinity acid water will be as follows (Cooke & Simmons 2000):

\[
\text{AuHS}_{(aq)} + 0.5 \text{H}_2(aq) \leftrightarrow \text{Au}_{(s)} + \text{H}_2 \text{S}_{(g)}, \\
\text{AuHS}_{(aq)} + 0.5 \text{[FeO]} \leftrightarrow \text{Au}_{(s)} + 0.5 \text{FeS}_2(s) + 0.5 \text{H}_2 \text{O}(l), \quad \text{and} \\
\text{AuHS}_{(aq)} + 4 \text{H}_2 \text{O}(l) \leftrightarrow \text{Au}_{(s)} + \text{SO}_4^{2-}(aq) + 2 \text{H}^+ + 3.5 \text{H}_2(aq).
\]

The removal of \( \text{H}_2 \text{S} \) from the mineralizing water is a highly effective method for depositing gold (reaction 1), as the boiling of low-salinity acid water (from 300º to 100ºC) will cause its pH to decrease due to the dissociation of ion pairs such as \( \text{HCl}_{(aq)} \) and \( \text{H}_2\text{SO}_4_{(aq)} \) and may lead to gold deposition in association with quartz, kaolinite and covellite (Reed 1992).

In a high-sulphidation environment there are no mineralogical indicators to prove that boiling occurs other than the presence of coexisting liquid-rich and vapour-rich fluid inclusions. By increasing the acidity of the residual liquid, boiling may cause the deposition of argillic and advanced argillic mineral assemblages, whereas the solubility of base metals may increase (Cooke & Simmons 2000).

Wall-rock sulphidation reactions (reaction 2) are probably not important for the precipitation of gold in the residual silica zones of high-sulphidation deposits. The initial stage of ore formation involves acid leaching and the development of residual quartz and advanced argillic alteration assemblages. As iron is leached from the protolith, the residual siliceous rock is not a favourable site for later wall-rock sulphidation reactions. Gold, sulphides and sulphosalts generally occupy secondary pore spaces, indicating that the metals and sulphur were probably precipitated directly from stage two fluids rather than via water-rock interactions. (Cooke & Simmons 2000).

The high-sulphidation ore fluid typically deposits sericite, dickite or kaolinite in places where silicate minerals are able to form adjacent to the vuggy quartz rock that typically hosts the ore. These minerals indicate a lower pH than is typical of low-sulphidation fluids that deposit adularia and calcite, about pH 4 to 5 versus pH 6 to 7 (Hedenquist et al. 1998). In a more acid high-sulphidation environment the dominant gold complex may be bisulphide (Benning & Seward 1996, Giggenbach 1997), even though the fluid is relatively well oxidized and of moderate salinity (Hedenquist et al. 1998). By contrast, within an epithermal environment the
Boiling may not be the sole mechanism for the precipitation of sulphides and gold in high-sulphidation deposits, but it must occur at some stage in the life of such systems, as indicated by the common occurrence of hydrothermal breccias, some of which predate mineralization, and evidence of steam-heated alteration blankets prior to erosion (Sillitoe 1999). Most researchers concerned with high-sulphidation deposits advocate fluid mixing as their preferred ore deposition mechanism (see Izawa & Cunningham 1989, Losada-Calderon & McPhail 1996, Jannas et al. 1999). In contrast to many indications for boiling in low-sulphidation deposits, evidence for mixing can be derived from the O and H isotope data for some high-sulphidation ones (Hedenquist et al. 1998). Oxygen and deuterium isotopes have been used to demonstrate effectively that the water which precipitates silica and advanced argillic alteration assemblages represents a mixture of magmatic and meteoric sources. Trends in the isotopic data indicate a groundwater diluent during alunite formation (Arribas 1995). Even though some high-sulphidation deposits contain unequivocal isotopic and fluid inclusion evidence for fluid mixing it remains unclear for many deposits whether mixing is responsible for ore deposition or gangue deposition and it is possible that it may be responsible for increasing the solubility of gold by the adoption of ligands and/or metals from magmatic brines or vapours (Cooke & Simmons 2000).

Some high-sulphidation deposits contain fluid inclusions as evidence of high-salinity brines. In these systems, acid brines could transport gold as AuCl₂⁻ under oxidized or, above a temperature of ~300°C, reduced conditions. For any given high-sulphidation deposit, if acid brines transport gold as a chloride complex, then dilution, cooling, reduction and/or an increase in pH can cause deposition of gold, possibly during mixing with groundwater. In contrast, if gold is transported in high-sulphidation environments as an AuHS(aq) complex within acidic, reduced, low-salinity water, then boiling or mixing-induced oxidation may promote gold deposition, but cooling, dilution and changes in pH are unlikely to be effective depositional mechanisms. (Cooke & Simmons 2000).

7.15 Duration of hydrothermal alteration and ore deposition

Estimates of the duration of mineralizing processes have been made concerning recent hydrothermal systems and ones in which activity is still going on, e.g. the
Kuroko strata-bound sulphide deposits in Japan that are being mined for lead, zinc, copper, silver and gold. Most of the Kuroko deposits are associated with felsic volcanism of Middle Miocene age (14 Ma) and are considered to have been formed within a short time. Ueno (1975) has made an attempt to determine the length of the time interval involved from the magnetic polarity of the associated rocks, and regards it as reasonable to conclude on the grounds of reversals in magnetization polarity, combined with a knowledge of the stratigraphy, that the deposits were formed and precipitated within less than 0.2 Ma.

A good example of a high-sulphidation epithermal gold deposit is Rodalquilar in Spain. According to Rytuba (1994) the mineralization is closely related to caldera collapse and resurgence, and the development of two nested calderas and regional faulting was followed by late post-caldera intrusion and gold mineralization with acid sulphate alteration. The post-caldera stock provided the heat and SO$_2$ necessary to form the large areas of acid sulphate alteration. All these events occurred within 2.0 Ma, but it is obvious that the acid sulphide alteration and the mineralization process driven by the post-caldera stock required only a fraction of this time span.

Investigations into active geothermal systems provide an insight into the mechanism by which epithermal precious metal ore deposits were formed. The time necessary for this is naturally dependent on the outflow rate and the gold content of the fluids, but various estimates of the time taken to transport a million ounces of gold have been made. Weissberg et al. (1979) calculated a time of 57,000 years to transport this amount of gold at an outflow rate of $1.6 \times 10^6$ kg/hr. According to investigations made by Brown (1986) at Ohaaki in New Zealand with the same outflow rate but an aquifer gold concentration of 1.5 µg/kg, it would take only 1,500 years. Surface heat flow measurements suggest an upflow of $\sim 100 \pm 50$ kg/sec. At this rate, using the minimum estimate of 1.5 µg/kg for the aquifer concentration, 4,700 g of gold/year would be transported by the deep system fluid. At this rate a concentration of one million ounces would take approximately 6,600 years to accumulate. In the native pre-exploitation state of the Ohaaki field the discharge at the surface via the Ohaaki Pool was only 10 to 20 kg/sec, so that the time span required for a million ounces could be at least 30,000 years.

Judging by these investigations, the ore formation process is of relatively short duration, and an approximate time span of some tens of thousands of years would be sufficient to create a considerable deposit.
7.16 Age of epithermal gold deposits

Epithermal deposits are essentially associated with volcanic-plutonic arcs in a subduction zone setting, particularly on the back arc side, within 100 km from the active volcanic front. They are hosted especially by recent (Cenozoic, Late Miocene, Pliocene and Quaternary) subaerial volcanic rocks of andesitic to rhyolitic composition belonging mostly to a calc-alkaline suite. In the opinion of some researchers their abundance in recent formations is in fact an artifact due to their formation near the surface, which makes them particularly susceptible to erosion and leads to difficulties in identifying them in ancient metamorphic rocks. It is very likely that epithermal deposits were equally as common in older formations. (Marcoux & Milési 2000).

Epithermal Au deposits occur in island and continental arcs, but erosion rates are high in island arcs owing to rapid uplift and high rainfall, and vary between 0.1 mm/a and almost 1 mm/a being highest of all in mountainous regions. Conversely, erosion rates in continental settings are lower, so that rates of about 0.01 mm/a are common (Hedenquist et al. 1996).

Epithermal deposits are mostly situated completely within the uppermost 1000 m of the earth’s crust. Thus, at the erosion rates noted in recent island arcs, the entire deposit could be eroded away in 1 to 10 Ma. It is evident that some radical events are necessary very quickly after mineralization to preserve the ore deposit from erosion and ensure its preservation during subsequent orogenic development, and that it is possible to find such deposits in ancient formations. Radical episodes of this kind could be a new volcanic eruption of lava flow to cover the deposit or initial swift orogenic folding to hide the deposit from erosion.

Most of the known high-sulphidation epithermal gold deposits in Canada, for instance, are Tertiary in age or younger, but the Hope Brook gold deposit differs from the others by virtue of its Late Proterozoic age, the mineralization and alteration having occurred in a relatively short time interval between 574 and 578 Ma. Its preservation is likely to have been due to its early tilting and a significant extensional event which most probably buried the deposit and saved it from erosion. (Dubé & Dunning 1998).
7.17 Examples of Precambrian epithermal deposits

7.17.1 Archaean examples

Epithermal gold deposits are well known in relatively young formations, and of the two types – low-sulphidation and high-sulphidation – the former is the more common, accounting for approximately six out of every ten deposits. Additional epithermal deposits have been identified in Proterozoic and even Archaean bedrock during the last decade of the past millennium.

On the basis of his research into high alumina mineral assemblages in association with the Haile and Brewer gold deposits in the Palaeozoic Carolina slate belt and into the Sons of Gwalia and Mount Celia gold deposits in the Archaean Norseman-Wiluna greenstone belt in Western Australia, Ririe (1990) concluded that the spatial association of the acid sulphate alteration with gold mineralization, together with a comparison of analogous alteration-associated gold deposits with younger, unmetamorphosed acid sulphate ones, suggested that at least some of the gold within the Sons of Gwalia and Mount Celia deposits was introduced during an early premetamorphic alteration event. Since the present location of the gold in each deposit is a result of local changes brought about by later metamorphic and deformational events, there must exist some examples of Archaean high-sulphidation epithermal gold deposits.

Some mentions and published investigations also exist concerning Archaean deposits of the low-sulphidation type. The Campbell mine and adjacent Red Lake mine in Northwestern Ontario, Canada, are based on deposits hosted by volcanic rocks of the Archaean greenstone belt and contain considerable gold resources. The age of the gold mineralization and associated hydrothermal alteration is constrained between 2722 and 2710 Ma by the age of the pre-ore and post-ore felsic intrusions. The style of the hydrothermal alteration, including the zoned aluminous alteration, the textures of the main-stage veins, the sheeted veinlet zones and stockwork fracturing, the multiple phases of hydrothermal breccias and the anomalous Au-Ag-As-Sb-Hg-Zn-K are characteristics similar to those of Phanerozoic low-sulphidation epithermal deposits. The mineralized environment was deformed and metamorphosed to a middle to upper greenschist facies during the Kenoran orogeny (ca. 2650 Ma), but the preservation of vein textures with well-defined compositional zoning and the presence of veinlet-controlled alteration and mineralization indicate that element mobilization during the metamorphism was negligible. The main difference between the Campbell-Red
Lake deposits and low-sulphidation epithermal deposits in general is the presence of biotite and the absence of adularia (or a metamorphosed equivalent) at Campbell, and the high Au/Ag ratio there is another characteristic which differs from the situation in most low-sulphidation epithermal deposits, which are typically richer in silver. (Penczak & Mason 1997).

The Taivaljärvi Ag-Zn deposit in the Archaean greenstone belt of eastern Finland is associated with felsic metavolcanics (Taipale 1983, Kopperoinen & Tuokko 1988, Papunen et al. 1989). With the cut-off grade at 100 g/t Ag, this deposit contains 3.45 million metric tons of ore with 1.27% Zn, 0.57% Pb, 160 g/t Ag, and 0.37 g/t Au. The sulphide ore mineral content is low, however, only 5% on average. The main ore minerals are sphalerite, galena, chalcopyrite and pyrite, with some pyrrhotite, dyscrasite, freibergite, electrum and native gold. The deposit displays disseminated and vein-type ore textures.

The host rocks of the deposit are hydrothermally altered felsic metavolcanics containing abundant garnet and biotite in some places, especially close to the footwall, in addition to ubiquitous quartz and sericite. Accessory minerals are chlorite, tourmaline, rutile, barite, epidote and tremolite. Abundant carbonate minerals, in veins and as disseminations, are characteristic of the ore zone. Microprobe analyses show that the carbonate is commonly manganoankerite and manganocalcite.

Both the volcanic host rocks and the deposit are deformed and metamorphosed in high amphibolite facies, and the metamorphic mineral assemblages reflect the variations in chemical composition in the rock sequence. The intensive hydrothermal alteration associated with the mineralization predated the regional metamorphism, but it can still be recognised in the chemical compositions of the rock sequence.

The kyanite-quartz rock located 90 m above the ore layer is depleted in all elements except SiO₂ and Al₂O₃. The probable pre-metamorphic minerals, silica and kaolinite were deposited as a result of intense leaching by a liquid with a high H⁺ concentration, while the ore deposit is a product of hydrothermal activity associated with volcaniclastic felsic eruption and deposition is assumed to have been epithermal in character. (Papunen et al. 1989).

Several features refer to an epithermal deposit associated with a geothermal system and related to precious metal deposition obviously of the intermediate sulphidation type. According to Kopperoinen & Tuokko (1988), the deposit is associated with an explosive volcanic centre, as indicated by coarse pyroclastics and by the great thickness and areal extent of the fragmentary volcanics. This
displays close association with a volcanic centre, and does not preclude the possibility that the occurrence was even situated in a caldera, a situation commonly associated with recent epithermal low-sulphidation deposits. It is possible that the kyanite-quartz rock may have been a result of advanced argillic steam-heated alteration.

7.17.2 Palaeoproterozoic examples

The Palaeoproterozoic bedrock is similarly not totally lacking in epithermal high-sulphidation deposits. Some such deposits are known of today in Scandinavia, where at least two deposits in Sweden, the Enåsen Au-Cu deposit and the Boliden Cu-Au-As deposit, have been interpreted as high-sulphidation Cu-Au deposits.

Total production from open-pit and underground mining in the Enåsen Au-Cu deposit in central Sweden in 1984–1991 was 1.7 Mt of 3 g/t Au and 0.5% Cu (Hallberg 1994). This deposit consists of fine-grained gold together with disseminated chalcopyrite, pyrrhotite and pyrite hosted by topaz-bearing quartz-sillimanite gneiss. The gold occurs mainly as native gold and in tellurides (Nysten & Annersten 1985), and the sillimanite-bearing quartz rock has been found to carry high amounts of Bi, Sb and Se besides Au and Te. The ore-bearing quartz-rich rock is not homogeneous but contains argillitic lenses that are rich in cordierite and biotite.

The gold-hosting sillimanite-quartz gneiss is of an unusual composition, with quartz and sillimanite accounting for more than 95% of its bulk. Topaz, rutile, zircon and sulphides are present as accessories, and the footwall is occupied by a cordierite and sometimes magnetite-bearing biotite gneiss, most likely an altered andesitic volcanic rock. The hanging wall consists of veined sedimentary gneisses and amphibolites. The position of the deposit at a major break in the stratigraphy is expressed by the lack of hydrothermal alteration in the hanging wall and also by isotopic differences between the supracrustal units. Lead isotopes from galena within the deposit indicate an age of around 1.88 Ga, while the isotope ratios of the sulphides indicate a magmatic source of sulphur. Deformation and metamorphism to an upper amphibolite-granulite facies produced the present shape and mineralogy of the deposit, but textural observations and the sulphur isotopes indicate that the bulk chemistry of the rocks, including the sulphide and gold mineralizations, remained unchanged during the regional metamorphism. The obvious conclusion to be drawn from the above data is that the Enåsen gold deposit represents a metamorphosed counterpart to a Tertiary or younger.

The Enåsen deposit is very much analogous to the Orivesi deposit in other respects as well as its mineralogy. They are almost synchronous in age, the both lie within the realm of the Svecofennian orogeny, and even the suite of trace elements is practically identical; only the metamorphic facies is different.

The Boliden Cu-Au-As deposit, dating from around 1.87 Ga, belongs to the same orogeny and age group. This deposit is located in northern Sweden and produced 8.4 Mt of ore containing an average of 15.5 g/t Au and 1.43% Cu from 1925 to 1967 (Bergman Weihe et al. 1996). The surrounding Skellefte district possesses more than 85 pyritic volcanic-hosted massive sulphide deposits, and the Boliden Au-Cu-As deposit was one of the first to be discovered there. During the last two decades the massive sulphide ores have been collectively interpreted as volcanic exhalative formations. The Boliden ore can be divided into massive ore, pyrite and arsenopyrite-dominated lenses and vein ore, which comprises a quartz-chalcopyrite-sulphosalts-dominated assemblage occurring in brecciated parts of the arsenopyrite bodies and quartz-tourmaline veins mainly in the host rocks below the massive ore. The gold is regularly associated with the vein ore.

The hydrothermal alteration around the Boliden deposit is different from that in the majority of the massive sulphide deposits in the neighbourhood, and forms a characteristic zoned pattern with an andalusite-bearing assemblage below the massive ore in the centre, followed outwards by a sericite zone and a chlorite zone. The central andalusite zone occurs immediately below the massive ore and consists of andalusite-sericite-quartz rock and bodies of almost monomineralic andalusite. Corundum has been observed locally within the andalusite zone, and rutile-rich rock and apatite-rich rock have also been documented. (Ödman 1941, Nilsson 1968, Grip & Wirstam 1970).

The massive ore is enveloped by a sericite zone in which the dominant assemblage is sericite and quartz with minor amounts of andalusite, and substantial arsenopyrite and chalcopyrite impregnations are present adjacent to the ore. Also, an extremely silicic variety of this sericite rock with up to 80% SiO₂ has been distinguished by Nilsson (1968) in the deeper part of the sericite zone. The outermost chlorite zone is dominated by quartz, sericite and chlorite, commonly with impregnations of pyrite. Apart from Au, Cu and As, the Boliden deposit contains minor amounts of Bi, Sb, Se, Te, Mo and Co.

The massive pyrite and arsenopyrite ore in the Boliden deposit is overprinted by a later gold-quartz-copper-rich system with associated sulphosalts (Bergman
Weihed et al. 1996). This sequence of mineral deposition and the common occurrence of Bi, Se, Mo, Zn, Pb and Sn-bearing minerals in the Boliden deposit are features similar to those described for modern high-sulphidation deposits. Early leaching in the Boliden ore system was followed by the deposition of massive pyrite and Cu-As ore, typically followed later by Au ore. The timing of mineralization might have been associated with a period of shallow-water or subaerial conditions in an otherwise submerged area some 30 to 40 Ma after the main period of deposition of the exhalative volcanic-hosted massive sulphide deposits. The ore was emplaced prior to the deformation and metamorphism in a lower amphibolite facies which later remobilized the ore minerals to their present distribution. In its present setting, the ore zone is in a more or less vertical position and oblique to the lithological contacts.
8 Discussion

The nature of epithermal gold deposits, especially of the high-sulphidation type, has been discussed widely in the previous section, and comparison of the characteristics of the Orivesi deposit with the general global features of high-sulphidation deposits reveals many points in common.

The geodynamic setting of the Orivesi area meets the empirical restrictions concerning epithermal deposits, which form in shallow parts of magma-related hydrothermal systems, commonly in volcanic arcs, both island arcs and intra-continental arcs, associated with subduction zones. Geologically, the environment of the Orivesi deposit represents an arc-type system with sub-aerial calc-alkaline volcanism (Kähkönen 1999; 2005). The volcanic rocks hosting the deposit range in composition from felsic to intermediate.

The source of metals and energy for circulation is assumed to be the Pukala porphyry, an intrusion emplaced in a varied, unconsolidated volcano-sedimentary sequence. This idea is not at variance with the age determinations, because the age of the porphyry (1896 ± 3 Ma) is slightly younger than that of the protolith suite (1900 ± 4 Ma). The comb quartz layering provides powerful evidence for the ancient hydrothermal potential of the Pukala porphyry. According to many authors (Moore & Lockwood 1973; Shaver 1984; Kirkham & Sinclair 1988) comb quartz layers are an expression of the generation and evolution of ore-forming magmatic-hydrothermal fluids. Quartz crystals probably nucleate on solid or semi-consolidated substrates along the roof and walls of the magma chamber but precipitate in the presence of a separate aqueous phase. The aqueous fluid separates from the silicate melt as a product of resurgent boiling and migrates to the roof and walls of the chamber. The pressures then increase to the point that they exceed the lithostatic pressure and the tensile strength of the enclosing rocks, resulting in extensive fracturing of the surrounding rocks. Hydrothermal fluids charged with Si, K, Na, ore and gangue constituents then escape into dilatant zones in the surrounding rocks, where they precipitate their dissolved constituents in the form of veins, stockworks, breccias and replacement bodies and cause extensive hydrothermal alteration. The sudden drop in pressure also causes pressure quench crystallization of aplite or porphyritic aplite on the comb quartz layers. The hydrothermal system then seals itself by mineral precipitation and the process is repeated time and time again, as signified by the numerous rhythmic repetitions of layers. If the process is sufficiently intense and repeated many times, and if the
evolved aqueous fluids contain sufficient dissolved ore constituents, economic ore bodies may form.

Most of the intrusions hosting comb quartz layering are evidently Phanerozoic in age. This type of layering is relatively common in the apical parts of felsic intrusions related to porphyry deposits. Although the relationship between high-sulphidation gold deposits and porphyry deposits is well known, no signs of porphyry-type mineralization occur in the neighbourhood of the Au deposit at Orivesi.

The porosity and permeability of the protoliths are very important factors for the circulation of fluids, as they govern the location of the flow channels in the hydrothermal system, given that fluids always choose the passage of least resistance. The pore space within pyroclastics and volcaniclastics would have been very important in facilitating fluid circulation. It is hardly pure chance that the Orivesi ore deposit is situated at the present site, where the proportion of felsic volcanic rocks is greater than is usual within the Koskuenjärvi formation, so that the protolith was more porous and permeable due to the fragility of the high-viscosity crystallizing felsic magma. Some units of the protoliths were also capable of creating rheological barriers. It should be noted that the most prominent exceptional protolith of intermediate lava, currently containing magnetite porphyroblasts, was coherent and impermeable (Fig. 11). A protolith of this kind could have had far-reaching effects in focusing hydrothermal fluids, because the ore pipes and the proximal alteration halo are located to the south of it.

Observations on recent deposits indicate that a silicic core is the principal host of the high-sulphidation ore (Stoffregen 1987). The silicic core at Orivesi, the main host of the ore, contains up to 97% SiO₂ as its highest (Table 3). The advanced argillic alteration zone that envelopes the quartz core contains not only an abundance of quartz but also andalusite and sericite, and hence slightly less silica, but still mostly more than 80%. Advanced argillic alteration is followed by argillic alteration (sericite-dominant rocks), and in the extreme distal direction by propylitic alteration (chlorite-dominant rocks). This zonal alteration pattern is typical of high-sulphidation deposits (Steven & Ratté 1960; Stoffregen 1987; Hedenquist et al. 2000).

The occurrence of fine-grained topaz fragments is a striking feature within the alteration halo at Orivesi. The zone where topaz occurs is undoubtedly one of intense leaching by acidic, corrosive fluids. The author suggests that the reason for the occurrence of breccia-like fragments of topaz is the collapse of the hydrothermal system as a consequence of leaching. This is a common
phenomenon in association with the evolution of typical gold-rich porphyry systems, for instance (Sillitoe 2000). The fragmentation is not limited strictly to topaz-bearing rocks, however, as other lithologies in the vicinity also contain fragments of varying size and display a structure that is different from the ubiquitous primary volcaniclastic pattern (Fig. 15). One significant additional piece of evidence for hydrothermal collapse is the observation of some xenolithic blocks in the mine located near the ore pipes, the most prominent of these, approximately 100 m$^3$ in volume and consisting of a sericite-quartz schist, being situated in an entirely alien setting within andalusite-quartz rocks.

Since the protoliths in hydrothermal deposits are affected by feldspar-destructive alteration events, the chemical changes in which include a depletion in alkalis and an addition of silica, a common hydrothermal origin has been suggested for rocks of this kind, on the grounds of the striking similarity in their whole-rock chemistry. As a result of such an alteration process the ratio of aluminium to total alkali (A/CNK) is greater than 3 in the proximity of the deposit, as in the proximity of the Orivesi deposit (Fig. 37). Some researchers consider that epithermal high-sulphidation hydrothermal alteration is responsible for producing an alteration chemistry of this type (e.g. Ririe 1990).

The two end-member styles of epithermal gold deposits, low-sulphidation and high-sulphidation, can generally also be clearly distinguished on the basis of their hypogene alteration mineralogy. Quartz-adularia-carbonate veins with sericitic or clay halos commonly host low-sulphidation ore, while the silicate alteration mineralogy of intermediate sulphidation-state deposits is broadly similar but includes the feature that they also form from an ore fluid of near-neutral pH. The distinctions include an abundance of rhodochrosite and anhydrite in intermediate sulphidation-state deposits versus chalcedony and adularia in end-member low-sulphidation deposits. No clues to the original occurrence of adularia, calcite and minerals containing manganese exist within the Orivesi alteration halo.

The primary hydrothermal minerals that are most frequently mentioned as belonging to high-sulphidation epithermal deposits are quartz, alunite, kaolinite, pyrophyllite, diaspore, illite and barite, while pyrite and enargite-luzonite are the most common ore minerals (Hedenquist et al. 1996). In the case of the Orivesi deposit, these hydrothermal minerals evidently recrystallized in the course of the prograde metamorphism as other species corresponding to the prevailing pressure and temperature, while later retrograde metamorphism created some of these minerals anew, e.g. pyrophyllite, diaspore and kaolinite.
The ore minerals share a common metamorphic history with the silicates. Geothermobarometric studies of arsenopyrite inclusions in pyrite and pyrrhotite and of Fe-rich sphalerite inclusions in pyrite in the Orivesi deposit indicate temperatures below 350ºC and a pressure of about 2.4 kb (Luukkonen 1994), evidently recording conditions involving the re-equilibration of sulphides. One further piece of evidence supporting this view is the occurrence of enargite-luzonite (Cu₃AsS₄) and jordanite-geocronite (Pb₁₄(As,Sb)₆S₂₃) in a vein or vein-like setting (T. Törmänen, unpublished data).

One feature peculiar to the Orivesi deposit is that the gold occurs in a very fine-grained form. The native gold is poor in silver and electrum is uncommon. The copper content of the ore is low, as also is that of zinc and lead, whereas tellurium and bismuth occur in abundance. Many of these features are characteristic of high-sulphidation deposits. Low-sulphidation deposits contain more silver, zinc and lead, and the same is especially true of intermediate sulphidation-state deposits, which are often extremely rich in silver, zinc and lead by comparison with low-sulphidation ones, and particularly with end-member high-sulphidation ones (cf. Cook & Simmons 2000; Hedenquist et al. 2000).

As a whole, the data indicate that the Au-content of the ore pipes in the Orivesi deposit decreases downwards. Peak values become more infrequent, and the mean level is definitely reduced. The trend is quite similar with Te and Bi, and the inevitable conclusion is that the decrease in these elements with depth follows that of gold. This trend, together with the decrease in cross-sectional area, illustrates the primary situation in deformed up-flow channels and predicts withering of the ore formation at a greater depth. This shape is characteristic of a high-sulphidation deposit.

Different groups of elements exist within the alteration halo of an epithermal deposit in relation to their behaviour in the alteration process. Some elements are considerably enriched and others are depleted. In addition to Au and Ag, many elements, such as Cu, Zn, Pb, Mo, Te, Bi, Sn, As, Sb, Se, Hg ± Tl and Ba, are typically enriched during epithermal ore deposition in general (Hedenquist et al. 2000), but it is not possible to define a suite of elements that is always anomalous, either in epithermal high-sulphidation deposits themselves or in the associated alteration halos. The elements that are enriched within the deposit and/or alteration zone at Orivesi are Au, Ag, Te, Bi, Sb, As, Se, Zn, Pb, Cu, Sn and Mo, a combination that matches up well with the expected assortment of elements that are usually enriched during epithermal ore deposition.
The characteristics noted in the Orivesi deposit are compatible with those commonly discovered in epithermal high-sulphidation deposits according to the literature, and there is every reason to suggest that the deposit is of the epithermal high-sulphidation type. Some differences do exist between the Proterozoic Orivesi deposit and other recently described high-sulphidation deposits, but every deposit is individual and unique in some respects.

In the light of observations on the Orivesi deposit, and applying published information on other well-known deposits, it is possible to create the following schema for its geological evolution and ore formation (cf. Table 16).

Volcanic activity initiated via an accumulation of volcanic rocks belonging to the Koskuenjärvi formation and emplacement of the hypabyssal porphyry occurred below and within the volcanic pile some time after the eruptions had commenced. It is highly possible that local faults and fractures were created almost simultaneously by the emplacement of the porphyry, through associated earthquakes or the formation of a caldera shaped by explosive volcanic eruption, and that hydrothermal alteration started in the most porous and permeable pyroclastics shortly after that. The pore space within the pyroclastics and volcaniclastics was very important for facilitating the fluid circulation required in the later mineralization process. The location of the deposit is thus partly related to the morphology of the volcanic centre. The prime expression of the major role of the Pukala porphyry in initiating hydrothermal fluids and subsequent ore formation is the occurrence of comb quartz layering. The hypabyssal porphyry was the source of metalliferous fluids and the heat source for the hydrothermal circulation.

At the beginning of circulation the hydrothermal fluids were acid, with a pH possibly under 2, resulting in intense leaching at the core of the alteration zone. In a subsequent stage of leaching the corroded zone collapsed in some places, as indicated by the occurrence of fragments of fine-grained topaz. In the course of time the fluids gradually became more basic, the pH increasing to around 3 to 5, and the fluid circulation resulted in the transport of gold and associated elements and in silicification.

The hypogene acid fluid generated in the volcanic-hydrothermal environment had an influence on certain characteristics of the deposits. The low pH prevented the deposition of carbonates and adularia, for instance, which are therefore not found in this type of mineralization. The characteristic zoning, both lateral and vertical, illustrates a decrease in temperature and rise in pH with increasing distance from the hydrothermal pipes.
It is well established that the predominant Au(I)-bisulphide complexes are AuHS⁺ at acidic pH, and Au(HS)₂⁻ at weakly acidic to weakly basic pH (Benning & Seward 1996; Cook & Simmons 2000), and it is conceivable that these bisulphide complexes played an important role with respect to gold in the context of ore transportation at Orivesi. There is a possibility, however, that gold may have been transported in chloride complexes (AuCl, AuCl₂⁻, AuCl₄⁻) under conditions of high chloride activity and oxidization at various temperatures.

The most common conceivable causes of gold precipitation in the case of high-sulphidation deposits are fluid mixing and boiling (Cooke & Simmons 2000), and it is obvious that related phenomena also promoted gold deposition at Orivesi. One conceivable piece of evidence for boiling as the important factor in precipitation is the overwhelming abundance of Te within the Orivesi deposit, particularly in the upper part. It is recognized that the transport of tellurium in a gas phase is indicated over a wide range of temperature and fluid compositions (McPhail 1995; Cook & McPhail 1995; Cook et al. 1996).

After a relatively short-term episode of hydrothermal activity, volcanic eruptions commenced again, producing the suite of unaltered volcanic rocks located south of the alteration halo, and the associated earthquakes may possibly have brought the hydrothermal circulation to an end by closing the hydrothermal channels. The volcanic activity in the TSB was of fairly short duration according to the age determinations, covering the period 15–20 Ma altogether. The volcanic activity was followed by deformation and metamorphism, with pressure and temperature reaching peak conditions within the lower part of the amphibolite facies, and the primary hydrothermal minerals were recrystallized or replaced by metamorphic mineral species according to the prevailing pressure and temperature. No signs of significant enrichment or remobilization of gold are visible on any major scale, however.

The initial folding tilted the hypabyssal intrusion steeply to the south, and the upward-flaring alteration halo and the ore pipes were reshaped as a result of the multiphase deformation. The lineation is very pronounced, the pyroclasts are markedly stretched in a vertical direction, and the pipes became at least double their primary length. The known parts of the pipes are nearly vertical in attitude, dipping to the NE after deformation. The position of the eroded parts of the pipes and of the continuations of the fluid channels at greater depths are obscure. Finally, the subsequent retrograde metamorphism created some of the primary hydrothermal minerals anew.
Table 16. Geological evolution and characteristic features of the Orivesi gold deposit and its host rocks.

<table>
<thead>
<tr>
<th>Tampere Schist Belt</th>
<th>time span of subduction-related volcanic activity</th>
<th>1904 ± 4 to 1889 ± 5 Ma</th>
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</thead>
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<td>Protolith</td>
<td>part of the Koskuenjärvi formation, age</td>
<td>1904 ± 4 Ma</td>
</tr>
<tr>
<td></td>
<td>sub-aerial calc-alkaline volcanic rocks, mainly felsic in the proximity of the deposit</td>
<td></td>
</tr>
<tr>
<td>Pukala porphyry</td>
<td>crystallization</td>
<td>1896 ± 3 Ma</td>
</tr>
<tr>
<td></td>
<td>hydrothermal alteration and ore deposition</td>
<td></td>
</tr>
<tr>
<td>Ore deposit</td>
<td>high-sulphidation type</td>
<td>150-300°C</td>
</tr>
<tr>
<td></td>
<td>characteristic elements: Au, Te, Bi, F</td>
<td></td>
</tr>
<tr>
<td></td>
<td>low Cu</td>
<td></td>
</tr>
<tr>
<td>Metamorphism</td>
<td>prograde metamorphism</td>
<td>460-560°C, 3-4 kbar</td>
</tr>
<tr>
<td></td>
<td>culmination</td>
<td>ca. 1880 Ma</td>
</tr>
<tr>
<td></td>
<td>retrograde metamorphic events</td>
<td>250-370°C, 0.5-2.5 kbar</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ca. 1500 Ma and ca. 1000 Ma</td>
</tr>
</tbody>
</table>
9 Conclusions

The Tampere Schist Belt is one prominent subarea of the Palaeoproterozoic Svecofennian domain in Finland. Within this volcanic-sedimentary belt the chemical composition of the volcanic rocks, resembling those of present day volcanic arcs, varies greatly from rhyolites to basalts, from subalkaline to alkaline, from low-K to very high-K or shoshonitic, and from calc-alkaline to tholeiitic, although intermediate rocks with calc-alkaline affinities predominate or are abundant in most areas. The metamorphosed Palaeoproterozoic Orivesi gold deposit is located within the central part of the Tampere Schist Belt, about 25 km NE of the city of Tampere. The Orivesi mine was in production from 1994 to 2003, run by Outokumpu Mining Oy and a total of approximately 1.7 million tons of ore was extracted, with an Au content of 9.31 g/t, yielding a total output of gold in concentrate of 13.115 tons.

The central part of the Tampere Schist Belt comprises an EW-striking major synform with subvertical axial planes and subhorizontal fold axes, due to the main phase of folding. The lowermost unit at Orivesi, on the northern limb of the major synform, is the Koskuenjärvi formation, in which felsic rocks are relatively abundant but mafic volcanics are rare. The volcanic rocks are largely of pyroclastic origin, as is typical of sub-aerial and shallow-water arc environments. The host rock of the Orivesi Au deposit belongs to this Koskuenjärvi formation, which is 1904 ± 4 Ma in age.

According to observations made in the mine, the mining area and its vicinity, the Koskuenjärvi formation consists of two members of slightly different age: the southern hydrothermally unaltered intermediate volcanic rocks and the northern protolith suite which embraces the alteration halo. The former has not encountered any hydrothermal alteration at all. The sharp contact and an abrupt change in mineralogy denote that these rocks that remained unaltered are slightly younger than the protolith suite and thus erupted after the period of hydrothermal alteration. The protolith suite is mainly pyroclastic in origin and rhyolitic in composition, but intermediate volcanic rocks also occur within it.

The hypabyssal intrusion known as the Pukala porphyry is emplaced immediately north of the alteration halo and is separated from it by a narrow transition zone. The EW-striking intrusion is ca. 18 km long and 1–2 km wide, bordering on the protolith suite. The Pukala porphyry is a quite homogeneous, light-coloured rock consisting mainly of porphyritic granodiorite and trondhjemite, and comb quartz layering has been discovered in the transition zone.
at its contact. The Pukala porphyry is the obvious source of both hydrothermal fluids and metals. The zircon U-Pb age for the granodiorite variant is 1896 ± 3 Ma.

The lithologies within the alteration halo may be divided according to their prevailing minerals (from the outer zone inwards) into chlorite-dominant, sericite-dominant and quartz-dominant rocks. The host rocks of the ore are mainly quartz-dominant, comprising quartz rocks, andalusite-rich quartz rocks and a minor proportion of andalusite-dominant rocks.

The quartz-dominant rocks contain approximately 97% SiO₂ at most, and are considered equal to the silicic core that normally occurs in high-sulphidation epithermal deposits. The advanced argillic alteration zone enveloping this quartz core also contains andalusite and sericite, and hence slightly less silica. This advanced argillic alteration is followed by argillic alteration (sericite-dominant rocks), and in the extreme distal direction by propylitic alteration (chlorite-dominant rocks).

Topaz-bearing and topaz-dominant lithologies are met with in the inner zone of the alteration halo. A minor proportion of the topaz-bearing rocks consist of almost monomineralic topaz, while the rest are topaz-bearing quartz rocks in which the topaz occurs in kaolinized spot-like fragments varying in size from 5 mm to a few decimetres. These fragments are mostly roundish but locally also angular, and contain tiny cracks and splits which are usually filled with quartz. The grain size of the topaz is very fine, and the fragments were originally almost monomineralic. It is possible to infer from these features that the fragmentation was a relatively early phenomenon, and that the most likely reason for the occurrence of topaz in breccia-like fragments is collapse of the hydrothermal system as a consequence of acid leaching. The fragmentation also extends beyond the topaz-bearing rocks in places.

The deposit consists of five nearly vertical pipes, with Pipe 5 constituting a direct downward continuation of Pipe 4, cut only by a basic dyke with a minor fault relevant to the dyke intrusion. Each pipe has its own particular characteristics in addition to common features. The volcaniclastic structure is equally prominent everywhere, but the amount of matrix between the clasts varies to some extent. Another common feature is that the pyroclasts are markedly deformed, with a length at least double their primary length. This means that the vertical length of the pipes must also be at least double what it was before deformation.

The matrix in Pipe 1 consists mostly of pyrite, although sericite and andalusite also occur. For that reason the feed mined from Pipe 1 contains 3.4% S while that from Pipe 5 contains only 0.3% S and that from Pipes 2 and 3 remains at much the same level. Apart from the pyrite-rich matrix, Pipe 1 contains semi-massive to
massive pyrite lenses. The matrix in Pipe 2 is composed mainly of a thin sericite film with a few andalusite grains between tightly packed clasts, while the host rock in Pipes 3 and 5 contains more andalusite and the matrix is rich in andalusite and sericite. The clasts are more loosely packed and the amount of matrix increases noticeably.

The well-known part of the pipe system is almost vertical, migrating slightly to the NE with increasing depth, and the hydrothermal channel follows approximately the same direction as it continues to a depth of at least 1200 m. This continuation was noticed in connection with diamond drilling, but its gold content is lower and sporadic.

The most common sulphides in the alteration halo, and especially in connection with the Orivesi gold deposit, are pyrite, pyrrhotite, sphalerite, chalcopryite, galena and arsenopyrite, and the sulphosalts identified are tetrahedrite ((Cu,Fe)\textsubscript{12}Sb\textsubscript{4}S\textsubscript{13}), boulangerite (Pb\textsubscript{5}Sb\textsubscript{4}S\textsubscript{11}), bournonite (PbCuSbS\textsubscript{3}) and meneghinit (Pb\textsubscript{13}CuSb\textsubscript{2}S\textsubscript{23}). The most important group of minerals is the tellurides, which occur in many species within all the ore pipes. The most common gold, gold-silver and silver tellurides are calaverite (AuTe\textsubscript{2}), montbrayite ((Au,Sb)\textsubscript{2}Te\textsubscript{3}), kostovite (CuAu\textsubscript{4}Te\textsubscript{4}), petzite (Ag\textsubscript{3}AuTe\textsubscript{2}), sylvanite ((Au,Ag)\textsubscript{2}Te\textsubscript{4}) and hessite (AgTe\textsubscript{2}), while other known tellurides include tellurobismuthite (Bi\textsubscript{2}Te\textsubscript{3}), altaite (PbTe), melonite (NiTe\textsubscript{2}), frohbergite (FeTe\textsubscript{2}), tsumoite (BiTe), tetradymitie (Bi\textsubscript{2}Te\textsubscript{2}S) and rucklidgeite ((Bi,Pb)\textsubscript{2}Te\textsubscript{4}). The minor tellurides identified are josiite (Bi\textsubscript{4}(S,Te)\textsubscript{3}), volynskite (AgBiTe\textsubscript{2}) and tellurantimony (Sb\textsubscript{2}Te\textsubscript{3}). Tiny crystals of krennerite (AuTe\textsubscript{2}) and coloradoite (HgTe) have been identified occasionally.

Native gold particles occur either as single grains or in groups of grains and are unevenly distributed. The grain size of the gold is usually very small, in general less than 20 µm and almost half of the gold grains are inter-grown with various tellurides, while approximately one quarter of the gold is bound to gold tellurides. The rest of the gold consists of native grains, either as inclusions in quartz or between grains of quartz and other silicates. A small minority of the native gold occurs as inclusions in sulphides and arsenides.

A volcaniclastic structure is ubiquitous within the deposit, with the clasts or fragments varying greatly in size from centimetre to decimetre-scale. The proportion of matrix also varies greatly. The distribution of gold between the matrix and clasts is unbalanced, as the gold content of the clasts is low, approximately 5% of the gold being situated in the clasts and the rest in the surrounding matrix. The distribution of tellurium and bismuth between the clasts and matrix is of the same order.
The gold correlates very well with Te and also with Bi within the ore pipes. The highest values for Te and Bi are about 1 wt %. The values for Bi exceed those for Te on a very few occasions, and the Bi content is usually at most equal to that of Te or commonly lower. The average Te content of 2439 samples is 196 ppm and the Bi content 99 ppm (Au 11.3 ppm). The Te/Bi ratio is almost exactly 2:1.

The elements existing within the alteration halo can be grouped in relation to their behaviour during the alteration process. Some of them, Ag, Te, Bi, Sb, As, Se, Zn, Pb and S in addition to Au, are considerably enriched, while others such as Cu, Sn and Mo are enriched slightly or to some extent. In the case of elements such as Cr, U, Th, P, Ba and La it is hard to see any enrichment or depletion.

Whole-rock analyses show the ratio $\text{Al}_2\text{O}_3 / (\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$ (A/CNK) in the proximity of the ore pipes to be 3 or more. Hydrothermal alteration of the epithermal high-sulphidation type was responsible for producing an alteration chemistry of this kind. The zonal structure of the alteration halo, the silicic core with quartz-rich rocks and the signs of hydrothermal collapse caused by conceivable corrosive acid fluids support this view.

After its formation, the Orivesi deposit encountered deformation and metamorphism, reaching lower amphibolite facies conditions. The deformation reshaped the deposit, which is now located on the northern limb of a major synform where the directions at the top of strata point south. These observations support the view that all volcanic formations are turned into an upright position. It is consistent to think that the porphyry has also been turned, as its roof now faces south, lying edgeways in its present position. The alteration halo has also been moved into its present position, and has been bent and flattened against the porphyry. Judging by the conditions of entrapment of fluid inclusions and the reappearance of some hydrothermal minerals typical of high-sulphidation epithermal deposits, this deposit has encountered retrograde metamorphism. One particular type of retrograde rock, andalusite-muscovite breccia, contains pyrophyllite, diaspore, gem topaz, apatite, lazulite and thucholite, and other late and mostly porphyroblastic minerals that support the idea of retrograde metamorphism are phlogopite, muscovite, late andalusite several centimetres in length and rich in inclusions, spessartite, kaolinite and fluorite.

Conventional U-Pb dating of the brannerite/thucholite for determination of the junctures of retrograde metamorphic events gave an age approximation of 1.5 Ga for the andalusite-muscovite breccia and an even younger age of ca. 1.0 Ga for the muscovite-rich vein.
References


Sederholm JJ (1891) Studien über archäische Eruptivgesteine aus dem südwestlichen Finnland.


Appendices
Appendix 1 Whole-rock analyses of hydrothermally altered rocks.

Results in wt %, X-ray fluorescence analyses by the VTT Geoanalytical Services

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rock type</th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>Fe₂O₃</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
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<th>TiO₂</th>
<th>P₂O₅</th>
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<td>0.14</td>
<td>0.07</td>
<td>0.20</td>
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<td>0.01</td>
<td>0.19</td>
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<td>0.00</td>
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Abbreviations
- QzSerS: quartz-sericite schist
- AdlMuBrc: andalusite-muscovite breccia
- QzR: quartz rock
- ToQzR: topaz-quartz rock
- AdToQzR: andalusite-topaz-quartz rock
- AdChlSerS: andalusite-chlorite-sericite schist
- ChiSerS: chlorite-sericite schist
Appendix 2 Whole-rock analyses of the protolith suite.

Results in wt %, X-ray fluorescence analyses by the VTT Geoanalytical Services

<table>
<thead>
<tr>
<th>Sample</th>
<th>SiO₂</th>
<th>TiO₂</th>
<th>Al₂O₃</th>
<th>FeOtt</th>
<th>MnO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
<th>P₂O₅</th>
<th>OxSum</th>
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Appendix 3. Whole-rock analyses of unaltered volcanic rocks

Results in wt %, X-ray fluorescence analyses by the VTT Geoanalytical Services

<table>
<thead>
<tr>
<th>Sample</th>
<th>SiO₂</th>
<th>TiO₂</th>
<th>Al₂O₃</th>
<th>Fe₂O₃</th>
<th>MnO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
<th>P₂O₅</th>
<th>OxSum</th>
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494. Syrjänen, Anna-Liisa (2007) Lay participatory design: A way to develop information technology and activity together

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Aulis Kinnunen

A PALAEPROTEROZOIC HIGH-SULPHIDATION EPITHERMAL GOLD DEPOSIT AT ORIVESI, SOUTHERN FINLAND